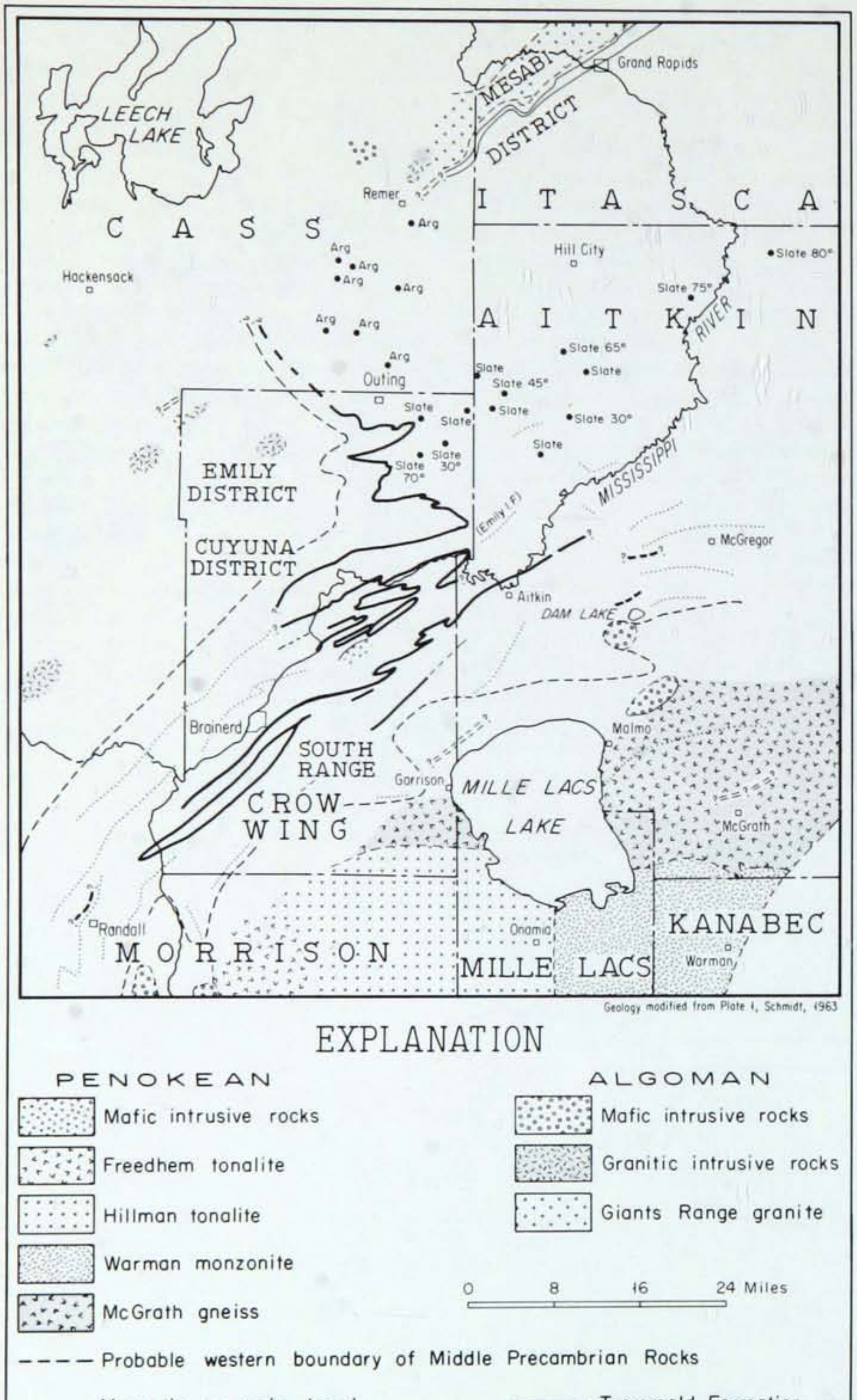

Figure IV-17. Generalized geologic map of the Cuyuna district and adjoining areas showing the locations of the Emily, North Cuyuna, and South Cuyuna ranges (modified from Schmidt, 1963).

CUYUNA DISTRICT

Ralph W. Marsden

The Cuyuna district is about 100 miles west-southwest of Duluth, in Aitkin, Cass, Crow Wing, and Morrison Counties, and is defined here to include the Emily, North, and South iron ranges, inasmuch as the rocks in the three areas are lithologically, stratigraphically, and structurally similar (fig. IV-1). The proposed boundaries for the Cuyuna district differ from those used by Schmidt (1963), but are consistent with those of Harder and Johnston (1918). The Emily range extends from the Mississippi River northward to the north line of Crow Wing County and into southern Cass County, and comprises an area of about 450 square miles. The North range includes the principal iron ore-producing area of the Cuyuna district in the vicinity of Crosby, Minnesota. The South range includes an area of northeast-ward-trending, generally parallel belts of iron-formation extending from near Randall northeast for about 60 miles. In addition to the three named ranges, several linear magnetic anomalies occur within the Cuyuna district that may indicate other, as yet unexplored, areas of iron-formation (fig. IV-17).

The geologic relations of the Precambrian rocks are obscured by a nearly complete blanket of glacial drift, which, together with local Cretaceous strata, is from 20 to 450 feet thick. Except in the mine areas of the North range, the geology of the Cuyuna district is pieced together from data obtained from a number of explorations for iron and manganiferous ores made during the past 70 years. Early exploration work consisted of magnetic surveys followed by

Table IV-3. Stratigraphic sequences in the Cuyuna district and westernmost Mesabi range.

	CUYUNA DISTRICT	*WESTERNMOST MESABI RANGE*
Pleistocene	Des Moines drift	Des Moines drift
	unconformity	unconformity
Upper Cretaceous	Coleraine Formation	Coleraine Formation
	unconformity	
Keweenawan ?	acidic volcanic rocks?	———
	unconformity	unconformity
Middle Precambrian		
Animikie Group		
	Rabbit Lake Formation	Virginia Formation
	Upper Member	argillite
	Emily Member	ferruginous slate and iron-formation
	Lower Member	argillite
	Trommald Formation	Biwabik Iron-formation
	Mahnomen Formation	Pokegama Quartzite
	possible unconformity	
pre-Animikie	Trout Lake formation	———
	slate and quartzite?	
	unconformity	unconformity
Lower Precambrian	granite and greenstone	granite and greenstone

drilling to determine the cause of magnetic anomalies. More recent work in the 1940's and 1950's utilized detailed gravity surveys to supplement airborne and ground magnetic surveys and drilling. The geologic studies show complex folding and some local faulting, marked lateral variations in the magnetic character and the facies of the iron-formations, and longitudinal variations in the lithology of rock units. These factors alone, in the absence of glacial drift and Cretaceous sediments, would require careful geologic mapping to show the geologic relationships. Much interpretation of limited information is required to give continuity to the geology, so the geologic maps represent an approximation of the distribution of formations and of the rock structures. Different interpretations of the geology can be expected if additional geological, geophysical, and exploration work is done.

This report has utilized data from all published and unpublished sources to which I have had access. In places where data are in conflict, I have exercised my judgment in accepting or rejecting interpretations and information; for example, some of the early reports appear to use the term "slaty iron formation" for oxidized, red-brown slate having a low iron content. The Trommald Formation is the main marker bed used in interpreting the structure and stratigraphy of the Cuyuna district. Inasmuch as iron-formations commonly are identified in drill samples without information on the associated rocks, an iron-bearing member that occurs in the Rabbit Lake Formation locally may be mistakenly identified as the Trommald Formation.

STRATIGRAPHY

The generalized sequence of Middle Precambrian rocks in the Cuyuna district is rather simple (table IV-3). It consists of a central iron-formation that is underlain by clastic strata and a dolomite formation and is overlain by clastic, locally carbonaceous strata that include an intercalated ferruginous slate and iron-formation unit. Details of the lithologic, stratigraphic, and sedimentational relationships within the formations are poorly known, for much of the available information is from geophysical surveys and exploration drilling designed to discover iron or manganiferous iron ores. As most drill holes were located to check magnetic or gravity anomalies or to outline areas of potential ore, drilling was done largely along the iron-formation zones. Determination of the stratigraphy was a secondary consideration.

The stratigraphic terminology defined by Schmidt (1963) for the Cuyuna North range can be applied with some modification throughout the district. The rock sequence, from oldest to youngest, includes (1) Lower Precambrian granites and greenstones, and (2) Middle Precambrian rocks consisting of a possible lower unit of clastic strata, a dolomite here informally termed the "Trout Lake formation," the Mahnomen Formation, the main iron-formation (Trommald Formation), and the Rabbit Lake Formation. The Rabbit Lake Formation includes a lower clastic and volcanic member, a ferruginous slate and iron-formation member here termed the "Emily member," and an upper slate, graywacke, and argillite member. The stratigraphic sequences in the Cuyuna district and the Mesabi range are shown in Table IV-3.

Knowledge of the pre-Middle Precambrian rocks in the Cuyuna district is limited. Lower Precambrian rocks (fig. IV-18) include granitic rocks exposed near the Pine River

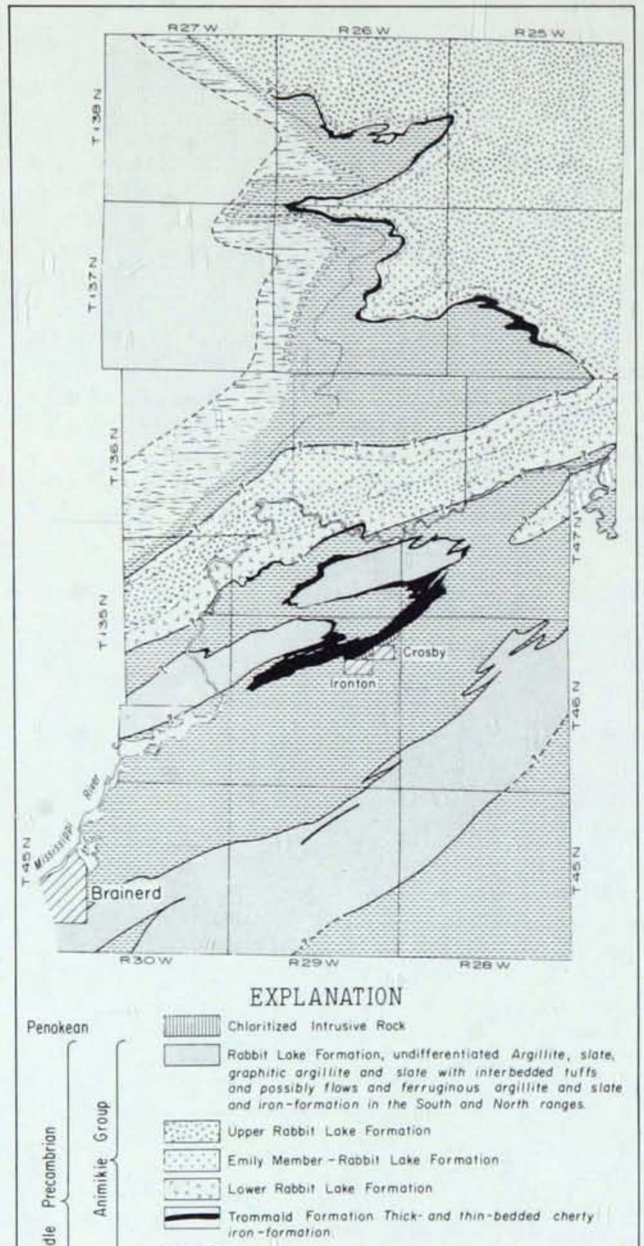

Figure IV-18. Geologic map of the Cuyuna district (compiled from various sources including Schmidt, 1963).

Table IV-4. Chemical analyses, in weight percent, of selected samples of dolomite from the Trout Lake formation.

Location	Drilled Thickness (in feet)	Depth (in feet)	CaO	MgO	CO_2	SiO_2
Sec. 6, T. 137 N., R. 26 W.	38	555-560	25.96	18.62		2.95
Sec. 31, T. 137 N., R. 26 W.	120	350-355	10.53	9.40	21.64	50.56
		445-450	13.55	18.08	36.12	14.86
Sec. 32, T. 138 N., R. 26 W.	124	396	28.72	12.98	43.87	4.20
		420	29.04	19.16	45.14	1.98
		420	29.86	18.76	45.96	1.40

and in the southeastern part of T. 138 N., R. 28 W., mafic intrusive rocks exposed in the central part of T. 138 N., R. 29 W., and greenstone exposed in the northern part of T. 136 N., R. 29 W. In addition, it is reported that a drill hole in the SW¼ NE¼ sec. 14, T. 137 N., R. 27 W. intersected greenstone, two drill holes in the NW¼ SE¼ sec. 35, T. 138 N., R. 27 W. cut medium- to fine-grained gabbro, and a drill hole in SE¼ sec. 10, T. 138 N., R. 27 W. cut schist. The widely scattered information suggests that many of the rock types found north of the Mesabi range also occur in the Lower Precambrian sequence of this region.

Trout Lake Formation

The name "Trout Lake formation" is used herein to designate a fine-grained lithographic, locally granular, massive, gray to buff, somewhat cherty dolomite. Dolomite was encountered in drilling in SW¼ NE¼ sec. 6 and SE¼ NW¼ sec. 31, T. 137 N., R. 26 W.; SE¼ NE¼ sec. 2, T. 137 N., R. 27 W.; NE¼ NE¼ sec. 32 and NW¼ NE¼ sec. 33, T. 138 N., R. 26 W., and in sec. 31, T. 138 N., R. 27 W. (fig. IV-19). A glacial boulder about 30 feet long and 15 feet high that is in the NW¼ SE¼ sec. 29, T. 137 N., R. 28 W. is similar to the dolomite cut in the drill holes and is the only known probable exposure of the dolomite in the region. The glacial erratic was first described by Harder and Johnston (1918, p. 63). Some of the dolomite contains siliceous layers and patches that have a granular texture and may be either quartzite lenses or recrystallized chert.

Chemical analyses of available samples (table IV-4) indicate that the Trout Lake formation varies from a dolomite to a dolomitic limestone. Analysis of one sample shows an excess of MgO, suggesting that magnesite may be present. The silica content of the dolomite ranges from 1.40 to 50.56 percent.

The distribution, thickness, and position of the Trout Lake formation relative to adjacent formations are poorly known, for none of the drill holes that penetrated the dolomite cuts either the top or the bottom of the formation. Judged from drill hole and gravity information, the dolomite possibly has a strike length of more than 15 miles. A stratigraphic thickness in excess of 90 feet is indicated by drilling; the total thickness is unknown, but judged from gravity data probably is considerably greater than drilling indicates.

The Trout Lake formation lies beneath the Mahnomen

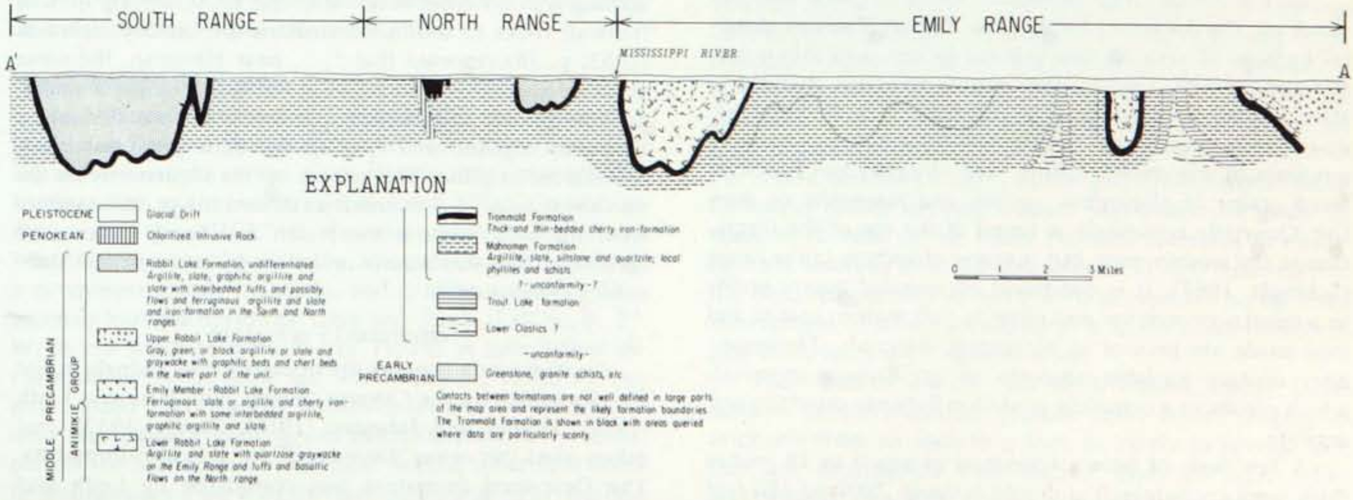

Figure IV-19. General north-south cross-section showing the inferred structure of the Cuyuna district. Location of cross-section shown in Figure IV-18.

Formation and appears to be underlain by slate and quartzite. The latter observation is based on limited information from drill holes in the SW¼SE¼ sec. 35 and in the NW¼SE¼ sec. 36, T. 136 N., R. 27 W., which penetrated rocks described as slate, quartzite, and paint rock. The dolomite is inferred from relationships of similar lithologic units observed on the Penokee and Gogebic ranges in Wisconsin and Michigan to lie unconformably beneath the Mahnomen Formation, but without marked discordance. In Wisconsin and Michigan, the Middle Precambrian Bad River Dolomite is separated from the overlying Palms Quartzite by an unconformity that lacks discordance. The unconformity between the units marks a period of erosion during which the Bad River Dolomite and the underlying Sunday Quartzite were completely removed over large areas. The strikes and dips of the formations above and below the unconformity, however, are similar, suggesting uplift and erosion without marked deformation. A similar interpretation of the relationship between the Trout Lake formation and the Mahnomen Formation is tenable, for the known structural trends and projected distribution of dolomite, based on gravity data, indicate that the dolomite has the lateral continuity shown in Figure IV-18. Although available information does not require an unconformable relationship with the Mahnomen Formation, this interpretation is preferred inasmuch as it is consistent with the known Middle Precambrian history in Wisconsin and Michigan.

Mahnomen Formation

The Mahnomen Formation consists of a clastic sequence that includes argillite, slate, siltstone, graywacke, and quartzite, and lesser amounts of schist, phyllite, and limestone. The formation was named by Schmidt (1963) to include the thick clastic sequence that underlies the Trommald Formation in the North Cuyuna range. Similar clastic rocks underlie the main iron-formation in the South Cuyuna and Emily ranges. In the Emily and North range areas, the formation appears to have a minimum thickness of 1,000 feet and a maximum thickness of at least 2,000 feet. A similar range in thickness is inferred in the South range.

In the North range (Schmidt, 1963), argillite and siltstone are the dominant lithologies. The argillaceous material consists of sericite, fine-grained quartz, and minor biotite and chlorite. Locally, the argillite has been changed to slate, schist, or phyllite. The siltstone is a fine-grained, quartz-rich, massive or bedded rock that contains variable amounts of muscovite, biotite, and chlorite, and has scattered grains of plagioclase, pyrite, and magnetite or martite. Quartzite commonly is found at the top of the formation in the western part, but is sparse elsewhere in the range (Schmidt, 1963). It is composed of rounded quartz grains in a quartz- or sericite- and chlorite-rich matrix; calcite and iron oxide are present as cementing materials. The quartzites contain variable amounts of argillaceous material, which results in a complete gradation between quartzite and argillite.

A few beds of brown limestone as much as 18 inches thick were encountered at depths between 200 and 450 feet below the top of the Mahnomen Formation in a drill hole in sec. 20, T. 47 N., R. 28 W. Apparently this rock type has a local distribution, for it was not encountered in drilling elsewhere in the Mahnomen Formation.

In the Emily range, the Mahnomen Formation consists of argillite-quartzite, siliceous argillite, slate, siltstone, and quartzite. Argillite-quartzite, the most abundant rock type, consists of intercalated layers of argillite and fine-grained quartzite. Commonly, argillite and quartzite are present in about equal amounts, but in some parts of the formation argillite is dominant, whereas in others quartzite is dominant. Elsewhere in the Lake Superior region, this lithology has been termed "quartz-slate." Where fresh, the rock is light gray to light greenish gray, but where oxidized it ranges in color from yellow and brown to brick red. The rock generally has a prominent lamination, but is massive where either argillite or quartzite is dominant. Commonly, the argillitic layers have a slaty cleavage and the coarse-grained quartzite layers are cross-bedded.

The argillite is composed of about 70 percent fine-grained quartz and sericite and 30 percent rounded, medium-size quartz grains. In the quartzite layers, sand-size quartz grains are dominant in a sericitic matrix containing associated scattered grains of magnetite or martite and pyrite. Magnetite is relatively abundant in an argillite and quartzite unit in the lower part of the Mahnomen Formation, possibly about 1,000 feet below the upper contact. A prominent linear magnetic anomaly that can be traced for 14 miles is associated with this unit. A continuation of this unit may be the cause of a magnetic anomaly that trends from sec. 20, T. 138 N., R. 27 W. to sec. 35, T. 140 N., R. 28 W. (fig. IV-18). Thus, except where the Mahnomen Formation is oxidized, a magnetite-bearing marker zone may extend throughout the Emily range from near Lower Mission Lake in T. 136 N., R. 27 W. to near Wabedo Lake in T. 140 N., R. 28 W., a strike distance of about 45 miles. The possible occurrence of a similar magnetite-bearing argillite and quartzite unit on the North range is suggested by magnetic anomalies in those parts of the range that appear to be underlain by the Mahnomen Formation.

Little reliable information concerning the Mahnomen Formation is available in the South range. As most of the drilling was done between 1900 and 1920, descriptions of footwall rocks to the iron-formation are sketchy. Schmidt (1963, p. 38) reported that ". . . near Hassman, the same general sequence was found as in the North range: a weakly metamorphosed thin-bedded iron-formation is underlain by light-gray argillite and overlain by dark gray and black titaniferous argillite. The bottom of this sequence is on the northwest side. . . ." Schmidt examined many drill samples from the area between the North and South ranges and concluded that this area is underlain by rocks of the Mahnomen Formation.

Trommald Formation

Geologic data support the occurrence of a single major iron-formation in the Cuyuna district. Van Hise and Leith (1911), Harder and Johnston (1918), Zapffe (1933), and others used the name "Deerwood" for this iron-formation. The Deerwood formation was considered by Leith and others (1935) to be an iron-bearing member of the Virginia Formation. Schmidt (1963) concluded, however, that the iron-formation in the North range should have formational status, and proposed the name "Trommald Formation" for the iron-formation found near Trommald in Crow Wing County.

Thickness

The Trommald Formation ranges in thickness along strike from about 45 to 500 feet in the North range and from about 10 to 600 feet in the Emily range. Data on the thickness on the South range are limited, but a maximum of about 300 feet seems likely. Concerning the thickness of the Trommald Formation in the North range area, Schmidt stated (1963, p. 31), "Where only the thick-bedded facies is present, the iron-formation is thinner, and it is possible that the Trommald Formation wedges out entirely if traced westward from the North Range. The formation is thickest where only the thin-bedded chert-carbonate-siderite-magnetite facies was deposited."

A direct relationship between iron-formation thickness and facies also exists on the Emily range. Where only the thick-bedded cherty iron-formation occurs, the Trommald Formation is thin. The Trommald Formation is reported to be from 65 to 100 feet thick in sec. 17, T. 138 N., R. 26 W., and to be generally thin in secs. 10, 11, 15, and 20, T. 137 N., R. 26 W. The iron-formation appears to be thin along the southern edge of the Emily range, from near Lower Mission Lake, T. 136 N., R. 27 W. to sec. 2, T. 136 N., R. 25 W. The iron-formation in this area is projected to follow an indistinct magnetic trend and a weak gravity anomaly, and was cut by only two drill holes. A drill hole in lot 4, sec. 24, T. 136 N., R. 27 W. cut 59 feet of iron-bearing strata (Emily member) and 312 feet of gray, graphitic slate (both of the Rabbit Lake Formation), 11 feet of chert-carbonate iron-formation (Trommald), and 39 feet of interbedded quartzite and argillite. The sequence dips about 40°. Another drill hole in SW¼NW¼ sec. 9, T. 136 N., R. 25 W. cut 126 feet of thin-bedded graywacke, 45 feet of cherty iron-formation (Trommald), 19 feet of gray argillite, and 47 feet of massive quartzite. These drill holes penetrated thin Trommald Formation that typically is overlain by graywacke, graphitic slate, and the Emily member, and is underlain by quartzite and argillite.

North and Emily Ranges

Iron-formation that is correlated with the Trommald Formation has been traced by geophysical surveys and drilling north from the mines on the North range through the Emily range to the north line of Crow Wing County, as shown on Figure IV-17. It may extend north from Crow Wing County into Cass County along the northeast side of a prominent linear, magnetic, and gravity anomaly. This anomaly extends northwest from sec. 32, T. 139 N., R. 27 W. to just south of Wabedo (fig. IV-18). A hole drilled on the anomaly in the SE¼ sec. 11, T. 139 N., R. 28 W. cut thin-bedded gray argillite, which tentatively is correlated with the magnetic argillite and quartzite zone in the lower part of the Mahnomen Formation. Several other drill holes in southeastern Cass County have penetrated gray argillite that is considered to be a part of either the Virginia or the Rabbit Lake Formation. Although the Biwabik and Trommald formations are considered equivalent and possibly continuous, no attempt is made in Figure IV-17 to project the iron-formation into southeastern Cass County.

The Trommald Formation can be divided into five distinct facies: (1) thin-bedded; (2) thick-bedded, commonly granule-textured and cherty; (3) mixed thin- and thick-bedded; (4) algal chert; and (5) quartzitic iron-formation. Of these, the thin- and thick-bedded facies comprise a significant proportion of the iron-formation, and are lithologically similar to "slaty iron-formation" and "cherty iron-formation," respectively, on the Mesabi range. Schmidt (1963) gave a detailed description of the "thin-bedded" and "thick-bedded" units.

Thin-bedded Facies. The thin-bedded facies of the Trommald Formation in the North range is characterized by having individual bedding laminae that are less than half an inch thick, as well as having many laminae that are less than an eighth of an inch thick. Chert beds that are more than a quarter of an inch thick are common; some chert lenses are several inches thick and sparse beds are as much as 10 feet thick.

Unoxidized, thin-bedded iron-formation consists of iron- and silica-rich layers that range in color from gray to dark gray and greenish gray; the darker layers have a higher silicate content. The rock may be characterized as carbonate-chert, carbonate-silicate, silicate-carbonate, or as magnetite-rich iron-formation. The unoxidized, thin-bedded facies is composed of quartz, siderite, magnetite, stilpnomelane, minnesotaite, and chlorite. Grunerite, acmite, and tourmaline occur locally. Chemically, the thin-bedded facies commonly contains 10 to 15 percent CO_2, indicating a content of 25 to 40 percent siderite. Silica generally ranges from 30 to 35 percent and attains a maximum of 59.3 percent. Most of the silica is cherty quartz or is in iron silicates; some is present as clastic quartz. Manganese ranges from 2 to 5 percent, and probably occurs in carbonate minerals combined with iron, calcium, and magnesium.

Oxidation and leaching have resulted in a marked change in the physical appearance as well as in the chemical composition of the thin-bedded facies. Siderite and iron-silicates are converted to goethite and hematite, and less commonly to fine-grained manganite, pyrolusite, and cryptomelane. Much or all of the ferrous iron is oxidized to ferric iron, yielding a rock that is yellow, brown, or red and has a weathered appearance. As a result of leaching, the rock is depleted in CO_2, Ca, Mg, SiO_2, and is enriched in Fe, Al_2O_3, and H_2O relative to the unaltered material. Iron-formation that is strongly leached may attain ore grade by removal of most of the silica. Porosity increases as oxidation and leaching proceed, but Schmidt (1963) reported that the observed porosity is less than that expected solely from the removal of silica; accordingly, it seems likely that some iron oxide has partly replaced silica.

Thick-bedded Facies. The thick-bedded, cherty facies consists of wavy-bedded, granule chert layers ranging in thickness from an inch to a foot, or rarely to several feet, that are intercalated with thinner iron-rich layers. The wavy-bedded character results from variations in the thickness of individual layers. Schmidt (1963) reported that the cherty layers commonly are lenticular. They are composed of ovoid granules of cherty quartz, carbonate, silicate, or magnetite and, where oxidized, of cherty quartz and iron oxides in a matrix of the same composition. The granule cherts are similar in mineralogy and in texture to the granule taconites of the Mesabi range.

In the Emily area, cherty iron-formation occurs that is similar to the thick-bedded facies on the North range, but unoxidized, thick-bedded iron-formation has not been found. This apparent lack of unoxidized iron-formation possibly reflects the relatively shallow drilling. The thick-bedded, cherty facies commonly occurs as two members that are separated by a thin-bedded or by a mixed thin- and thick-bedded facies. This occurrence of thick-bedded iron-formation in two zones differs from the North range, where only one thick-bedded facies occurs. In addition, it always overlies the thin-bedded facies. In both areas, where the iron-formation is thin it consists only of the thick-bedded facies.

Mixed Thin- and Thick-bedded Facies. A transitional lithology of intercalated thin- and thick-bedded iron-formation occurs locally in the central Emily range, between Ruth Lake and Ross Lake. Commonly, this unit is 50 to 150 feet thick and is gradational into thin-bedded iron-formation below and thick-bedded iron-formation above. The facies is characterized by 1- to 6-inch-thick granule chert layers and layers of fine-grained, laminated iron-formation. In secs. 20 and 21, T. 138 N., R. 26 W., the mixed facies is 100 to 165 feet thick, is between quartzitic facies below and thick-bedded facies above, and contains well-rounded, detrital quartz grains. The mixed facies containing sand grains appears to be laterally equivalent to non-clastic, fine-grained, and mixed facies iron-formation. The occurrence of the mixed facies iron-formation suggests that both the thick- and the thin-bedded facies may have been deposited under generally similar sedimentational conditions and that both could have been deposited in areas where sand-size clastic material was being deposited.

Quartzitic Facies. An iron-rich quartzitic facies of the Trommald Formation occurs west of Ruth Lake in secs. 20, 21, 22, and 23, T. 138 N., R. 26 W. This facies consists of abundant clastic quartz in a matrix of cherty quartz, iron oxides, and manganese oxides. Megascopically, the material appears to be a quartzite that has small amounts of chert and jasper as nodules and layers. An algal chert zone 5 to 10 feet thick occurs in the upper part of the unit, and may be equivalent to part of the lower, thick-bedded member elsewhere in the Emily range. Oolites and pisolites occur with jaspery chert nodules and layers in the lower part of the quartzitic unit. The occurrence of chert as nodules and layers with algal chert in sandy beds containing iron and manganese oxides suggests a gradual change from clastic deposition to iron-formation deposition. The rock is oxidized and leached so that the nature of the original sediment is not known. However, the quartzitic facies appears to be similar to the sandy facies at the base of the Biwabik Iron-formation in the Eveleth area on the Mesabi range (White, 1954), where interbedded chloritic sandstone, jaspery chert, and algal chert occur.

Algal Chert Facies. Beds of algal chert as much as 10 feet thick occur locally on the Emily range at the base of the lower thick-bedded cherty member. The algal chert facies consists of algal, jaspery chert, and in secs. 20, 21 and 23, T. 138 N., R. 26 W. it contains ubiquitous clastic quartz. The algal chert appears to be similar in character and to occur at the same stratigraphic position as the lower algal chert layer on the Mesabi range.

Iron-formation on the South Range

The main iron-formation of the South range is considered a part of the Trommald Formation. The unoxidized material was described by Harder and Johnston (1918) as a medium- to fine-grained, greenish-gray to black, commonly laminated "magnetitic slate" and "amphibole magnetite rock," which appears to represent metamorphosed thin-bedded facies. At the Adams mine, the iron-formation is intruded by a dark green, coarse-grained diabase, and is interlayered with green chloritic schist. Harder and Johnston (1918, p. 162 and 163) stated, "For a short distance, the main drift is in diabase beyond which it penetrates in succession a layer of amphibole-magnetite rock and magnetitic slate, a layer of green schist and then another layer of manetitic slate and amphibole-magnetite rock of considerable thickness. . . . The typical magnetitic slate consists of interlaminated light-green, finely crystalline amphibole and black, siliceous or argillaceous, fine-grained magnetite. . . . The proportion of chert or quartz present is small in the magnetitic slate and with increasing siliceous material and coarser layering the slate grades into amphibole-magnetitic rock." With respect to the amphibole-magnetite rock of the Cuyuna district, which includes the North range, Harder and Johnston (1918, p. 117-118) stated, "The typical amphibole-magnetite rock of the Cuyuna district is a finely-banded rock consisting of alternating bands of magnetite and amphibole with a minor amount of quartz." On the South range, the amphibole-magnetite rock is a greenish-gray to black, laminated rock that may contain layers of dark green to black chert or argillaceous, iron-rich layers several inches thick.

Oxidation and leaching of the iron-formation results in a bedded, limonitic or hematitic ferruginous chert, ferruginous slate, paint rock, and yellow, brown, red, or black ore.

Rabbit Lake Formation

The name "Rabbit Lake Formation" was proposed by Schmidt (1963, p. 11) for a thick sequence of gray to black argillite, graywacke, iron-formation, and ferruginous slate that overlies the Trommald Formation at Rabbit Lake. It is variable in lithology, and is known only in a general way because most available information is from short drill holes and mine exposures on the North range. Information on the rocks in the upper part of the formation is particularly sparse. Three informal members are recognized: (1) a lower member that includes argillite, slate, and graywacke and, on the North and South ranges, tuffaceous sediments, tuffs, and flows; (2) a ferruginous slate-iron-formation member, which is herein termed the "Emily member"; and (3) an upper member consisting of argillite, slate, graphitic slate, and graywacke.

Lower Member

On the North range, the lower member of the Rabbit Lake Formation consists of argillite, slate, and tuffaceous beds and, in the western part, also of local basalt flows (Schmidt, 1963). At the Maroco mine in secs. 3 and 4, T. 46 N., R. 29 W., 250 feet of chloritized basalt either directly overlie the Trommald Formation or are separated from it by a few inches to 20 feet of gray slate. Similar chloritized basalt occurs throughout much of the southwestern part of the North range, west and northwest of Ironton, where it is separated from the Trommald Formation by 10 to 40 feet of slate and argillite. Associated with the basalts, and dispersed elsewhere in the lower 100 feet of the Rabbit Lake Formation, are bedded, gray-green chloritic tuffs, locally interlayered with gray and black argillite or slate. The tuffs are composed of flattened and elliptical or spindle-shaped grains that range from fragments half an inch across to silt- and clay-size particles. Graded bedding occurs but cross-bedding is lacking.

In the Emily range, a quartzose, argillite-graywacke unit overlies the Trommald Formation. Commonly, the argillite-graywacke member is 200 to 300 feet thick, but it is as much as 500 feet thick and locally may be missing. It is thin (25 feet thick) in sec. 3, T. 137 N., R. 26 W. and probably is locally absent in secs. 20 and 21, T. 137 N., R. 25 W. In areas where drilling suggests its possible absence, problems of poor sample recovery were encountered, which made identification of the rocks overlying the Trommald Formation uncertain. Where unoxidized, this lower unit of the Rabbit Lake Formation is a gray to light gray, thin-bedded to massive, fine-grained to locally conglomeratic rock composed of 10 to 50 percent sand-size quartz grains, 50 to 80 percent matrix consisting of sericite and fine quartz, 1 to 5 percent chert and iron-formation fragments, less than 1 percent feldspar, and variable but small amounts of iron oxides and other materials. Some thin conglomeratic layers contain abundant fragments of chert and iron-formation. Locally, coarser grained layers are cross-bedded.

The common occurrence of fragments of chert and iron-formation in the argillite-graywacke unit indicates a source area containing cherty iron-formation. Possibly, the detritus was derived from erosion of the Trommald Formation in areas adjacent to the Emily range. If such an erosional surface extended into the Emily area, it could explain in part the observed variations in thickness of the Trommald Formation and the argillite-graywacke unit in that area. On the other hand, areas of thin Trommald Formation coincide with the thick-bedded facies, and accordingly it seems likely that the thickness of the iron-formation is in part a direct result of sedimentation. The suggested erosion episode, after the deposition of the Trommald Formation, may coincide with the time of volcanism in the North and South ranges.

Emily Member

A ferruginous slate and iron-formation unit within the Rabbit Lake Formation was encountered in many drill holes in the Emily range, and following Zapffe's (1933) original designation is termed herein the "Emily member." Zapffe used the term "Emily member" for the rocks in the lower part of the rock sequence now assigned to the Rabbit Lake Formation. He stated (Zapffe, 1933, p. 76), "Iron-bearing lenses in the Emily member are numerous but appear to be scattered and short. Although some are little oxidized, most of them are heavily oxidized and one near Emily Village is almost rich enough to constitute an ore deposit." Even though Zapffe may have included segments of Trommald Formation in the Emily member, his intent was to recognize an iron-rich unit in what is now termed the "Rabbit Lake Formation."

The Emily member constitutes a distinct stratigraphic unit throughout the Emily range and extends, at least locally, into the North range and possibly into the South range. In the Emily range, the thickness and extent of the Emily member justifies its being designated a formation. However, because of its rather indistinct boundaries, the presence of intercalated argillite, slate, and graphitic slate, and the possible lenticularity of the cherty iron-formation, the unit is included as a member of the Rabbit Lake Formation. The thickness of the Emily member is uncertain; several drill holes have penetrated 200 or more feet of iron-rich strata, and it locally may be more than 1,000 feet thick in the central part of the Emily range. In this area, shallow drilling indicates that the Emily member underlies an area a mile or more wide, perhaps as a consequence of thickening by complex folding.

The Emily member consists of gray to black, fine-grained, locally graphitic, iron- and carbonate-rich iron-formation having sparse but characteristic white or light gray chert in pea-size nodules and one-eighth- to three-inch-thick beds. Some beds can be designated iron-rich, graphitic argillite or slate. Intercalated argillite, graphitic argillite, and slate commonly contain minor disseminated grains and veinlets of pyrite. In the Ross Lake area, a 100-foot-thick section contains about 10 percent pyrite.

In thin section, the iron-formation is seen to consist of at least 50 percent fine-grained cherty quartz and 10 to 40 percent iron carbonate. In sec. 2, T. 137 N., R. 25 W., there is a siliceous, dolomitic limestone that appears to be equivalent to the Emily member. Chemically, the member commonly contains from 10 to 20 percent iron, about 50 percent silica, 1 to 3 percent manganese, as much as 6 percent alumina, and 0.05 to 0.4 percent phosphorus. Both the manganese and phosphorus vary in amount, and neither can be used to characterize the iron-formation. Much of the Emily member that has been penetrated on the Emily range is oxidized, but it approaches ore grade only locally. Commonly, the upper 50 to 100 feet is oxidized, but in some areas oxidation appears to extend considerably deeper. Commonly, the oxidized material is brown, red, or brownish gray, and contains goethite and hematite.

Iron-formation lenses as much as several hundred feet thick and which are continuous for several miles occur within the Rabbit Lake Formation in the North range, and were referred to by Schmidt (1963) as the "upper iron-formation." There is little doubt that these lenses are stratigraphically about 500 to 2,000 feet above the base of the Rabbit Lake Formation; only thin lenses or beds of iron-formation occur in the lower part of the formation.

The iron-formation in the Emily member differs from the Trommald Formation by having gradational contacts, by commonly containing interbedded argillite or slate, and by generally being very siliceous. Where the Emily member is less argillaceous and siliceous, it is similar to the thin-bedded facies of the Trommald Formation.

The Emily member may be present in the South range, but available descriptions of iron-formation and associated hanging wall and footwall rocks leave major uncertainties as to possible correlations. However, near Clear Lake in secs. 28 and 29, T. 46 N., R. 25 W. and south of Dam Lake in secs. 9, 19, 28, 29, and 30, T. 46 N., R. 25 W., graphitic slates, pyritic slates, and cherty rocks are reported (fig. IV-17). Inasmuch as graphitic slates are not mentioned in descriptions of other South range rocks, but do occur in the Rabbit Lake Formation in the Emily and North ranges, these rocks may represent a sulfide-rich facies that is equivalent to the Emily member.

Upper Member

A very thick succession of clastic strata apparently overlies the Emily member, and here is referred to as the "upper member." Widely scattered, shallow drilling in the area east of the Emily range and northeast of the east end of the North and South ranges indicates that argillite or slate underlies a very large area. These rocks are described as gray, green, or black argillite, gray and green slate, graphitic slate and, rarely, as graywacke. Commonly, they are thin-bedded, locally massive, fine- to medium-grained, and contain sparse pyrite. Chert beds, intercalated with beds of graphitic slate and argillite, seem to be characteristic of the lower part of the member.

Drilling in Cass County, south and southwest of Remer, in Ts. 139, 140, and 141 N., Rs. 26, 27, and 28 W., encountered gray to light gray, very fine- to medium-grained, thin-bedded argillite that has sparse pyrite and lacks graphite. Drilling in T. 136 N., Rs. 25, 26, and 27 W. and T. 137 N., Rs. 25 and 26 W. also encountered fine-grained clastic rocks, including thin-bedded gray slate or argillite, massive graphitic-pyritic argillite or slate, and greenish-gray argillite and slate. Some of these rocks are reported to be cut by veinlets of carbonate, pyrite, and "asbestos."

IGNEOUS ROCKS

Chloritized igneous rocks of both extrusive and intrusive origin occur on both the North and South ranges but have not been reported from the Emily range. On the North range, the volcanic rocks include basaltic flows and tuffaceous strata that are in the lower member of the Rabbit Lake Formation; on the South range, rocks of similar lithology are present. However, in the area southeast of Brainerd, probably younger felsic flows are reported to overlie folded and eroded rocks of the Animikie Group. Mafic dikes and sills locally cut the Trommald Formation and adjacent Mahnomen and Rabbit Lake Formations.

Volcanic Rocks

Conformable beds of chloritized basalt and laminated, gray to green and dark green chloritic tuff, tuffaceous slate, and gray-green schist, as much as 300 feet thick, have been reported by Schmidt (1963) from the lower part of the Rabbit Lake Formation. The basalt is similar in chemical composition to the average Hawaiian basalt. Both the tuffaceous strata and basalt commonly contain 1.0 to 4.0 percent titania. Mineralogically, they are composed of a fine-grained intergrowth of chlorite and lesser clinozoisite, calcite, and leucoxene. Schmidt (1963) considered the basaltic rocks as lava flows, although he recognized that they might be sills. The tuffaceous rocks contain chloritic fragments that are half an inch or more in size, and appear to represent water-laid pyroclastic material. Chlorite schists are reported in the South range, and may represent altered basalt and tuff. Schmidt (1963, p. 38) reported the occurrence of dark gray to black titaniferous argillite overlying the Trommald Formation on the east end of the South range near Hassman. It seems likely that these rocks are similar to those described from the North range.

Van Hise and Leith (1911, p. 215) reported the occurrence of volcanic rocks at three localities south and east of Brainerd on the South range that overlie and probably are younger than the eroded Middle Precambrian strata. They stated: "An acidic extrusive rock with amygdaloidal texture in beds 15 to 25 feet thick has been found by drilling to rest across the edges of the Virginia slate and Deerwood iron-bearing formation member in section 2, T. 44 N., R. 31 W., section 6, T. 44 N., R. 30 W.; and section 7, T. 45 N., R. 29 W." They suggested a Keweenawan age for these rocks. Harder and Johnston (1918) and Schmidt (1963) also referred to the presence of flat-lying, volcanic rocks of possible Keweenawan age.

Intrusive Rocks

Basic intrusive rocks occur in the North and South Cuyuna ranges. Schmidt (1963, p. 41) reported that dikes of intermediate or mafic composition are locally abundant in an extensive belt along the southeastern side of the North range. He stated: "All the rock in the belt is thought to be the same general intrusive body, although specimens obtained from different places appear to have been derived from the metamorphism of diorite and gabbro. All specimens examined were thoroughly chloritized, and some were slightly or extensively sheared. The alteration products are masses of fine intergrown minerals, the relative volumes of which cannot be readily estimated by grain-count method. The common minerals are chlorite, epidote, clinozoisite, albite or oligoclase, calcite and sphene or leucoxene."

Intrusive rocks may be more common on the South range than on the North range. For example, Harder and Johnston (1918, p. 123) reported intrusive rocks from the Adams mine, sec. 30, T. 46 N., R. 28 W., and the Barrows mine, sec. 10, T. 44 N., R. 31 W. They described these rocks as altered diorite, diabase, and gabbro that have a granular or ophitic texture. Harder and Johnston (1918, p. 162) described the intrusive rocks at the Adams mine as follows: "The diabase through which the shaft passes is dark green and consists of a groundmass of fine grained, dark green chlorite, in which are abundant, long, narrow laths of pink feldspar. . . . Near the bedrock surface the diabase is light brownish green and thoroughly decomposed. On nearing the contact of the magnetic slate along the main drift, the diabase loses its porphyritic texture and becomes very fine-grained. However, it retains a fine ophitic texture. This clearly indicates that the rock is intrusive." Intrusive rocks also occur in: (1) sec. 17, T. 45 N., R. 28 W., south of Clearwater Lake; (2) Long Lake (or Lac Wiben), T. 46 N., R. 25 W.; (3) secs. 9, 17, and 19, T. 45 N., R. 28 W.; (4) secs. 19, 29, and 30, T. 46 N., R. 25 W.; and (5) on both sides of the iron-formation in sec. 17, T. 45 N., R. 28 W. These rocks generally are referred to as greenstone, diorite, chlorite schist, basic intrusives, or altered basic intrusives; the intrusion near Long Lake was called "basic hornblende gneiss" by Thiel (1947, p. 47).

The intrusive rocks in the Cuyuna district have not been dated by radiometric methods, but their geologic relationships suggest that they are Penokean in age. The rocks are intrusive into strata assigned to the Animikie Group, and they were altered and sheared with these strata during the deformation assigned to the Penokean.

CRETACEOUS ROCKS

Exploration drilling in northern Crow Wing and southern Cass Counties in the Emily range encountered from 50 to 280 feet of flat-lying, well-bedded, light-colored shale or mudstone beneath glacial drift and above the Precambrian bedrock that locally contains lignitic material. In the North and South range areas, Van Hise and Leith (1911, p. 215) and Harder and Johnston (1918) reported the occurrence of a ferruginous conglomerate, which locally contains iron-formation pebbles in a shaly matrix; it lies on Animikie rocks. The conglomerate was well exposed in an exploration shaft in the SW¼SE¼ sec. 8, T. 45 N., R. 25 W., east of Brainerd. Although no fossils were found, it seems probable that the conglomerate, shale, and mudstone are of Cretaceous age.

STRUCTURE

The structural pattern of the Middle Precambrian rocks in the Cuyuna district is dominated by folds that trend N. 55°-65° E. The South range appears to be characterized by isoclinal folds, the North range by doubly-plunging, tight folds, and the Emily range by open eastward-plunging folds (fig. IV-18). This pattern indicates that the intensity of deformation decreased from south to north. Because most of the structural relationships on the North range are a result of folding and the geologic relationships on the Emily and South ranges can be explained by folding, faulting is considered relatively unimportant. Faulting may be more common than generally recognized, however.

Because of the sparse knowledge of the stratigraphy, only the general structural configuration of the South range is known. The structure is strongly linear, possibly indicating isoclinal folds, and is dominated by a large northeastward-trending syncline that extends from northeast of Randall to beyond Hassman, a distance of about 60 miles. Magnetic trend lines over the Trommald Formation appear to define the flanks of the synclinal axis (fig. IV-19). The south side of the syncline is inferred to be south of Clearwater Lake, and marked by an iron-formation that extends from sec. 18, T. 46 N., R. 27 W. to sec. 26, T. 45 N., R. 29 W. The north side of the syncline is complicated by a second-order anticline that brings iron-formation to the bedrock surface in secs. 2, 9, and 10, T. 45 N., R. 29 W., secs. 26, 27, 32, and 33, T. 45 N., R. 30 W., and southwestward into T. 44 N., Rs. 30 and 31 W. Additional second-order folds are suggested by the distribution of iron-formation in the area embracing sec. 36, T. 46 N., R. 29 W. and secs. 2 to 16, T. 46 N., R. 28 W., where the iron-formation is compressed and crumpled. Within the major syncline, bedding is nearly vertical or overturned along the northern belt of iron-formation, and the strata dip northward near Clearwater Lake. Probably, the syncline has a nearly horizontal axis and a northwestward axial plane.

The North range is characterized by a complex fold pattern consisting of several doubly-plunging anticlines and synclines and abundant second-order folds (figs. IV-18 and 19). It appears to be separated from the South range by a major anticline. Schmidt (1963, p. 45-46) described the rocks as being ". . . tightly folded into several large doubly-plunging folds. Bedrock surrounding most of the North range is Mahnomen formation, the oldest of the three formations. The Trommald and Rabbit Lake formations occur as irregular ellipses and long narrow areas that are synclines flanked by and underlain by Mahnomen formation. The axes of the major folds are not ordinarily mappable, except where indicated by the areal pattern of the Trommald and Rabbit Lake formations and only the synclines containing Trommald and Rabbit Lake formations stand out as distinct folds; the intervening anticlinal areas lack distinctive form, and their major axes can generally be only approximated. The bedrock pattern of the district is, in effect, one of major synclines and relatively inconspicuous major anticlines. . . . The axial planes of almost all the folds dip steeply southeastward, and the southeast limbs of the synclines are overturned in most places. . . . Drag folds are abundant in all sizes and bear a normal and systematic relation to the principal folds. . . . No major faults were definitely identified in this study of the Cuyuna district although two have been suspected on the basis of stratigraphic and structural evidence. Both are perhaps a mile or less in length. . . . Many minor faults were observed during the detailed mapping of the district. Some of them cut sharply across the strike of fold limbs but most were strike faults of little stratigraphic throw, probably not more than 20 to 30 feet. Several of the faults observed were related to the broken crests of anticlines."

The Emily range, to the north line of Crow Wing County, is characterized by a generally continuous but crinkled subcrop pattern of the Trommald Formation, which suggests rather open eastward-plunging folds. Because the Trommald Formation is the principal marker zone, its trend shows the configuration of the middle part of the Animikie strata. A less complex structural configuration is suggested for the lower part of the Animikie rocks by a prominent, linear, magnetic anomaly that extends from west of Lower Mission Lake in T. 135 N., R. 27 W. to sec. 10, T. 137 N., R. 26 W.; from this point, the anomaly can be traced less distinctly into Cass County (figs. IV-17 and 18). There is a marked difference in the tightness of the folds indicated by this anomaly and by the trend of the Trommald Formation in the northern part of T. 136 N. and the southern part of T. 137 N. The Trommald Formation is folded into a major anticline, whereas the unit giving the magnetic anomaly in the Mahnomen Formation extends north across the western end of the major fold without appreciable curvature. In the southern part of T. 137 N., however, the two structural markers are subparallel.

In the vicinity of the village of Emily, the strata are folded into a prominent syncline and anticline and dip nearly vertically in secs. 3, 4, 5, and 21, T. 137 N., R. 26 W. Along the south side of the Emily range, in Ts. 135 and 136 N., the subcrop of the Trommald Formation can be traced along strike for about 18 miles in a N. 65° E. to N. 80° E. direction. This trend is generally parallel to that of the northernmost iron-formation on the North range, and appears to define the north side of a major syncline.

Although somewhat contorted, the Trommald Formation trends northwestward from near the southwest corner of T. 137 N., R. 25 W. into Cass County. Because of the thick deposits of glacial drift and Cretaceous strata in this area, little drilling has been done to prove the extension of the Trommald Formation northwest from T. 138 N., R. 26 W. Judged from gravity and magnetic data, however, the western edge of the Animikie rocks extends to the vicinity of Wabedo Lake (fig. IV-18). Neither the extent of the Animikie rocks nor the trace of the iron-formation can be determined between Wabedo Lake and the westernmost end of the Mesabi range, a distance of about 20 miles, from the available magnetic and gravity data. Drilling has shown that argillite, possibly belonging to the Rabbit Lake or Virginia Formations, occurs beneath 350 to 450 feet of glacial drift and Cretaceous strata in the area between Remer and Wabedo Lake, suggesting that the north edge of Middle Precambrian strata probably lies northwest of a line between these two localities.

South of a line that trends N. 60° E. from Ruth Lake in the Emily range to T. 52 N., R. 23 W.—a distance of about 30 miles—the Rabbit Lake Formation is folded and commonly dips 65°-80°. Where penetrated by drilling, the rocks have a well-developed slaty cleavage. Northwest of the line, in southeastern Cass County, however, rocks considered to belong to the upper Rabbit Lake Formation dip gently and lack slaty cleavage. Thus, it seems probable that deformation a short distance north of the arbitrary, northeastward-trending line was not sufficiently intense to develop slaty cleavage. It is of interest that the trend of the fold axes in the Emily range area is subparallel to that of fold axes in the "Virginia horn" on the Mesabi range.

METAMORPHISM

The Middle Precambrian rocks of the Cuyuna district are metamorphosed to the chlorite zone of the greenschist facies, and some to the biotite or garnet zone. The general metamorphic grade is suggested by the widespread occurrence of biotite-chlorite schists and, at the Milford mine in sec. 23, T. 47 N., R. 29 W., by iron-formation that contains grunerite, which may be equivalent in rank to the garnet isograd. The common metamorphic minerals are chlorite and sericite; some biotite and magnetite occurs in the argillite and slate; magnetite, minnesotaite, stilpnomelane, recrystallized cherty quartz, and carbonate occur in the iron-formations; and chlorite, epidote, clinozoisite, albite or oligoclase, calcite, and sphene or leucoxene occur in the igneous rocks. The fine-grained clastic rocks of the Mahnomen and Rabbit Lake Formations have a metamorphic foliation commonly given by chlorite and sericite, and the rocks can be classified as slate, phyllite, or schist, depending on the degree of recrystallization. Coarser grained rocks are commonly massive and lack a megascopically visible preferred mineral orientation. Commonly, slate is reported to be present in the Cuyuna district as far north as the north line of Crow Wing County and in drill holes from Ruth Lake northeast to the east edge of T. 52 N., R. 23 W. In Cass County and on the Mesabi range, equivalent rocks lack a slaty cleavage, and are classed as argillites. Despite their marked deformation, rocks in the South range have approximately the same metamorphic rank as those farther north. Harder and Johnston (1918) used the term "amphibole-magnetite rock" or "magnetite slate" for parts of the iron-formation in this area, but inasmuch as the associated rocks were described as chlorite schists or slates, and the equivalent rocks in the North range were described in the same terms, it is probable that all the rocks belong to the greenschist facies.

Drilling near Clearwater Lake in secs. 9, 17, and 19, T. 45 N., R. 28 W. and in the Glen area in secs. 19, 29, and 30, T. 46 N., R. 25 W. encountered rocks that appear to be in the chlorite zone of metamorphism. An iron-formation at Clearwater Lake is associated with rocks described as gray slates, pyritic slates, diorites, and altered basic intrusions.

Near Randall, at the southwest end of the Cuyuna district, the iron-formation is associated with rocks classified as slates, schists, and diorites. The descriptions do not aid in classifying the metamorphic rank.

CORRELATION WITH THE MESABI RANGE

During the early part of this century, any correlation of rocks of the Cuyuna district and the Mesabi range was debatable inasmuch as the stratigraphic successions in the two areas were believed to be dissimilar. However, knowledge of the geology of the Cuyuna district was expanded greatly by exploration during the 1940's and 1950's in the Emily range, and as a result it was clearly demonstrated that the stratigraphic succession in the Emily range is similar to that of the westernmost Mesabi range (table IV-4). Results of the more recent work in the Emily range, summarized in this report, indicate that: (1) the Trommald Formation extends continuously from Lower Mission Lake in sec. 6, T. 135 N., R. 27 W. to east of Blue Lake in sec. 17, T. 138 N., R. 26 W. but ranges in thickness within that area from about 10 to possibly 600 feet; (2) the iron-formation (Trommald) is underlain by clastic strata of the Mahnomen Formation, which includes quartzite, slate, and argillite, and is overlain by the Rabbit Lake Formation, which includes a lower graywacke, slate and argillite member from 25 feet to 500 feet thick, the Emily ferruginous slate and iron-formation member, which possibly is more than 1,000 feet thick, and an upper argillite and slate member, the lower part of which is graphitic (table IV-4).

Rocks of the Animikie Group on the westernmost Mesabi range differ markedly in thickness and lithology from those on the main Mesabi range. These differences were described by White (1954), and are summarized by Morey (this chapter). The principal differences are: (1) a marked thinning of the Biwabik Iron-formation westward, from about 450 feet at Grand Rapids in R. 25 W., to 200 feet in the central part of R. 26 W., to 150 feet in R. 27 W., and to 20 feet in eastern Cass County; and (2) the occurrence on the westernmost Mesabi of an iron-bearing member in the Virginia Formation, which separates this formation into a lower argillite member, an iron-bearing member, and an upper argillite member. A drill hole in lot 2, sec. 21, T. 54 N., R. 27 W., near the Cass County line, penetrated the following section: surface to 260 feet, surficial materials; 260-301 feet, mixed quartzite and taconite; 301 to 367 feet, slate (argillite) and paint rock; 367-448 feet, taconite; and 448 to 459 feet, quartzite. This drill hole cut the iron-bearing member and the lower argillite member of the Virginia Formation, the entire Biwabik Iron-formation, and the upper 11 feet of the Pokegama Quartzite; the stratigraphic sequence is comparable to the sedimentary strata associated with the Mahnomen, Trommald, and Rabbit Lake Formations on the Emily range.

Commonly, it is assumed that the Pokegama Quartzite on the westernmost Mesabi range is relatively thin and similar in lithology to that farther east on the range. The main difference between the Pokegama and the Mahnomen formations, which are clastic units that contain quartzite, appears to be the thickness of the units.

Sparse information from drill holes in southeastern Cass County suggests that the Virginia and Rabbit Lake Formations continue throughout the area between the Mesabi range and the Cuyuna district.

MINERAL DEPOSITS

The Cuyuna district is a part of the major iron province of the Lake Superior region. It differs from other Lake Superior iron districts in having extensive layers of unoxidized and unleached iron-formation that contain 2 to 6 percent manganese and in producing ores that commonly are manganiferous. Since the first ore was shipped in 1911, approximately 104 million gross tons have been mined (Alm and Trethewey, 1970). The reserves of natural ore have decreased markedly, and the annual production has declined from about 3.6 million gross tons in 1953 to 474,000 gross tons in 1970. Except for 913,690 gross tons that were produced from the Emily member of the Rabbit Lake Formation, all the ore was produced from the Trommald Formation. Today, commercial-quality iron and manganiferous ores are largely exhausted; current small-scale mining operations produce only marginal-grade ore. Although present economic conditions do not justify mining of the low-grade manganiferous materials, the Cuyuna district represents one of the largest, if not the largest, manganese reserves in the United States. In addition, the Cuyuna district may have considerable potential as a source for sulfur from the pyritic-graphitic slate of the South Cuyuna range, in T. 46 N., R. 25 W. in Aitkin County (see discussion of Aitkin County sulfide deposits, this chapter).

All iron and manganiferous ores and potential ores are associated with cherty iron-formations or highly ferruginous slates that contain variable amounts of chert. Exploration has been concerned primarily with the discovery of ore in the oxidized and leached parts of the iron-formations, where much of the silica, magnesium, calcium and carbon dioxide have been removed to form deposits of shipping-grade or concentrating-grade ore.

Ore Bodies

Oxidation and leaching of the iron-formation at the bedrock surface are widespread, and the oxidized and leached zone locally extends to depths of more than 1,000 feet. Zapffe (1933, p. 81) reported the occurrence of ore to a depth of 1,020 feet. The deepest mined ore came from the 800-foot level of the Armour No. 1 mine in sec. 10, T. 46 N., R. 29 W. In the Emily area, oxidized iron-formation was penetrated by drilling to depths of more than 800 feet, and many drill holes encountered oxidized material to the maximum depths drilled. In some parts of the Emily range, as for example in sec. 26, T. 138 N., R. 26 W. and sec. 21, T. 137 N., R. 25 W., however, unoxidized carbonate-facies iron-formation occurs locally within 50 to 60 feet of the bedrock surface. Very little information is available on the depth of oxidation and leaching on the South range. All attempts to mine iron ore were at shallow depths—207 feet at the Adams mine, 120 feet at the Hobart exploration, 203 feet at the Wilcox mine, 160 feet at the Barrows, and 164 feet at the Brainerd-Cuyuna mine. Even at these shallow depths, "amphibole-magnetite rock" was reported from the Adams and Hobart workings.

Oxidation and leaching of the Trommald Formation in the North and Emily ranges appear to have taken place in pre-Cretaceous or Early Cretaceous time, for poorly consolidated, light-colored, thin-bedded mudstone and interbedded sandstone and lignite-bearing strata, believed to be of Cretaceous age, lie on oxidized and partly leached iron-formation. This occurrence of Cretaceous strata on oxidized iron-formation and ore is similar to that on the Mesabi range.

The pattern of oxidation and leaching in the Cuyuna district is difficult to define, but exploration and mining indicate that areas of tight folding and associated fracturing in the Trommald Formation are favorable for the occurrence of ore. Schmidt stated (1963, p. 60), "No relationship was found between ore bodies and folds in the thin-bedded facies. Many ore bodies occur in synclines because more iron-formation remains uneroded in the synclinal troughs. Where anticlinal folds of iron-formation remain uneroded, they seem as subject to alteration to ore as the synclines. Many ore bodies are located on steep virtually monoclinal limbs of large folds and do not extend downward into the adjacent synclinal trough." Concerning ore development in the thick-bedded facies, Schmidt (1963, p. 61) stated, ". . . the position of several large ore deposits in areas of intense drag folding suggests that folding favored the formation of ore. Folding has also increased the width of the exposure and likewise the mineable width of any ore available."

In the South range, ore occurs in steeply-dipping iron-formation as narrow, shallow bodies; in the North range, relatively large, moderately deep ore bodies are developed on any part of the folds; and in the Emily range, the most intense oxidation and leaching occur in areas of tight folding. As a generalization, areas in the Cuyuna district having relatively large and wide outcrop belts of iron-formation and having secondary folding are favorable for the occurrence of ore. The iron-formation facies does not seem to be important to the development of ore. On the North range, Schmidt (1963) reported that, of the ore mined, about 86 percent is from thin-bedded facies and 14 percent is from thick-bedded facies, which is about the same as the percentage of each facies exposed at the bedrock surface. Available information concerning the Emily range area suggests a similar relationship between strong oxidation and leaching and lithology of the iron-formation. Because the thin-bedded or mixed thick- and thin-bedded facies form the thicker parts of the Trommald Formation, a greater proportion of strongly altered iron-formation ore occurs in these lithologies.

Ore Grades

Several grades of iron and manganiferous ores have been mined on the North Cuyuna range. Schmidt (1963) classified the ores by texture and structure as laminated ore, gnarled ore, bedded-wash ore, specular ore, cindery and massive ore, and solution-banded ore. The several classes of ore are indicative of the wide range in physical properties and appearance of ore grade materials. In most mines, there is a mixture of various textures and ore types; however, the laminated, gnarled, and cindery and massive ores are characteristic of the thin-bedded facies, whereas specular, bedded-wash, and solution-banded ores occur in the thick-bedded facies. The structure of the ore varies from soft and friable to sandy, compact, and massive.

Ores in the Cuyuna district also are grouped by color into brown goethitic ores, red-brown hematitic ores, dark brown or black manganiferous-iron ores, and black high-manganese ores. Concentrating-grade ores commonly contain both goethite and hematite and have layers of sandy quartz, forming an irregularly banded ore having dark iron oxide layers and lighter siliceous layers.

The commercial classification of Cuyuna ores is based on chemical composition and structure, and ores are classified as bessemer, low phosphorus-non-bessemer, high phosphorus-non-bessemer, and manganiferous-iron ore. The manganese content of manganiferous-iron ore is more than two percent. Some ore shipped from the Cuyuna district contained from 15 to 22 percent manganese. In addition to shipping-grade ores, some manganiferous lean-ore containing from 7 to 15 percent manganese was supplied to the Manganese Chemical Corporation at Riverton, Minnesota during the period 1953 to 1961, and was concentrated to a high-grade manganese product.

Ores of the South range are reported to occur as narrow, rather shallow, small deposits scattered along the outcrop of the iron-formation. Grout and Wolff (1955, p. 29) reported that the South range ores ". . . range in character from soft high-moisture limonitic ores to medium hard red-brown hematite. . . ." Shipments of ore from the South range contained from 54 to 57 percent iron on a dry basis, 0.25 to 0.45 percent phosphorus, and 2.9 to 3.1 percent Al_2O_3. Zapffe (1933, p. 82) reported that the average South range ores contain more than 0.40 percent phosphorus.

Origin of the Ores

The iron and manganiferous ores of the Cuyuna district were formed by the removal of silica, calcium, magnesium, and carbon dioxide by solutions and the concentration of iron and manganese oxides. There was some movement of iron within the iron-formation during the ore-forming process. Schmidt (1963, p. 62) postulated two stages of ore development, one related to rising hydrothermal solutions that ". . . stimulated circulation of ordinary ground waters and accelerated their oxidizing and leaching capacity perhaps partly by increasing their temperature but probably mostly by increasing their rate of circulation. The resultant ore bodies are large, deep, tabular and hematitic, for example, those of the red-brown type." Schmidt cited the presence of boron, in tourmaline, and the occurrence of deep ores as evidence for a hydrothermal stage. Schmidt further stated, "The second stage of ore development took place by the relatively more irregular and widely distributed oxidation and leaching of the iron formation by ordinary weathering processes." The two-stage development of the Cuyuna district ores seems feasible. The extensive oxidation and leaching of iron-formation suggest, however, that there was a long stable period of weathering in pre-Cretaceous or Early Cretaceous time, and it is possible that weathering alone developed the ores. Assuming a deep water table, a warm climate, moderate rainfall, and a long period of oxidation and leaching, normal weathering processes could have developed all of the oxidized iron-formation and ore.

Future Potential

Most of the commercial-grade ore that can be extracted from the Cuyuna district under present economic conditions already has been mined. The State of Minnesota estimated (1970) that the ore reserves of the Cuyuna district are 26,868,173 gross tons, of which about 15,675,000 gross tons or 58 percent are on the South range. The North range is estimated to contain about 11,000,000 gross tons of unmined ore, which is distributed in 44 properties; the largest single reserve is about 934,172 gross tons. The present economic potential for natural ores in the Cuyuna district is small. The most important potential for future production of iron or manganiferous ores is the concentrating-quality, low-grade, iron- and manganese-bearing materials. So far, attempts to concentrate the iron-bearing and manganiferous-iron-bearing materials into an iron ore or a manganiferous-iron ore have been only partly successful. Manganese ore and iron ore can be produced from certain of the low grade, iron-bearing materials using available concentration methods, but at a relatively high cost. Low-grade, iron-bearing materials in other districts—such as the Mesabi, Michigan, and Wisconsin magnetite-taconite ores, the Michigan jaspers, and the Michigan and Mesabi oxidized iron-formations—offer more promise for economic success than do materials from the Cuyuna district. A future strategic demand for manganese, even at high cost, may be required to revive large-scale mining in the Cuyuna district.

The technical feasibility of manganese recovery was demonstrated by the Manganese Chemical Corporation, but the costs of the product were high. Lewis (1951, p. 37-43) estimated that 500,000,000 gross tons of manganiferous iron-formation containing from 2 to 10 percent manganese are available to open-pit mining to a depth of 150 feet. This may be a conservative figure for the potential reserve of the Cuyuna district inasmuch as many areas having a potential for low-grade manganiferous materials are obscured by cover of 200 or more feet of glacial drift. Little systematic exploration has been done to prove tonnages of iron-formation that contain 5 to 10 percent manganese in the Trommald Formation and the Emily member. Although the North range has been fairly well explored, the Emily range has been only partly explored because of the thick glacial drift. If a need for high-cost manganese comes about, past work in the Emily and North ranges will serve to localize initial exploration efforts, but a systematic geologic, geophysical, and drilling program will be needed to determine adequately the distribution and quality of manganiferous iron-formation.

ACKNOWLEDGMENTS

This paper and the accompanying figures are based on cited published reports and on many unpublished company reports that present the results of investigations that have extended over a period of about 70 years. I am indebted to various mining companies for access to exploration data, and particularly to John S. Owens and Neil Walker for their help in acquiring information and for their assistance in the interpretation of the geology of the Cuyuna district.

EAST-CENTRAL MINNESOTA

C. W. Keighin, G. B. Morey, and S. S. Goldich

The age and relationships of the igneous and metamorphic rocks that crop out over a wide area in east-central Minnesota (fig. IV-1) have been debated since the time of the earliest field investigations. N. H. Winchell (*in* Winchell and Upham, 1884) first suggested that granitic rocks, which intrude metasedimentary rocks now assigned to the Thomson Formation, were post-Animikie in age, but he later concluded (*in* Winchell and Upham, 1888) that they were correlative with the Lower Precambrian rocks in the northern part of the state. Much of our current knowledge of the field relationships comes from the work of Margaret Skillman Woyski (1949), who divided the granitic rocks into four groups, and summarized their history as follows: (1) intrusion of the McGrath Gneiss and regional dynamic metamorphism (Early Algoman); (2) intrusion of a series of intermediate igneous rocks (Late Algoman); (3) intrusion of the Stearns magma series (Middle Keweenawan), and intrusion of felsite and basalt dikes (Middle Keweenawan). However, Goldich and others (1961) showed that the igneous rocks are neither Algoman (2,600-2,700 m.y.) nor Middle Keweenawan (1,000-1,200 m.y.) in age; hence, they redefined the name "Penokean orogeny"—which was used first by Blackwelder (1914) for what he thought was a post-Keweenawan period of folding and mountain building in the Lake Superior region—and applied it to the events which occurred 1,600 to 1,800 m.y. ago in east-central Minnesota, northern Wisconsin, and Michigan. Subsequently, the name "Hudsonian" was introduced by Stockwell (1964, p. 5) for a similar orogenic period in the Canadian Shield having a mean K-Ar mica age of 1,735 m.y. and a range from 1,640 to 1,830 m.y.

The Precambrian rocks of east-central Minnesota are being restudied presently, and this paper is a progress report of the work being carried on jointly by staff members of the Minnesota Geological Survey and of the department of geology of Northern Illinois University. Because outcrops are sparse, field observations must be supplemented by indirect methods, the most important of which are radiometric age determinations.

NOMENCLATURE

Names such as the Thomson Formation (Schwartz, 1942b) and the McGrath Gneiss (Woyski, 1949) are well established, but many of the other names that have been used for the igneous rocks are local or trade names associated with the quarrying industry. In this category are the Hillman tonalite, the Freedhem tonalite, the Warman quartz monzonite, and the St. Cloud gray and red granites. These names are used informally in this paper because their ages and correlations are not well established.

PREVIOUS WORK

The geologic studies of Hall (1901b), Schwartz (1942a), and Woyski (1949) served as the basis for early radiometric studies by Goldich and others (1961). Micas, principally biotite, separated from the coarser grained igneous and metamorphic rocks, were dated by the K-Ar and Rb-Sr methods, and whole-rock samples of slate and phyllite from the Thomson Formation and from the metasedimentary rocks of the Cuyuna district were dated by the K-Ar technique.

The radiometric ages quoted in this paper have been recomputed using decay constants for K^{40} and Rb^{87} that are now commonly accepted and which differ slightly from the values used in the original calculations. The basic analytic data and the recomputed ages are given in Tables IV-5 and IV-6. K-Ar determinations by Peterman (1966), Hanson (1968), and Hanson and Malhotra (1971) are included in Table IV-5, and three new Rb-Sr determinations are given in Table IV-6. The ages, for the most part, fall within the range from 1,600 to 1,800 m.y., and when first published provided the first widely accepted evidence for an orogeny that had not been recognized previously in Minnesota (Goldich and others, 1957, p. 547).

Both the K-Ar and Rb-Sr decay systems in micas, however, are now known to be sensitive to low-temperature metamorphic events, and mica ages generally are interpreted as minimum dates, commonly reflecting younger events that are difficult to recognize geologically. For example, samples of the metasedimentary rocks from the Cuyuna district as well as from the Thomson, Virginia, and Rove Formations were determined by Peterman (1966), using the whole-rock Rb-Sr technique, to have an isochron age of 1,850 m.y., whereas earlier K-Ar determinations on the same rocks gave ages ranging from 1,540 to 1,670 m.y. Similarly, Hanson (1968) obtained a K-Ar age on hornblende from the granite at Rockville of 1,800 m.y., whereas earlier K-Ar and Rb-Sr determinations on biotite from the same rock gave ages of 1,640 and 1,730 m.y., respectively (tables IV-5 and IV-6). These and other investigations are referred to in a later section of the paper, and serve to demonstrate that mica ages must be interpreted with care.

McGRATH GNEISS

Although known outcrops of the McGrath Gneiss are relatively sparse and small, the area underlain by the gneiss appears to be extensive (fig. IV-20). At the type locality southwest of McGrath, the rock is a coarse-grained, pinkish-gray biotite gneiss containing large crystals of microcline. Some of the large crystals are rounded, giving the appearance of augen, but many are euhedral or subhedral and are oriented obliquely to the foliation. These crystals

Table IV-5. Summary of K-Ar age determinations of Middle Precambrian rocks.

Sample No.	Map No. (fig. IV-20)	Description	K (pct.)	*Ar40 (ppm)	Age[1] (m.y.)
		(A) *Thomson Formation*			
35 wr[2]	a	Slate, Thomson	3.00	0.553	1630
38 wr	b	Phyllite, Atkinson	1.03	0.191	1630
39 mu	c	Phyllite, Barnum	4.28	0.803	1650
40 mu	d	Phyllite, Moose Lake	3.79	0.685	1610
96 wr	y	Phyllite, Little Falls	1.90	0.336	1580
232 wr		Slate, Duluth	3.55	0.433	1220
233 wr		Slate, Duluth	3.34	0.387	1170
		(B) *Rove Formation*			
131 wr		Argillite, Gunflint Trail	4.06	0.410	1060
		(C) *Virginia Formation*			
137 wr		Argillite, Virginia	3.40	0.539	1470
212 wr		Argillite, West Mesabi	3.38	0.396	1180
		(D) *Mahnomen Formation, Cuyuna district*			
33 wr	o	Argillite	4.60	0.878	1670
132 wr	m	Phyllite	4.09	0.698	1540
134 wr	k	Argillite	3.84	0.677	1580
215 wr	n	Argillite	3.36	0.644	1670
408 (90-100)		Argillite	4.69	0.800	1550[3]
410 (100-110)		Argillite	3.11	0.548	1580[3]
410 (180-200)		Argillite	4.28	0.680	1470[3]
		(E) *McGrath Gneiss*			
41 bi	e	West of Denham	7.16	1.38	1670
43 bi	f	SW. of Denham	6.29	1.19	1650
164 bi	h	McGrath	6.82	1.36	1710
63 bi	g	Pliny (Dad's Corner)	7.38	1.21	1500
		(F) *Intermediate granitic and related rocks*			
1 bi	i	Quartz monzonite, Warman	6.91	1.44	1760
62 bi	j	Quartz monzonite, Isle	6.07	1.18	1680
64 bi	r	Tonalite, Hillman	6.44	1.35	1770
60 bi	p	Gneiss, Freedhem	6.47	1.28	1710
59 bi	p	Schist, Freedhem	6.18	1.14	1630
61 bi	q	Quartz monzonite, Pierz	6.96	1.41	1730
10 bi	x	Granodiorite, St. Cloud	5.04	1.06	1770
		(G) *Granites*			
6 bi	u	Porphyritic granite, Rockville	6.23	1.16	1640
RH-21 ho	t	do.	0.893	0.191	1800[4]
58 bi	s	Red Granite, St. Cloud	5.46	1.02	1640

Table IV-5—Continued

Sample No.	Map No. (fig. IV-20)	Description	K (pct.)	*Ar[40] (ppm)	Age[1] (m.y.)
		(H) *Basaltic dikes*			
M8300 wr		Basaltic dike, St. Cloud	0.776 0.770	0.0991 0.101	1280[4]
M8301 wr		Basaltic dike, St. Cloud	2.00 1.96	0.309	1460[4]
M8302 wr		Basaltic dike, Rockville	0.568	0.0982 0.0986	1570[4]
MN-15 wr		Basaltic dike, St. Louis River	1.023 1.011	0.102	1050[5]

* Radiogenic Ar^{40}
[1] All ages from Goldich and others (1961) have been recomputed, using
$\lambda\epsilon = 5.84 \times 10^{-11}\ yr^{-1}$
$\lambda\beta = 4.72 \times 10^{-11}\ yr^{-1}$
$K^{40}/K = 0.0119$ (atomic ratio)
[2] Abbreviations: wr, whole rock; mu, muscovite; bi, biotite; ho, hornblende
[3] Peterman (1966)
[4] Hanson (1968)
[5] Hanson and Malhotra (1971)
Analytical uncertainty in the ages is about 5 percent.

Table IV-6. Summary of Rb-Sr age determinations on biotites from Middle Precambrian rocks.

Sample No.	Map No. (fig. IV-20)	Description	Rb (ppm)	Sr_n (ppm)	Sr^{87}/Sr^{86} [1] (Atomic)	Rb^{87}/Sr^{86} (Atomic)	Age[2] (m.y.)
164	h	McGrath Gneiss, McGrath	515	10.5	4.21	142	1750
41	e	McGrath Gneiss, W of Denham	413	5.40	6.11	221	1740
1	i	Quartz monzonite, Warman	427	16.9	2.57	73.2	1810
64	r	Tonalite, Hillman	484	17.2	2.73	81.5	1760
58	s	Granite, St. Cloud	886	8.48	8.06	302	1730
10	x	Granodiorite, St. Cloud	372	53.0	1.23	20.3	1820
532	v	Quartz monzonite, Rockville	571	4.29	10.1	385	1730
536	w	Quartz monzonite, SW. of St. Cloud	617	6.09	7.78	293	1720

[1] Calculated from spiked runs
[2] Computed with initial ratio = 0.710
$\lambda\beta = 1.39 \times 10^{-11}\ yr^{-1}$; $RB^{87}/Rb^{85} = 0.386$ (atomic)

have the appearance of porphyroblasts. A conspicuous foliation is given by biotite, which is wrapped around the large microcline crystals. At Dad's Corner (fig. IV-20), the gneiss is similar in appearance to that near McGrath. Small discontinuous pegmatites and quartz veins cut the foliation, and pale green epidote and pyrite occur on joint surfaces. Small flat outcrops of gneiss along the road 2 miles south of Arthyde and approximately 9 miles east of Dad's Corner are more variable in lithology. Here, foliation is well developed, and there is some mineral layering. The large microcline crystals generally appear deformed. In the vicinity of Denham, however, the McGrath Gneiss is distinctly red and more sheared than at other localities. Muscovite is prominent, and the large microcline crystals are strongly de-

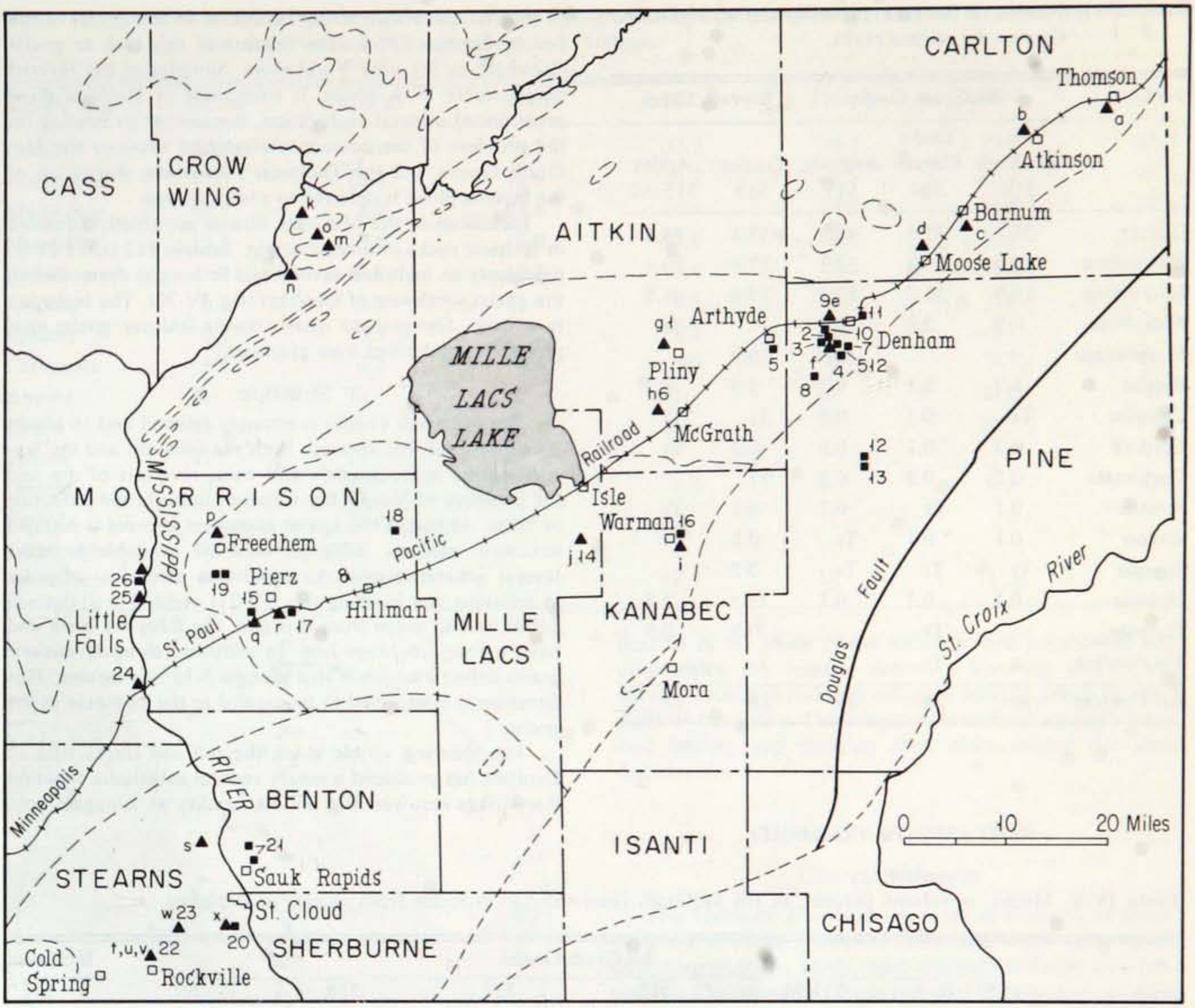

Figure IV-20. Generalized geologic map of east-central Minnesota (modified from Woyski, 1949, and Goldich and others, 1961) showing the location of samples cited in the text. ● 8, sample location and number; ▲g, sample location with age data.

formed and locally granulated. Pegmatites and quartz veins which cut the foliation have, like the gneiss, undergone mechanical deformation.

Petrography

Three phases of the McGrath Gneiss are distinguished: (1) relatively unsheared gneiss at McGrath, Dad's Corner, and Arthyde (table IV-7); (2) sheared gneiss in the vicinity of Denham (table IV-8); and (3) highly sheared, fine-grained, layered gneiss in the contact zone between the gneiss and the Thomson Formation west of Denham. A fourth phase, of apparent local extent, is designated herein the gneiss of Bremen Creek (table IV-7).

The modes of the gneiss (phase 1) from McGrath, Dad's Corner, and Arthyde, together with a chemical analysis of the rock from Dad's Corner (Sandell and Goldich, 1943, p. 110), indicate that the original rock was a porphyritic quartz monzonite. The primary essential minerals are quartz, andesine, microcline, and biotite. Mechanical deformation, accompanied by low-temperature metamorphic recrystallization, accounts for the textural, structural, and mineralogical variations. Small amounts of chlorite, epidote, calcite, and muscovite occur in all samples, but the sheared phase in the vicinity of Denham contains from 2 to 18 percent muscovite (table IV-8). The gneiss along Bremen Creek (table IV-7) resembles typical McGrath Gneiss (phase 1) in hand specimen. In thin section, however, it is seen to be granular and recrystallized rock, and much fresher appearing than phase 1; the plagioclase is less highly sericitized and there is an appreciable amount of fluorite, which is found only in trace amounts in other samples of the gneiss. Fluorite is abundant in a fine-grained aplite dike (table IV-

Table IV-7. Modes, in volume percent, of the McGrath Gneiss and related rocks.

	McGrath Gneiss			Bremen Creek	
	McGrath 518	Dad's Corner 501	Arthyde 517	Gneiss 515	Aplite 515-A2
Quartz	38.7	32.6	34.8	37.4	26.5
Plagioclase	33.0	33.3	38.3	17.6	29.1
Microcline	19.9	23.2	12.8	37.6	41.8
Muscovite	0.9	2.0	2.7		Tr
Hornblende				0.9	
Biotite	6.7	8.1	9.2	5.3	0.2
Chlorite	Tr	0.5	0.5	Tr	Tr
Epidote	0.3	0.1	0.8	0.3	Tr
Carbonate	0.2	0.3	0.2	Tr	
Apatite	0.1	Tr	0.2	0.1	Tr
Zircon	0.1	0.1	Tr	0.1	Tr
Sphene	Tr	Tr	Tr	0.2	
Opaque	0.1	0.2	0.1	Tr	1.5
Fluorite		Tr		0.5	0.9
Tourmaline		Tr		Tr	
An Content	34	34	36	28	33

7) that is concordant to the structure of the gneiss in the bed of Bremen Creek. The relation of this rock to gneiss elsewhere in the area is unknown. Samples of the layered gneiss (table IV-9, phase 3) from west of Denham show pronounced mineral segregation. Because of its bearing on the problem of the contact relationships between the McGrath Gneiss and the Thomson Formation, discussion of the layered phase is deferred to a later section.

Inclusions in the McGrath Gneiss are chiefly schistose or gneissic rocks of diverse origin. Sample 512 (table IV-8) represents an inclusion several feet in longest dimension in the gneiss southwest of Denham (fig. IV-20). The inclusion is a dark, fine-grained quartz-biotite-feldspar gneiss that probably was derived from graywacke.

Structure

The McGrath Gneiss is strongly foliated and in places also is layered and sheared. Both the foliation and the layering strike approximately east-west; reversals of dip and the presence of locally flat foliation indicate the existence of folds. Although the sparse exposures prevent a detailed structural analysis, sufficient data are available to make several generalizations. An equal-area projection of poles to foliation and layering (fig. IV-21) yields a well-defined girdle, and to judge from this plot, the folds are open and have a rounded hinge line. In addition, elongate mineral grains define a lineation that plunges 5-15° to the east. This direction is interpreted to be parallel to the fold axes in the gneiss.

The shearing visible along the railroad tracks west of Denham has produced a nearly vertical cataclastic foliation that strikes east-west (fig. IV-20, locality a). Elongate bou-

Table IV-8. Modes, in volume percent, of the McGrath Gneiss and an inclusion from vicinity of Denham.

	McGrath Gneiss						Inclusion
	511-B	511-M	513	537	538	Av	512*
Quartz	32.2	34.8	33.3	33.2	37.6	34.2	32.4
Plagioclase	12.7	10.8	34.2	14.9	23.1	19.1	13.5
Microcline	34.9	29.7	14.7	32.7	28.8	28.1	9.4
Muscovite	8.8	17.5	5.0	18.1	2.1	10.3	0.9
Biotite	9.4	6.1	11.1	0.6	6.0	6.6	37.6
Chlorite	Tr	0.4	0.4	Tr	1.4	0.4	2.0
Epidote	Tr	0.1	0.1	Tr	0.9	0.4	1.1
Carbonate	1.8	0.6	1.2		Tr	0.7	3.1
Apatite	Tr	Tr	Tr	Tr	.Tr	Tr	Tr
Zircon	Tr	Tr	Tr	Tr	Tr	Tr	Tr
Sphene					Tr	Tr	Tr
Opaque	0.2			0.5	0.1	0.2	Tr
Fluorite							
Tourmaline	Tr					Tr	Tr
An content			25	30	30		

* NE¼SW¼ sec. 35: 45-21

Table IV-9. Modes, in volume percent, of samples from fine-grained layered gneiss in contact zone between the McGrath Gneiss and the Thomson Formation west of Denham.

Sample No.	510-M	510-T	539-1	539-2A	539-2B	539-3	539-4
Quartz	37.5	44.1	36.8	35.3	33.7	30.0	33.2
Plagioclase	17.3	15.3	19.9	24.7	25.5	13.5	28.3
Microcline	30.1	8.9	21.8	10.4	21.5	22.3	27.8
Muscovite	10.3	12.6	8.0	17.6	10.5	10.5	2.7
Biotite	3.0	13.6	12.7	9.4	8.0	20.0	7.0
Chlorite	0.3	5.0	Tr	0.5	0.7	1.3	0.4
Epidote	0.1	0.3	0.1	0.2	Tr	1.1	0.2
Carbonate	1.3	0.2	0.4	0.6	0.1	0.7	Tr
Apatite	Tr	Tr	Tr	Tr	Tr	0.1	0.1
Zircon	Tr	Tr	Tr	Tr	Tr	Tr	Tr
Sphene	Tr	Tr	Tr				
Opaque	0.1	Tr	0.3	1.3	Tr	0.5	0.3
Fluorite		Tr					
Tourmaline	Tr	Tr					
An content		28	13	12	12	11	10

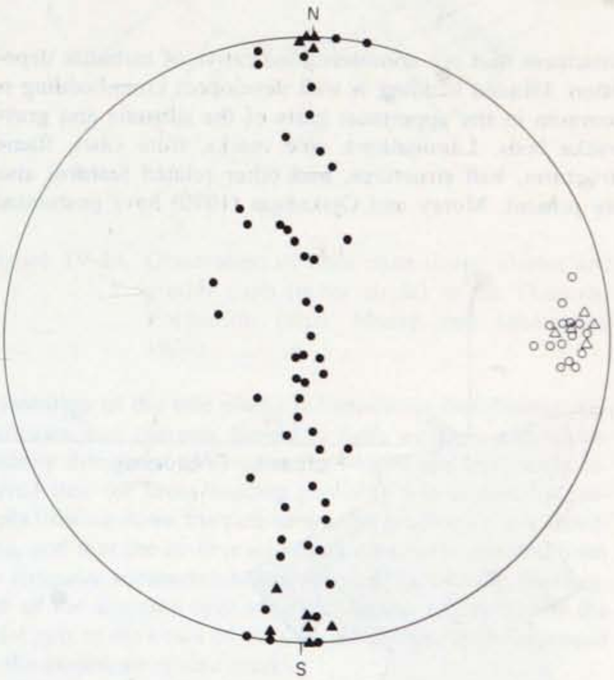

Figure IV-21. Lower hemisphere equal area diagram summarizing various structural elements in the McGrath Gneiss. ●, pole to foliation planes as defined by biotite layering; ▲, pole to shear planes as defined by mineral layering in the sheared layered gneiss west of Denham; O, lineations defined by biotite streaking in mafic layers and by elongate microcline grains in felsic layers; △, lineation defined by hornblende rodding in boudins of the layered gneiss west of Denham.

dins lie in the plane of the foliation, and hornblende rodding within the boudins defines a lineation that is subparallel to that defined by elongate mineral grains in other parts of the gneiss. The congruent structures strongly imply that folding and shearing took place during the same deformation.

THOMSON FORMATION

General Statement

The Thomson Formation consists dominantly of intercalated graywacke, siltstone, shale, and lesser amounts of quartzite, graphitic black slate, and sulfide facies iron-formation. The formation crops out sporadically over an area of about 500 square miles in parts of Carlton, Pine, Aitkin, and St. Louis Counties, and is best exposed in the valley of the St. Louis River from just west of Duluth, where it is overlain by Keweenawan sandstone (Morey, 1967b), to the vicinity of Cloquet and Carlton, 20 miles west of Duluth. Southward, the formation crops out locally along several abandoned drainage channels that served as outlets for Glacial Lake Upham during Pleistocene time (see Wright and others, 1970).

Near Thomson, the formation consists of graywacke, siltstone, and slate which contain primary sedimentary textures and structures; but southwest of Thomson it is metamorphosed to higher ranks, and the metamorphic grade progressively increases southwestward.

Petrography

At its type locality, the Thomson Formation consists of about 46 percent graywacke, 27 percent siltstone, and 27 percent slate (Schwartz, 1942a; Wright and others, 1970). Two stratigraphic sections measured by Morey and Ojakangas (1970) in the vicinity of Carlton have somewhat

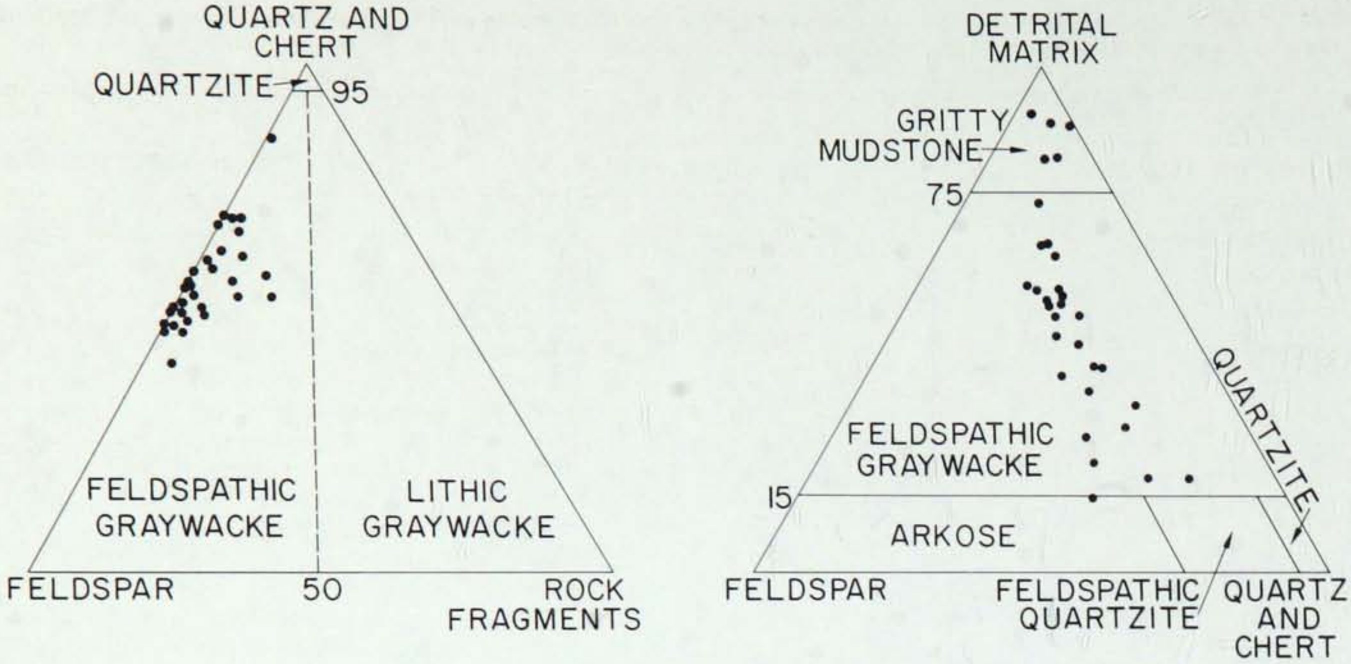

Figure IV-22. Summary of mineral composition in rocks of the Thomson Formation from near the type locality (after Morey and Ojakangas, 1970).

different proportions, averaging 34 percent graywacke, 39 percent siltstone, and 27 percent slate. These data indicate that the abundance of graywacke, siltstone, and slate varies without regard to stratigraphic position. The total thickness of the formation is not known inasmuch as neither the top nor bottom is exposed. The lack of continuous exposure and suitable marker beds and the presence of folds contribute to the difficulty of making estimates of thickness. Accordingly, estimates of the probable thickness range from at least 3,000 feet (Wright and others, 1970) to as much as 20,000 feet (Schwartz, 1942b).

From detailed petrographic studies, summarized in Figure IV-22, Morey and Ojakangas (1970) have shown that the Thomson Formation is similar lithologically to the Rove Formation (Morey, 1969). The major framework constituents of the graywacke are quartz, sodic plagioclase and, in the coarser-grained beds, granitic rock fragments and trace amounts of microcline and orthoclase. Except for clasts of slate, the framework grains rarely exceed 1 mm in diameter, and most are 0.1 to 0.5 mm. The matrix is composed of chlorite, muscovite, and calcite in grains finer than 0.03 mm. The shape and size of both the framework grains and the matrix particles have been modified by recrystallization. However, the grain size and mineralogy are closely related, inasmuch as samples having a small component of detrital matrix are coarser grained than those having a large component of detrital matrix. Abundant carbonate concretions characterize the formation; they were first studied by Schwartz (1942c) and more recently by Weiblen (see Morey and Ojakangas, 1970, p. 13-15).

Sedimentological aspects of the Thomson Formation have been described by Morey and Ojakangas (1970). Graywacke beds range in thickness from 1 inch to 14 feet, and commonly display a wide variety of internal sedimentary structures that are considered indicative of turbidite deposition. Graded bedding is well developed; cross-bedding is common in the uppermost parts of the siltstone and graywacke beds. Laminations, sole marks, flute casts, flame structures, ball structures, and other related features also are present. Morey and Ojakangas (1970) have postulated

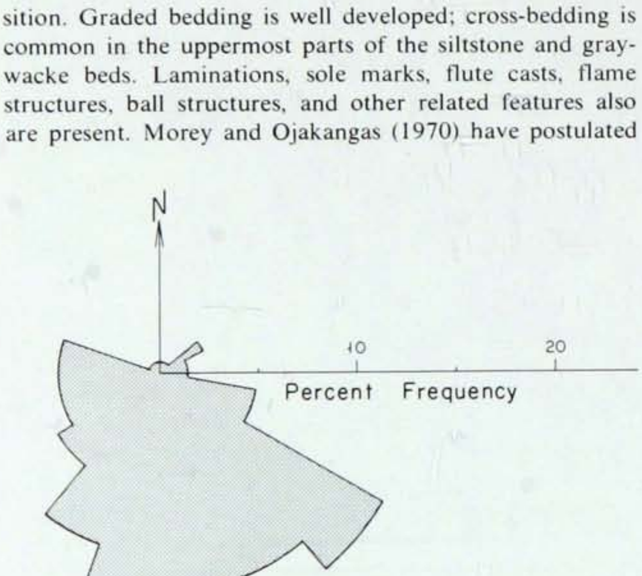

Figure IV-23. Rose diagram summarizing orientation of cross-bedded units in the Thomson Formation (after Morey and Ojakangas, 1970).

that the laminated black muds, now slate, accumulated slowly in quiet water and were not reworked subsequently. Deposition of this material was interrupted periodically by the influx of silt and sand beds deposited by turbidity currents entering the area. The dip azimuth directions of cross-bedding show consistent current flow from north to south (fig. IV-23). In contrast, the directions of current flow indicated by sole marks, including flute and groove casts, are randomly scattered (fig. IV-24). Considered by itself, the

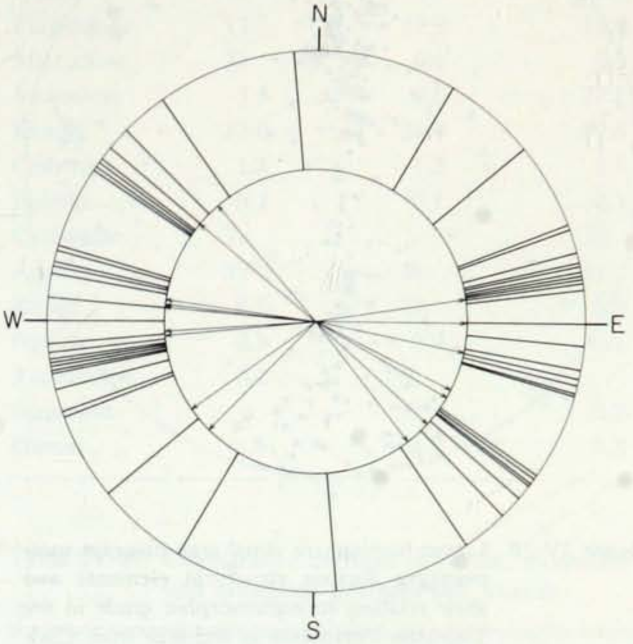

Figure IV-24. Orientation of flute casts (inner circle) and groove casts (outer circle) in the Thomson Formation (after Morey and Ojakangas, 1970).

orientation of the sole marks is bimodal in distribution and indicates that currents flowed in both westerly and southeasterly directions. Consequently, Morey and Ojakangas inferred that the cross-bedding probably was caused by currents flowing down the paleoslope, perpendicular to a shoreline, and that the diverse sole mark directions resulted from an irregular submarine topography. Alternatively, fanning-out of the currents over a gently-sloping paleoslope in the axial part of the basin could account for the observed spread in the directions of sole marks.

Because the mineralogy of the Thomson Formation is similar to that of the Rove Formation and inasmuch as the paleocurrent indicators in the Thomson Formation imply a northerly source, it is inferred that the detritus was derived from a Lower Precambrian terrane composed in large part of granitic rocks.

Metamorphism

There is a progressive increase in metamorphic grade of the Thomson Formation between Thomson and Denham (Hall, 1901b; Schwartz, 1942a; and Weiblen, 1964, unpub. M.S. thesis, Univ. Minn.). Near Thomson, the formation contains quartz, albite (An_5), chlorite, sericite, and calcite, whereas near Denham it contains quartz, andesine (An_{32}), biotite, muscovite, staurolite, and garnet. The former assemblage is characteristic of the lower part (chlorite zone) of the greenschist facies, and the latter assemblage represents Winkler's (1967) B 2.1 staurolite-almandine subfacies of the Barrovian-type facies series.

Incipient phyllite first occurs about 4 miles southwest of Carlton (fig. IV-20). The outcrops consist mainly of metagraywacke and black slate, but locally contain beds of crumpled and folded phyllite. Asymmetric folds are numerous, and muscovite is abundant. At Atkinson, phyllite is well developed and muscovite flakes as long as 0.1 mm are common. Within these rocks, detrital quartz and feldspar grains are still discernible, but twinning in plagioclase is not apparent. Dominant mineral assemblages within the phyllitic rocks consist of quartz, albite (An_5), chlorite, muscovite, and calcite or dolomite. Biotite first appears in the pelitic rocks near Barnum, and from there southward chlorite occurs only as a late alteration product. Quartz and feldspar are somewhat recrystallized and together with the micas define a crude foliation. Dominant mineral assemblages contain quartz, albite (An_{10}), muscovite, biotite, and calcite. At Moose Lake, muscovite- and biotite-bearing schist is well developed. Garnet occurs in the schist and metagraywacke beds as does hornblende in more calcareous units. Dominant mineral assemblages in the schist contain quartz, oligoclase, biotite, muscovite, hornblende, and apatite. Staurolite first appears east of Denham in biotite- and muscovite-bearing schists. Quartz and feldspar are extensively recrystallized and elongate in the foliation plane. Andesine occurs as small nodules, and garnet metacrysts are abundant and well developed. The metagraywacke units consist of quartz, plagioclase (An_{27}), biotite, and garnet. The garnet is characterized by a helicitic texture having contorted lines of inclusions that can be traced from one garnet metacryst to another. Biotite and muscovite are wrapped around the garnet, indicating rotation of the garnet in a late stage of recrystallization. The calcareous concretions are zoned, and are characterized by rims of hornblende, garnet, plagioclase, and quartz, and by cores of epidote, quartz, plagioclase, and calcite (Weiblen, 1964, *op cit.*).

Small isolated outcrops of slate, phyllite, and metagraywacke also occur along the Mississippi and Little Elk Rivers in the vicinity of Little Falls in Cass County (fig. IV-20). North of Little Falls, the outcrops are characterized by intercalated beds of gray slate, incipient phyllite, and graywacke that strike northeastward and dip variably either northwest or southeast. South of Little Falls, however, the grain size of these rocks becomes markedly larger, and the rocks are characterized by large (2 cm) metacrysts of staurolite in a granular matrix consisting of quartz, untwinned plagioclase (andesine or oligoclase near andesine), biotite, and garnet; muscovite and K-feldspar are minor constituents. Thus, as in the Carlton-Denham area, the metamorphic grade progressively increases southward, from the greenschist facies to the amphibolite facies in proximity to the granitic intrusions near St. Cloud.

Except for a greater abundance of staurolite, the rocks in the Little Falls area are similar in composition to the Thomson Formation between Denham and Thomson. Therefore, because of the similar lithology and the presence of similar-appearing carbonate concretions and quartz veins, Hall (1901b), Harder and Johnston (1918), Schwartz (1942b), and Goldich and others (1961) correlated the metasedimentary rocks near Little Falls with those of the Thomson Formation. Also the K-Ar age of 1,580 m.y. determined on phyllite from an outcrop on Little Elk River is, within analytical error, comparable to the ages for samples from the Thomson Formation (table IV-5). Although the K-Ar ages are metamorphic ages and do not give the time of deposition, there is no reason to refute this correlation, and it appears that the Thomson Formation once formed a continuous belt between Duluth and Little Falls, a distance of approximately 140 miles.

Elsewhere in east-central Minnesota, large inclusions of metasedimentary rock that resemble metamorphosed Thomson Formation are found in various granitic rocks, particularly in the region between McGrath, Little Falls and St. Cloud (fig. IV-20). Inclusions in the Hillman tonalite of Woyski (1949) most typify rocks resembling the Thomson and are described in the section on the tonalite.

Structure

The Thomson Formation has been strongly deformed, but a lack of exposures in critical areas and a general absence of key marker beds have prevented a detailed structural analysis. Mattson (in Wright and others, 1970), however, demonstrated that the structure in the Cloquet-Carlton area is dominated by large open synclines and anticlines that have many second-order folds on their limbs. The wave lengths of the second-order folds range from a few inches to several hundred feet, but are mostly a few tens of feet. The cross-section in Figure IV-25 shows the major folds to

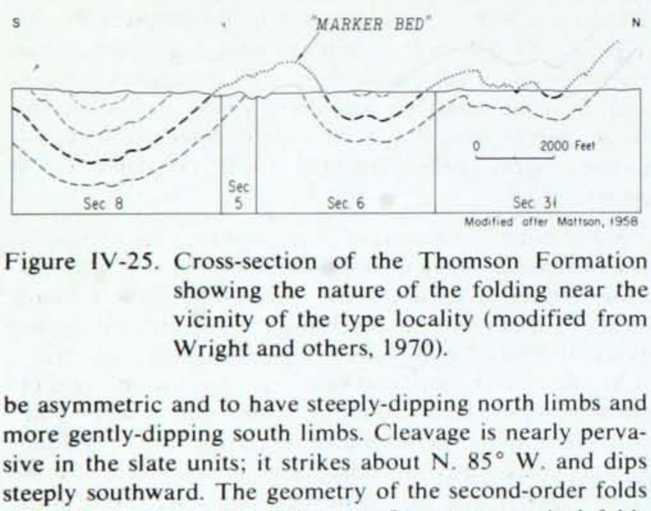

Figure IV-25. Cross-section of the Thomson Formation showing the nature of the folding near the vicinity of the type locality (modified from Wright and others, 1970).

be asymmetric and to have steeply-dipping north limbs and more gently-dipping south limbs. Cleavage is nearly pervasive in the slate units; it strikes about N. 85° W. and dips steeply southward. The geometry of the second-order folds depends somewhat on rock type. Open symmetrical folds are characteristic of slate-bearing units. Axial planes of the second-order folds are nearly vertical and trend generally within 15° of east. Fold axes of both major and second-order folds plunge both eastward and westward at angles up to 20°.

To judge from the available data, the structure of the Thomson Formation does not change appreciably as the metamorphic grade increases (fig. IV-26). Furthermore,

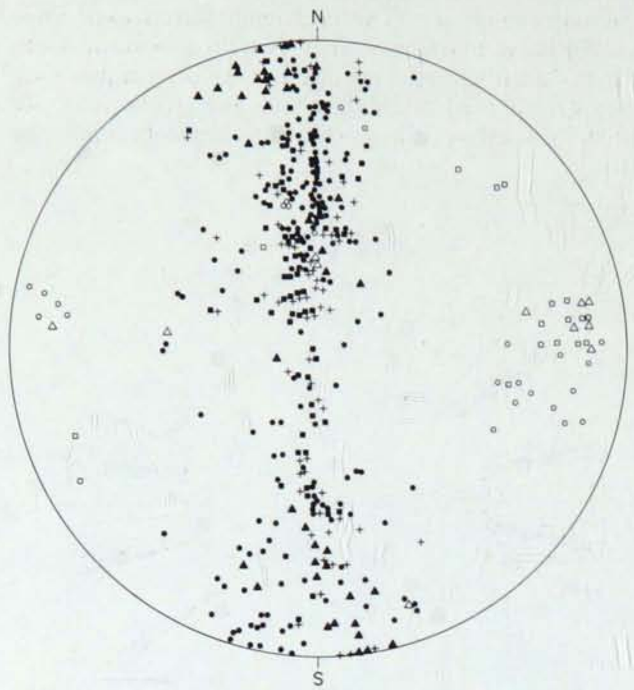

Figure IV-26. Lower hemisphere equal area diagram summarizing various structural elements and their relation to metamorphic grade in the Thomson Formation in the area from Carlton to Denham. Closed symbols, poles to bedding, open symbols, lineations. Circles, Carlton to Atkinson; triangles, Atkinson to Barnum; squares, bedding Barnum to Denham; pluses, foliation Barnum to Denham.

foliation in the metagraywacke and schist in the area from Moose Lake to Denham is parallel to original bedding, suggesting that metamorphism may have occurred before deformation.

RELATIONSHIP OF THE McGRATH GNEISS TO THE THOMSON FORMATION

Disagreements in the past concerning the age of the McGrath Gneiss (see Woyski, 1949, p. 1001) have centered largely on the interpretation of the relationship of the gneiss to the Thomson Formation and on the age of the latter. However, it generally was accepted that the McGrath Gneiss is intrusive into the Thomson Formation. Woyski described the relationship as a *lit-par-lit* injection of the Thomson by the McGrath quartz monzonite, which subsequently was dynamically metamorphosed to the present gneiss. Woyski assigned the McGrath Gneiss an Algoman age in accordance with Schwartz's (1942b) correlation of the Thomson Formation with the Knife Lake Group of northeastern Minnesota. Later, Goldich and others (1961) correlated the Thomson with the Virginia and Rove Formations of the Animikie Group, and on the basis of mica ages (tables IV-5 and 6) considered the McGrath Gneiss a Penokean intrusive rock. G. M. Schwartz (1961, oral comm.) accepted this correlation of the Thomson Formation, but suggested that the McGrath probably was older than the Thomson. Subsequently, T. W. Stern of the U.S. Geological Survey confirmed this suggestion; lead-alpha determina-

Table IV-10. Modes, in volume percent, of samples from the Thomson Formation northeast of Denham.

Sample No.	540	540-1	540-2	540-3	540-4	540-5	540-6
Quartz	53.8	43.8	28.2	32.3	58.1	48.0	48.0
Plagioclase	13.7	19.2	12.8	16.2	13.3	20.0	11.8
Microcline	Tr	0.1	0.3	0.3	0.2	Tr	Tr
Muscovite	3.5	4.6	27.1	17.4		Tr	15.8
Biotite	22.0	24.7	19.6	22.4	20.3	21.8	16.0
Chlorite	1.8	1.2	5.8	4.8	0.7	1.4	1.8
Epidote	0.2	0.1	0.3	0.2	Tr	Tr	Tr
Carbonate	Tr		Tr	Tr			
Apatite	Tr	Tr	Tr	Tr	Tr	Tr	Tr
Zircon	0.1	Tr	0.2	Tr	Tr	0.2	Tr
Opaque	0.1	0.4	0.3	0.1	0.7	Tr	0.4
Tourmaline	0.2						
Staurolite		Tr	0.2	0.6		0.2	Tr
Garnet	4.6	5.9	5.2	5.7	7.0	8.4	6.2

Table IV-11. Comparative averages of modes, in volume percent, of McGrath Gneiss, layered gneiss, Thomson Formation, and associated quartzite and marble.

	McGrath Gneiss			Thomson Formation		
	Relatively Unsheared	Sheared Phase	Contact Zone	Metagraywacke and Schist	Quartzite	Marble
	Table IV-6	Table IV-7	Table IV-8	Table IV-9		
Quartz	35.4	34.2	35.8	44.5	72	34
Plagioclase	34.9	19.1	20.6	15.3	4	2
Microcline	18.7	28.1	20.4	Tr	16	12
Muscovite	1.9	10.3	10.3	9.8	6	1
Biotite	8.0	6.6	10.5	21.0		2
Chlorite	0.3	0.4	1.2	2.5		
Epidote	0.4	0.4	0.3	Tr		
Carbonate	0.2	0.7	0.5	Tr		49
Apatite	0.1	Tr	Tr	Tr	Tr	
Zircon	0.1	Tr	Tr	0.1	Tr	
Sphene	Tr	Tr	Tr			
Opaque	0.1	0.2	0.4	0.3	Tr	Tr
Fluorite	Tr		Tr			
Tourmaline	Tr	Tr	Tr	Tr		
Staurolite	None	None	None	0.1		
Garnet	None	None	None	6.1		
Rock fragments					1	Tr

tions on zircon concentrated from the McGrath Gneiss give a minimum age of 2,400 m.y. (written comm. from Stern to Goldich, 1961).

In the Denham area, where both the Thomson Formation and the McGrath Gneiss are moderately well exposed, the Thomson Formation consists of intercalated beds of schist and metagraywacke. In addition, a fairly thick succession (>250 feet) of interlayered marble and quartzite is present at one locality. Just northwest of Denham (fig. IV-20), the Thomson Formation (table IV-10) consists mainly of fine-grained garnet-staurolite metagraywacke and biotite-garnet schist; coarse-grained quartz-plagioclase-biotite rocks, together with large, discontinuous veins or pods of milky quartz, also are present locally. The pegmatites and quartz veins are interpreted as having developed through local mobilization of the felsic fraction in the graywacke during metamorphism; Schwartz (1942b) suggested that some of the small pods or nodules of quartz were formed by the replacement of carbonate concretions.

Re-examination of the contact zone between the Thomson Formation and the McGrath Gneiss west of Denham has shown that rocks previously considered as a product of *lit-par-lit* injection of the Thomson by the McGrath are actually sheared McGrath (see table IV-11 and appendix IV-A). Garnet and staurolite, which characterize the Thomson Formation, are absent, and the rocks are highly fractured and granulated; accordingly, they resemble a fine-grained, layered gneiss. To judge from the accordant structural data (compare figs. IV-21 and 26), it seems likely that deformation of the Thomson Formation and the McGrath Gneiss took place during the same orogenic event.

The mineralogy of the quartzite and marble exposed southeast of Denham substantiates a pre-Middle Precambrian age for the McGrath Gneiss, inasmuch as the beds contain detritus derived from the gneiss. The quartzite consists of sand-size framework grains admixed in a somewhat finer grained matrix. The framework grains are dominantly monocrystalline and polycrystalline quartz, highly sericitized and cloudy, gridiron-twinned microcline, and lesser amounts of "granitic" rock fragments consisting of interlocking quartz and microcline grains (table IV-11). The matrix is extensively recrystallized, and consists of fine sand-size quartz and sodic plagioclase and lesser amounts of interstitial muscovite and opaque minerals. Intercalated beds of marble vary widely in composition (table IV-11), and generally consist of inequigranular grains of calcite that enclose scattered detrital grains of quartz, microcline, and plagioclase. Locally, small clusters of muscovite and biotite exist.

The geologic relationships in the Denham area are consistent with the conclusion that the detritus in the quartzite and marble was derived from the McGrath Gneiss. However, the precise stratigraphic position of the quartzite and marble relative to the Thomson Formation is uncertain. The mineralogy differs from that of typical Thomson Formation in having microcline as the dominant feldspar phase. Similarly, the quartzite and marble appear to have been deposited under shallow-water conditions, whereas the bulk of the Thomson Formation was deposited under deeper water conditions. Because the quartzite at Denham resembles quartzite exposed near Dam Lake in T. 47 N., R. 25 W., which is inferred to be part of the Mahnomen Formation (see Marsden, fig. IV-17, this chapter), it may be equivalent to it and therefore much older than the Thomson Formation.

IGNEOUS ROCKS

General Statement

A variety of igneous rocks, ranging in composition from gabbro to granite and in texture from coarse porphyritic phanerites to aphanites, has been described by Woyski (1949). On the basis of field relationships she recognized two main episodes of batholithic intrusion younger than the McGrath Gneiss. The principal rocks assigned to the older group are intermediate or basic in composition, and from east to west include the quartz monzonites of Warman, Isle, and Pierz, the tonalites in the vicinity of Hillman and Freedhem, and the granodiorite of St. Cloud (gray) and related types. The younger group, which Woyski named the "Stearns magma series," includes the augite-hornblende granite of St. Cloud (red), the porphyritic quartz monzonite of Rockville, and similar intermediate to silicic rocks. Representatives of the two main groups of Woyski are discussed briefly below.

Quartz Monzonite

The quartz monzonite quarried near Warman (fig. IV-20) is a gray, medium-grained, massive rock that has relatively few inclusions. Woyski correlated similar gray quartz monzonitic rocks exposed in quarries south of Isle and south of Pierz (fig. IV-20) with the rocks at Warman. Two types of quartz monzonite are exposed in the quarry south of Isle, a medium-grained phase and a coarse-grained and porphyritic phase. Both phases are cut by aplite dikes. Modal analyses of samples from the three localities are given in Table IV-12.

Tonalite and Granodiorite

Hillman Tonalite

The Hillman tonalite of Woyski (1949), a gray, medium- to coarse-grained, slightly foliated rock, crops out along the Skunk and Little Skunk Rivers and Hillman Creek in eastern Morrison County and along the Rum River in northern Mille Lacs County (fig. IV-20). The principal minerals are quartz, andesine, and biotite, but the proportions of these minerals vary considerably; accordingly, the modes in Table IV-13 are averaged.

Locally, the tonalite contains numerous inclusions of metasedimentary rocks, which Woyski (1949) interpreted as having been derived from the Thomson Formation. At locality 548 (fig. IV-20), the inclusions are exceptionally abundant and range in length from a few inches to several feet. Foliation in the tonalite in large part is given by schlieren, but foliation in the inclusions commonly differs from that in the tonalite by as much as 40°. This discordance suggests that foliation in the inclusions was developed prior to their entrainment in the tonalite magma. Large (2 cm) garnet crystals are abundant in the biotite-rich inclu-

Table IV-12. Modes, in volume percent, of quartz monzonites in east-central Minnesota.

Locality	Warman	Isle	Pierz
Sample No.	520	500	504
Quartz	31	30	22
Oligoclase-andesine	34	25	47
Microcline	20	32	21
Biotite	14	11	8
Accessories			
Apatite	x	x	x
Opaque	x	x	x
Zircon	x	x	x
Secondary			
Carbonate	x	x	x
Chlorite	X	X	X
Epidote	x	x	x
Sericite*	X	X	X

X, generally less than 1 percent
x, generally less than 0.5 percent
*, includes muscovite

Table IV-13. Modes, in volume percent, of tonalite and related rocks in east-central Minnesota.

		Freedhem		
	Hillman tonalite	granodiorite	quartz monzonite	St. Cloud granodiorite
Quartz	27	7	21	16
Andesine	51	43	27	45
Microcline	4	13	35	12
Biotite	13	18	10	7
Hornblende	2	12	2	12
Accessories				
Pyroxene	X	2	2	3
Apatite	x	x	X	x
Zircon	x	x	x	x
Sphene	X	x	X	X
Opaque	X	X	X	X
Secondary				
Carbonate	x	X	X	x
Chlorite	x	X	x	x
Epidote	X	X	x	x
Sericite	X	X	X	X

X, generally less than 1 percent
x, generally less than 0.5 percent

sions, which contain quartz, andesine, biotite, cordierite, garnet, and lesser amounts of other minerals. The other minerals include muscovite, microcline, pyroxene, magnetite, apatite, zircon, chlorite, and epidote (table IV-12). The assemblage quartz-andesine-biotite-cordierite-garnet represents the upper level of the amphibolite facies (Winkler, 1967). DeWaard (1965) has suggested that cordierite-bearing assemblages are transitional from the amphibolite to the granulite facies; and Gable and Sims (1969) estimated that similar cordierite-bearing assemblages in the Front Range, Colorado were formed at 3-5 kb and 620-710° C.

The cordierite-bearing inclusions in the tonalite appear to be local in distribution, and are interpreted to represent Thomson Formation regionally metamorphosed to the staurolite-andesine-almandine subfacies and then prograded by the engulfing tonalite magma. Hence, the cordierite-bearing assemblage is considered a result of local contact metamorphism rather than regional metamorphism. The cordierite is partially altered to an unidentified micaceous mineral, and muscovite, chlorite, and epidote are retrograde minerals.

Freedhem Tonalite

The Freedhem tonalite of Woyski (1949) is a variable rock ranging in color from black to gray and in grain size from fine to medium. The darker phase is massive except adjacent to dikes of quartz monzonite, where it is deformed. Woyski (see Skillman, 1946, unpub. Ph.D. thesis, Univ. Minn.) noted the range in composition from melatonalite to granodiorite. She found that the most common type was tonalite consisting of 20 percent quartz, 40 percent plagioclase, 35 percent hornblende and biotite, and less than 5 percent K-feldspar. The average modal composition of three samples from the quarries south of Freedhem exhibits somewhat more microcline (13 percent), indicating that the rocks are granodiorite (table IV-13).

The granodiorite is cut by numerous dikes of quartz monzonite, which contain abundant schlieren of the dark granodiorite. The abundance and the deformation of the inclusions suggest that the granodiorite was not completely solidified at the time of intrusion of the quartz monzonite. Microcline (table IV-13) is abundant in the quartz monzonite, and may have been introduced into tonalite to produce the granodioritic composition.

St. Cloud Gray Granodiorite

Granodiorite, commonly referred to as St. Cloud gray granite, has been quarried for many years in the vicinity of St. Cloud (fig. IV-20). The rock is dark gray, but has a slight reddish hue imparted by K-feldspar. It is medium grained and massive, but contains numerous small inclusions which define a vague foliation locally. The granodiorite is cut by dikes of aplite and red granite, and Woyski (1949, p. 1012) described a pink phase of the granodiorite which she attributed to alteration (granitization) by the younger red granite of the area. The mode of the granodiorite in Table IV-12 is comparable to the composition given by Woyski (1949).

Younger Granitic Rocks (Stearns Magma Series)

A red, medium-grained augite-hornblende granite, commonly called the St. Cloud red granite, is quarried at several localities in Benton and Stearns Counties in the vicinity of St. Cloud. Outcrops of similar granitic rocks also are known in western Mille Lacs County and in central Morrison County. Woyski (1949, p. 1006) considered the granite one of the younger, larger intrusive masses, for it cuts and contains inclusions of the granodiorite of St. Cloud, the tonalites near Freedhem and Hillman, and the Thomson Formation. Woyski (see Skillman, 1946, *op. cit.*, p. 33) noted that east of the Mississippi River the granite is pink, whereas west of the river it is red. She also noted that the rock is darker red or green along joints, owing to the presence of chlorite- and epidote-bearing veins, and that the dark red color can be related to alteration spatially associated with the veins. She described cataclastic textures in the rocks, and attributed the red color to hydrothermal activity rather than to surficial weathering.

At several localities the augite-hornblende granite clearly transects the St. Cloud gray granodiorite and contains inclusions of it, but contacts with the coarse porphyritic quartz monzonite in the vicinity of Rockville are lacking; hence the age relations of these two granitic rocks are not known. Woyski considered that the two rocks probably are similar in age, but noted that the mineralogical compositions (table IV-14) differ somewhat, especially with respect to the ratio of plagioclase to microcline.

Table IV-14. Modes, in volume percent, of augite-hornblende granite of St. Cloud and quartz monzonite of Rockville.

Locality	St. Cloud	Rockville
Quartz	29	23
Oligoclase-andesine	10	36
Microcline	52	24
Biotite	4	
Accessories		
Hornblende	3	x
Pyroxene	x	
Apatite	x	x
Zircon	x	x
Opaque	x	x
Secondary		
Carbonate	x	x
Chlorite	x	x
Epidote	X	x
Fluorite	x	x
Sericite	X	x
Sphene	x	x
Pyrite	x	x

X, generally less than 1 percent
x, generally less than 0.5 percent

AGES

K-Ar and Rb-Sr ages of rocks from east-central Minnesota (tables IV-5 and 6) range from 1,051 to 1,820 m.y. The younger ages are related, in part indirectly, to Keweenawan igneous activity (Goldich and others, 1966). The Rb-Sr ages given in the tables are approximately 6 percent greater than the dates originally reported by Goldich and others (1961) and by Peterman (1966).

McGrath Gneiss

K-Ar ages on biotite from the McGrath Gneiss (table IV-5) range from 1,500 to 1,710 m.y. and two Rb-Sr determinations (nos. 164 and 41) give ages of 1,750 and 1,740 m.y. These are interpreted as metamorphic ages related to the Penokean orogeny, for radiometric studies now in progress show beyond any doubt that the McGrath Gneiss was emplaced during the Algoman orogeny, in Early Precambrian time, which in northern Minnesota is dated at 2,700-2,750 m.y. ago (Hanson and others, 1971b; Catanzaro and Hanson, 1971; Prince and Hanson, in press; Peterman and others, in press).

Middle Precambrian Metasedimentary Rocks

K-Ar ages for several of the Animikie formations are given in Table IV-5. With respect to their stratigraphic positions, the oldest unit is the Mahnomen Formation in the Cuyuna district, which is equated with the Pokegama Quartzite of the Mesabi range and the Kakabeka Quartzite of the Gunflint range (Schmidt, 1963; Peterman, 1966). As discussed previously, the Thomson Formation is correlated with the Virginia Formation of the Mesabi range and the Rove Formation of the Gunflint range. The available radiometric ages on these formations differ considerably; hence, it is more expedient to discuss the data as a whole.

The K-Ar ages (table IV-5) for slate and phyllite from the Thomson Formation are similar and average 1,630 m.y. Samples of argillite from the Mahnomen Formation (table IV-5) show a greater range and a somewhat lower average age of 1,580 m.y. A single sample of phyllite from the Thomson Formation near Little Falls (sample 96, table IV-5) gave a similar age of 1,580 m.y.

Samples 232 and 233 from the Thomson Formation, taken near the contact with the gabbro at Duluth, have K-Ar ages of approximately 1,200 m.y., and a sample of the Rove Formation (no. 131, table IV-5) from near the contact of a diabase sill assigned to the Logan intrusions (Goldich and others, 1961, p. 176) gave an age of 1,060 m.y. Two samples of argillite from the Virginia Formation were dated at 1,470 and 1,180 m.y. These ages and the K-Ar ages as a whole are difficult to explain, but in no way do they date the exact or relative time of deposition. Tentatively, we conclude that, except for those rocks affected by Keweenawan igneous activity, the K-Ar system was stabilized in the Mahnomen and Thomson Formations approximately 1,600 m.y. ago. The factors involved in that stabilization are not known. Possibly the last cataclastic deformation of the rocks occurred approximately 1,600 m.y. ago but, alternatively, this date may represent the time at which thermal stability of the K-Ar system in the metasedimentary rocks was achieved, perhaps as a consequence of erosion of much of the overlying succession of rocks.

Peterman (1966) made whole-rock Rb-Sr measurements on the same samples from the Cuyuna district that Goldich had used for K-Ar age measurements, and in addition made the same measurements on samples obtained from drill cores. The isochron age for fresh samples is 1,850 m.y. and for altered samples 1,540 m.y. Peterman (1966, p. 1041) concluded that the 1,850 m.y. date represents the time of weathering or of hydrothermal alteration that may have coincided with development of the natural iron ores.

Seven samples from the Thomson Formation, four from the Virginia Formation, and one from the Rove Formation were plotted by Peterman (1966, fig. 4) in a Rb-Sr diagram with a resulting isochron age of 1,660 m.y. Peterman's data for the seven samples from the Thomson Formation were used by us in constructing Figure IV-27. Samples 232 and

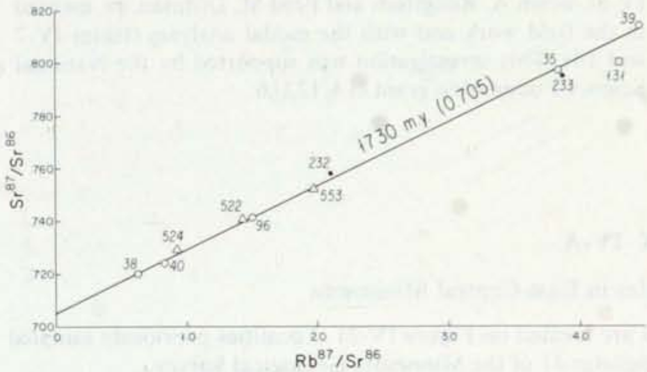

Figure IV-27. Rb-Sr isochron diagram for the Thomson Formation (modified from the data of Peterman, 1966).

233, having K-Ar ages of 1,200 m.y., do not fall on the isochron, suggesting that processes which lowered the K-Ar ages also affected the Rb-Sr isotopic system in these rocks. Three additional samples from the Thomson Formation near Little Falls also are shown on the diagram. A simple linear regression for the eight points representing the Thomson Formation, excluding samples 232 and 233—which have been affected by contact action of the Duluth Complex—gives an isochron age of 1,730 m.y., with an initial ratio of 0.705.

The Rb-Sr isochron age of 1,730 m.y. for the Thomson Formation is similar to the Rb-Sr ages on biotite from the McGrath Gneiss. Biotite from the layered gneiss west of Denham (table IV-6, no. 41) is dated at 1,740 m.y., and this age may be considered to represent a minimum age for the time of shearing of the McGrath Gneiss and for the time when the gneiss was brought into juxtaposition with the Thomson Formation.

Peterman's data for four samples from the Virginia Formation define an isochron age of approximately 1,660 m.y. (fig. IV-28). Faure and Kovach (1969) obtained a similar isochron age of 1,635 m.y. for the Gunflint Iron-formation in the vicinity of Kakabeka Falls, Ontario. In summary, the isochron ages of the various Animikie rocks are: Mahno-

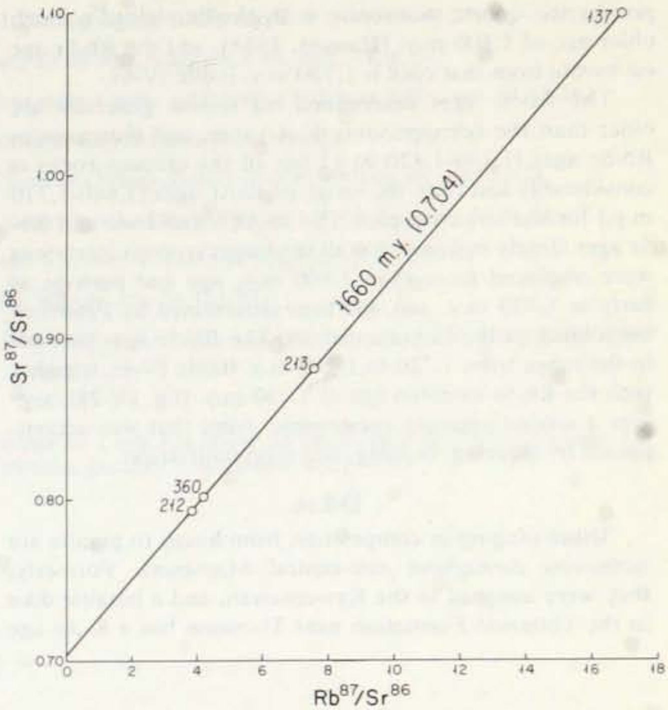

Figure IV-28. Rb-Sr isochron diagram for the Virginia and Rove Formations (modified from the data of Peterman, 1966).

men Formation, 1,850 m.y.; Thomson Formation, 1,730 m.y.; Virginia Formation, 1,660 m.y.; and Gunflint Iron-formation, 1,635 m.y.

Faure and Kovach have argued that the 1,635 m.y. date on the Gunflint Iron-formation represents the time of deposition and diagenesis. If this age is accepted as the time of deposition, the Gunflint Iron-formation was deposited after the metamorphism in east-central Minnesota. The geology and particularly the correlations of the formations, however, contradict this contention. More logically, the different ages are attributable to processes that affected the Rb-Sr system, as Peterman (1966, p. 1041) suggested previously. He pointed out that the linearity of points defining an isochron requires a mechanism that affects all the samples in a regular fashion. In such a model, it is easier to visualize a loss of radiogenic Sr rather than an addition of Rb, which would require that the amount added be proportional to the amount originally present. Peterman also suggested that the retentivity of metasedimentary rocks for radiogenic strontium may be related to the grade of metamorphism. Thus, the relatively unaltered Virginia and Gunflint formations would have remained open systems for a longer time than did the Thomson and Mahnomen Formations.

Igneous Rocks

In accord with inferred geologic age relationships, the K-Ar ages on biotite from the igneous rocks of intermediate composition described by Woyski (1949) range from 1,680 to 1,770 m.y. (table IV-5), whereas biotite from the younger granites of the Stearns magma series have lower K-Ar ages (1,640 m.y., table IV-5). However, hornblende from the porphyritic quartz monzonite at Rockville yielded a much older age of 1,800 m.y. (Hanson, 1968), and the Rb-Sr age on biotite from that rock is 1,730 m.y. (table IV-6).

The Rb-Sr ages determined on biotite generally are older than the corresponding K-Ar ages, and the range in Rb-Sr ages (1,730-1,820 m.y.) for all the igneous rocks is considerably less than the range of K-Ar ages (1,640-1,770 m.y.) for the same samples. The K-Ar hornblende and Rb-Sr ages clearly indicate that all the larger igneous intrusions were emplaced more than 1,800 m.y. ago and perhaps as early as 1,850 m.y. ago, the time determined by Peterman for folding in the Cuyuna district. The Rb-Sr ages that fall in the range from 1,720 to 1,760 m.y. (table IV-6), together with the Rb-Sr isochron age of 1,730 m.y. (fig. IV-28), suggest a second younger epeirogenic event that was accompanied by shearing, faulting, and recrystallization.

Dikes

Dikes ranging in composition from basalt to granite are numerous throughout east-central Minnesota. Formerly, they were assigned to the Keweenawan, and a basaltic dike in the Thomson Formation near Thomson has a K-Ar age of 1,050 m.y. (table IV-5, no. MN-15), consistent with this period of igneous activity. Hanson (1968), however, reported K-Ar ages of 1,280 and 1,460 m.y. on whole-rock samples of two basaltic dikes that cut red granite in the St. Cloud area. A third dike cutting the quartz monzonite at Rockville was dated at 1,570 m.y. He considered these ages as minimum dates, and suggested that all basaltic dikes in the St. Cloud area may have been emplaced during a single period, at least 1,570 m.y. ago.

ACKNOWLEDGMENTS

We are indebted to John R. Richards of the Australian National University who, while Visiting Professor at Northern Illinois University, did the isotopic analyses of samples 522, 524, and 553 (fig. IV-27) and to Gilbert N. Hanson of the State University of New York who generously assisted in the analyses of biotite samples 10, 532, and 536 (table IV-6). Brian A. Rongitsch and Fred M. Dittman, Jr. assisted in the field work and with the modal analyses (tables IV-7 and 14). This investigation was supported by the National Science Foundation grant GA 12316.

APPENDIX IV-A

Location and Description of Samples in East-Central Minnesota

All samples collected for the present study are listed here and are located on Figure IV-21. Localities previously sampled by Goldich and others (1961) are identified by page reference to Bulletin 41 of the Minnesota Geological Survey.

Map No.	Original No. and Description
1	500: McGrath Gneiss; 2.4 miles west of Pliny (Dad's Corner); NE cor. sec. 1: 44-23. Locality KA-63, Bull. 41, p. 177.
2	511B: McGrath Gneiss; 1.2 miles south of railroad crossing west of Denham; SW cor. sec. 27:45-21; sheared pink to gray gneiss.
3	511M: McGrath Gneiss; 0.2 mile south of 511B; NE cor. sec. 33:45-21; granulated pink to dark-gray gneiss.
4	513: McGrath Gneiss; 0.1 mile north of Bremen Town Hall; from ditch on west side of county road; SE¼SW¼ sec. 2:44-21; pink to gray, coarse-grained gneiss.
5	517: McGrath Gneiss; along road 2 miles south of Arthyde; SW cor. sec. 35:45-22; pink to gray gneiss.
6	518: McGrath Gneiss; type locality southwest of McGrath; NE cor. sec. 12:43-24. Locality KA-164, Bull. 41, p. 177.
7	537: McGrath Gneiss; 1.5 miles south of Denham; SW¼ sec. 36:45-21; sheared pink to gray gneiss.
8	538: McGrath Gneiss; 6 miles southwest of Denham; SW¼NW¼ sec. 17:44-21; coarse-grained pink porphyritic gneiss.
9	510: McGrath Gneiss; outcrop blasted along railroad track west of Denham; SE¼NE¼ sec. 21:45-21. Locality KA-41, Bull. 41, p. 177. 510M is sheared red-colored gneiss and 510T is dark-colored phase, layered gneiss in contact zone between the McGrath Gneiss and the Thomson Formation.
10	539: McGrath Gneiss; layered gneiss just south of 510.
11	540: Thomson Formation; railroad cut northeast of Denham; NE¼ sec. 19:45-20. 540 collected from flat outcrop north of road and west of railroad tracks. The 540-1 to 540-6 series was collected from cuts along the railroad north of the road intersection: No. 1 is approximately 2,025 feet north of the intersection; No. 2 is 75 feet south of No. 1; No. 3 is 185 feet south of No. 2; No. 4 is 140 feet south of No. 3; No. 5 is 200 feet south of No. 4; and No. 6 is 200 feet south of No. 5 and approximately 1,225 feet north of road intersection.
12	515-A2: aplite concordant with gneiss in bed of Bremen Creek; middle west section line, sec. 20:44-21; pink, fine-grained dike.
13	515: McGrath Gneiss (?), Bremen Creek; 200 feet south of bridge on west side of road; SE¼ sec. 19:44-21.
14	500: quartz monzonite; quarry 5 miles south of Isle; SE¼NE¼ sec. 3:41-25. Locality KA-62, Bull. 41, p. 177.
15	504: quartz monzonite; quarry 2 miles southwest of Pierz; SE cor. sec. 13:40-31. Locality KA-61, Bull. 41, p. 177.

16 520: quartz monzonite; quarry north of Warman SE¼ sec. 5: 41-23. Light-gray, fine-grained.

17 543: tonalite, approximately 2 miles southeast of Pierz; SW¼ sec. 10:40-30. Locality KA-64, Bull. 41, p. 177.

18 548: tonalite with inclusions of the Thomson Formation; approximately 5 miles northeast of Hillman; NE¼ sec. 34:42-28.

19 544: granodiorite and quartz monzonite; quarries 2.5 miles south of Freedhem; secs. 23-24:41-31.

20 535: granodiorite (St. Cloud gray); quarry 3 miles south of St. Cloud on County Road 136; SW¼NW¼ sec. 34:124-28.

21 526-528: augite-hornblende granite (St. Cloud red); abandoned quarry north of Sauk Rapids; NE¼NW¼ sec. 11:36-31.

22 532: porphyritic quartz monzonite; quarry in Rockville; SW¼ sec. 9:123-29. Locality of KA-6, Bull. 41, p. 178.

23 536: porphyritic quartz monzonite; quarry southwest of St. Cloud; NE cor. sec. 26:124-29.

24 522: Thomson Formation; approximately 8 miles south of Little Falls; below dam and Soo Line Railroad bridge; staurolite schist.

25 524: Thomson Formation; below dam in Little Falls; slate.

26 553: Thomson Formation; 1.5 miles north of Little Falls at bridge on Little Elk River on Highway 213; SW¼SE¼ sec. 31: 130-29; slaty or phyllitic unit interlayered with fine-grained graywacke. Locality KA-96, Bull. 41, p. 177.

DIABASE DIKES IN NORTHERN MINNESOTA

P. K. Sims and M. G. Mudrey, Jr.

Diabase dikes of Middle Precambrian age are present at many places in the Lower Precambrian rocks north of the Mesabi range. Most are in a northwestward-trending belt, about 60 miles wide, that extends from the western part of the Mesabi range into Ontario, Canada (fig. IV-29). This belt is referred to herein as the main dike belt. A few dikes occur east of the main belt, in the Tower area. The diabases have K-Ar whole-rock and mineral ages ranging from 1,395 to 2,240 m.y. and most are interpreted to have been intruded prior to 2,200 m.y. and at about 2,000 m.y.; the age of others is uncertain because of intense alteration (Hanson and Malhotra, 1971).

MAIN DIKE BELT

The dikes in the main belt trend north-northwest or northwest and are steeply inclined, are as much as 200 feet thick, and have chilled borders. They transect Lower Precambrian metavolcanic, metasedimentary, and granitic rocks, and occupy at least three sets of fractures that were formed in Early Precambrian time as part of the regional fracture pattern. None of the dikes is known definitely to cut rocks of Middle Precambrian age, but some of those in the Nashwauk area (loc. 12, fig. IV-29) may cut the Biwabik Iron-formation as well as the Lower Precambrian rocks.

The dikes have fine- to medium-grained diabasic textures and range in composition from gabbro to hornblende diorite and, locally, to hornblende tonalite (table IV-15). Apparently, most of the dikes initially had the primary mineral assemblage clinopyroxene-plagioclase-opaque oxides. Olivine was a local, additional phase. Subsequent to crystallization, the rocks were altered to different degrees, yielding their present diverse mineralogy. Hornblende formed abundantly in places by pseudomorphously replacing clinopyroxene, and subsequently was partially altered, and plagioclase was moderately to extensively saussuritized. Quartz crystallized late, probably nearly contemporaneously with the hornblende. The rocks were further modified later by mild low-grade metamorphism.

Compositional data were obtained on two samples of diabase from the main dike belt, to aid in understanding the alteration. The sample from locality 2 (fig. IV-29) is a rela-

Table IV-15. Estimated modes, in volume percent, of representative samples of diabase dikes in Lower Precambrian rocks north of the Mesabi range (No. 9 by M. G. Mudrey, Jr.; all others by Ramesh Malhotra).

	Main dike belt					Tower area
	2[1]	7	6	3	4	9
Quartz	Tr	4	7	8	20	Tr
Plagioclase	57	47	43	63	63	54
Clinopyroxene	32	33	33	Tr	Tr	24
Hornblende	2	3	7.5	24	12	
Olivine	Tr					
Opaques	6	4.5	5	3	4	5
Apatite	1	1.5	0.5	1	1	
Alteration minerals	2	7	4	1	C	17
TOTAL	100	100	100	100	100	100
Composition of plagioclase[2]	An_{51-58}	An_{58-70}	An_{53-70}	An_{33-64}	An_{23-47}	Core, An_{57-58} Rim, An_{47-51}

[1] Location of samples given on Figure IV-29
[2] Composition determined by microprobe for samples 2 and 9; all others by Michel-Levy method

(Mn9) 2 Olivine-bearing gabbro; alteration slight; another sample from same locality contains about 5 percent olivine
(Mn12) 7 Gabbro
(Mn8) 6 Gabbro
(Mn6) 3 Hornblende diorite, from quarry west of Cook
(Mn7) 4 Hornblende tonalite
(T24, 217) 9 Gabbro; average of 4 modes from two similar dikes

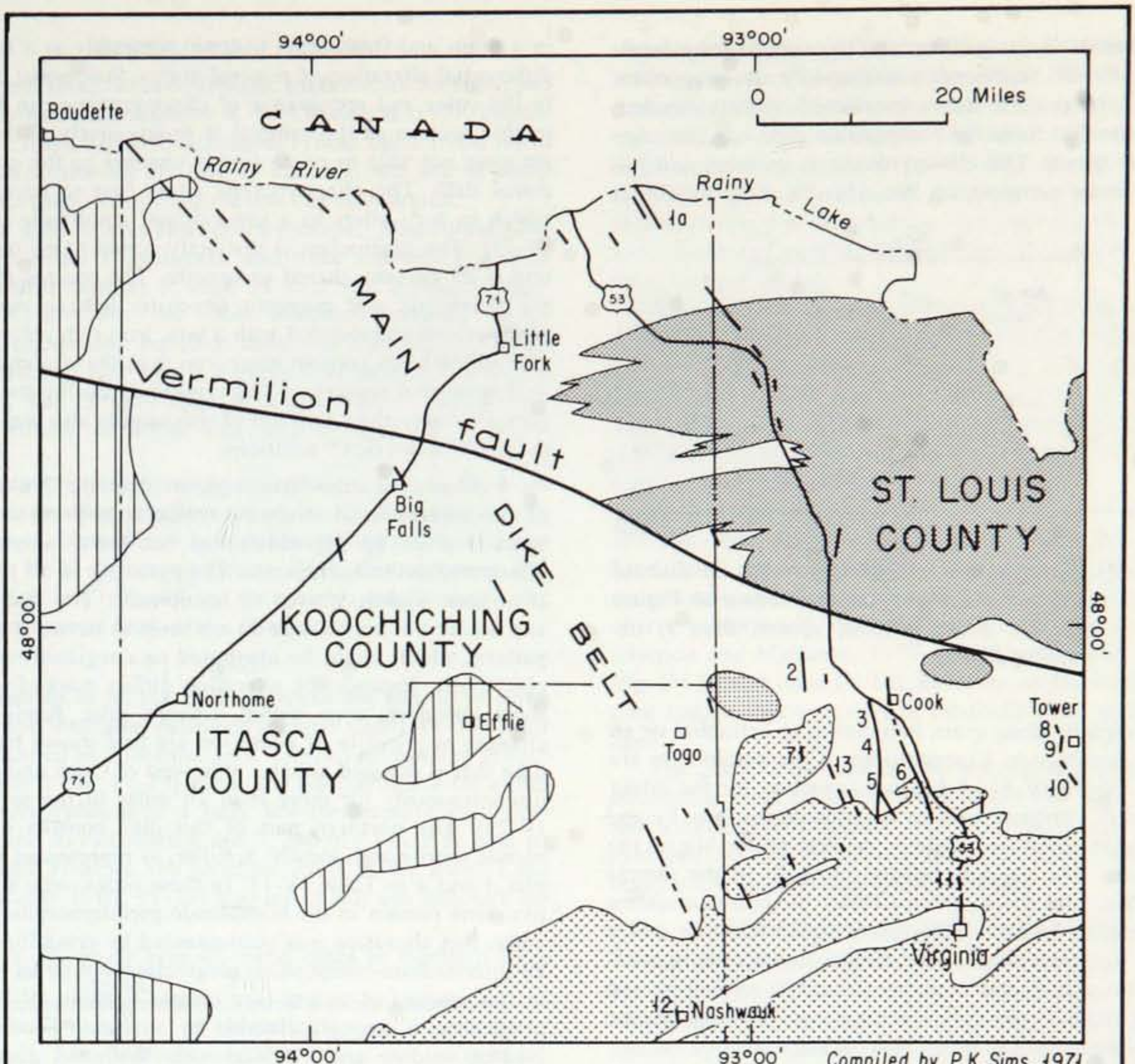

Figure IV-29. Map showing major known diabase dikes of Middle Precambrian age in northern Minnesota.

tively unaltered olivine gabbro, and the sample from locality 13 (fig. IV-29) represents a moderately altered gabbro. In sample 2, plagioclase forms unoriented, slightly clouded, twinned laths that have the composition An_{51-58}. The crystals are not zoned. The clinopyroxene is uniform and has the approximate composition $Wo_{42}Hy_{47}Fs_{11}$ (fig. IV-30). It

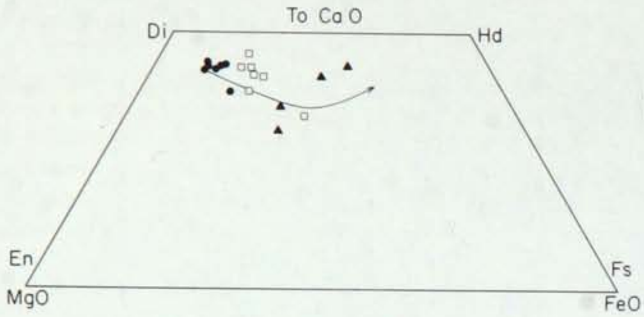

Figure IV-30. Composition of clinopyroxenes in diabase dikes. Location of samples shown on Figure IV-29; circle, dike 2; square, dike 9; triangle, dike 13.

is altered slightly along grain boundaries to actinolite or an actinolitic hornblende. Compositions of the amphiboles are shown on Figure IV-31. A balanced reaction for the direct conversion of clinopyroxene to hornblende involves Fe and H_2O on the reactant side and Ca, Al, and minor Mg on the product side. The primary opaque minerals in the sample are pyrrhotite and ilmenite ($Ilm_{94}Hm_6$), the latter consisting of two separate phases, a dominant ilmenite phase and a dominant hematite phase. The olivine is uniform in composition ($Fo_{60}Fa_{40}$), and is partly altered to serpentine and dusty magnetite. From compositional data on these phases, a balanced equation can be written for the reaction:

$$1.8\ \text{ol} + 2.4\ H_2O + 0.10\ O_2 = 1.2\ \text{serp} + 0.25\ \text{mt}$$

If the alteration of clinopyroxene and olivine were approximately contemporaneous, the iron consumed in the breakdown of clinopyroxene might be provided by this reaction. These reactions are consistent with a late stage, deuteric (?) alteration.

In the more altered sample (locality 13, fig. IV-29) the clinopyroxene varies in composition (fig. IV-30), both with-

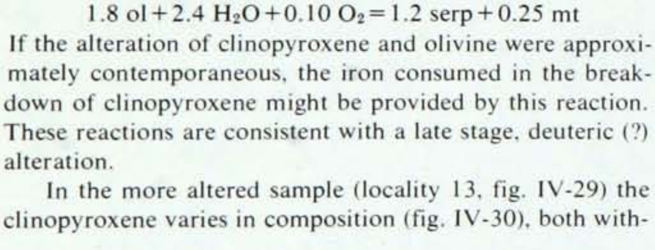

Figure IV-31. Composition of amphiboles in diabase dikes. Location of samples shown on Figure IV-29; circle, dike 2; square, dike 9; triangle, dike 13.

in a grain and from grain to grain, probably as a result of differential alteration of mineral grains. Successive changes in the color and appearance of clinopyroxene can be seen in thin section as the mineral is progressively altered, but we were not able to relate visible changes to the compositional data. The clinopyroxene alters first to hornblende, which in turn alters to a low-calcium amphibole (see fig. IV-31). The plagioclase is distinctly zoned (An_{52} to An_{61}), and is 25 percent altered to sericite. The opaque minerals are pyrrhotite and magnetite-ilmenite. Sphene occurs as small granules associated with a late, iron-rich chlorite. All alteration phases contain more iron than the clinopyroxene, indicating that some iron was consumed during the alteration. Probably the alteration of this sample also was caused mainly by deuteric (?) solutions.

Evidence for autometamorphism (deuteric (?) alteration) of the dike rocks at moderate temperature-pressure conditions is given by the widespread but spotty alteration of clinopyroxene to hornblende. The pyroxene in all the rocks is at least slightly altered to hornblende. The distribution and amount of hornblende do not seem to have a systematic pattern, which might be attributed to a regional metamorphic event. Instead, the alteration differs markedly within short distances, even within a single dike. Rapid lateral changes in intensity of alteration are best shown by a long dike that is exposed about a mile west of Cook and extends discontinuously for more than 10 miles to the south (fig. IV-29). The northern part of this dike consists of hornblende diorite and, locally, tonalite, as represented by samples 3 and 4 in Table IV-15. In these rocks, only shreds of pyroxene remain in the hornblende pseudomorphs. Apparently, this alteration was accompanied by crystallization of an intermediate-composition plagioclase and by an increase in the amount of quartz (see sample 4, table IV-15). The plagioclase is strongly clouded by alteration products, including epidote and a white mica. Within a distance of three miles from the southernmost sample of hornblende diorite (loc. 4, fig. IV-29), the pyroxene in the rock (loc. 5, fig. IV-29) is only moderately altered to hornblende.

Most of the gabbroic rocks contain evidence of further alteration at lower temperatures. Commonly, a green or distinctly bluish-green actinolitic hornblende or actinolite occurs as overgrowths on hornblende or as homoaxial intergrowths with hornblende in the cores of pseudomorphs. Also, dioritic rocks in the dikes near Cook contain a reddish-brown biotite that cuts and is later than the quartz, inasmuch as some of it occurs in fractures in the quartz. Concomitantly, the plagioclase becomes increasingly clouded with secondary products. Whether or not these minerals represent a still lower temperature phase of autometamorphism or are a result of a later, mild retrogressive metamorphism, or both, is not known.

Most samples also contain patchy intergrowths of very fine-grained chlorite and actinolite and lesser sphene, hydrogarnet, leucoxene, and epidote; associated with these minerals are prehnite, calcite, albite, and secondary quartz. Hanson and Malhotra (1971, p. 1110) interpreted these minerals as products of a low-grade metamorphism typical of the prehnite-pumpellyite facies (Coombs, 1960).

DIKES NEAR TOWER

Several small dikes of diabasic gabbro have been mapped in the Tower quadrangle by R. W. Ojakangas (1971, unpub. geologic map, Tower quadrangle). These dikes trend north or north-northwest, are steeply inclined, and are as much as 15 feet thick. They have narrow chilled margins.

The gabbro contains clinopyroxene, plagioclase, and pyrite as major constituents, and trace amounts of quartz and ilmenite (sample 9, table IV-15). The texture is subophitic. Plagioclase, the most abundant mineral, forms cloudy, unoriented, euhedral laths that are twinned, zoned concentrically, and have a composition ranging from An_{58} to An_{47}. The clinopyroxene is a pale brown, faintly pleochroic variety of augite (see fig. IV-30). The pyrite is skeletal.

Alteration of these rocks differs from that in the main belt in consisting almost entirely of the low-temperature assemblage. Hornblende is lacking and actinolite is sparse (see fig. IV-31 for composition). Both clinopyroxene and plagioclase are replaced by the patchy intergrowths; the pyroxene appears fresh where unaltered.

AGE

Inasmuch as all the diabasic rocks are altered to some extent, the available radiometric dates (table IV-16) must be considered as minimum ages. The olivine-bearing gabbro at locality 2 (fig. IV-29) shows the least evidence for either autometamorphism or a later low-temperature metamorphism, and its radiometric age (2,240 m.y.) may be near its actual age. Possibly, the gabbroic rocks in the Rainy Lake area (locality 1, fig. IV-29) also are of this age although interpretation of the ages of these rocks is complicated by the local presence of composite dikes. D. L. Southwick (1971, oral comm.) reported that the dike at the west end of Kabetogama Lake (loc. 1a, fig. IV-29), which appears to be the southeasterly continuation of the dike in the Rainy Lake area that was dated (Hanson, 1968) is composite—a hornblende diorite was intruded into the central part of an older hornblende diorite dike. The younger diorite is more altered than the older diorite.

Hanson and Malhotra (1971) interpreted the hornblende diorite dikes in the Cook area, represented by samples 4 and 6 (fig. IV-29 and table IV-16), probably as having been emplaced about 2,000 m.y. ago. Alternatively, they may belong to the older dike set, for they have been thoroughly modified by autometamorphism (deuteric (?) fluids) and, later, also by mild cataclasis.

Aside from samples from localities 11 and 12 (fig. IV-29), the other diabases from the main dike belt that have been dated are moderately or intensely altered, and their radiometric ages ". . . may more closely represent the time of recrystallization than the time of emplacement . . ." (Hanson and Malhotra, 1971, p. 1111). Samples 11 and 12 (fig. IV-29 and table IV-16), however, are little altered, and their radiometric ages may be relatively close to their actual ages.

The dikes in the Tower area differ petrographically and chemically from those in the main dike belt and may represent a distinct set emplaced about 1,700 m.y. ago. They lack evidence for autometamorphism and have been affected only by low-temperature metamorphism.

Table IV-16. K-Ar ages of mafic dikes in northern Minnesota.

Map No.[1]	Sample No.[2]	Material Analyzed	Age (in m.y.)
	Main Belt		
[3]1	M7052c	Whole-rock	2,130
		do.	2,040
2	Mn9	do.	2,240
4	Mn7	Hornblende	1,980
		Whole-rock	1,980
6	Mn8	Pyroxene+ Hornblende+ Biotite	1,950
		Whole-rock	1,770
7	Mn12	Whole-rock	1,740
11	Mn13	Whole-rock	1,630
12	Mn14	Whole-rock	1,395
	Tower Area		
8	Mn3	Whole-rock	1,685
9	Mn4	do.	1,570
10	Mn5	do.	1,520

[1] Location of samples given on Figure IV-29
[2] Refers to sample no. in original report
[3] Source: Hanson, 1968; all others from Hanson and Malhotra (1971)

DISCUSSION OF METAMORPHISM

As pointed out by Hanson and Malhotra (1971, p. 1110), all the diabases in northern Minnesota having radiometric ages greater than 1,300 m.y. show some evidence of metamorphism; diabase and basaltic dikes having younger ages lack such features. They (p. 1110) suggested that the ". . . low-grade metamorphism typical of the prehnite-pumpellyite facies may have been associated with burial by the overlying sediments of the Animikie Group." More probably, this metamorphism was related to one or more episodes of mild deformation and metamorphism that occurred in Middle Precambrian time. That mild deformation was involved is indicated by the local cataclasis in the diabases. Neither the extent of the cataclasis and related (?) alteration nor its exact age (or ages) is known, but cataclasis seems to be more intense in dikes near the Mesabi range and to decrease in intensity northward. Possibly, it was in part related to the same mild deformation—manifested by slickenside striae and corrugations along bedding planes—that affected the Biwabik Iron-formation, described previously by Morey (this chapter).

Zeck (1971) has described similar prehnite-pumpellyite facies metamorphism from Precambrian quartzofeldspathic gneisses and granitoid rocks in Sweden. From this occurrence and other described occurrences in Precambrian basement rocks, he has demonstrated that the prehnite-pumpellyite facies is not exclusively developed by burial metamorphism. Apparently, it is a common type of low-grade retrogressive metamorphism in older Precambrian rocks.

MINNESOTA RIVER VALLEY

G. B. Morey

Several small plutons, ranging in composition from gabbro to granite, and a variety of aphanitic dike rocks were emplaced in the older terrane of the Minnesota River Valley (Goldich and others, 1970; Grant, this volume) during Middle Precambrian time. In addition, about 1,850 m.y. ago a low-grade metamorphic event of sufficient intensity to recrystallize biotite in the older rocks occurred over a wide area between Granite Falls and Ortonville in the valley. Goldich and others (1970, p. 3690) have pointed out that this event ". . . should not be equated with the Penokean orogeny to the north and northeast in the sense of a period of folding and metamorphism. . . ." However, the discussion is included here because these events fall within the time-span that defines the Middle Precambrian.

DIKE ROCKS

Three distinct kinds—tholeiitic diabase, hornblende andesite, and olivine diabase—of dike rocks are exposed in the vicinity of Granite Falls and Montevideo (Himmelberg, 1968). All the dike rocks transect the foliation of the older rocks, which were metamorphosed about 2,650 m.y. ago (Goldich and others, 1970). The rocks were intruded preferentially along a nearly vertical fracture set that strikes about N. 55° E.; a few occupy a nearly eastward-trending set (fig. III-80). At several localities, older tholeiitic diabase dikes are crosscut by different varieties of hornblende andesite dikes that commonly have chilled margins against the older dikes. Furthermore, the older tholeiitic diabase dikes are cut by shear zones, but the hornblende andesite dikes cut across the shear zones (Himmelberg, 1968). Age relationships of the olivine diabase dikes were not determined in the field.

Tholeiitic Diabase

The tholeiitic diabase dikes are as much as 75 feet thick, are dark gray and medium grained, and have fine-grained borders. Principal minerals include plagioclase (An_{50}), intergranular augite, and opaque oxides. Minor amounts of apatite, quartz, and epidote are present, and semi-fibrous green hornblende commonly mantles or completely replaces augite. Hanson and Himmelberg (1967) suggested that the dikes may have been emplaced about 2,080 m.y. ago, but that date also may represent an intermediate age as a result of partial loss of argon during a younger 1,800 m.y. thermal event.

Hornblende Andesite

Hornblende andesite dikes are generally gray or black, aphanitic, porphyritic rocks of variable texture and composition. The phenocrysts are generally plagioclase and/or anhedral quartz. Principal matrix minerals include plagioclase (An_{40-60}), hornblende, and biotite. Minor amounts of oxides, apatite, interstitial quartz, and potassium feldspar are present as are secondary chlorite, epidote, calcite, and sericite. Himmelberg (1968) observed one dike that contained microcline. Hornblendes from these dikes give K-Ar ages mostly between 1,690 and 1,730 m.y., but one age of 1,930 m.y. was obtained (Hanson and Himmelberg, 1967). Biotites give K-Ar ages of 1,770 and 1,800 m.y. These values confirm field evidence indicating that these dikes are younger than the tholeiitic diabase dikes.

Olivine Diabase

Olivine diabase dikes are black aphanitic rocks composed of olivine, plagioclase (An_{65-70}), and augite microphenocrysts in a matrix of plagioclase microlite, granular pyroxene, and sparse opaque oxides. Coronas of colorless and green amphiboles occur around the olivine, and some coronas contain green spinel and orthopyroxene. The olivine diabase dikes have not been dated.

PLUTONIC ROCKS

A small, structurally discordant adamellite pluton, the "late granite" of Lund (1956, p. 1489) or the "late granite of section 28" of Goldich and others (1961, p. 140), crops out just north of Granite Falls (fig. III-80), where it crosscuts granitic gneiss and a hornblende andesite dike. Also, in the area south and southeast of Franklin, several small, discordant, stock-like plutons of gabbro, diorite, and granite intrude gneissic country rocks. The largest of these crops out as a small knob, locally known as Cedar Mountain, along the N.-S. section line between sections 14 and 15, T. 112 N., R. 34 W.

Adamellite

The adamellite is a pink, medium-grained hypidiomorphic-granular rock that locally contains xenoliths of hornblende andesite (Himmelberg, 1968). Principal minerals include plagioclase, microcline, quartz, and biotite. The plagioclase is euhedral to subhedral and has marked zoning and polysynthetic twinning. Microcline is anhedral, relatively unaltered and slightly microperthitic. Myrmekite is present but not abundant. Biotite is dark brown to yellowish brown and commonly is altered to green chlorite.

Zircons from this body give a Pb^{207}/Pb^{206} age of 1,825 m.y. (Catanzaro, 1963); Rb-Sr analyses of the two whole rocks and their respective K-feldspar samples define a $1{,}830 \pm 160$ m.y. isochron (Goldich and others, 1970, fig. 12). K-Ar and Rb-Sr mineral ages on micas range from 1,650 m.y. to 1,990 m.y. (Goldich and others, 1961). Radiometric ages for the hornblende andesite and the adamellite agree within the analytical uncertainty, indicating that the adamellite body was emplaced shortly after the andesite dikes.

Gabbro, Diorite, and Granite of Cedar Mountain Type

Several small stock-like bodies of gabbro, diorite, and granite occur in the Franklin area (fig. III-80). The Cedar Mountain complex, the largest of these bodies, is slightly less than half a mile in diameter and has been described in detail by Lund (1956), Bury (1958, unpub. M.S. thesis, Univ. Minn.) and Goldich and others (1961, p. 131-134).

Gabbro diorite comprises the bulk of the rock in Cedar Mountain and forms an outer shell 500 feet thick surrounding a core of granite, which is considered to be a differentiate; however, the gabbro-granite contact has not been observed. The border phase of the intrusion consists of chilled, fine-grained olivine trachybasalt porphyry. The trachybasalt grades inward, with increasing grain size, to a granophyric gabbro diorite. Well-developed, nearly vertical flow structures consisting of alternating dark and light layers characterize this part of the complex. The layers are composed of various proportions of andesine, augite, biotite, hornblende, and abundant quartz and orthoclase in granophyric intergrowth.

Granite from the core of the complex consists of albite, orthoclase, quartz, biotite and accessory minerals. The quartz is largely in a micrographic intergrowth with the orthoclase.

Small dikes of granite cut the gabbro and represent magma that extended outward from the central part of the intrusion along fractures, suggesting that the core was still fluid after the rim crystallized. Because of the relatively large amount of granite in the core, Goldich and others (1961) concluded that the complex formed by multiple intrusions of magma which differentiated at depth.

The Cedar Mountain rocks have been dated by the K-Ar method; the ages on biotite (Goldich and others, 1961, p. 135) and hornblende (Hanson, 1968, p. 5) are both 1,750 m.y. This age may not be the actual time of emplacement, but it is probable that these rocks are ". . . nearly contemporaneous with the hornblende-andesite dikes near Granite Falls" (Hanson, 1968, p. 14).

AITKIN COUNTY SULFIDE DEPOSITS

G. B. Morey

Since the early 1900's, significant concentrations of sulfides have been known to be present in Aitkin County. Interest in these deposits as a possible source of sulfur has been sporadic since that time, but no commercial operation has resulted. Although there are several known deposits, those in sections 14, 15, 28, and 29, T. 47 N., R. 25 W. (Glen Township) have been explored most intensively (fig. IV-32). The Glen Township deposit has been studied by Thiel (1924), Schwartz (1951), and Han (1968). The earlier workers concluded that the sulfides were of epigenetic origin related to nearby metadioritic intrusions. Han (1968), however, demonstrated that the sulfides primarily are syngenetic, a conclusion in accord with Schwartz's later opinions.

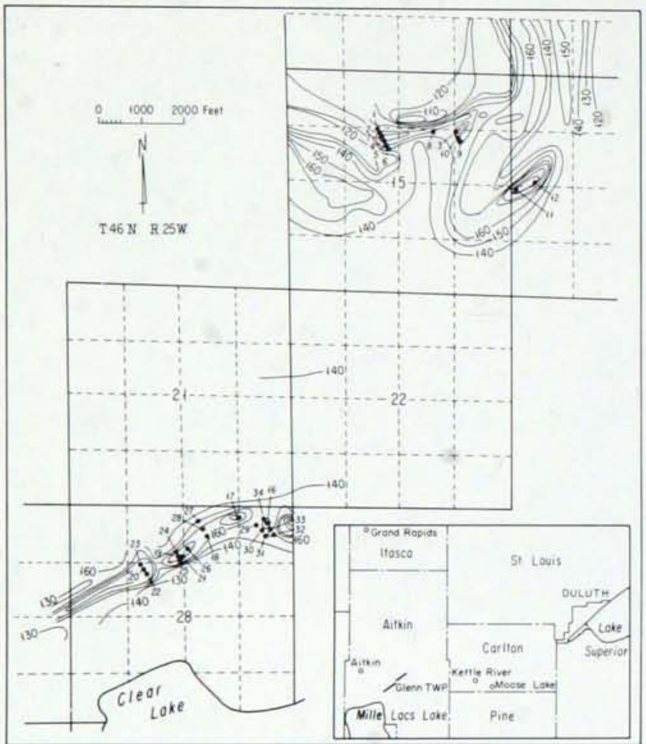

Figure IV-32. Generalized map showing the magnetic anomalies and location of drill hole in the Glen Township sulfide deposit (modified from Han, 1968).

The iron sulfides in Glen Township occur in two separate tabular bodies that have been delineated by ground magnetic surveys (fig. IV-32). The deposits, which originally consisted of a pyritic and carbonaceous shale and intercalated carbonate facies iron-formation, are within a thick succession of Animikie shale and graywacke assignable to either the Rabbit Lake or Thomson Formations. Subsequently, the rocks were locally modified by the intrusion of small plutons, dikes, and sills of diorite, and were regionally metamorphosed during the Penokean orogeny. They now consist of slate, metagraywacke and, locally, phyllite or schist, and indicative mineral assemblages are assignable to the lower part of the greenschist facies of metamorphic grade. The rocks, including the diorite intrusions, have been deformed and have a northeastern trend in accord with the regional structure of east-central Minnesota.

Within the Glen Township deposit, sulfide-bearing black slate and recrystallized carbonate or magnetite facies iron-formation are interbedded on all scales; however, the black slate is predominant in the upper part and the iron-formation in the lower part of the section, as shown in Figure IV-33. According to Han (1968, p. 111), the sulfide-bearing rocks originally consisted of carbonaceous material and admixed pyrite, whereas the iron-formation consisted of iron-rich carbonates and chert. The ore mineralogy has been complicated, however, by metamorphism and subsequent supergene alteration. Pyrite and pyrrhotite are the

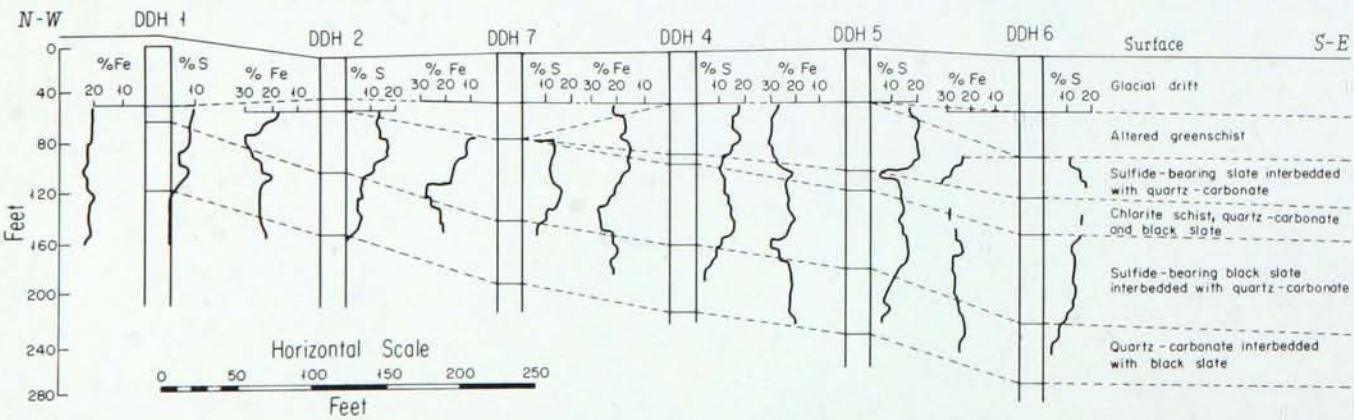

Figure IV-33. Geologic cross-section across the north ore body showing the relationship between the stratigraphic succession and the distribution of iron and sulfur (modified from Han, 1968).

major ore minerals; marcasite and magnetite are subordinate, and chalcopyrite, sphalerite, arsenopyrite, covellite, hematite, and goethite are rare (fig. IV-34). Although all the

<table>
<tr><th>IRON-FORMATION</th><th>PRE-METAMORPHIC</th><th>METAMORPHIC</th><th colspan="3">POST-METAMORPHIC (Supergene)</th></tr>
<tr><td rowspan="9">SULFIDE</td><td rowspan="9">PYRITE I in Carbonaceous Sediments</td><td rowspan="6">PYRRHOTITE</td><td>PYRITE III</td><td colspan="2">MARCASITE II</td></tr>
<tr><td>PYRITE III</td><td colspan="2">Magnetite III</td></tr>
<tr><td colspan="3">MAGNETITE II Unchanged</td></tr>
<tr><td rowspan="2">MARCASITE I</td><td>PYRITE IV</td><td>MARCASITE II</td></tr>
<tr><td colspan="2">Magnetite III
Iron Carbonate
Goethite</td></tr>
<tr><td>Goethite</td><td colspan="2">Unchanged</td></tr>
<tr><td>PYRITE II
Sphalerite
Arsenopyrite</td><td colspan="3">Unchanged</td></tr>
<tr><td>Chalcopyrite</td><td colspan="3">Covellite (in highly oxidized ore)</td></tr>
<tr><td>Ilmenite</td><td colspan="3">Leucoxene</td></tr>
<tr><td rowspan="4">CARBONATE</td><td rowspan="3">IRON-RICH CARBONATE</td><td>Pyrite II</td><td colspan="3">Unchanged</td></tr>
<tr><td>Pyrrhotite</td><td colspan="3">Same type of alterations as shown above</td></tr>
<tr><td>Magnetite I</td><td colspan="3">Hematite</td></tr>
<tr><td>Carbonate-Chert</td><td>Carbonate</td><td colspan="3">Goethite</td></tr>
</table>

Figure IV-34. Genetic relationship of ore minerals in the Glen Township sulfide deposit (modified from Han, 1968). Note that all major ore minerals are shown in capital letters.

rocks contain sulfides, the black slate appears to be the chief host for the sulfide minerals.

These sulfide deposits are considered a potential source of sulfur for use in the manufacture of sulfuric acid. Bleifuss and others (1963) concluded that one ton of sulfuric acid requires 1,600 pounds of 40 percent sulfur or 4,750 pounds of crude ore averaging 13.5 percent recoverable sulfur. An annual ore production of nearly 475,000 tons of crude ore of that grade would be required to support production of 200,000 tons of acid per year. If a mine operates five days a week or 250 days a year, the mine capacity would have to be 1,900 tons per day, and if a 20-year life operation is assumed, an ore reserve of at least 9,500,000 tons must be available.

Drilling in Glen Township has outlined a quantity of ore sufficient to meet the above requirements. Needham (1955) estimated that approximately 36 million tons of crude ore averaging 23 percent iron and 15 percent sulfur were present. Schwartz (1965, files of the Minn. Geol. Survey) refined these estimates somewhat by defining two separate ore bodies and calculating grade and tonnage individually for each body, using an ore cutoff of 8 percent sulfur. The north ore body, consisting of a simple, tabular ore zone dipping about 20° SE., was estimated to contain 13,766,000 tons of measured, indicated, and inferred ore averaging 13.9 percent sulfur. The depth to which ore was estimated is about 360 feet; undoubtedly, the body extends some distance further downdip. Estimation of the ore reserves in the south ore body was more difficult than for the north ore body because the structure appears to be complicated and because there is considerable variability in the apparent grade. Schwartz estimated approximately 25 million tons of ore averaging 13.8 percent sulfur in the plus 8 percent ore zone. Thus, in Glen Township there are about 39 million tons of proven ore reserves averaging 13.8-13.9 percent sulfur.

Metallurgical testing of the sulfide ores has been summarized by Pennington and Davis (1953) and by Bleifuss and others (1963). Using a crude ore containing 32.71 percent iron and 19.9 percent sulfur, Bleifuss and others (1963) obtained a magnetic concentrate containing 57.35 percent iron and 37.92 percent sulfur. Subsequent roasting of the magnetic concentrate yielded an iron-rich concentrate containing approximately 66 percent iron, 1.5 percent silica, and 0.008 percent phosphorus. However, the magnetic separation technique that was used concentrates only the pyrrhotite, and because the crude ores have pyrrhotite/pyrite ratios ranging in value from 1:1 or 2:1 to 4:1, only a small part of the potential sulfur-bearing minerals were separated. Using a crude ore containing an average of 21 percent iron and 13.2 percent sulfur, Pennington and Davis (1953) obtained a flotation concentrate containing 36.40 to 46.6 percent sulfur, 46.96 to 53.37 percent iron, and 1.36 to 8.16 percent silica. Sulfur and iron recoveries averaged 87 and 76 percent of the original ore, respectively. Thus, metallurgical testing indicates that a suitable sulfur concentrate can be obtained from the Glen Township ores. In addition, a high-grade iron oxide by-product would be available after the sulfur has been recovered.

Because chalcopyrite, sphalerite, and arsenopyrite are present in small amounts (fig. IV-34), these ores were evaluated as a potential source of copper and zinc. Analyses of crude ore show that the zinc content ranges from 0.003 percent to 0.107 percent, and averages 0.029 percent ($N=63$). After flotation, zinc averaged 0.086 percent ($N=29$) in the concentrate, with values ranging from 0.022 to 0.251 percent. Copper was not analyzed in the crude ore, but averages 0.16 percent ($N=8$) in the flotation concentrates, with values ranging from 0.04 to 0.47 percent. Microscopic analyses of the flotation concentrate indicate that the sphalerite is present either as a free mineral or in combination with gangue. In contrast, chalcopyrite occurs in combination with iron sulfides even where the concentrate is finer than 200 mesh. Attempts to remove zinc and copper from the concentrates have not as yet been successful.

EVIDENCES OF PRECAMBRIAN LIFE IN MINNESOTA

David G. Darby

Four principal types of more or less unequivocal fossils have been found in strata of Precambrian age: algal stromatolites, algal cells, bacterial cells, and impressions of metazoa. All four types have been reported from Precambrian rocks of Minnesota. However, not all the structures reported are of reasonably certain biologic origin. Such problematica as "worm tubes," "fucoids," and other megascopic trace fossils attributed to animal activity, to my knowledge, have not yet been reported from the Precambrian rocks of this state. Some nearby occurrences of such supposed traces of metazoans have been reported from strata of Middle Precambrian age from Michigan (Faul, 1950) and Ontario (Hofmann, 1967). Since few geologists working on Precambrian rocks are well acquainted with the nature of the organisms most commonly responsible for fossils of that age, a brief explanation seems appropriate.

ALGAL STROMATOLITES, ALGAE, AND BACTERIA

Probably the most fundamental division between groups of living organisms is on a cellular level. Many unicellular and all higher plants and animals are composed of eukaryotic cells. The principal visible characteristic of a eukaryotic cell is a nucleus enclosed within a membrane. Two large groups of organisms which do not have this type of cell, and which are now placed within a separate Kingdom, Monera, are the true bacteria and the blue-green algae (Cyanophyceae or Myxophyceae, various authors). These organisms have a prokaryotic cell in which there is no membranated nucleus or chromosomes. Prokaryotic organisms are generally regarded as more primitive and as evolutionarily conservative. Although the word "primitive" may be applied to their cellular form, it does not imply simplicity. Many are capable of photosynthesis (although the bacteria produce no oxygen), many are capable of motion, and their habitats include a wide range of terrestrial, freshwater and marine environments. In general, prokaryotic organisms have a higher tolerance of environmental extremes than other organisms, and their growth has been observed under temperature ranges of less than 0° C to more than 90° C. All the reasonably certain fossils of Precambrian age from Minnesota involve prokaryotes.

Algal Stromatolites

In 19th century publications, occasional references can be found to strangely shaped, concentrically laminated structures in rocks of Precambrian and later age. The first of these references in North America appears to be that of Steele (1825), who reported ". . . calcareous concretions of a most singular structure. . . ." In the same publication Steele described the overall morphology and the nature of the concentric laminae of the structures in rocks from New York state. Later, Bell (1870) reported ". . . vertical cylinders of chalcedony, transverse sections of which shew fine concentric rings resembling agate." In the 19th century, some of these laminated structures found in rocks of Paleozoic and Precambrian age were considered possibly the result of the metabolic activity of animals of a coralline or foraminiferal nature and were given taxonomic designations, *e.g.*, *Cryptozoon* (Hall, 1884) and *Eoozoon* (Dawson, 1865; Matthew, 1890). Subsequently, opinions concerning the structures began to shift from "probably inorganic" to "possibly organic" in origin. In 1903, Leith commented on and illustrated some of these curious structures from the basal part of the Biwabik Iron-formation in Minnesota. He made no mention of a possible organic origin, but compared them with ". . . contorted lines resembling the flowage lines in the matrix of a vitreous lava." The figures of Steele and Leith leave no doubt that they found what now would be termed algal stromatolites (fig. IV-35A). As early

Figure IV-35. Microstructures from the Pokegama Quartzite and Biwabik Iron-formation and pseudofossils from the Sioux Quartzite. A, reproduction of the first published figure of an unequivocal fossil from the Precambrian rocks of Minnesota (Leith, 1903). Algal stromatolites from the basal part of the Biwabik Iron-formation, x0.9. B, "fossils" from the Sioux Quartzite. Reproduced from Winchell (1885). "Brachiopods" x2, "trilobite" x0.5. C, Winchell's specimen 5559 from which the "brachiopods" in Figure A were drawn, x0.7. D, algal stromatolites of the "gymnosolen type" from the Biwabik Iron-formation; Mary Ellen mine, x1. *cf.* Form B, Hofmann (1969); *Gruneria biwabikia* Cloud and Semikhatov (1969); *Collenia* (?) *ferrata*, Grout and Broderick (1919b). E, horizontal thin section of same type as D, showing links between columns, x3. F, vertical thin section of stromatolite from the Gunflint Iron-formation near Kakabeka Falls, Ontario, showing layering, oncolites in interspaces and links between columns, x5. G, thin section showing oncolites, partially replaced by hematite, from the Biwabik Iron-formation, x30. H, microstructures as figured by Gruner (1922, 1924), "Algae resembling Microcoleus." In chert from the Pokegama Quartzite, x250. I, microstructures, same source as H, "Blue-green algae." Possibly trails of migrated pyrite crystals, x2000. J, trail of migrated pyrite grain in chert from Biwabik Iron-formation. Courtesy E. S. Barghoorn, x1500. →


J
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C
1 cm.
1 in.
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D
A

work progressed, studies on modern calcareous algal structures by biologists gave the geologists their first insights into the nature and paleoenvironmental implications of the structures they had seen and described. Not only were the algal origins recognized, but Walcott (1914) illustrated retouched photomicrographs of some of the individual algal cells. Despite the work of a few geologists such as Dawson in Canada, and Walcott and later Gruner in the United States, reports of fossils in Precambrian rocks were, for the most part, relegated to the "probably inorganic" realm until the latter half of this century.

The term "stromatolite" derives from "stromatolith," originally proposed by Kalkowski (1908). The structures were known in Germany long before the name was proposed; a local name, "Napfstein" or "bowlstone," had been in use because the structures in limestone were split along their concentric laminae, inverted, and used in households. Kalkowski's figures leave no doubt that he was describing a structure we now would consider as reflecting organic activity, but the term "stromatolite" also has been used for structures of inorganic origin. To clarify any interpretative problems, many authors have used the term "algal stromatolite," which Logan and others (1964) defined as follows: "Algal stromatolites are laminated structures composed of particulate sand, silt, and clay-size sediments, which have been formed by the trapping and binding of detrital sediment particles by an algal film." This definition generally is sufficiently accurate for modern stromatolites that commonly form in the following way: algae, most frequently blue-green, adhere to a solid surface in the littoral zone. As particulate-bearing waters periodically cover the algal "mat," some of the sediment settles on it. As the water withdraws, the particles are trapped and retained among the algae which then grow or work their way through to the new surface. In this way a laminated structure forms and grows. The gross form of the structure has been found to be more indicative of the environment rather than of the species of algae involved (Logan, 1961). This observation is of value in attempting to understand the paleoenvironment, but puts stringent theoretical limitations on the use of stromatolites for age correlations.

Most ancient algal stromatolites were formed of carbonate rocks; however, in Minnesota, the only known ones of Precambrian age are in the cherty members of the Biwabik Iron-formation. There is little evidence that the algae trapped individual particles of sediment. Also there is little evidence of the algal cells themselves within the stromatolitic structure. The best preserved cellular remains are in the more normally bedded cherts associated with the stromatolites. It is clear that algal stromatolites are not fossilized organisms, nor do they appear to be confined to sites of clastic particulate deposition. A definition somewhat revised from those expressed in some recent works is: algal stromatolites are three-dimensional, laminated, organo-sedimentary structures generally undulose, domal, or columnar in cross-section, which are formed through the influence of algae, predominantly blue-green; the algae exert control on deposits of clastic sediments or precipitates, and may also contribute material through their own biochemical activity.

Some stromatolites form under the influence of one species of alga, whereas others may have many species within the algal "mat." In the absence of remains of the algae themselves, it has been and is a problem whether or not algal stromatolites should receive generic and specific identification. Even in a moderately strict sense, it is clear that they should not. Yet, this has been done. Today, workers still are divided in their acceptance of any particular form of nomenclature along the lines of the Linnean system. Some have proposed alphabetical systems; others have used "generic" and "specific" names, terming them "group" and "form," respectively, or the binomial terms may be called simply "form-taxa." The reader is referred to Logan and others (1964), Raaben (1969), Cloud and Semikhatov (1969), and Hofmann (1969) for an excellent summary of the present nomenclatural positions.

The conclusion, from the studies of modern algal stromatolites, that environment is far more significant than the algal species in producing the structural form, may not always have been the case. Soviet geologists have been successful, as indicated by radiometric concordant ages, in correlating late Precambrian stromatolites, especially the columnar forms, between regions of the U.S.S.R.

Algal Cells

The principal species of blue-green algae active in stromatolite development are filamentous in form. Cells, generally 10 microns or less in diameter, join together in a chain, termed a "trichome," which may be hundreds of microns long. A second general cell form is the individual spherical cell (coccoid form) which also is generally 10 microns or less in diameter. Either form may be surrounded by a sheath of mucilaginous material, which if permineralized by opaque minerals might indicate a diameter in excess of the cell or trichome diameter.

I have measured mats of living filamentous algae more than two cm thick near thermal springs; some mats in marine environments are of this magnitude (Davies, 1970), but they may be only a millimeter or so thick (Ginsburg, 1955; Logan, 1961). The number of individual algae involved in the formation of a stromatolite is immense. In modern environments the coccoid forms apparently are of lesser importance in forming the stromatolites; perhaps the intertwined filamentous species act as a more efficient sediment trap.

Bacteria

Bacteria commonly are found in association, apparently symbiotic, with modern blue-green algae (Castenholz, 1969; Brock, 1969). Work on modern marine stromatolites generally has not included data on the bacterial population. Bacteria are most commonly 1 to 3 microns in size, although much larger ones are known, and they have three general cell forms: rod-shaped (bacilli), spherical (cocci) and, less commonly, spiral (spirilla). The first two forms have been reported from Precambrian rocks by Walcott (1915), Schopf and others (1965), and Barghoorn and Schopf (1966). Some bacteria, like blue-green algae, are capable of motion, can form elongate cell-chains, and can be enclosed within an elongated sheath. These sheaths may become iron-encrusted through the precipitation of ferric hydroxide during the oxidation of ferrous iron by certain sulfur and iron bacteria (Brock, 1970), and thus could resemble fossilized filamentous blue-green algae. The mineralized sheaths are stable, and therefore could be preserved as hollow filaments, the bacterial cells having decayed.

The prokaryotic cell types of bacteria and blue-green algae have led to the theoretical conclusion that these organisms evolved on earth at a very early time. To date, this conclusion has been borne out by the fossil evidence. The reports of bacteria of Precambrian age are more debatable than those of algae inasmuch as the bacteria generally are much smaller and little more than their gross morphology can be examined. Nevertheless, their presence as fossils is generally accepted.

PRECAMBRIAN FOSSILS IN MINNESOTA

Upper Precambrian

Sioux Quartzite

The first reference to Precambrian "fossils" in Minnesota was by N. H. Winchell (1885). He reported the presence of molds and impressions in the catlinite (mudstone/argillite) from the lower part of the Upper Precambrian Sioux Quartzite at Pipestone (see Austin, this volume). He illustrated and identified these "fossils" as a linguloid inarticulate brachiopod (*Lingula calumet* n. sp.) and a trilobite (*Paradoxides barberi* n. sp.). Probably being uncertain of their organic nature, Winchell reinforced his opinions by quoting from letters of two other geologists who had examined the material and had substantially supported his views. In addition, a chemist also reported calcium phosphate from what appeared to be white shell remnants. C. D. Walcott, who at that time was the most eminent invertebrate paleontologist in the United States, commented on Winchell's find in 1899: "The latter [the trilobite impression] I have examined carefully and have concluded that it is of inorganic origin. The Lingula-like forms are so obscure that it is difficult to tell whether they are of organic origin or not. The weight of eidence is in favor of their being small flattened concretions that in some specimens have the appearance of a crushed Obolus or Acrothele."

Judged from our current knowledge of the evolution of metazoans and the age of the Sioux Quartzite, which is at least 1.2 b.y. old (Goldich and others, 1959), Walcott's view probably is correct. I have examined the original specimens from which Winchell's figures were derived (Univ. Minn. Paleo. Coll. nos. 5555 and 5559). Figure IV-35B is a photographic reproduction of Winchell's plate (1885) showing the "fossils." The concentric "growth lines" shown in his drawings taken from specimen #5559 (fig. IV-35C) are overemphasized from any similar lines that now appear on the specimen. Those concentric lines that are present seem to be irregular, and resemble small exfoliations on the slightly convex surfaces. Specimen #5555 looks no more like a trilobite than Winchell's illustration of it. Although the "trilobite" probably is a current feature or load structure, the circular to ovoid structures cannot be dismissed as readily. They could be organic in origin, but no firm decision can be made on the basis of the known specimens. Miller (1961, unpub. M.S. thesis, Univ. Minn.) reported ripple marks and cross-beddings in the catlinite, and it is possible that the structures are compressed mud pellets formed by wave or current action.

Unless one accepts Winchell's specimens as fossils, direct evidence of Precambrian animal life is not known from Minnesota. The significant reports of metazoans from other areas, such as Australia (Glaessner and Wade, 1966) and Newfoundland (Misra, 1969), have been from uppermost Precambrian strata. Such metazoans could occur in the Upper Keweenawan formations of this state, *e.g.*, the Hinckley Sandstone.

Middle Precambrian

Biwabik Iron-formation

Algal stromatolites from the Biwabik Iron-formation were figured first in 1903 by Leith (fig. IV-35A), but were not attributed to organic processes. Grout and Broderick (1919b) illustrated two forms of algal stromatolites from the upper and lower cherty members of the Biwabik Iron-formation. These were considered organic in origin and were given taxonomic identities. The larger columnar forms, 2 inches to 2 feet in diameter, from the lower member were referred to *Collenia (?) biwabikensis* (n. sp.) and those from the upper member, a more finger-like columnar form, were referred to *Collenia (?) ferrata* (n. sp.). Probably, the finger-like forms would have been referred to the "genus" *Gymnosolen*, erected by Steinmann (1911), had this Finnish publication been available to Grout and Broderick. *Collenia* had been erected by Walcott (1914) on material from the Precambrian Belt Series in Montana. Both forms of algal stromatolite have been reported by Hofmann (1969) to occur in both the upper and lower algal chert facies of the Gunflint Iron-formation in Ontario, a unit correlative with the Biwabik Iron-formation. The form most common to the upper zone of the Biwabik (fig. IV-35D), and found in both facies of the Gunflint—the *Collenia ferrata* of Grout and Broderick (1919b)—apparently is the same form as *Gruneria biwabikia* (n. group, n. form) of Cloud and Semikhatov (1969) and Form B of Hofmann (1969).

The Biwabik and Gunflint Iron-formations also contain numerous spheroidal, generally concentrically laminated, "oolitic" structures. These commonly are found between the columns of algal stromatolites (figs. IV-35F and G). If these structures are the result of algal activity rather than inorganic processes, they are referred to as oncolites, and may be thought of as small "free" stromatolites. Twenhofel (1919) erected the genus *Osagia* to identify similar structures occurring in rocks of Paleozoic age. Distinguishing oolites from oncolites is a problem. Both tend to form in agitated shallow waters of the littoral zone, and both may have a detrital grain as a nucleus. The spherules of the Biwabik and Gunflint Iron-formations generally are siliceous and the silica may be primary; at least many have no evidence of complete replacement. This would indicate a possible paleoenvironment not extant today, for neither siliceous oolites nor oncolites are known to be forming at present. No firm criteria to distinguish oncolites from inorganic spherules are yet known, but oncolites seem to be more elliptical rather than spherical and the laminae tend to be overlapping and of more variable thickness (Logan and others, 1964; Hofmann, 1969). In the Biwabik Iron-formation, the form of the spherules and their common association with algal stromatolites suggest that they are oncolites.

Cellular remains from the Biwabik Iron-formation were first reported by Gruner in 1922. He illustrated several possible fossil bacteria and algae, which were found in Precambrian rocks near Eveleth, Minnesota (figs. IV-35H and I; IV-36A and B). He reported similar cellular structures from cherts of the stromatolitic zones of the Biwabik and Gunflint Iron-formations, but his figured specimens come from chert pebbles and fracture fillings in the directly underlying Pokegama Quartzite. Cloud and Licari (1972) subsequently have found well-preserved microfossils in the Pokegama. The figures in Gruner's 1922 paper were republished in 1924. The magnifications indicated in the two reports differ, and herein, I have used the magnifications given in the 1924 report. Cloud (1965) has commented that some of the structures figured by Gruner may be trails produced by pyrite crystals which migrated through the chert. Such trails had been shown by Tyler and Barghoorn (1963) to occur in the age-equivalent cherts of the Gunflint Iron-formation (fig. IV-35J). In cases where structureless threads or tubules are found, their production by crystal migration must be considered.

Impetus for a renewed search for Precambrian microfossils came with the publication of the paper by Tyler and Barghoorn (1954). They reported and illustrated microscopic structures ". . . representing both blue-green algae and simple forms of fungi . . ." from cherts of the Gunflint Iron-formation in Ontario. The same authors later expanded this study and produced what appears to be unequivocal evidence of blue-green algae in Precambrian strata (Barghoorn and Tyler, 1965). Since the Gunflint and Biwabik Iron-formations appear to be correlative, the same fossils might be expected to occur in the latter. LaBerge (1967b) and Cloud and Licari (1968) reported almost unquestionable evidence of fossil algae in chert from the Biwabik Iron-formation; the genera are the same as those from the Gunflint Iron-formation (figs. IV-36C, D, and E). However, the microfossils that have been reported from the Biwabik Iron-formation are not as well preserved as those from the Gunflint. The metamorphism of the Biwabik may have altered the state of preservation of the microfossils. In this regard, it may be significant that the best preservation on the Gunflint cherts is in material from outcrops farthest from the Keweenawan igneous bodies, *i.e.*, near Schreiber, Ontario (figs. IV-36F, G, H, I, J, and K).

Most of the best preserved microfossils from the Schreiber locality have been found in bedded cherts, generally black, which occur in the stromatolitic zones. They are best observed in uncovered thin sections using an oil-immersion lens at about 1,000x. The above-mentioned papers, as well as those by Licari and Cloud (1968), Cloud (1965) and LaBerge (1967b) will give the reader an excellent review of the microfossils and possible microfossils found in the Biwabik and Gunflint Iron-formations.

Pokegama Quartzite

Gruner (1922, 1924) first reported the presence of microorganisms in the cherts of the Pokegama Quartzite. Cloud and Licari (1972) found well-preserved filamentous forms of blue-green algae in cherts from the base of the formation (figs. IV-36L and M). The algae have been assigned to the same species, *Gunflintia minuta*, Barghoorn (Barghoorn and Tyler, 1965) as those forms from the younger Gunflint Iron-formation.

Lower Precambrian

Knife Lake Group

Gruner (1923) reported and illustrated algal-like filaments from two chert pebbles from the Ogishke conglomerate of Winchell (1887), a unit now assigned to the "younger Knife Lake succession." Hawley (1926) doubted their algal affinities, but most of his objections can be countered. One reason for his doubts was the apparent size differential be-

Figure IV-36. Microstructures from the Biwabik, Gunflint, and Soudan Iron-formations and the Pokegama Quartzite. A, microstructure as figured by Gruner (1922, 1924), "iron bacteria." In chert from the Pokegama Quartzite, *cf.* J, x700. B, microstructures, same source as A, "iron bacteria resembling Chlamydothrix." Possibly filamentous blue-green algae, x250. C, spheroidal microstructure from the Biwabik Iron-formation, possibly *Huroniospora* sp. Courtesy G. L. LaBerge, x1160. D, spheroidal microstructures from the Biwabik Iron-formation. Courtesy G. L. LaBerge, x800. E, filamentous and spheroidal microstructures resembling blue-green algae, from the Biwabik Iron-formation. Courtesy G. L. LaBerge, x1160. F, Probable sheath of filamentous blue-green algae. Gunflint Iron-formation, Schreiber, Ontario, x1000. G, filamentous blue-green alga, probably *Gunflintia grandis*. Gunflint Iron-formation, Schreiber, Ontario, x1000. H, unicellular blue-green alga, with bud-like projection; *Huroniospora* sp. Gunflint Iron-formation, Schreiber, Ontario, x2000. I, unicellular blue-green alga, possibly with a sheath surrounding the cell body. Gunflint Iron-formation, Schreiber, Ontario, x2000. J, Filament of blue-green alga showing central trichome surrounded by sheath. Gunflint Iron-formation, Schreiber, Ontario, x1000. K, electron micrograph showing rod-shaped bacteria in a surface replica of black chert from the Gunflint Iron-formation, scale=1 micron. Courtesy J. W. Schopf. L, transmission electron micrograph of a filamentous blue-green alga, probably the sheath of *Gunflintia minuta*, from the Pokegama Quartzite. From Cloud and Licari (1972). M, filament of *Gunflintia minuta* from the Pokegama Quartzite. From Cloud and Licari (1972). N, electron micrograph of structures from within a pyrite nodule from the Soudan Iron-formation. Possibly bacteria. From Cloud and others (1965).


A
N
1 µm
B
I
H
L
2 µm
M
10 µm
J
K
G
C
F
D
E

tween the forms illustrated by Gruner and modern blue-green algae. The structures of Gruner's figure 1 range in diameter from less than 5 to more than 20 microns, and this is within the size range of the sheathed filament of either of the two freshwater forms of algae to which they were compared, *i.e.*, *Inactis* and *Microcoleus*. This size range of filaments within the chert is greater than one would expect for a single species, however. Tyler and Barghoorn (1963) attributed the structures, which are similar to those shown here in Figure IV-35H, to pyrite crystal trails. These microstructures are here considered to be of problematic origin.

Soudan Iron-formation

Gruner (1925) reported and illustrated what he considered to be cellular remains of blue-green algae from the Soudan Iron-formation. The photomicrographs show opaque structures resembling algal filaments. The material was found in one thin section of black chert from a sample of banded black "jasper" collected north of Armstrong Lake, between Tower and Ely, Minnesota. The following year, Hawley (1926) argued against the organic origin of these and other reported Precambrian microfossils. He experimentally produced filaments in solutions of ferrous silicates, etc., and stated that the size of the structures in the rock was inconsistent with that of modern blue-green algae. Gruner's photomicrographs indicate filament diameters of approximately 4 to 15 microns, which is not contradictory to the structures being fossil algae. Again, this type of structure has been attributed to trails made by pyrite migrating through chert. Gruner's figures do not permit a reasonably firm decision to be made either for or against a biological origin.

Cloud and others (1965) reported and illustrated microstructures as much as 1.5 microns across from within pyrite balls from the Soudan Iron-formation (fig. IV-36N). These were interpreted as possible remnants of bacteria or blue-green algae. Later, Cloud and Licari (1968) illustrated spheroidal structures that have a size range from 4 to 10 microns from the same slides. In both cases the biological origin of the structures is questionable.

To my knowledge, no further work has been reported which describes fossils from the Soudan Iron-formation. If any of the reported structures are indeed algae or bacteria, they are the oldest cellular remains found outside of Africa.

CHEMICAL FOSSILS

A large number of hydrocarbons have been reported from rocks of Precambrian age. Papers by Schopf (1970), Calvin (1969), Margulis (1970), and Eglinton and Murphy (1969) provide an excellent review of current data and problems. Hoering (1962), Belsky and others (1965) and Meinschein (1965) were among the first to report occurrences of hydrocarbons of possible biological origin from the Soudan Iron-formation. Hydrocarbons in the Gunflint Iron-formation in Ontario, including amino acids, have been reported by several workers (Barghoorn and Tyler, 1965; Oró and others, 1965; Schopf and others, 1968; Calvin, 1969). Geochemical methods of study are obviously valid ones, but several factors interfere with attempts to clarify the evolution of biochemical compounds. The most serious criticism of much of the data is contamination of the samples. Abelson and Hare (1968) concluded that the amino acids reported for the Gunflint Iron-formation are of recent origin. Schopf (1970), after reviewing the data, concluded that many, perhaps all, of the biologically significant hydrocarbons reported from the Soudan Iron-formation are of recent origin.

If some hydrocarbons are proved to be syngenetic with Precambrian sediments, another factor must be considered. The quantity of amino acids and other organic molecules that have been abiologically synthesized (ignoring the human organism involved) in the laboratory is large. Although the produced compounds are nearly always racemic mixtures, little data are available concerning the optical activity of the compounds reported from Precambrian rocks. This problem is further complicated since biological amino acids, such as L-isoleucine, have been shown to undergo racemization to an equilibrium of D and L forms within a geologically short period of time (Wehmiller and Hare, 1971). Since abiogenic synthesis of hydrocarbons is thought to have occurred in the early pre-free oxygen earth—indeed as it must have if current ideas on the evolution of life are correct—complex hydrocarbons that are not products of organisms probably exist in Precambrian rocks.

Living photosynthetic organisms preferentially select the lighter of the stable C^{12}/C^{13} isotopes of carbon, and this ratio should differ in some compounds of biological origin as compared to the ratio in an inorganic fraction, such as in carbonates, from the same strata. This C^{13} deficiency has been reported by Calvin (1969) in both soluble and insoluble organic extracts from the Soudan Iron-formation. The ratios for the soluble and insoluble fractions differ, and this difference could indicate that the soluble fraction is not syngenetic but instead entered the system later. The data are not clear in this regard. The fact that a C^{13} deficiency in the insoluble fraction does exist is, however, indicative of biological activity in Minnesota about 2.7 b.y. ago.

Perry and Tan (1970) reported C^{13} values from samples of the Biwabik Iron-formation as much as 18 per mil lower than the accepted Cretaceous belemnite standard. Although this study was done on carbonates, rather than on hydrocarbons, they attributed the C^{13} deficiency to the influence of organic carbon on the carbonate reservoir. Their study also indicated an essentially reducing atmosphere at the time of deposition of the Biwabik Iron-formation.

All organic chemical studies yet conducted on rocks of Precambrian age are hampered in that whole-rocks rather than discrete fossil material must be analyzed. Still, the available data support rather than conflict with ideas and suppositions concerning the presence of organisms in the Precambrian rocks of this state.

CONCLUSIONS

New discoveries of fossils will be made in the Precambrian rocks of Minnesota. That they exist is almost unequivocally accepted. The paleoenvironmental implications and their uses in correlation have only begun to be understood. Ideas on the origin of life and of an oxygenated atmosphere have been greatly furthered within the last decade because of fossil discoveries. The microfossils from the Middle Precambrian rocks of Minnesota are not as well preserved as those from the more pristine cherts of the Gunflint Iron-formation in Canada. Also, the chances are even less for finding well-preserved microfossils in the Lower Precambrian rocks. Probably, some microstructures that in their gross morphology resemble bacteria or algae are of inorganic origin. Modern blue-green algae have been experimentally "fossilized" in silica (Oehler and Schopf, 1971), and such changes as take place still leave them comparable in morphology to natural fossils in chert. Pyrite spherules, often with framboidal structure, can be formed by biological activity (Lougheed and Mancuso, 1971). Perhaps some of the spheroidal structures reported from the Soudan and Biwabik Iron-formations by Cloud and Licari (1968), LaBerge (1967b) and Cloud and others (1965) are related more closely to such biogenic mineral forms than to algal cells. The possibility exists that abiogenic coacervates or other similar spontaneously formed microstructures are preserved in rocks of Precambrian age. Some of these produced in the laboratory (see Margulis, 1970; Calvin, 1969; Keosian, 1968; Fox, 1965) are remarkably similar in form to modern algal cells. Even though the identification of microstructures from Precambrian rocks is difficult and errors in interpretation will be made, the search cannot be left to the paleobiologist. The rare and often chance occurrences make it necessary for every geologist concerned with Precambrian sedimentary rocks to look for and make known what may be additional evidence of Precambrian life.

Schopf (1970) has summarized the significant fossil discoveries from rocks of Precambrian age. Sedimentary rocks of similar age and lithology are not, of course, always present in Minnesota, but these reports do serve as guides for further search. The data below indicate some of the significant finds that were not discussed in this paper.

(1) Onverwacht Series, South Africa, older than 3.2 b.y.; cherts and argillites have filamentous and spheroidal microstructures of possible biological origin (Engel and others, 1968).

(2) Fig Tree Series, South Africa, approximately 3.1 b.y. old; cherts and shales with bacterium-like as well as filamentous and spheroidal alga-like microstructures, some of probable biological origin (Barghoorn and Schopf, 1966; Schopf and Barghoorn, 1967; Pflug, 1967).

(3) Bulawayan Group, Zwankendaba Limestone, Rhodesia, Africa, approximately 2.8 b.y. old; stromatolites in limestone (MacGregor, 1940).

(4) Belcher Group, Hudson Bay area, Canada, older than 1.6 b.y.; chert has bacterium-like as well as filamentous and spheroidal alga-like microstructures, some of probable biological origin; possible earliest eukaryotes (Hofmann and Jackson, 1969).

(5) Nonesuch Shale, Michigan, approximately 1.1 b.y. old; shale with disc-like siliceous microstructures of unknown biological affinities and rare filamentous and spheroidal alga-like microstructures of probable biological origin (Meinschein and others, 1964; Jost, 1968).

(6) Bitter Springs Formation, Australia, approximately 0.9 b.y. old; cherts have various types of algal microfossils, some apparently green algae, and therefore the earliest probable eukaryotes (Barghoorn and Schopf, 1965; Schopf and Blacic, 1971).

(7) Pound Quartzite, Ediacara area, Australia, approximately 0.65 b.y. old; argillaceous partings in sandstone have impressions of various soft-bodied animals (Glaessner and Wade, 1966).

AMINO ACIDS IN SOME MIDDLE PRECAMBRIAN ROCKS OF NORTHERN MINNESOTA AND SOUTHERN ONTARIO

James R. Niehaus and Frederick M. Swain

Organic geochemical studies of Minnesota rocks have resulted in the recognition of a variety of hydrocarbon and carbohydrate compounds (table IV-17). These results suggest that by 2,000 m.y. ago life on earth had already evolved to a complex state, a conclusion in agreement with evidence derived from the study of microfossils (see Darby, this chapter). Of the many organic compounds presently recognized, amino acids are particularly good indicators of the nature of Precambrian life, if they can be demonstrated to be indigenous to the rocks. However, because they occur in small amounts, contamination by younger organic matter presents a serious problem. Amino acids have been reported from the Middle Precambrian Gunflint Iron-formation of Ontario (Schopf and others, 1968). Subsequently, Abelson and Hare (1968) and Smith and others (1970) concluded that these amino acids are of recent origin.

So far as the authors can determine, all of the previous studies have dealt with samples obtained from natural or artificial exposures, which are highly susceptible to recent contamination. However, the possibility of recent contamination seems much less likely for samples from deep drill cores, particularly where the rocks have low permeability. Therefore, this study was undertaken to compare the kinds and concentration of amino acids observed in surface and subsurface samples of similar lithology and subsequent geologic history, and to evaluate any observed differences with respect to possible contamination.

DESCRIPTION AND LOCATION OF SAMPLES

Samples of various rocks from the Middle Precambrian Animikie Group were selected for their high content of carbonaceous material, and include both surface and subsurface (drill core) material. The formations that were sampled include the Biwabik Iron-formation, the overlying Virginia Formation, and its northeastern equivalent, the Rove Formation. Table IV-18 gives descriptive data on the samples. The stratigraphy and geologic history of the rocks that were sampled will be described briefly later in a discussion of the contained amino acids.

Table IV-17. Summary of investigations of Precambrian rocks of northern Minnesota for organic substances.

Authors	*Rocks Investigated*	*Substances Reported*
Swain, Blumentals, and Prokopovich (1958)	Thomson Formation, Rove Formation, "Cuyuna" formation	Hydrocarbons, humic acids, ninhydrin reacting (amino-N) substances, carbohydrates
Meinschein (1965)	Soudan Iron-formation	Isoprenoid and N-alkane hydrocarbons
Swain, Pakalns, and Bratt (1970)	Soudan Iron-formation, Thomson Formation, "Cuyuna" formation, Biwabik Iron-formation, Rove Formation	Carbohydrates
Swain, Bratt, and Kirkwood (1970)	Soudan Iron-formation, "Cuyuna" formation, Biwabik Iron-formation, Virginia Formation, Rove Formation	Carbohydrates (monosaccharides and polysaccharides)
Oró and Nooner (1970)	Soudan Iron-formation	Aliphatic hydrocarbons (norpristane, pristane, and phytane)
Hoering (1967)	Soudan Iron-formation	Methane from samples heated at 266° C.
Hoering (1967)	Soudan Iron-formation, Thomson Formation	δC^{13} −36.36 and −29.44 respectively
French (1964)	Biwabik Iron-formation	Carbonaceous material gives broad diffuse X-ray diffraction, except near Duluth Complex, where well-defined lines of graphite occur

Table IV-18. Description, location, and source of samples.

Formation	*Sample Type*	*Lithology*	*Location and Source*
Rove	surface	graphitic slate and graywacke	Hanging wall of diabase dike; Silver Islet, Ontario. Collection, Dept. of Geology and Geophysics, University of Minnesota.
Rove	surface	anthraxolite	Silver Islet, Ontario. Collection, Dept. of Geology and Geophysics, University of Minnesota.
Virginia	subsurface (667-668H.)	dark gray argillite	Mesabi Deep Drilling Project, drill hole no. 2., SW-SE 22-58-16, south of Biwabik, Minn.
Virginia	subsurface (677-678H.)	dark gray argillite	
Virginia	subsurface (1444-1446H.)	black argillite	
Virginia	subsurface (1464.5-1465.5H.)	black argillite	
Biwabik	subsurface (approx. 1150H.)	anthraxolite	Mesabi Deep Drilling Project, drill hole no. 5., SE-NW 36-58-20, south of Buhl, Minn.
Biwabik	surface	algal chert	Mary Ellen Mine, Biwabik, Minnesota. Collection, Dept. of Geology and Geophysics, University of Minnesota.

RESULTS

Rock Samples

The samples were analyzed using techniques outlined in Appendix IV-B of this chapter. Small concentrations of amino acids were detected in all the rock samples (table IV-19), in amounts ranging from insignificant traces (10^{-10} mole/g) in the anthraxolite from the lower cherty member core sample of the Biwabik Iron-formation to slightly more than 10^{-7} mole/g in a graphitic sample of the Rove Formation. Most of the amino acids occur in the water hydrolyzates, and therefore probably exist in the free state in the rocks, or at least, in a loosely combined state. Both thermally stable and unstable varieties are reported. This and the generally uncombined nature of the amino acids suggest fairly recent contamination, a possibility discussed previously by one of us elsewhere (Niehaus, 1969, unpub. M.S. thesis, Univ. Minn.). Also, a marked difference in concentration level exists between the surface and subsurface samples, with the former being much greater.

The organic carbon and nitrogen content of the samples from the Virginia Formation is given in Table IV-20. Both greater variety and higher concentrations of amino acids were noted in the more carbon-rich samples. Three unidentified, presumably non-protein, amino compounds were detected in the samples. Their locations on the chromatograms are shown in Figure IV-37.

STABILITY OF THE AMINO ACIDS

A linear regression analysis (Niehaus, 1969, *op. cit.*) was performed on published data recording the concentration and geologic age of a reportedly stable amino acid (glycine) and a reportedly unstable amino acid (serine). The linear regression analysis indicated very little direct correlation between geologic time and the two amino acids. The regression lines are nearly horizontal, and the correlation coefficients have insignificant values at the 10 percent probability level.

Table IV-19. Amino acid analyses. (Values are in 10^{-9} mole/gram of dry sample.)

Sample	State of Amino Acids	Neutral and Acidic Amino Acids										Basic Amino Acids					
		Asp	Ser	Thr	Glu	Pro	Gly	Ala	Cys	Val	Met	Lys	His	Arg	U-1	U-2	U-3
Rove Fm.	F	1.5		4.0	56.0			67.0				6.0	1.5				
graphite wall rock	C		5.0														
Rove Fm.	F	5.6	9.0				5.0					2.5	2.0				
anthraxolite	C																
Virginia Fm. argillite (core)																	
667-668	F	0.1										Tr					
	C		Tr				Tr										
677-678	F														Tr		
	C																
1444-1446	F		0.3				1.4					0.4		Tr		Tr	Tr
	C																
1464.5-1465.5	F						Tr?										
	C						0.2	0.2	Tr	Tr	Tr	Tr?					
Biwabik Iron-fm. (core)	F		Tr				Tr										
anthraxolite	C																
Biwabik Iron-fm.	F			8.0			0.2										
algal chert	C																

State of Amino Acids column:
F, free amino acids; C, combined amino acids

Names of amino acids:
Asp, aspartic acid; Ser, serine; Thr, threonine;
Glu, glutamic acid; Pro, proline; Gly, glycine;
Ala, alanine;
Cys, cystine; Val, valine; Met, methionine;
Lys, lysine; His, histidine; Arg, arginine;
U-1, unknown no. 1; U-2, unknown no. 2;
U-3, unknown no. 3.

Concentrations of amino acids:
Tr, trace amount, less than 0.1 x 10^{-9} mole/g.;
Tr?, questionable trace;
Tr, amount greater than 0.1 x 10^{-9} mole/g., but not computable, since there is no standard for comparison.

Despite the many assumptions that were made and the high variability of the data in the statistical tests, the analysis strongly suggests that amino acids decompose differently under geologic conditions (including complex interplay of amino acids with clay particles, humic acids, sugars, etc.) than under conditions of laboratory studies conducted to date.

STRATIGRAPHIC RELATIONSHIPS OF THE AMINO ACIDS

Biwabik Iron-formation Samples

Samples of algal chert and anthraxolite from a drill hole that penetrated the lower cherty member of the Biwabik Iron-formation were analyzed. The chert is from the basal part of the member (see Morey, this chapter), whereas the anthraxolite is from the uppermost part, immediately beneath the so-called intermediate slate; the anthraxolite occurs as small patches of black, vitreous material in ferruginous chert. Only traces of amino acids were found in the anthraxolite sample. Temperature gradient measurements made in the drill hole (T. Baldwin, 1969, oral comm.) from which the sample was obtained, indicate that ground water is now moving through the Biwabik Iron-formation within a distance of about 30 feet from the sampled section. Apparently the ground water is moving through a brecciated zone (Pfleider and others, 1968) that was penetrated in drilling. Apparently, no contamination has resulted, however, for there are no significant concentrations of amino acids in the sample.

The algal sample is a stromatolitic ferruginous chert—a shallow-water deposit—as indicated by oolitic hematite, intraformational conglomerates, and algal structures—which accumulated slowly under oxidizing conditions during a time of tectonic stability (White, 1954). Because sig-

Table IV-20. Nitrogen and organic carbon content of Virginia samples from formation in drill hole 2, Mesabi Deep Drilling Project.

Sample interval	C	N
depth in ft.	percent	
667-668	0.36	0.29
677-678	0.62	0.09
1444-1446	1.22	0.04
1464.4-1465.5	1.04	0.05

Analyses by Microanalytical Laboratory, University of Minnesota

nificant concentrations of amino acids were not likely to be preserved under such an oxidizing environment, we believe that the amino acids found in this surface sample are relatively recent.

Rove Formation Samples

Two samples of the Rove Formation were analyzed. One is a graphitic (Ingall, 1887) slate and graywacke from the hanging wall of a 300-foot-thick diabase dike exposed on tiny Silver Islet. The location of the anthraxolite sample is not known exactly, but it must have been taken close to the dike. Accordingly, it must be concluded that the samples have been metamorphosed. Probably, the graphitic material was mobilized and concentrated from carbonaceous material finely disseminated in the Rove strata at the time the dike was intruded. Because of the thermal metamorphism, it is unlikely that either sample contains original amino acids. Those that were detected must be of recent origin.

Virginia Formation Samples

Two samples of Virginia Formation from a drill hole near Biwabik, Minnesota were analyzed. Judged from oxygen isotope fractionation data on coexisting quartz and magnetite pairs from the underlying Biwabik Iron-formation, the temperatures to which the rocks were subjected probably never exceeded about 100° C (Perry and Morse, 1967).

Temperature measurements made in the Biwabik drill hole within a few months after its completion showed no major fluctuations in the thermal gradient (T. Baldwin, 1969, oral comm.); thus, there is no apparent movement of ground water through these rocks.

We suggest that the probable reducing nature of the depositional environment of the Virginia Formation was conducive to the preservation of organic matter, including amino acids, and that, owing to the mild thermal history of the rocks, small concentrations of the amino compounds remained in the rocks until the present time. The preponderance of neutral and basic amino compounds over acidic constituents in these rocks is not inconsistent with a reducing environment. Euxinic environments generally are slightly acidic at the sediment-water interface, but are commonly neutral or slightly basic immediately below the sediment surface (Krauskopf, 1967). Basic and neutral amino acids accumulating under such conditions would be near their isoelectric points at the higher pH, and thus would be diagenetically more stable, whereas the acidic amino acids would not be near their isoelectric points, would be less stable diagenetically, and therefore are more likely to be destroyed. The net result would be an original sediment rich in basic and neutral amino acids. Owing to the complexities concerning the stability of amino acids, however, one cannot expect the relationship, as deduced from the present small concentrations of amino acids, to be that simple.

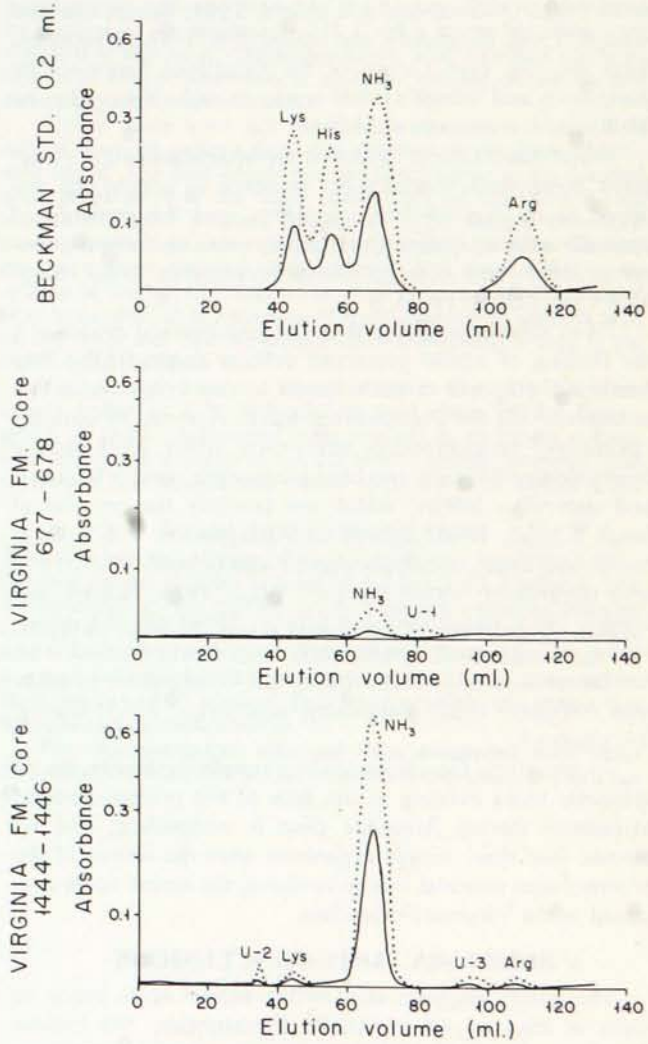

Figure IV-37. Location of unidentified amino compounds (U-I, etc.) compared to basic amino acid parts of a standard chromatogram. See Table IV-19 for explanation of abbreviations.

ORIGIN OF THE AMINO ACIDS CONTAINED IN THE VIRGINIA FORMATION

As indicated above, the Virginia Formation samples contain amino acids that we believe are probably native to the rock. For these amino acids, a Precambrian biogenic source is postulated.

The existence of a simple biota in Precambrian time is now believed to be unquestionable, as a result of studies of both indigenous organic compounds and of microscopic and sub-microscopic fossils in Precambrian rocks (see Darby, this chapter). Numerous classes of organic compounds of probable biological origin, other than amino acids, have been extracted from Precambrian rocks. The isoprenoid hydrocarbons phytane and pristane are believed to be derived from the chlorophyll of photosynthetic organisms and, when found in sedimentary rocks, have been used as evidence of biogenic activity. Phytane and pristane have been isolated from organic extracts of numerous Precambrian rocks (Meinschein and others, 1964; Barghoorn and Schopf, 1966; Barghoorn and others, 1965). Those reported from the Fig Tree Group of the Swaziland Sequence by Barghoorn and Schopf (1966) are from rocks believed to be 3 billion years or more old.

Other geochemical evidence for biological origin of organic compounds includes the presence of porphyrins, believed to be derived from pigments, and the presence of optically active alkanes. Compounds meeting these requirements have been isolated from Precambrian rocks (Barghoorn and others, 1965).

Even more convincing than the geochemical evidence is the finding of actual preserved cellular forms in the Precambrian. Reports of such fossils in the Precambrian are numerous. All the Precambrian fossils reported to date are "primitive" prokaryotic or eukaryotic forms. They include forms similar to green and blue-green algae, simple bacteria, and spore-like bodies, which are possibly the remains of fungi (Cloud, 1965; Schopf and Barghoorn, 1967; Barghoorn and Tyler, 1965; Barghoorn and Schopf, 1965; Tyler and Barghoorn, 1954; Gruner, 1922, 1923; Schopf and others, 1965; Barghoorn and Schopf, 1966). Similar occurrences of algae and iron-bacteria have been reported from the Biwabik Iron-formation and the Gunflint Iron-formation (Gruner, 1922; Barghoorn and Tyler, 1965; Darby, this chapter).

In summary, the evidence for a simple, primarily photosynthetic biota existing in the area of the present State of Minnesota during Animikie time is compelling, and we assume that these simple organisms were the source of the protenaceous material that gave rise to the amino acids contained in the Virginia Formation.

SUMMARY AND CONCLUSIONS

This study suggests that native amino acids occur in rocks at least as old as Middle Precambrian. We believe that the probability of relatively recent geologic contamination or of laboratory contamination of deep core samples of the Virginia Formation that contain the amino acids is minimal.

Those amino acids that are found are not only combined amino acids but also are free constituents. Amino acids presently considered thermally unstable as well as stable varieties theoretically can be and apparently are preserved in rocks as old as the Middle Precambrian. The interaction of amino acids with other organic and inorganic substances apparently has a stabilizing effect which commonly overshadows the degradative effect of temperature. Thus, the inherent stabilities of amino acids, or at least of some amino acids, as they occur under geologic conditions, are not those of the individual compounds in dilute solution but rather are those of the compounds in association with their geochemical environment.

An attempt was made to correlate the amino acid content of the uncontaminated Virginia Formation samples with their probable depositional environment. A reasonable correlation can be made, but the uncertainties connected with the stability relations of amino acids prohibit a high degree of confidence in the correlation. We believe that this degree of uncertainty will remain in all such studies, at least of Precambrian rocks, until the stability relationships are better understood. Lastly, this investigation shows the value of using drill core material, as opposed to outcrop samples, wherever possible, in organic geochemical studies where very small concentrations of compounds are encountered.

ACKNOWLEDGMENTS

Gratitude is extended to Miss Theodora Melone, former Librarian, Winchell Library of Geology, for assistance in locating the references, to G. B. Morey for providing core samples, and to Mrs. Judy M. Bratt, for advice on laboratory techniques.

APPENDIX IV-B

Laboratory Procedures

Samples initially were scrubbed with a wire brush to remove outer weathered material, or for the drill cores, to remove any adhering drilling mud. This was followed by washing in water and then in chromic acid solution. The latter solution was used to degrade any surficial organic matter present. The samples were then rinsed with double-distilled water and air-dried, thus preparing them for the crushing operation.

The cleaned samples were broken down to small pebble size in a jaw crusher. The pebble-size material then was crushed to 100 mesh or less in a disc grinder. Both the crusher and grinder were previously cleaned with a wire brush and rinsed with acetone. Care was taken to avoid touching the cleaned samples during the process, and the material was handled only with clean rubber gloves or forceps. The crushed samples were stored in sealed plastic bags while awaiting hydrolysis.

Hydrolysis and Reduction

Samples from the Virginia Formation

Both water and hydrochloric acid hydrolyzates were prepared. Sample weights ranged from 500 to 1,000 grams, depending upon the amount of material available after cleaning and grinding. For water hydrolysis, the sample was mixed with an equal volume of double-distilled water and boiled under reflux conditions for 24 hours. The acid hydrolysis was carried out on the same material after removal of water. Each ground sample consisting of 4-10 grams was hydrolyzed with 100 ml of standardized 0.555$\underline{N}$ sulfuric acid, prior to the actual hydrochloric acid hydrolysis, to determine the amount of acid needed to neutralize the carbonates in the larger sample. Titration of the standardized sulfuric acid with standardized NaOH, after 24-hr. hydrolysis, allowed calculation of the amount of concentrated hydrochloric acid needed to neutralize the actual sample. This amount of the concentrated acid was added to the sample, and an additional 300 ml of 6 N hydrochloric acid was added. Refluxing for 24 hours completed the acid hydrolysis. Hydrolyzates were removed from sediments with the aid of a Buechner funnel. Analytical grade filter paper was used. Samples were next reduced to near dryness by vacuum distillation at 50° C and desalted on a freshly prepared column of DOWEX 50W-X8 cation exchange resin. 2 N NH_4OH was used to elute the amino acids from the ion-exchange resin. Vacuum distillation at 50° C reduced the eluant to dryness. The sample was then taken up in exactly 5 ml of 10 percent isopropanol, labeled, and placed under refrigeration to await analysis.

Samples from the Rove Formation and Biwabik Iron-formation

These samples were prepared specifically to analyze for carbohydrates, and accordingly the hydrolysis and subsequent preparation differed from those for the samples of the Virginia Formation. The cleaning and grinding procedures are identical for all samples studied. For water hydrolysis, 500 ml of double-distilled water was added to 70 to 350 grams of cleaned and crushed sample. Forty-eight hour hydrolysis was carried out under reflux conditions. The sediment was centrifuged from the resulting solution and the solution was reduced to a small volume under vacuum distillation, and filtered through glass filter paper, using a Buechner funnel. The filtrate was further reduced, and the small volume remaining was dialyzed to separate monosaccharides from polysaccharides. Both fractions were taken up in exactly 5 ml of double-distilled water. The undesalted monosaccharide fraction, containing the bulk of any amino acids present, was analyzed for amino acids. Before sulfuric acid hydrolysis, a few grams of the wet sediment remaining after water hydrolysis were weighed and then oven-dried to determine moisture content. The bulk of the wet, water-extracted sediment was then weighed and its dry weight calculated on the basis of the water content of the oven-dried aliquot. Fifteen to 100 ml of 72 percent sulfuric acid were added to the sediment and the mixture was allowed to stand for 1 hour at room temperature. Sufficient double-distilled water was then added to reduce the acid concentration to 0.2 N and hydrolysis was continued under reflux conditions for 24 hours. The extract thus obtained was neutralized with $CaCO_3$ which previously had been maintained at 400° C for 4 hours to reduce the possibility of organic contamination in the reagent. The solution was filtered through a fritted disc, reduced to a small volume by vacuum distillation, and desalted with 95 percent ethanol. The precipitated salts were centrifuged from the clear solution, and the solution reduced to dryness. The precipitate remaining was taken up in exactly 5 ml of double-distilled water, and the extract then was ready for analysis for combined amino acids.

Analytical Blanks

To evaluate laboratory contamination, both a water and a hydrochloric acid blank were prepared. The procedure was to take 500 ml of double-distilled water and 500 ml of the 6 N hydrochloric acid used for hydrolysis and to put each of the liquids through the entire analytical procedure used in preparation of the Virginia Formation samples, from filtration through desalting to final reduction.

A fingerprint blank was prepared to ascertain whether fingerprint contamination was prevalent. Data are available on the amino acids contained in wet fingerprints (Hamilton, 1965; Hare, 1965; Oró and Skewes, 1965). The fingerprint blank that was prepared was dry, rather than wet, and is perhaps more typical of contamination that is likely from insufficiently cleaned glassware than is a wet blank. The procedure was to pat the inside of a small, clean, dry beaker with the fingertips until fingerprints were visible on the glass. The fingertips had been previously wiped with clean tissue to remove any adhering grit. The inside of the beaker was then rinsed with 1 ml of 2.2 pH sodium citrate buffer which dissolved any free amino acids. The buffer was removed and put into a labeled vial, the beaker was lightly rinsed with double-distilled water and 2 ml of 6 N hydrochloric acid was placed in the beaker. This was followed by heating to approximately 75° C on a warm hot plate over a period of one hour. The hydrochloric acid was then removed from the beaker and both it and the sodium citrate sample were analyzed for amino acids.

Amino acids were not detected in the water blank, and only a trace of aspartic acid was detected in the hydrochloric acid blank. The acid blank was the first sample to be desalted after preparation of the water hydrolyzate of the 667-668-foot sample from the Virginia Formation. This sample contained some aspartic acid, and was the only sample of those desalted on this column that did. It is believed that insufficient regeneration of the ion exchange resin prior to desalting of the blank may have resulted in retention of trace amounts of aspartic acid on the resin, and that the amino acid subsequently eluted in the desalting of the blank.

Amino acids were not found in either the 2.2 pH sodium citrate buffer or 6 N hydrochloric acid rinses of dry fingerprints. A faulty photometer lamp resulted in an irregular baseline during the analysis of the 2.2 buffer sample, however, and a rerun of this sample may reveal some traces of amino acids.

Amino Acid Determinations

A 0.1 to 1 ml aliquot of each sample was analyzed with a Phoenix Model K-5000 Automatic Amino Acid Analyzer. The determinations were carried out with the ion exchange columns maintained at 50° C. For a neutral and acidic analysis, elution was with 3.25 pH sodium citrate buffer for 9 hours and 15 minutes, followed by elution with 4.25 pH sodium citrate buffer for 4 hours and 35 minutes. In a basic analysis, the eluant was 5.25 pH sodium citrate buffer.

The chromatograms obtained were compared with those from analyses of 0.2 or 0.3 ml aliquots of Beckman Calibration Mixture Type 1.

Chapter V

LATE PRECAMBRIAN

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General Geology, Northeastern Minnesota, J. C. Green
North Shore Volcanic Group, J. C. Green
Duluth Complex, History and Nomenclature, William C. Phinney
Northwestern Part of Duluth Complex, William C. Phinney
Northern Prong, Duluth Complex, William C. Phinney
Eastern Part of Duluth Complex, Donald M. Davidson, Jr.
Southern Part of Duluth Complex, Bill Bonnichsen
Sulfide Minerals in the Duluth Complex, Bill Bonnichsen
Logan Intrusions, P. W. Weiblen, E. A. Mathez, and G. B. Morey
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Magmatic Sulfides and Associated Fissure Vein Deposit at the Green Prospect, Cook County, M. G. Mudrey, Jr.
Puckwunge Formation of Northeastern Minnesota, Allen F. Mattis
Keweenawan Geology of East-Central and Southeastern Minnesota, Campbell Craddock
Keweenawan Volcanic Rocks in East-Central Minnesota, G. B. Morey and M. G. Mudrey, Jr.
Sedimentation and Petrology of the Upper Precambrian Hinckley Sandstone of East-Central Minnesota, A. D. Tryhorn and Richard W. Ojakangas
Petrology of Keweenawan Sandstones in the Subsurface of Southeastern Minnesota, G. B. Morey
The Sioux Quartzite, Southwestern Minnesota, George S. Austin

REGIONAL GEOLOGIC SETTING

Campbell Craddock

The Upper Precambrian rocks of the Minnesota region are of interest and importance for both economic and scientific reasons. Copper was obtained from these rocks by unknown miners in prehistoric times, and the Middle and Upper Keweenawan strata of Michigan still comprise an important copper-producing district. Some Upper Precambrian igneous rocks of the region contain minerals rich in copper, nickel, and other metallic elements. Rocks of the Baraboo district in Wisconsin that may be Late Precambrian in age have been mined for iron ore. Porous Upper Precambrian sandstones are an important source of ground water in the Twin Cities area, and these sedimentary strata also are possible reservoirs for the injection and summer storage of natural gas. Upper Precambrian sedimentary and igneous rocks have been quarried throughout the region for use as dimension stone. Red argillaceous beds in the Sioux Quartzite have been a source of pipestone for centuries.

The Midcontinent Gravity High, the most prominent gravity anomaly in the United States, extends from Kansas across southeastern Minnesota to Lake Superior and beyond (pl. 2). Studies in Minnesota and Wisconsin (Thiel, 1956; Craddock and others, 1963; Craddock and others, 1970) have shown that the Midcontinent Gravity High is closely related to the major structural features developed in Upper Precambrian rocks. During part of the Late Precambrian this linear zone was the site of extensive mafic igneous activity. Thousands of feet of lava flows accumulated by fissure eruptions, and these and older rocks were intruded by gabbroic and anorthositic plutons. The Midcontinent Gravity High overlies these dense mafic rocks, which commonly occur in an elevated crustal block flanked by basins containing less dense sedimentary rocks. This narrow belt cuts across the grain of older Precambrian rocks, and it appears to represent a rift zone of continental dimension, probably caused by crustal extension.

Precambrian fossils are rare but of great importance in tracing the early history of life on Earth, and several fossil localities have been described from Upper Precambrian rocks near Lake Superior. Stromatolites have been found in dolomite beds of the lower Upper Precambrian Sibley Group in the Thunder Bay district of northwestern Ontario. Algal heads occur in biostromal limestones in the Copper Harbor Conglomerate at several localities on the Keweenaw Peninsula in Michigan. The overlying Nonesuch Shale contains alkanes, porphyrins, microorganisms, spore-like spherical bodies, plant tissue, and viscous black crude oil.

STRATIGRAPHIC CLASSIFICATION

The geology of the Lake Superior district was summarized in the classic monograph by Van Hise and Leith (1911), and this work was later revised by Leith and others (1935). These papers laid the foundation for a stratigraphic classification of the Precambrian rocks. Van Hise and Leith divided the Precambrian into a lower Archean system and an upper Algonkian system, and the latter was further divided into a lower Huronian series and an upper Keweenawan series. Within the Keweenawan they recognized a lower sedimentary unit, a middle volcanic unit, and an upper sedimentary unit. Later, Grout and others (1951), from work in Minnesota, divided the Precambrian into three eras—Earlier, Medial, and Later Precambrian. The Later Precambrian was considered to be represented by a lower Animikie Group and an upper Keweenawan Group. The Keweenawan was divided into a lower sedimentary unit, a middle igneous unit, and an upper sedimentary unit. On the basis of numerous radiometric age determinations, Goldich and others (1961) and Goldich (1968) proposed a revised Precambrian classification for Minnesota. A three-fold division into Early, Middle, and Late Precambrian eras was adopted, and these eras are bounded by the Algoman orogeny and the Penokean orogeny. The Animikie Group was reassigned to the Middle Precambrian; it was deformed during the Penokean orogeny. The Upper Precambrian consists of the Sioux Quartzite and the younger Keweenawan sequence.

The stratigraphic classification and inferred correlations used in this paper are shown in Figure V-1. All Precambrian rocks which postdate the Penokean orogeny are assigned to the Upper Precambrian. The Sioux Quartzite, the Barron Quartzite, and the Sibley Group are considered about the same age, but probably older than similar rocks at the base of the Keweenawan sequence. The Keweenawan sequence is developed most fully in the Wisconsin-Michigan area, where it is divided into a lower sedimentary unit, a middle volcanic unit, and an upper sedimentary unit. The Upper Keweenawan is divided into a lower Oronto Group and an upper Bayfield Group, but the nature of the contact between these groups is not established. Possibly, the Bayfield Group is as young as Cambrian, but it is considered part of the Keweenawan sequence in this paper.

DISTRIBUTION OF UPPER PRECAMBRIAN ROCKS

Upper Precambrian rocks crop out extensively along the borders of Lake Superior. Upper Precambrian rocks considered here to predate the Keweenawan include the Sibley Group of northwestern Ontario, the Barron Quartzite of northwestern Wisconsin, and the Sioux Quartzite of southwestern Minnesota. It is possible that other Precambrian epicratonic sedimentary rocks of the Midcontinent region, such as the Baraboo Quartzite of southern Wisconsin (Dott and Dalziel, in press), are about this same age.

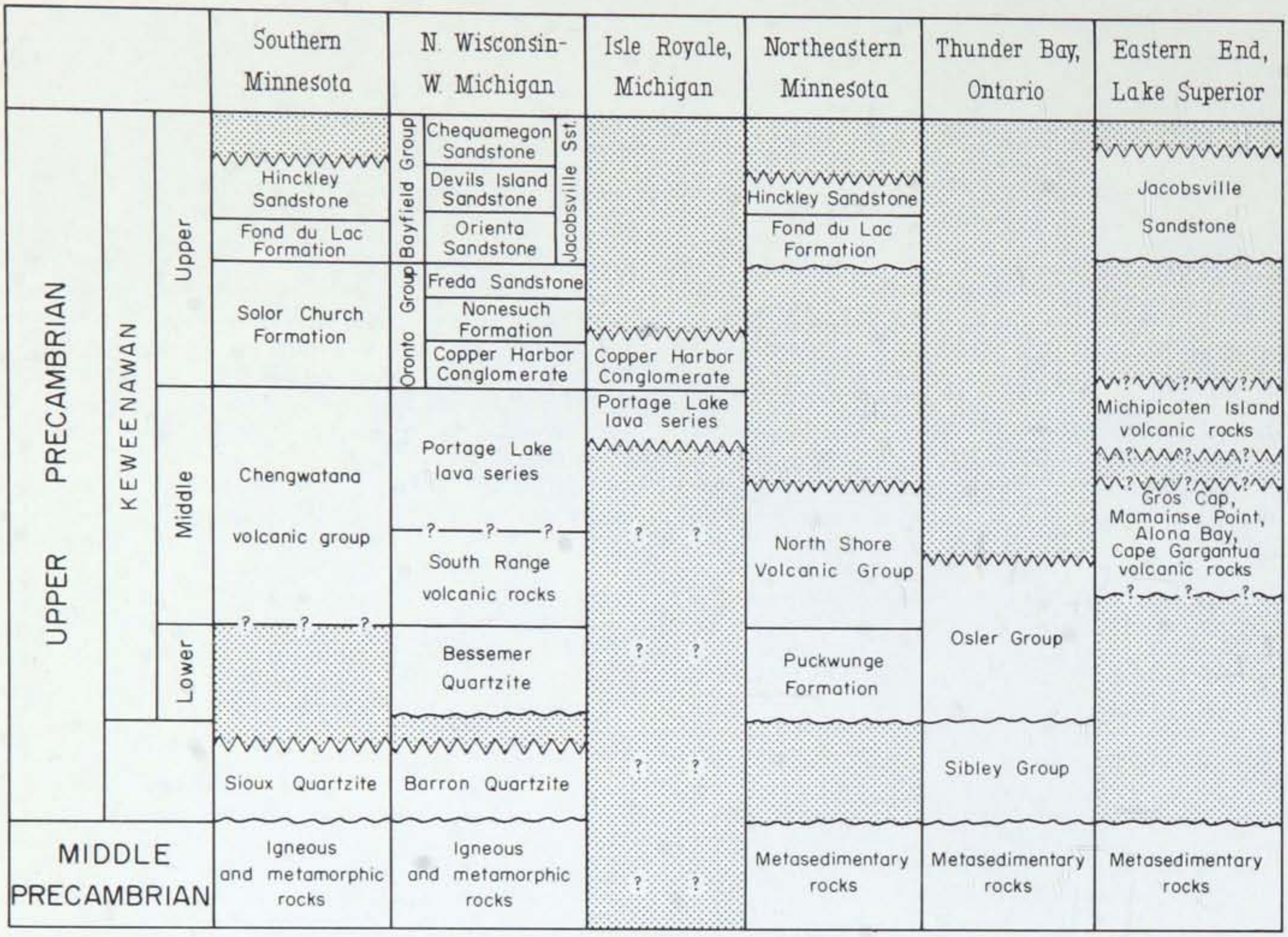

Figure V-1. Classification of Upper Precambrian stratified rocks, Minnesota and adjoining areas.

Rocks of the Upper Precambrian Keweenawan sequence are preserved most completely in the type area of western Michigan and northwestern Wisconsin, but they also crop out eastward to near the mouth of Lake Superior and southwestward into east-central Minnesota. Gravity and magnetic anomalies (Hinze, 1963; Hinze and Merritt, 1969) strongly suggest that these rocks continue southeastward from the east end of Lake Superior, beneath the Paleozoic strata of the Michigan basin, across lower Michigan at least to the buried continuation of the Grenville Front. Keweenawan rocks also can be traced by geophysical anomalies and subsurface geology to the southwest beneath younger rocks in southern Minnesota (Craddock and others, 1963), Iowa (Coons and others, 1967), and across Nebraska into Kansas (Woollard and Joesting, 1964; Bayley and Muehlberger, 1968; King and Zietz, 1971). The known and inferred extent of Upper Precambrian rocks in Minnesota and adjacent areas is shown on the accompanying geologic map (fig. V-2).

PREVIOUS STUDIES

Geologists have studied the Upper Precambrian rocks of Minnesota and adjacent areas for more than a century, and hundreds of reports have been published. Some of the more comprehensive regional reports—those found to be most useful in preparing this review—are summarized in this section. Many of the other papers are cited elsewhere in the text.

Among the earliest important work was that of Douglass Houghton, reported to the Michigan House of Representatives in 1840 and 1841, in which he called attention to the promising copper mineralization in the rocks of the Keweenaw Peninsula. David Owen conducted a reconnaissance geologic survey of Iowa, Wisconsin, and Minnesota for the U.S. Treasury Department in 1847-1850. The first systematic geologic survey of Minnesota was conducted during a ten-year period beginning in 1872 under the direction of N. H. Winchell; the last of six volumes describing this work was published in 1901. A similar pro-

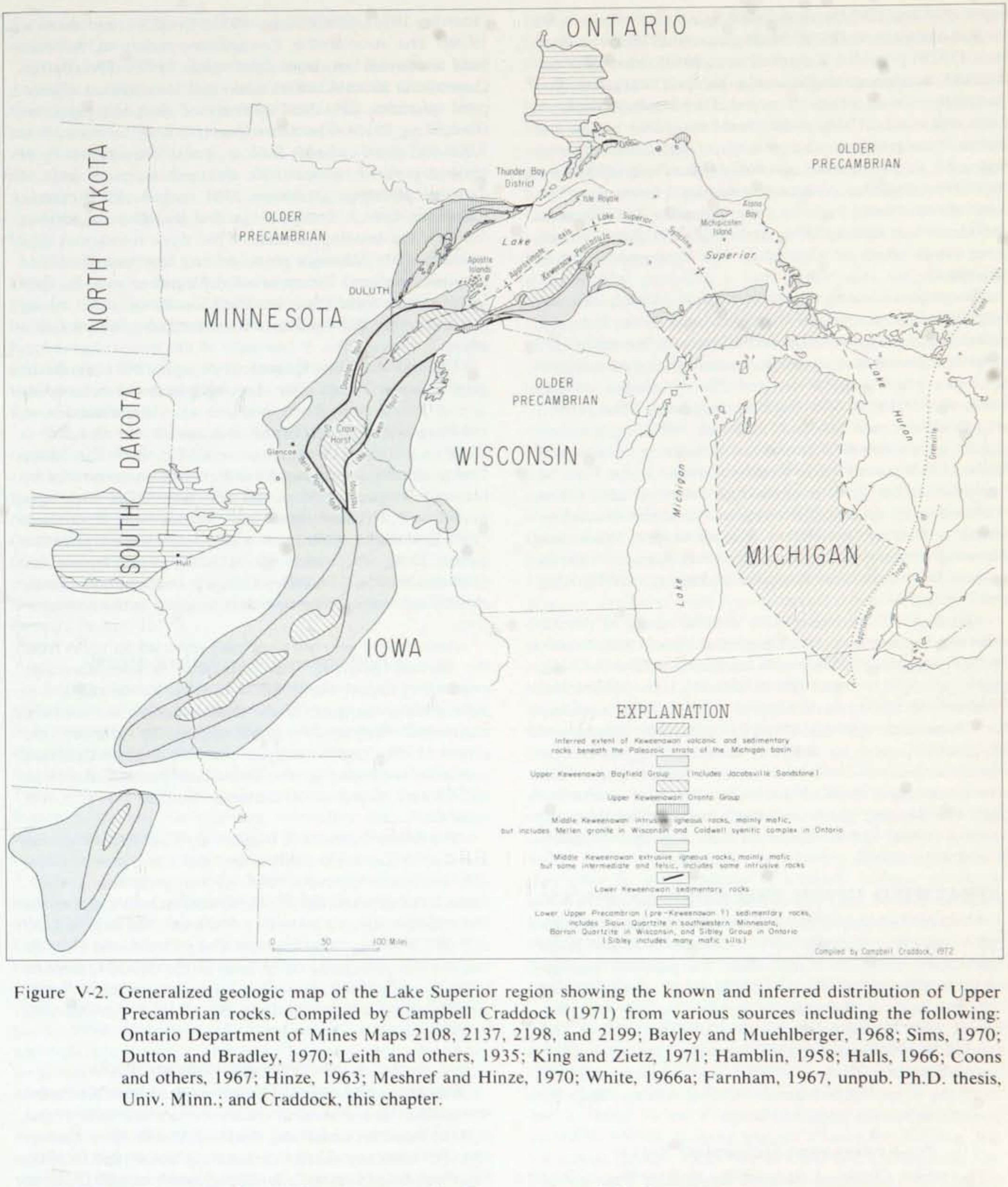

Figure V-2. Generalized geologic map of the Lake Superior region showing the known and inferred distribution of Upper Precambrian rocks. Compiled by Campbell Craddock (1971) from various sources including the following: Ontario Department of Mines Maps 2108, 2137, 2198, and 2199; Bayley and Muehlberger, 1968; Sims, 1970; Dutton and Bradley, 1970; Leith and others, 1935; King and Zietz, 1971; Hamblin, 1958; Halls, 1966; Coons and others, 1967; Hinze, 1963; Meshref and Hinze, 1970; White, 1966a; Farnham, 1967, unpub. Ph.D. thesis, Univ. Minn.; and Craddock, this chapter.

gram in Wisconsin during 1873-1879 was supervised by T. C. Chamberlin; four volumes treating the geology of the state appeared between 1877 and 1883. R. D. Irving (1883) presented a detailed summary of the Keweenawan rocks throughout the Lake Superior region. The comprehensive review of the Precambrian geology around Lake Superior by Van Hise and Leith (1911) includes a chapter on the history of geologic work.

Thwaites (1912) made a detailed study of the Upper Keweenawan sandstones along the Wisconsin shore of Lake

Superior. Lane (1911) summarized many years of work on the Keweenawan rocks of Michigan, and Butler and Burbank (1929) provided a detailed account of the copper deposits of the Keweenaw Peninsula. Tanton (1931) described the geology of the district around Thunder Bay, Ontario. Leith and others (1935) revised and brought up to date the earlier monograph on the geology of the Lake Superior region by Van Hise and Leith (1911). Sandberg (1938) discussed the sequence of Keweenawan lava flows along the north shore of Lake Superior from Duluth to Two Harbors, and Grout and others (1959) described the Upper Precambrian rocks of Cook County in the northeastern tip of Minnesota.

Geophysical surveys have yielded important information about the structure of the upper crust in the Lake Superior region. Hinze (1963) and Hinze and Merritt (1969) presented gravity and magnetic anomaly maps of southern Michigan which suggest the probable distribution of Keweenawan rocks beneath the Paleozoic beds. Thiel (1956), and Craddock and others (1963 and 1970) reported on gravity surveys in Wisconsin and Minnesota. Coons and others (1967) traced the Keweenawan rocks across Iowa by their distinctive gravity anomalies. Mooney and others (1970a and b) discussed the results of numerous shallow seismic refraction profiles in Minnesota and Wisconsin. Aeromagnetic surveys in the region from Kansas to northwestern Wisconsin were compiled and interpreted by King and Zietz (1971).

Hamblin (1958) presented a detailed study of the Jacobsville Sandstone of Michigan, and Hite (1968, unpub. Ph.D. dissert., Univ. Wisconsin) discussed the Oronto Group and Myers (1971, unpub. Ph.D. dissert., Univ. Wisconsin) the Bayfield Group in Wisconsin. Halls (1966) reviewed the Keweenawan geology of the Lake Superior region, and White (1966a and b) analyzed the structure of the Keweenawan basin at the western end of Lake Superior. Recent compilation maps of the United States of interest include the Bouguer gravity anomaly map by Woollard and Joesting (1964) and the basement rock map by Bayley and Muehlberger (1968).

STRATIFIED UPPER PRECAMBRIAN ROCKS

Most of the Upper Precambrian rocks of the Minnesota region are either sedimentary or volcanic in origin. Some sedimentary rocks in a few areas are probably pre-Keweenawan in age (see fig. V-1), but almost all the layered rocks are part of the very thick Keweenawan sequence. These Upper Precambrian strata will be discussed in five parts: pre-Keweenawan sedimentary rocks, Lower Keweenawan sedimentary rocks, Middle Keweenawan volcanic rocks, the Upper Keweenawan Oronto Group, and the Upper Keweenawan Bayfield Group.

Pre-Keweenawan Sedimentary Rocks

The Sibley Group of Ontario, the Barron Quartzite of Wisconsin, and the Sioux Quartzite of Minnesota are similar in composition, structure, and stratigraphic position, and possibly are about the same age. The Sibley Group is a red-bed sequence about 500 feet thick consisting mainly of sandstone, mudstone, conglomerate, and carbonate rocks (Tanton, 1931; Moorhouse, 1957; Franklin and Kustra, 1970). The stromatolite *Conophyton* occurs in dolomite beds at several localities (Hofmann, 1969). The Barron Quartzite is about 600 feet thick, and is composed of pale-pink quartzite and thin interbeds of dark-red pipestone (Hotchkiss, 1915). The Sioux Quartzite may be as much as 3,000 feet thick in South Dakota; it also consists mainly of gray to pink to red quartzite and includes some beds of dark-red pipestone (Baldwin, 1951, unpub. Ph.D. thesis, Columbia Univ.). Except for a few localities of possible faulting, the bedding in each of the three formations dips less than 25°. Although gentle folding has been postulated by some workers, the apparent deformation may be only initial depositional dips, modified locally by slight tilting accompanying the development of the Lake Superior syncline.

Each of the three formations occupies the same stratigraphic position, but their exact ages have not been determined. Whole-rock Rb-Sr isochron age determinations on red beds in the Sibley Group indicate an age of 1,265 to 1,409 m.y. (Franklin and Kustra, 1970). Both the Sibley Group and the underlying Rove Formation are intruded by the mafic Logan intrusions and associated dikes. K-Ar ages as old as 1,210 m.y. have been obtained on these dikes (York and Halls, 1969), and a sill in the Rove Formation gave a K-Ar whole-rock age of 1,300 m.y. (Hanson and Malhotra, 1971). The Sibley Group is overlain unconformably by the Osler Group, which is assigned to the Keweenawan.

Radiometric ages have not been reported on rocks from the Barron Quartzite. This formation is unconformably overlain by Upper Cambrian sandstone, but its relation to the Keweenawan rocks to the west and north has not been established. Refraction seismic studies by Mooney and others (1970a and b) suggest that the Barron Quartzite continues westward in the subsurface beneath the thick sedimentary sequence in the basin east of the St. Croix horst.

The Sioux Quartzite is overlain by Upper Keweenawan and Upper Cambrian sedimentary rocks in a well at Glencoe, Minnesota (Kirwin, 1963, unpub. M.S. thesis, Univ. Minn.). Goldich and others (1966) considered the age of the Sioux Quartzite to be between 1,200 and 1,700 m.y. A K-Ar age of 1,470 m.y. obtained on a slightly altered rhyolite (tentatively considered as a flow in the Sioux Quartzite) from a well at Hull, Iowa, has been interpreted as a possible minimum age for the deposition of the Sioux Quartzite by Lidiak (1971).

Lower Keweenawan Sedimentary Rocks

Van Hise and Leith (1911) defined the Lower Keweenawan as the thin sequence of clastic sedimentary rocks found in some localities underlying the thick Middle Keweenawan volcanic sequence. They did not designate a type locality, but these rocks are well developed north of the Gogebic iron range in Michigan and Wisconsin, where they are known as the Bessemer Quartzite. Lower Keweenawan rocks in Minnesota are assigned to the Puckwunge Formation, which crops out in northeastern Minnesota near Grand Portage. In this paper, the sedimentary rocks that comprise the lowest unit in the Osler Group of Ontario also are considered as Lower Keweenawan.

The Bessemer Quartzite in Michigan and Wisconsin has a maximum thickness of about 300 feet. Most of the formation is gray to pink, commonly laminated quartzite, but locally the basal unit is a conglomerate as much as 10 feet thick (Felmlee, 1970, unpub. M.S. dissert., Univ. Wisconsin). The Bessemer Quartzite dips steeply northward; it is structurally concordant with but stratigraphically unconformable upon the underlying Middle Precambrian Tyler Slate.

The Puckwunge Formation is exposed intermittently along a belt extending 25 miles westward from near Grand Portage. The formation consists of gray to light-buff, cross-bedded sandstone and quartzite at least 100 feet thick; a basal conglomerate can be observed in a few places (Grout and others, 1959). The Puckwunge Formation dips gently southward and rests with unconformity upon the underlying Rove Formation.

Although Van Hise and Leith (1911) assigned the Sibley Group to the Lower Keweenawan, those rocks are considered here to be probably pre-Keweenawan. In the Thunder Bay district of Ontario the Lower Keweenawan may be represented, however, by the lowest unit in the Osler Group. These beds consist of a basal conglomerate, 12 feet thick at one locality, overlain by about 200 feet of light-colored, cross-bedded quartz sandstone. These strata are nearly horizontal and rest unconformably upon the Sibley Group (Tanton, 1931).

Middle Keweenawan Volcanic Rocks

A thick sequence of Middle Keweenawan volcanic rocks underlies much of Lake Superior and crops out extensively in Minnesota, Ontario, Wisconsin, and upper Michigan: in the subsurface these rocks continue southwestward to Kansas and southeastward probably across lower Michigan. These volcanic rocks are mainly plateau basalts formed by fissure eruptions. Geophysical anomalies and thickness variations along the strike suggest that these flows accumulated in several major basins or troughs (White, 1966a).

The total thickness of the Middle Keweenawan volcanic sequence has been estimated at 30,000 feet by Tyler and others (1940) and by White (1966a). On the Keweenaw Peninsula the exposed volcanic rocks are at least 15,000 feet thick; the flows average 43 feet thick, but the greenstone flow locally exceeds 1,400 feet in thickness (White, 1960, 1966a and b). Recent work in western upper Michigan by White and others (1971) indicates that the Keweenawan volcanic sequence above the Bessemer Quartzite may be as much as 40,000 feet thick; this is similar to the thickness of 42,200 feet reported by Gordon in Van Hise and Leith (1911). Other estimates of thickness include 4,000 feet exposed in the St. Croix valley (Berkey, 1897 and 1898), nearly 20,000 feet in east-central Minnesota (Hall, 1901a), about 36,000 feet (Sweet, 1880) and more than 20,000 feet (Aldrich, 1929) in northwestern Wisconsin, about 25,800 feet along the north shore in Minnesota (Sandberg, 1938), and 10,000 feet on Isle Royale (Huber, 1971). In Ontario, reported thicknesses are 6,000 to 10,000 feet in the Thunder Bay district (Tanton, 1931), 11,500 feet on Michipicoten Island (Annells, 1970), 2,700 feet near Cape Gargantua (Ayres, 1969a), and more than 12,000 feet at Mamainse Point (Thomson, 1954).

The Middle Keweenawan volcanic rocks are mainly lava flows of wide lateral extent; pyroclastic rocks are rare. Most of the flows are basalt, but many felsites and some intermediate varieties such as andesite are present. Chemical analyses of lava flows indicate ranges of silica from 42 to 77 percent in Minnesota (Ruotsala and Tufford, 1965; Green, 1971b), 45 to 75 percent in the Keweenaw Peninsula (Broderick, 1935), and 42 to 47 percent at Mamainse Point (Thomson, 1954). Many of the mafic flows are amygdaloidal, especially in their upper parts, and the filling minerals include zeolites, quartz, calcite, epidote, and chlorite. A small fraction of the volcanic sequence consists of interflow sedimentary rocks, mainly red sandstones and conglomerates; locally the sedimentary rocks exceed 100 feet in thickness.

The exact time span represented by the Middle Keweenawan volcanic sequence has not been established. Where the sedimentary units of Early Keweenawan age crop out, their upper contact with Middle Keweenawan rocks appears to be conformable; elsewhere, the Middle Keweenawan volcanic rocks lie unconformably on Lower or Middle Precambrian rocks. Goldich (1968) estimated that most Keweenawan igneous activity took place during the interval 1,000-1,200 m.y. ago, but very few reliable radiometric ages are available from Keweenawan extrusive rocks. One difficulty is the problem of distinguishing sills from flows; another is the alteration that characterizes many of the volcanic rocks. Chaudhuri and Faure (1967) reported Rb-Sr isochron ages of 1,107 and 1,180 m.y. on felsite pebbles from the Copper Harbor Conglomerate; if these represent the true ages of pebbles from Middle Keweenawan flows, then volcanism began prior to 1,180 m.y. ago. Chaudhuri and Faure (1968) gave an age of $1,130 \pm 35$ m.y. on a syenodiorite intrusion emplaced in the Middle Keweenawan Portage Lake Lava Series. Chaudhuri and Faure (1967) also reported an age of $1,100 \pm 25$ m.y. for the Chippewa felsite, a flow in the Portage Lake Lava Series. In the same paper they gave the indicated age of a quartz feldspar porphyry lower in the sequence as 978 ± 40 m.y., but Brooks and Garbutt (1969) have interpreted this porphyry as an extrusive rock. Chaudhuri and Faure (1967) estimated the age of the Nonesuch Shale at $1,075 \pm 50$ m.y.; this formation is younger than the uppermost Middle Keweenawan volcanic rock.

Recent studies of the paleomagnetism of Keweenawan rocks by DuBois (1962), Books and others (1966), Books (1968), Beck and Lindsley (1969), Beck (1970), and Palmer (1970) have yielded very interesting results. Southwestward drift of the indicated Precambrian magnetic pole positions and a change in the Precambrian magnetic field from reversed to normal polarity suggest a basis for dividing Keweenawan igneous rocks into an older and a younger group. Most of the Keweenawan flows are assigned to the younger group, but some of the lower flows of the South Range near Ironwood, Michigan, of the North Shore Volcanic Group, of the Osler Group, and of Cape Gargantua, Alona Bay, and Mamainse Point sections are classed with the older group. The lowest flows of the South Range, however, possess normal polarity and may represent a still older group. Books (1968) has suggested redefining the top of the Lower Keweenawan at the change from reversed to normal polarity; this would require reassigning all the older Middle Keweenawan volcanic rocks to the Lower Keweenawan, a practice followed by J. C. Green in this chapter.

The difficulties encountered in developing a geomagnetic polarity time scale for the last 10 m.y. (Watkins, 1972) indicate a need for caution in applying this approach to the Precambrian. If Keweenawan igneous activity lasted 100-200 m.y., many polarity reversals may have occurred. In addition, the drift curve for the magnetic pole is poorly defined because of the scatter in the calculated pole positions, but this is not surprising in view of uncertainties about the amount and timing of tectonic deformations in many localities. Furthermore, at present very few Keweenawan rocks are dated with much precision, so the time of polarity reversals is poorly known. Future work may allow the construction of a geomagnetic polarity time scale for the Keweenawan, and the tracing of isochronous surfaces throughout the province will be a great advance in understanding Keweenawan history. In my view, however, the development of a time scale and the recognition of isochronous surfaces should not cause redefinition of rock-stratigraphic units. Accordingly, the traditional definitions of Lower and Middle Keweenawan, based on lithologic criteria, are retained in this paper.

Upper Keweenawan Oronto Group

The "Lake Superior sandstones" of earlier workers were divided by Thwaites (1912) into a lower Oronto Group and an upper Bayfield Group; he considered the two groups conformable and assigned them to the Upper Keweenawan. The Oronto Group was studied in detail by Hite (1968, *op. cit.*), and its stratigraphy is summarized by White (1971). The type area for the Oronto Group is the outcrop belt extending southwestward from the Keweenaw Peninsula into northwestern Wisconsin; it also crops out on Isle Royale (Wolff, 1969, unpub. M.S. dissert., Univ. Wisconsin), and occurs in the subsurface beneath Paleozoic strata in east-central Minnesota (Morey, this volume). The Oronto Group is believed to underlie the Bayfield Group in the sedimentary basin west of the St. Croix horst, and similar rocks may occur in the companion basin to the east (Mooney and others, 1970a and b). The Oronto Group probably continues southward in the subsurface into Iowa, Nebraska, and Kansas. The total thickness of the Oronto Group was estimated at 21,550 feet by Thwaites (1912), at 13,550 feet by Tyler and others (1940), and at about 20,000 feet by White (1971); the upper contact is not known to be exposed.

The Oronto Group is divided, in ascending order, into the Copper Harbor Conglomerate, the Nonesuch Shale, and the Freda Sandstone. The Copper Harbor Conglomerate is 200-7,000 feet thick and consists mainly of red to brown conglomerate and sandstone; in the type area, two thin oolitic limestones contain heads of the alga *Collenia undosa* (Hedlund, 1953, unpub. M.S. dissert., Univ. Wisconsin; Spiroff and Slaughter, 1961). The Nonesuch Shale is 250-750 feet thick and consists of gray siltstone, shale, and sandstone; the lower 50 feet of the formation contains extensive copper mineralization at the White Pine mine (Brown, 1971), where organic compounds, microfossils, and crude oil also have been found (Eglinton and others, 1964; Meinschein and others, 1964; Barghoorn and others, 1965; Moore and others, 1969). The Freda Sandstone appears to be at least 12,000 feet thick and consists of red arkosic sandstone, siltstone, and micaceous shale.

The Oronto Group can be assigned to the Upper Keweenawan because it seems to represent the same cycle of accumulation as the underlying Middle Keweenawan volcanic sequence, but the actual age of the Oronto Group is poorly known. The contact between the Portage Lake Lava Series and the Copper Harbor Conglomerate is gradational and marked by interbedded lava flows and conglomerates. The contact between the Oronto Group and the overlying Bayfield Group was thought by Thwaites (1912) to be conformable, but it cannot be seen in the field. Chaudhuri and Faure (1967) reported a radiometric age of $1,075 \pm 50$ m.y. for the Nonesuch Shale. It is possible that the entire Oronto Group may be older than 1,000 m.y., but the true age of the uppermost beds of the Freda Sandstone has not been established.

The rocks of the Oronto Group have been deformed moderately strongly. Throughout much of the area these beds have been tilted, and the strata dip 80° in the Wisconsin-Michigan border area just south of Lake Superior. In Wisconsin, a reversal of dip direction defines the Ashland syncline, which represents the axis of the Lake Superior syncline in that area. In both Wisconsin and Michigan, the Oronto beds are involved locally in large- and small-scale folding.

Upper Keweenawan Bayfield Group

Thwaites (1912) defined the Bayfield Group, and his pioneering work has been extended by Tyler and others (1940) and Myers (1971, *op. cit.*). The type area for the Bayfield Group is along the Wisconsin shore of Lake Superior, from the Apostle Islands and the Bayfield Peninsula westward to near Superior; this belt continues into Minnesota, where the Bayfield rocks have been divided into the Hinckley Sandstone and the underlying Fond du Lac Formation. Bayfield beds probably extend southward across western Wisconsin and southeastern Minnesota in the subsurface, perhaps as far as Kansas (Scott, 1966). The Jacobsville Sandstone of Michigan is considered here a part of the Bayfield Group (Hamblin, 1961), although this correlation should not be considered certain (Ostrom and Slaughter, 1967). The Jacobsville Sandstone crops out along the Michigan shore to the east end of Lake Superior and into Ontario (Hamblin, 1958), and it probably extends to the southeast into lower Michigan in the subsurface (Ells, 1967). Thwaites estimated the thickness of the Bayfield Group at about 4,300 feet, but he emphasized the scarcity of outcrops in the type area. Geophysical surveys suggest that the Bayfield Group may be 7,000 feet or more thick near the Keweenaw and Douglas faults (Bacon, 1966; Mooney and others, 1970a and b).

The exposed Bayfield Group has been divided, in ascending order, into the Orienta Sandstone, the Devils Island

Sandstone, and the Chequamegon Sandstone. The Orienta Sandstone consists of red feldspathic sandstone and siltstone with a few thin beds of conglomerate. Thwaites (1912) estimated its thickness at about 3,000 feet, but he included some beds exposed at Middle River and intersected in a well at Ashland that may belong to the Oronto Group (Myers, 1971, *op. cit.*). The Devils Island Sandstone is about 300 feet thick and consists of well sorted, very pure quartz sandstone. The Chequamegon Sandstone underlies the Bayfield Peninsula and the Apostle Islands; it is a red feldspathic sandstone with a few thin interbeds of red shale, siltstone, and conglomerate. Thwaites (1912) estimated its thickness at 1,000 feet, but seismic refraction surveys (Mooney and others, 1970a and b) suggest that it may be only about 500 feet thick. These three formations comprise the Bayfield Group in outcrop in the type area, but it is probable that older Bayfield beds exist in the subsurface.

The Bayfield Group is younger than the Oronto Group and is overlain unconformably by Upper Cambrian beds; its exact age and proper classification, within these limits, have been long debated. Thwaites (1912) and Atwater and Clement (1935) argued that the Bayfield is Keweenawan, but Raasch (1950) and Hamblin (1958) preferred a Cambrian age. A critical question is the nature of the contact between the Bayfield Group and the Oronto Group. Wherever the basal contact is exposed, the Bayfield Group rests unconformably upon rocks ranging in age from Middle Keweenawan to Archean; Murray (1955) presented evidence suggesting that one of these surfaces is a pre-Jacobsville glacial pavement. Although no outcrop demonstrates their relationship, it is probable that the upper Bayfield Group at the surface is also unconformable upon the Oronto Group seen at the surface. However, geophysical surveys suggest that younger Oronto beds and older Bayfield beds may exist in the subsurface in the sedimentary basins flanking the axial horst of the Keweenawan province, and it is possible that the Bayfield and Oronto Groups may be conformable in those basins. DuBois (1962) observed that the magnetic pole positions obtained from the Orienta and Jacobsville specimens suggested a Keweenawan rather than a Cambrian age, although he found different positions from his Chequamegon specimens. Although the possible age ranges from about 550 to 1,000 m.y., the Bayfield Group is considered here to be uppermost Keweenawan.

Throughout the Lake Superior region the Bayfield Group is nearly flat-lying, and dips greater than 10° are rare. Steep dips occur at some localities in sedimentary beds immediately adjacent to the Douglas and Keweenaw faults, but some of these strata probably belong to the Oronto Group. The generally undeformed character of the Bayfield Group provides a striking contrast to the sharp tilting and folding noted in the Oronto Group. This contrast in structural style is evidence for an unconformity between these two groups, at least at the stratigraphic levels exposed at the surface.

UPPER PRECAMBRIAN INTRUSIVE ROCKS

The Upper Precambrian extrusive and intrusive rocks of the region constitute a major igneous province. These rocks occur in a linear belt that extends from Kansas to at least the east end of Lake Superior. Geophysical anomalies suggest that these igneous rocks continue to the southeast across lower Michigan (Hinze and Merritt, 1969). The Kapuskasing Gravity High, which extends to the northeast from the east end of Lake Superior, may be related to a buried branch of the same igneous province.

The Upper Precambrian intrusive rocks are commonly assigned to the Keweenawan igneous cycle, but some of the rocks probably are older than the type Lower Keweenawan of Michigan and Wisconsin. Most of the intrusive rocks have mafic compositions, but some felsic and intermediate rocks have formed by differentiation and perhaps other processes. The smaller intrusive bodies are mainly tabular, and the larger ones are either ring complexes or roughly stratiform. Hundreds of mafic dikes occur throughout the region, commonly parallel to the axis of the Lake Superior syncline, suggesting crustal extension during the Late Precambrian. Only the major Upper Precambrian intrusive groups and complexes are discussed here.

Logan Intrusions

The Logan intrusions as defined by Grout and others (1959) occur in northeastern Minnesota and in the Thunder Bay district of Ontario from the International boundary to north of Lake Nipigon. The bodies are mainly tabular in form, and numerous sills and dikes are emplaced in the Sibley Group, the Rove Formation, and older Precambrian rocks. Some of the sills in Canada exceed 500 feet in thickness (Moorhouse, 1957).

The Logan intrusions consist mainly of diabase, porphyritic diabase, and gabbro, but basalt, granophyre, and intermediate rocks also occur (Grout and others, 1959). However, chemical and mineralogic dissimilarities of all the Logan intrusions suggest that these rocks are the product of more than one igneous event. Geul (1970) recognized at least two periods of intrusion which he named the Early Mafic intrusions and the Pigeon River intrusions. The Early Mafic intrusions consist of diabase and porphyritic diabase of tholeiitic composition, whereas the Pigeon River intrusions are an equigranular olivine diabase. Geul's Early Mafic intrusions resemble the Logan intrusions in Minnesota which dominantly are sill-like in form. In contrast, the Pigeon River intrusions most commonly occur as dikes. However, the sill at Pigeon Point, Minnesota, which is famous for the following vertical zonation: (1) lower chilled diabase; (2) diabasic gabbro; (3) intermediate rock; (4) granophyre; and (5) upper chilled diabase with anorthosite masses and labradorite phenocrysts (Grout and others, 1959), mineralogically resembles the Pigeon River intrusions.

The emplacement of the Logan intrusions and associated dikes of the Pigeon River intrusions seems to have extended over a long period of time. Hanson and Malhotra (1971) reported an age of 1,300 m.y. on a sill in the Rove Formation, and ages of 1,100, 1,020, 955, and 920 m.y. on dikes. York and Halls (1969) put the ages of two dikes at 1,210 and 1,150 m.y. DuBois (1962) divided the Logan intrusions into two groups on the basis of contrasting magnetization directions and the subsequent petrologic studies by Geul (1970) suggest that these two groups of rocks also can be recognized on petrologic grounds.

Coldwell Complex

The Coldwell Complex, located along the shore of Lake Superior near Marathon, Ontario, is an almost circular body about 16 miles in diameter (Milne, 1967) of alkalic igneous rocks. The five main rock types occur in roughly concentric belts and are classed in two groups, an outer, older, main group and an inner, younger, secondary group (Puskas, 1970). The complex was formed by forceful injection into older, Archean rocks, and it transects the structural trends in the sedimentary-volcanic country rocks. The temperature of intrusion was sufficiently high to form a pyroxene hornfels facies in the contact zone. The complex contains a variety of saturated and undersaturated alkalic igneous rocks. The older main group consists of massive and layered gabbros and laurvikites. The secondary group is composed of syenodiorite, nordmarkite, and nepheline syenites. The complex has been examined as a potential source of iron, base metals, radioactive minerals, nepheline, feldspar, and building stone. Three radiometric ages suggest that the complex was emplaced in Late Precambrian time. Fairbairn and others (1959) reported Rb-Sr ages of 1,225 and 1,065 m.y. on two syenite bodies in the Coldwell Complex. Watkinson (1970) cited the age of the nearby Prairie Lake Complex, composed of ijolitic rocks and carbonatites, as 1,112 m.y.

Duluth Complex

The Duluth Complex defines an arcuate belt that trends northward from Duluth and eastward almost to the Lake Superior shore near the International boundary. It lies stratigraphically at or near the unconformable contact between the Keweenawan stratified sequence and the older Precambrian rocks; for the most part, the contacts and the internal layering dip eastward or southward at low angles. The complex was formed by multiple intrusions, and its component rocks apparently can be explained as resulting from differentiation of a primary basaltic magma (Taylor, 1964). The order of intrusion of the major rock types was 1) gabbroic anorthosite, and 2) troctolitic rocks and, locally, picrite, dunite, norite, olivine gabbro, and ferrogabbro. Blocks of hornfels derived from older rocks are found commonly in the troctolitic rocks. Felsic to intermediate rocks occur discontinuously along the top of the complex from Duluth to the eastern termination.

The Duluth Complex intrudes the lower part of the Middle Keweenawan volcanic sequence, and probably it is overlain unconformably by the Bayfield Group. Radiometric ages reported on rocks from the complex include 1,040-1,200 m.y. (Goldich and others, 1961), 1,115 m.y. (Silver and Green, 1963), 1,050 m.y. (Hanson and Gast, 1967), and 1,115 m.y. (Faure and others, 1969).

Mellen Complex

The Mellen complex extends for about 40 miles along the regional strike and is centered on Mellen, Wisconsin. Stratigraphically the complex lies mainly in the Middle Keweenawan volcanic sequence; Aldrich (1929, 1933) believed it was emplaced along the Keweenaw fault, but Felmlee (1970, *op. cit.*) considered this improbable. Just west of Mellen the complex appears to follow the unconformity at the base of the Keweenawan sequence. The enclosing Keweenawan strata dip northward at 60° (Leighton, 1954), and the complex is about 15,000 feet thick (Olmsted, 1968). The rocks of the complex are similar to those in the Duluth Complex. Olmsted (1968) showed that the main body changes from anorthositic olivine gabbro at the base through anorthosite and ferrodiorite to granite at the top. Leighton (1954) mapped the rocks as a lower gabbro and an upper granophyre. Near Mellen, the gabbroic rocks of the complex are cut by a younger body, the Mellen Granite. The complex intrudes the Middle Keweenawan volcanic sequence, and possibly the lower part of the Copper Harbor Conglomerate. Goldich and others (1961) reported a K-Ar age of 1,000 m.y. for biotite from the Mellen Granite. Chaudhuri and others (1969) gave the Rb-Sr age of the Mellen Granite as 940 m.y.

GEOLOGIC STRUCTURE

Early ideas on the structure of the Upper Precambrian province, based on many years of personal field experience, were summarized by Van Hise and Leith (1911) in their comprehensive monograph on the geology of the Lake Superior region; this report was modified by Leith and others (1935) to take into account the later work. In recent years new insights into the regional structure have followed from a few deep exploratory and water wells and extensive gravity, magnetic, and seismic surveys. The Upper Precambrian province is distinctly linear in plan and displays an overall synclinal structure, but along much of its length the central part of the syncline is the site of an elevated crustal block or horst. Recent discussions of regional structure include those by White (1966a and b) and by King and Zietz (1971). The major structural features of the Upper Precambrian province are treated in this section.

Lake Superior Syncline

Irving (1883) recognized the synclinal nature of the Lake Superior basin and illustrated his conception of the structure on a map with form line contours. Hotchkiss (1923) called this structure the Lake Superior geosyncline, but "Lake Superior syncline" is the name used by most workers at present. This large feature is typically developed under western Lake Superior, but it can be traced across northwestern Wisconsin (Dutton and Bradley, 1970) into east-central Minnesota. It extends at least to the east end of Lake Superior where it passes beneath the Paleozoic beds of the Michigan basin. The syncline is markedly asymmetrical, with steep dips on the south limb in Wisconsin and Michigan, and gentle dips on the north limb in Minnesota and Isle Royale.

The Lake Superior syncline probably started to form with the beginning of the eruption of the Middle Keweenawan volcanic sequence. Subsidence and accumulation were approximately in balance through the time of deposition of the Oronto Group. The Bayfield Group (including the Jacobsville Sandstone) is broadly synclinal, across the province, but the dips are slight and could be initial. The location of the Michigan and Twin City basins on the axis of the Lake Superior syncline, however, suggests that subsidence in some segments of the syncline may have continued into the Paleozoic. Total subsidence of the pre-Keweenawan unconformity surface beneath western Lake Superior appears to be at least 50,000 feet at the synclinal axis.

Folds

Most of the stratified Upper Precambrian rocks of the region display a homoclinal structure related to their position on one limb or the other of the Lake Superior syncline. Folds of various sizes are present in a few localities, however, and these are important in reconstructing the kinematic history of the province. Hall (1901a) illustrated some large folds in the Middle Keweenawan volcanic rocks of east-central Minnesota. Thwaites (1912) described several folds in the Oronto strata in the area north of Mellen. Myers (1971, *op. cit.*) interpreted the steeply-inclined strata along the Middle River north of the Douglas fault as Oronto beds; the great thickness of this section makes folding a more probable explanation than fault drag for the steep dip. Hubbard (1971) described the Porcupine Mountains area as an asymmetric anticline with more than 8,000 feet of structural relief. Spiroff and Slaughter (1961) and White (1971) portrayed smaller folds in the rocks of the Keweenaw Peninsula, and Thomson (1954) discussed small folds in the Middle Keweenawan volcanic rocks of Mamainse Point.

All of the folds are developed in Middle Keweenawan volcanic rocks or in sedimentary rocks of the Upper Keweenawan Oronto Group. Folds are not known to involve rocks of the Bayfield Group, although local flexures attributed to fault drag have been described (Thwaites, 1912). Accordingly, the formation of these folds can be dated as post-Oronto but pre-Bayfield. At least that part of the Bayfield Group visible at the surface has not been involved in the folding.

If the Bayfield Group is truly Keweenawan, as it is considered here, then this folding phase is Late Keweenawan. The orientation of the fold axes and the steepening of the southeast limb of the Lake Superior syncline argue for compression in a northwest-southeast direction, and these structures may be related to the roughly contemporaneous Grenville orogenic belt lying to the southeast (Muehlberger and others, 1967). However, Brown (1971) concluded that mineralization at the White Pine mine preceded this deformational phase, and Chaudhuri and Faure (1967) reported a very tentative age of 720 m.y. for this mineralization.

The Axial Horst

In northwestern Wisconsin and east-central Minnesota the axis of the Lake Superior syncline lies in a crustal block that appears to have been significantly uplifted relative to the blocks to the northwest and southeast. This structure was first clearly defined by Thiel (1956), and Craddock and others (1963) named it the St. Croix horst. At the surface this horst consists mainly of the Middle Keweenawan volcanic sequence, and these beds show moderate to steep inward dips near the borders of the horst. Rocks of the Oronto Group are present in some localities along the synclinal axis or in small grabens atop the horst. The basins adjacent to the horst are underlain by thick sections of Upper Keweenawan sedimentary rocks.

This axial horst may terminate in northwestern Wisconsin, but it probably continues to the northeast. If the Keweenaw fault represents the southern edge, then the horst may well extend into eastern Lake Superior. The distinctive gravity and magnetic anomaly patterns (Hinze, 1963) suggest, but do not prove, that the axial horst is present in lower Michigan.

The St. Croix horst is truncated to the southwest by the Belle Plaine fault in southern Minnesota (Craddock, this chapter), but a similar feature extends southward into Iowa. The axial horst structure appears to continue across Iowa into Nebraska and perhaps into Kansas (Coons and others, 1967; Yoho, 1967; King and Zietz, 1971). The fault-bounded Nemaha anticline of Nebraska-Kansas-Oklahoma lies 40 miles east of the Midcontinent Gravity High and formed during late Paleozoic time (Merriam, 1963); its origin may have been related to continued adjustments in the buried belt of Upper Precambrian rocks (Lyons, 1959).

Major Strike Faults

The most important faults in the Upper Precambrian province are the strike faults that bound the axial horst; most workers interpret them as steep, inward-dipping, dip-slip reverse faults. The Keweenaw fault of the Keweenaw Peninsula is the best known of these strike faults, and it brings Middle Keweenawan volcanics into contact with Jacobsville sandstones. Irving and Chamberlin (1885) believed that it represents a fault line unconformably buried by Jacobsville strata and then reactivated, but Lane (1911, 1916) reconstructed a more complicated history. Bacon (1966) estimated the throw at 10,000 feet, and White (1971) estimated it at thousands of feet. The fault can be traced westward at least to Lake Gogebic and perhaps into Wisconsin (Meshref and Hinze, 1970).

Aldrich (1929) extended the Keweenaw fault into Wisconsin north of the Gogebic Range, and also defined the Lake Owen fault lying almost along the strike to the west. The Lake Owen fault is marked by the apparent thrusting of Middle Keweenawan volcanic rocks over Upper Keweenawan conglomerates, but this interpretation has been challenged (Hubbard, 1968). The area between the Lake Owen fault and the Keweenaw fault has few exposures and a complicated structure, and it is possible that these two faults may be continuous. Geophysical surveys (Sims and Zietz, 1967; Craddock and others, 1970) show that this fault continues southwestward into Minnesota as the Hastings fault. Estimates of the throw on the Hastings fault range from 8,000 to 10,000 feet (Craddock, this chapter).

The Douglas fault can be traced from Douglas County, Wisconsin, southwestward to the Belle Plaine fault in Minnesota. Thwaites (1912) interpreted it as a reverse fault bringing the Middle Keweenawan volcanic rocks up against the younger Bayfield Group, but Raasch (1950) explained it as an unconformity with sandstone resting upon volcanics. Estimates of the throw on this fault range from 8,200 to 11,500 feet (Craddock, this chapter). The behavior of the Douglas fault beyond Douglas County is unknown, but it may continue northeastward into Lake Superior, perhaps through the Bayfield peninsula; Halls and West (1971a) have postulated the existence of faults north of Isle Royale and Michipicoten Island, which could be continuations of the Douglas fault.

Major Transverse Faults

The linear Upper Precambrian province is divided into three segments that may be separated from one another by important transverse faults. A separation and an apparent offset of segments in the province occur in southeastern Nebraska, but subsurface data are too sparse to determine if they are the result of a fault. Another dislocation between segments occurs in southeastern Minnesota; here, the Belle Plaine fault coincides with the dislocation. Sloan and Danes (1962) discovered the Belle Plaine fault by surface geologic mapping and a gravity survey. The fault cuts Paleozoic strata at the surface and has a throw of at least 700 feet, but the mechanical classification and movement history of this fault have not been determined. It can be interpreted as a transform fault, a strike-slip fault, or a dip-slip fault (Craddock, this chapter).

Crustal Structure

Seismic refraction surveys to assess the thickness of the crust have revealed an uncommonly great depth to the Moho under parts of the Upper Precambrian province, and especially in Lake Superior. Steinhart and Meyer (1961) calculated a crustal thickness of 35-37 km beneath the Keweenaw Peninsula, and they presented preliminary crustal thickness contours for much of Wisconsin. Berry and West (1966) reported Moho depths as great as 60 km beneath Lake Superior, and Smith and others (1966) found a maximum depth of 55 km in the same area. Cohen and Meyer (1966) found a crustal thickness of 46 km below the Midcontinent Gravity High in Wisconsin, and a thickness of 42 km along a parallel line over the deep basin to the east. Anzoleaga (1971, unpub. Ph.D. dissert., Univ. Wisconsin) reinterpreted the seismic data from western Lake Superior by also utilizing new gravity measurements; his model reduced the apparent crustal thickness to less than 45 km from an earlier value of 50 km based on the seismic data alone.

STRUCTURAL HISTORY

Major Events

The Upper Precambrian province is a narrow belt that cuts across the structural grain of an older continental platform. The major events in the structural evolution of the province are the following:

1. Early in the Late Precambrian, probably more than 1,300 m.y. ago, igneous activity began with the emplacement of numerous mafic dikes and sills. Many of the dikes tend to parallel the axis of the present Lake Superior syncline.

2. Somewhat later, perhaps about 1,200 m.y. ago, eruption of mafic lavas began from fissures near the axis of the present Lake Superior syncline. Volcanic activity was discontinuous in time and space, but the cumulative thickness of these plateau basalts is about 40,000 feet. Individual flows have great lateral continuity, and almost all were formed under subaerial conditions. Regional subsidence and the growth of the volcanic sequence were about in balance.

3. During the igneous cycle the emplacement of voluminous intrusive bodies, mainly into and at the bottom of the volcanic pile, added to the load accumulating on the surface of the older crust. Although most of these intrusives probably formed along with the volcanics in Middle Keweenawan time, at least some were earlier (see above) and some were later (one rhyolite body cuts the Freda Sandstone).

4. The Middle Keweenawan cycle of subsidence and surface accumulation continued into the Late Keweenawan, and the Oronto Group was deposited in the axial part of the Lake Superior syncline.

5. A deformational phase, affecting rocks as young as the Freda Sandstone, formed a variety of northeast-trending folds and caused local steepening of the southeast limb of the Lake Superior syncline. Some strike faults, such as the one cutting the large Porcupine Mountains anticline, probably formed at this time. Possibly the transverse Belle Plaine fault, if it is a strike-slip fault, had its inception during this phase.

6. During or after this deformational phase, the axial horst of the Lake Superior syncline began to form. The formation of this structural feature probably involved actual uplift of the horst coupled with subsidence in the adjacent basins. The time of formation of the major boundary faults is not clear, but the major movements on these faults seem to have been completed by the end of Bayfield time.

7. Readjustments continued along this linear belt during Paleozoic time, and these movements are recorded by structures in the Paleozoic strata. Younger sedimentary layers above the Upper Precambrian province show a broadly synclinal pattern, and the Twin City basin in Minnesota and the Michigan basin are both centered on this narrow belt. Many small faults in the strata overlying the main boundary faults indicate a moderate continued rise of the axial horst until at least late Paleozoic time.

Mechanical Interpretation

Early and Middle Keweenawan time probably was a period of modest crustal extension, leading to vertical fracturing, rising magmas, and downward bending of the crust. This belt of igneous activity and subsidence may represent an incipient rift of continental dimension, the development of which was arrested in an early stage. Green (1971b) has drawn attention to similarities between the Keweenawan flows and the Tertiary plateau basalts of Iceland, and Kristmannsdottir (1971) has described a mafic sill with large anorthosite inclusions emplaced in the Icelandic volcanic sequence. That the amount of crustal extension in the Lake Superior region was slight, however, is suggested by the absence of evidence for the normal faulting and graben formation so common, for instance, in the East African-Red Sea system (Lowell and Genik, 1972).

The deformational phase that followed deposition of the Oronto Group suggests that an important mechanical change occurred during the Late Keweenawan. The folds and faults formed during this phase indicate compression of the layered rocks in a northwest-southeast direction. Although some of these structures might be a local response to tilting from deep subsidence, a regional compression probably affected the rocks of the Lake Superior syncline. This new stress pattern may have been related to the roughly contemporaneous Grenville orogeny in the region to the southeast.

The nature of the forces that formed the axial horst is not clear, but at least two possibilities exist. Because the main boundary faults are closely related in space, and perhaps in time, to the folds, flexures, and faults formed during the post-Oronto deformational phase, the axial horst may be mainly the result of lateral compression. On the other hand, buoyancy forces and isostatic adjustments to the earlier deep subsidence may have played an important role. The latter explanation better accounts for the unfolded character of the Bayfield beds, the long duration of movements into the Paleozoic, and the sharp bend in the Upper Precambrian belt (probably primary) in eastern Lake Superior where it turns southeastward into lower Michigan.

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GENERAL GEOLOGY, NORTHEASTERN MINNESOTA

J. C. Green

A large part of northeastern Minnesota is underlain by Upper Precambrian rocks, which generally are designated as Keweenawan (see fig. V-3). These constitute the northwest branch of the Midcontinent Gravity High (see pl. 2), a positive gravity feature produced by a thick mass of mafic igneous rocks (Craddock and others, 1970). According to seismic studies (Smith and others, 1966), the crust is anomalously thin (<30 km), at least in the southwestern part (Duluth-Two Harbors). Interpretation of this structure as an abortive continental rift has been suggested by several workers, and King and Zietz (1971) have summarized many of the arguments.

In contrast to the Upper Precambrian of Michigan and Wisconsin, which is composed principally of lavas and overlying clastic strata, the Upper Precambrian of northeastern Minnesota consists primarily of intrusive rocks, which underlie and transect volcanic rocks. The majority of the intrusive rocks are assigned to the Duluth Complex; the volcanic rocks have been named the North Shore Volcanic Group. Descriptions and discussions of these major units as well as the minor intrusions are given below.

Recent mapping (for example, Green, 1971a and b; White and others, 1971), seismic (Halls and West, 1971a and b), gravity (White, 1966a; Craddock and others, 1970; Ikola, 1970), and paleomagnetic studies (for example DuBois, 1962; Books, 1968; Palmer, 1970; Beck and Lindsley, 1969) provide many new insights into the development of the Lake Superior syncline in Late Precambrian time. Whereas earlier it commonly has been assumed that all the Upper Precambrian volcanic rocks of the Lake Superior region accumulated during the same grand event (see Craddock, this chapter), the new data show that there was more than one major episode of volcanism and intrusion, and perhaps that the volcanic rocks accumulated in several separate basins.

Using a reversal of the Earth's magnetic pole (from a reversed to normal orientation) as a time-stratigraphic marker to separate Lower from Middle Keweenawan rocks, it has been shown that major volcanism took place during Early Keweenawan time on both the south shore ("Traps of the South Range," east and west of Ironwood, Michigan) and the north shore (Osler series, Ontario, and Grand Portage to Hovland, Minnesota and westward). Lavas at the base of the Upper Precambrian succession at Duluth also appear to have been erupted at this time (Green and Books, 1972). Some remarkable lithic similarities imply that these lavas may have been erupted during the same event (and possibly in the same basin) as those 150 miles away at Grand Portage, and that those at Ironwood similarly are correlative to the succession at Hovland, 100 miles away across the Lake Superior basin. However, after these Lower Keweenawan lava sequences were deposited—perhaps but not necessarily in a single broad basin—fragmentation began. A major intrusion of gabbroic magma, which produced the remarkable layered northeastern (or Gunflint) prong of the Duluth Complex, penetrated along the unconformity between the Upper and Middle Precambrian rocks in Cook County, Minnesota—still during the time of reversed magnetic polarity.

Subsequently, the magnetic pole inverted to its normal orientation, and after an unknown time interval volcanism recommenced in Minnesota, producing the remaining succession of the North Shore Volcanic Group. Volcanism did not occur in a simple, symmetrical basin, however, but instead apparently involved two major depositional sites, one on either side of an area near Beaver Bay. To the northeast, some of the volcanic units can be traced around at least half the basin, but the uppermost part of the sequence—near Tofte and Schroeder—is markedly asymmetric. Southwest of Beaver Bay, the lavas appear to be superposed continuously from Duluth to Split Rock River, but geophysical evidence (White, 1966a; Ikola, 1970) implies a rise in the basement near the mutual boundary of Lake and Cook Counties, which would in turn imply another strongly asymmetrical basin of accumulation. These rocks were intruded subsequently by the main mass of the Duluth Complex, the Beaver Bay Complex, and other smaller gabbroic intrusions. In fact, petrochemical evidence implies that the younger Middle Keweenawan part of the North Shore Volcanic Group may have been erupted from the same magma chamber that produced the bulk of the Duluth Complex (Phinney, 1970).

The main part of the Duluth Complex was intruded generally along the base of the Middle Keweenawan lavas, but locally transected rocks both above (Middle Keweenawan lavas) and below that zone (Lower Keweenawan lavas, Middle and Lower Precambrian rocks). In fact, xenoliths of a variety of rocks are abundant in many parts of the Duluth Complex, as discussed below by Phinney, Davidson, and Bonnichsen; many of these could be metamorphosed lavas.

One of the more intriguing problems of the Keweenawan geology of Minnesota is the three-dimensional shape of the system, and particularly of the Duluth Complex. The outlines of some of the local intrusive bodies, such as the Bald Eagle intrusion of Lake County and the layered rocks of the Gunflint prong in Cook County, can be inferred from the surface patterns of foliations and lithologic subunits. However, inasmuch as the densities of the gabbroic rocks and basaltic lavas are similar, the subsurface configurations of the intrusions cannot readily be determined by gravity measurements. Furthermore, the low relief of the area and the locally sparse outcrops preclude determining dips where contacts cannot be observed directly. Possibly, some of the large areas of mafic hornfels in northwestern Cook County

(Morey and others, 1969, geol. map of Long Island Lake quad., Minn. Geol. Survey open-file map) and in east-central Lake County (Davidson, this chapter) are erosional outliers of the roof of the complex. Much more mapping and geophysical work are needed along the upper contact of the Duluth Complex.

From considerations of lithic character, structure, and stratigraphy, as well as geomorphology and geophysics, it appears that the youngest major lavas of the Lake Superior basin—the Portage Lake Lava Series of the Keweenaw Peninsula and Isle Royale, Michigan—were deposited in a still later basin that occupied the central part of the present Lake Superior syncline. It was at this stage that the present configuration of the Lake Superior syncline was developed. These younger rocks are regionally discordant to the synclinal axis at both Grand Portage and Duluth, and may overlie the rocks of the North Shore Volcanic Group unconformably. Thus, the Portage Lake Lava Series has no equivalents in northeastern Minnesota. It was followed by deposition of the Upper Keweenawan clastic sediments, with local volcanic activity continuing for a time in the Porcupine Mountains area of Michigan (White and others, 1971). According to White (1966a), the straight Minnesota shoreline of Lake Superior may be a relic of a mid-Upper Keweenawan unconformity that formed on the earlier lavas and immature sediments and was overlain by younger mature sandstones. By the time of the Late Keweenawan sedimentation in Minnesota (for example at Fond du Lac, Duluth), the North Shore Volcanic Group had been buried, subjected to greenschist- and zeolite-facies metamorphism, tilted, and eroded.

NORTH SHORE VOLCANIC GROUP

J. C. Green

PREVIOUS WORK

The Upper Precambrian lavas of northeastern Minnesota (North Shore Volcanic Group of Goldich and others, 1961) have attracted the attention of geologists for well over a century, particularly since the pioneering work by Douglass Houghton on the native copper deposits in similar rocks on the Keweenaw Peninsula of Michigan in 1841. Notable early studies on the north shore were carried out by Irving (1883), Elftman (*in* Winchell, 1894; 1898), and N. H. Winchell (*in* Winchell and others, 1899; *in* Winchell and Grant, 1900). More recently, detailed mapping of outcrops along the Lake Superior shore was done by A. E. Sandberg (1938), who mapped the flows between Duluth and Two Harbors, by Grout and Schwartz (1939), who studied the intrusions and flows in the eastern part of Lake County, by R. M. Grogan (1940, unpub. Ph.D. thesis, Univ. Minn.), who mapped the lakeshore between Two Harbors and Split Rock River, by George Gryc (1942, unpub. M.S. thesis, Univ. Minn.), who mapped the Grand Portage area, by Schwartz (1949), who studied the rocks of the Duluth area, by Grout and others (1959), who mapped most of Cook County, and by Norris Jones (1963, unpub. M.S. thesis, Univ. Minn.), who mapped the Hovland complex. Starting in 1965, I mapped the shoreline between Silver Bay and Grand Portage, and have done considerable reconnaissance to the southwest as well as inland. My work has been supported in part by grants from the National Science Foundation, although the field work has been done as a part of the geologic mapping program of the Minnesota Geological Survey. A few preliminary reports have appeared (Green, 1968a and b, 1970a, 1971a and b) and studies are continuing.

REGIONAL RELATIONS

Relations to Underlying Rocks

In the northeastern corner of Minnesota the lowest exposed Upper Precambrian flows disconformably overlie a thin quartzite (Puckwunge Formation; see Mattis, this chapter), which in turn overlies, apparently disconformably, the Middle Precambrian Rove Formation; here, all the rocks strike eastward and dip approximately 10° S. (see fig. V-3). At the southwest end of the basin immediately west of Duluth, 155 miles away, the lowermost Upper Precambrian flows also conformably overlie a thin quartzite informally called the Nopeming formation (fig. V-4A). They overlie the vertically folded slates and metagraywackes of the Middle Precambrian Thomson Formation, which is correlated with the Rove (Goldich and others, 1961). At this locality, the flows strike north and dip about 15° E. The angular unconformity here reflects the diastrophism and subsequent erosion associated with the Penokean orogeny, which evidently did not affect the northeastern corner of the state. Across the axis of the Lake Superior basin in northern Wisconsin and Michigan, the lowermost flows also conformably overlie a quartzite known as the Bessemer, which in turn overlies Middle Precambrian shale and graywacke with only minor discordance.

Several types of intrusive rocks of Late Precambrian age cut the Lower and Middle Precambrian as well as older Upper Precambrian rocks.

Relations to Overlying Rocks

The youngest Precambrian rocks in northeastern Minnesota are lavas and intrusive bodies. On Isle Royale, Michigan, however, only 20 miles east of Grand Portage, Minnesota, as well as on the Keweenaw Peninsula across the lake (and across the synclinal axis; see fig. V-3), the lavas are overlain conformably by clastic sedimentary strata that form the core of the Lake Superior syncline. South of Duluth, the lavas and intrusions also are overlain, apparently unconformably, by immature clastic sedimentary rocks (Fond du Lac Formation). White (1966a) has suggested that the remarkably straight and smooth Minnesota shore of Lake Superior, which truncates the lavas at Duluth and Grand Portage, may be an exhumed mid-Upper Precambrian unconformity.

Exposures of Upper Precambrian bedrock are excellent along the Lake Superior shore and in the beds of major streams that flow into the lake, and are relatively good in a wide easterly-trending zone between Ely and the lakeshore.

Internal Relations

The Upper Precambrian of the Lake Superior district consists of a variety of volcanic successions, clastic sedimentary strata, and intrusions, whose relations to one another are rather obscure because of faulting, dislocations caused by intrusion, poorly exposed contacts, and a lack of precise geochronologic data.

In earlier reports on the Lake Superior district, all the Upper Precambrian rocks were considered to be correlative with rocks exposed on or near the Keweenaw Peninsula, Michigan, and accordingly have been called Keweenawan. A tripartite stratigraphic division, as follows, has been standard (Van Hise and Leith, 1911): Lower Keweenawan, which was defined as including only the basal sandstone (Puckwunge, Bessemer); Middle Keweenawan, including all the lava flows and associated interflow sedimentary strata

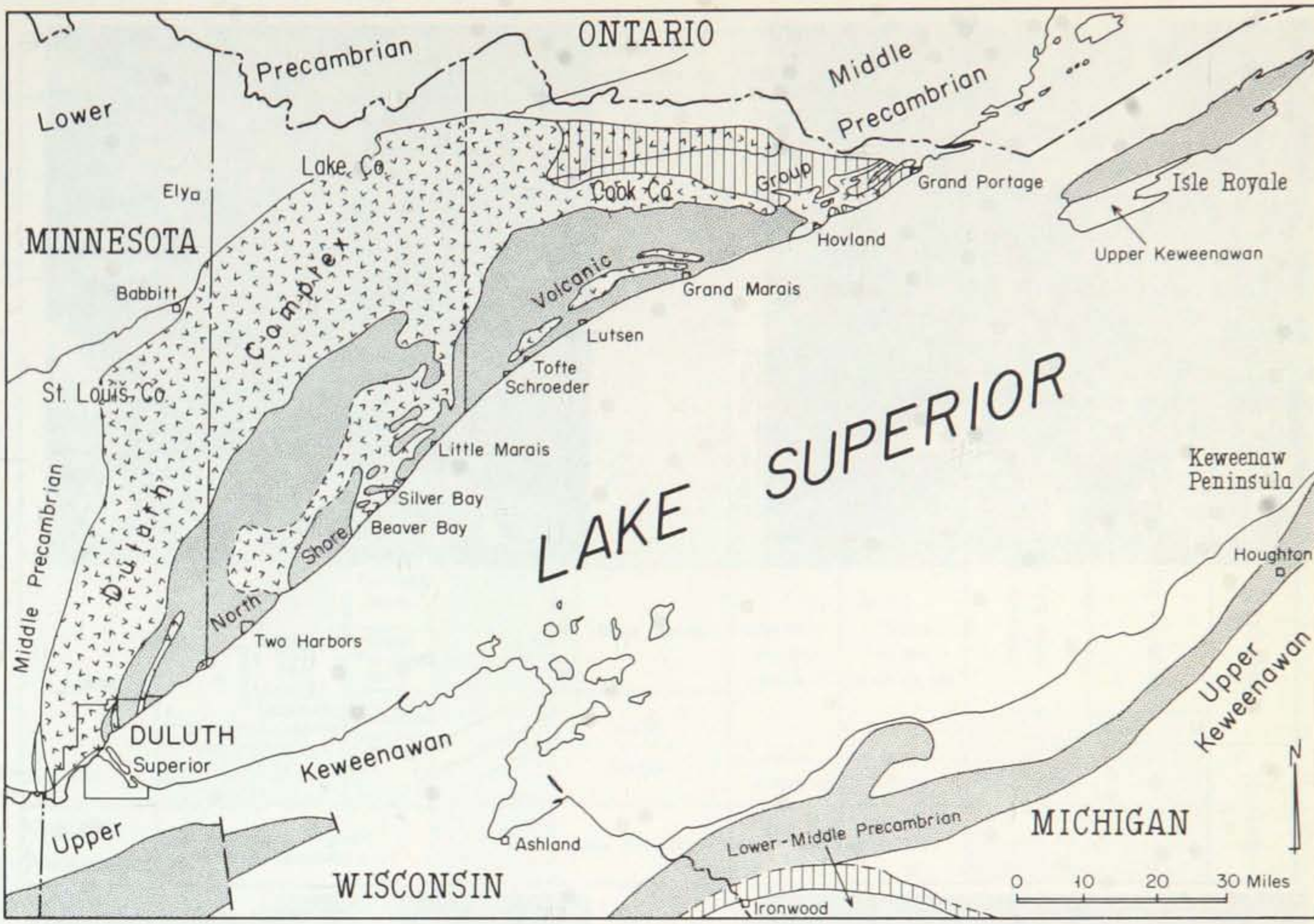

Figure V-3. Generalized geologic map showing the Keweenawan rocks of northeastern Minnesota and the western Lake Superior basin. Check pattern, intrusive rocks; vertical rule, Lower Keweenawan rocks (magnetically normal and reversed); stipple, Middle Keweenawan lavas (magnetically normal).

and most of the intrusions; and Upper Keweenawan, which includes all the clastic sedimentary rocks that overlie the lavas. Such a division was convenient and consistent for the major copper-bearing area of the Keweenaw Peninsula and for Minnesota. There always has been a question, however, about the contemporaneity of the various volcanic, intrusive, and depositional events in this unfossiliferous sequence across the width and length of the Lake Superior basin.

Recent paleomagnetic studies (DuBois, 1962; Beck, 1970; Books, 1968; Palmer, 1970) have shown that two reversals of magnetic polarity occur in the sequence (fig. V-5). These do not coincide with the established stratigraphic division, but they provide a means for time correlation of certain intervals throughout the basin. Books (1968), therefore has proposed that the boundary between the Lower and Middle Keweenawan be placed at the second magnetic reversal, where rocks of reversed magnetic polarity are overlain by rocks of normal polarity. The first reversal (normal to reversed) is situated near the base of the "Traps of the South Range" near Ironwood, Michigan and within the Sibley Series of the Thunder Bay district, Ontario, but has not yet been recognized in Minnesota; the oldest flows in Minnesota have reversed polarization.

The Lower Keweenawan, as defined magnetically, includes all the "Traps of the South Range," across Lake Superior and about 10,000 feet of lavas at the base of the section in the Grand Portage-Hovland area in northeastern Minnesota (see figs. V-3 and V-5). The magnetically defined Middle Keweenawan includes the remainder of the Upper Precambrian lavas of Minnesota, with the possible exception of those that underlie the Duluth Complex immediately west of Duluth; these are tentatively considered Lower Keweenawan because of their striking similarity in lithology and sequence to the flows at the Susie Islands near Grand Portage. The magnetically defined Middle Keweenawan also includes all the major intrusive rocks, except for some of the Logan intrusions and the northeastern part

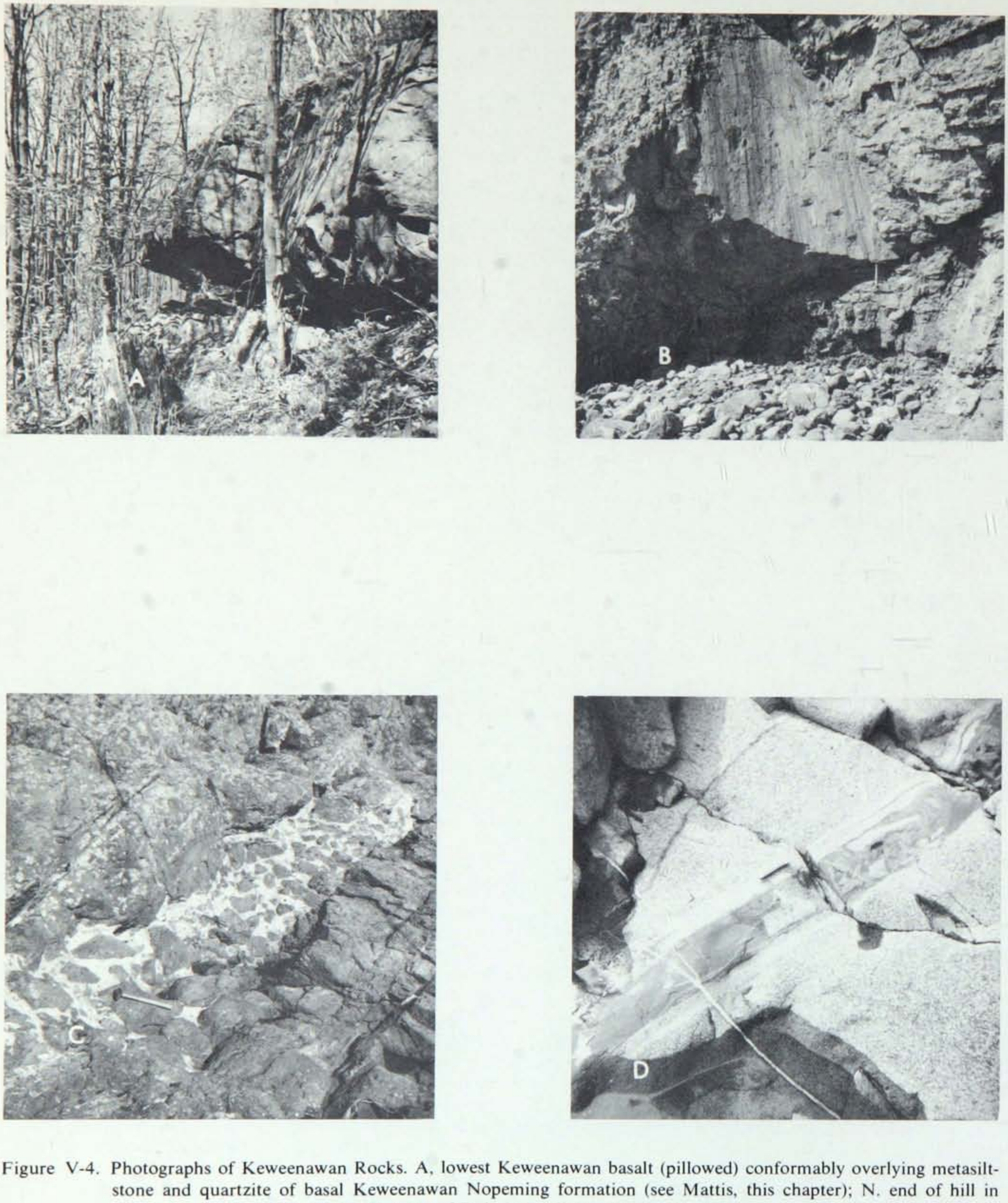

Figure V-4. Photographs of Keweenawan Rocks. A, lowest Keweenawan basalt (pillowed) conformably overlying metasiltstone and quartzite of basal Keweenawan Nopeming formation (see Mattis, this chapter); N. end of hill in SW¼ sec. 17, T. 49 N., R. 15 W., west of Duluth; B, slickensided fault surface. Lake Superior shore southwest of Little Marais, Lake County; C, fault breccia in basalt with calcite matrix, Lake Superior shore, northeast of Tofte, Cook County; D, red, cross-bedded sandstone with lava blocks filling open 5-inch fracture in ophitic basalt flow. "Clasolyte" is subsequently cut by white, calcite-zeolite vein (lower left). Lake Superior shore, NE. of Little Marais, Lake County.

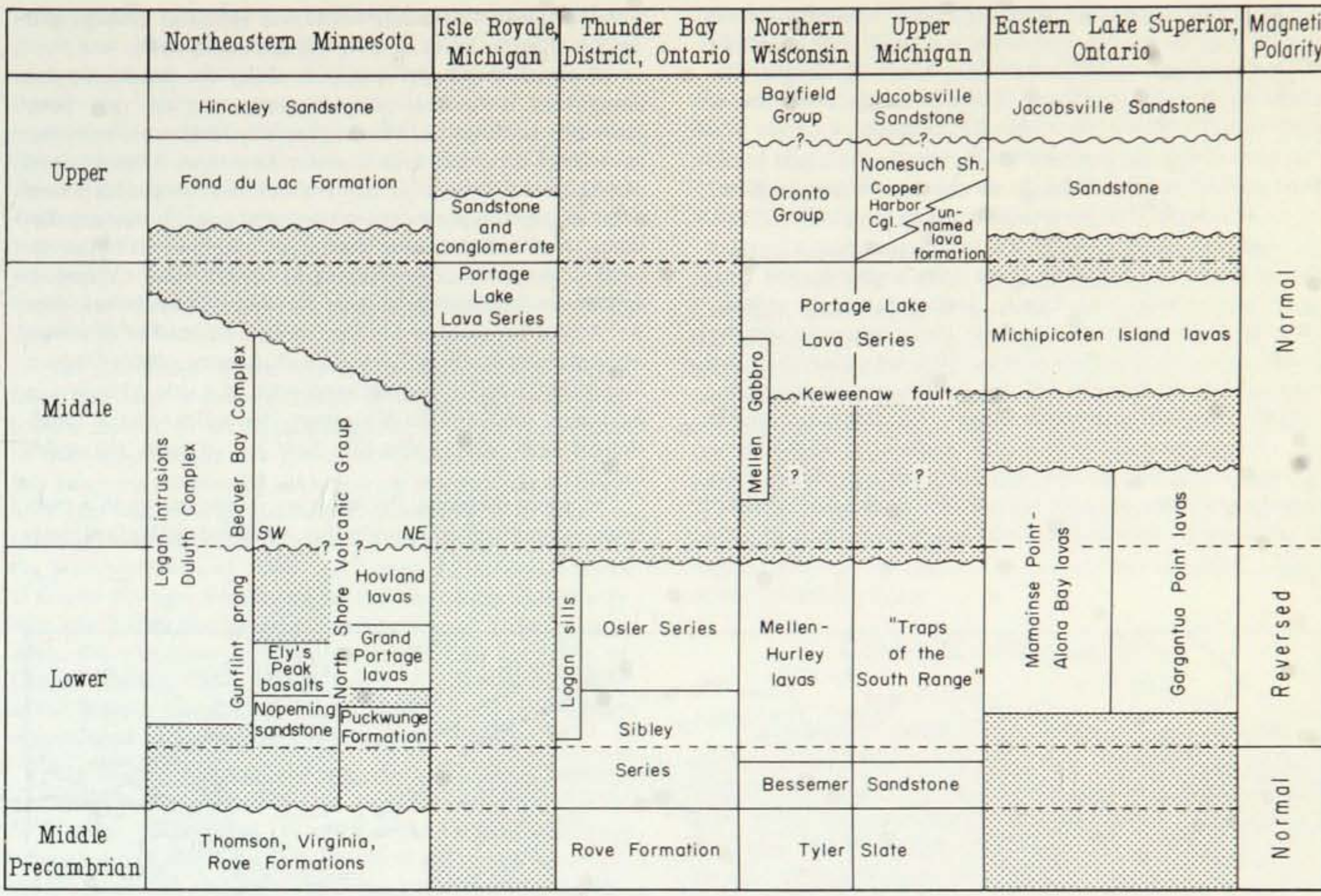

Figure V-5. Proposed correlations for the Keweenawan rocks of northeastern Minnesota. Solid areas indicate section missing or not exposed.

(Gunflint prong) of the Duluth Complex; these are magnetically reversed, and thus probably are Lower Keweenawan.

In northeastern Minnesota, intrusive rocks occupy a larger proportion of the area of Upper Precambrian rocks than do the lavas. The lavas were named the North Shore Volcanic Group by Schwartz (Goldich and others, 1961, p. 81), because they crop out principally along the Minnesota shore of Lake Superior; but they are segmented into several separate parts by major intrusions, and it is difficult to trace units across such discontinuities. The major intrusive unit is the Duluth Complex, an immense anorthosite-troctolite-gabbro-granophyre complex, described below, that was emplaced in several stages, generally along the unconformity between Middle and Upper Precambrian rocks or between Lower and Middle Keweenawan lavas. Other important intrusive units, which may or may not be related to the Duluth Complex, are the Beaver Bay Complex—which intrudes higher levels of the lavas—the Hovland diabase complex, the Reservation River diabase, and the Logan intrusions, a complex group of large sills and dikes in the northeasternmost part of the state.

STRUCTURE

In general, the North Shore Volcanic Group has the form of half a large, filled dish that is tilted southeastward toward Lake Superior (see fig. V-3). At its southwestward limit at Duluth, the flows strike north and dip about 15° E. Proceeding northeastward, the lakeshore in general intersects progressively higher level flows to Tofte, in southern Cook County, where the youngest flows in Minnesota are exposed. These strike approximately parallel to the shoreline and dip about 12° SE. Northeastward from Tofte, the flows strike more easterly than the lakeshore, so that successively lower flows are intersected. At the Onion River, a few miles northeast of Tofte, another slight change of strike takes place, and progressively higher flows are exposed again near Lutsen. For several miles northeast from Lutsen, high-level flows trend only slightly more easterly than the lakeshore, but the strike then swings more easterly and lower flows are encountered all the way to the base of the section at Grand Portage. Here, the flows strike slightly north of west and dip 10° S.; eastward across Grand Portage Island and Lucille Island, the base of the flows gradually changes in strike to about N. 75° E. and dips 16° S.

The lavas retain the same structural relations inland from the lakeshore, but exposures are much less abundant and flow contacts, and therefore dips, are rarely discernible. Several major flows or groups of similar flows can be traced inland from the shore for at least 25 miles (fig. V-6).

The flow trends, outlined above, are complicated locally by intrusions and faulting. In the area between Silver Bay and Little Marais, for example, many large dikes, plugs, sills, and less regular bodies of the Beaver Bay Complex intrude the lavas, breaking them into segments and producing many faults and, locally, major changes in attitude of the flows. In one area, the flows strike eastward and dip 55° N.; at another locality nearby, they are overturned and strike northward and dip 50° W. In this area, particularly, it is extremely difficult to trace individual flows and, therefore, to determine the stratigraphic succession of the lavas, although local successions can be mapped. Other areas where major intrusions complicate the correlation of flows are at Duluth, where the Duluth Complex intrudes the lavas, and near Hovland, where the Hovland diabase and the Reservation River diabase intrude the lavas.

Most of the faults mapped along the lakeshore are thought to have small displacements and to have been caused by adjacent intrusions; generally, they are transverse, have steep dips, and lack a preferred strike or direction of displacement. Typically, they are seen only on the lakeshore or in streambeds, and cannot be traced laterally for any distance (fig. V-4B). One of the larger longitudinal faults occurs northeast of Little Marais, in eastern Lake County; it appears to continue for at least 5 miles along strike and dips 30° NW. where exposed in the Caribou River. The direction of displacement was not determined. A more steeply-dipping longitudinal fault exposed southwest of Little Marais can be traced along strike for about 2½ miles; and a longitudinal fault near Tofte (fig. V-4C) is at least 1½ miles long.

In many outcrops, the lavas are transected by thin fractures as much as 5 cm wide that are filled with clastic ma-

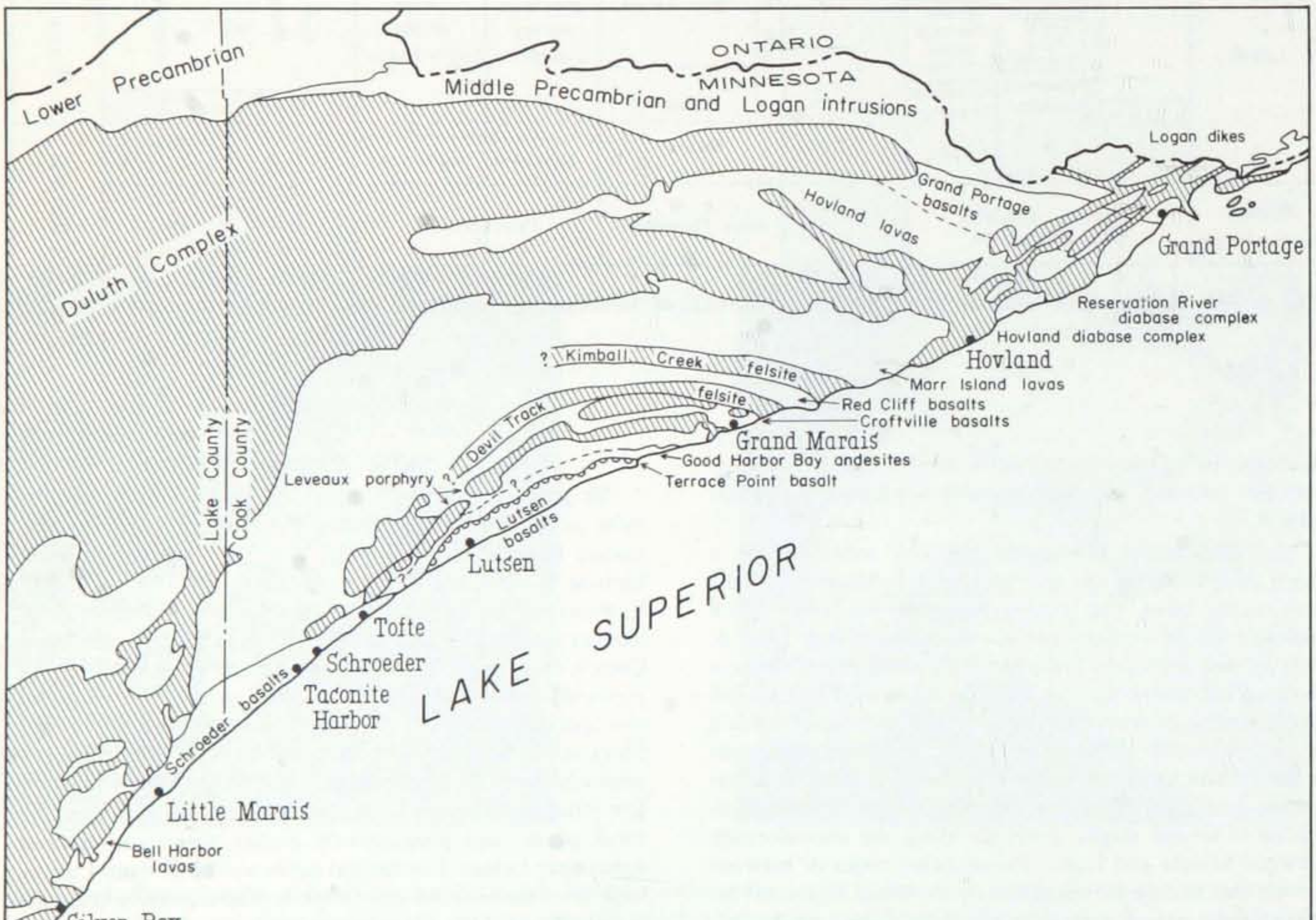

Figure V-6. Generalized geologic map of the North Shore Volcanic Group and associated intrusive rocks of northeasternmost Minnesota, showing continuity of some of the major lava units.

terial, mostly red sand and silt similar to the interflow deposits and referred to as "clasolytes" in some earlier reports. In some of the material, cross-bedding can be seen. Evidently the fractures were filled by sediments that were washed in from above (Fackler, 1941), implying that the fractures were open and were filled before the overlying flows were deposited (fig. V-4D). Although some of these clastic dikes follow cooling joints, others appear to occupy tectonic fractures that were formed contemporaneously with volcanism. Some fractures containing zeolites and associated hydrothermal minerals were not filled until after volcanism had ceased.

As mentioned above, the dip of the flows decreases from the base of the succession at Duluth northeastward toward the higher stratigraphic levels, in a manner similar to that observed on the Keweenaw Peninsula. Generally, this has been interpreted as indicating that the Lake Superior basin was sinking medially concurrently with lava extrusion. However, there is no significant change in dip from the youngest flows at Tofte and Lutsen to the oldest flows at Grand Portage, which implies that the tilting in that area took place after the last of the exposed lavas was extruded. Also, the rhythmic layering in the Duluth Complex at Duluth (Taylor, 1964) dips 20° to 35° E., at least as steeply as the flows it intrudes. This also implies tilting after extrusion of lavas and intrusion of the Duluth Complex.

The thickness of the lava succession has been measured and estimated in several ways. J. A. Kilburg (1972, unpub. M.S. thesis, Univ. Minn.) mapped about 1,000 feet of lavas beneath the Duluth Complex southwest of Duluth. Sandberg (1938) measured 21,330 feet of flows and minor interbedded sandstones along the Lake Superior shore between Duluth and Two Harbors, and Grogan (1940, *op. cit.*) measured 2,292 feet of flows between Two Harbors and Split Rock River, where the Beaver Bay Complex occurs, for a combined total of 24,620 feet. Although it is conceivable that such measurements, made at a relatively small angle to the strike of the strata, could result in an overestimate of the true thickness because of successive shingling of flows to the northeast, this would be quite fortuitous; and if the basin were sinking during extrusion, these estimates of thicknesses would be minimal. On the other hand, this part of the Keweenawan area is inferred by White (1966a) from geophysical data to correspond to a depositional basin; and if, as suggested by White, the basement rises in the vicinity of the Beaver Bay Complex, there must be considerable shingling, and accordingly the estimate of 24,620 feet of strata would be excessive. White (1966a, p. E12) estimated a total thickness of about 40,000 feet of lava in this basin, which is centered under northwestern Wisconsin; only a small amount of this total would underlie the Minnesota shoreline. Thus, the true thickness of this succession in Minnesota remains uncertain.

Reliable estimates of thickness cannot be made in the Beaver Bay-Silver Bay area because the lavas are highly deformed and separated by intrusions. For several miles northeast of Silver Bay, the flows are less deformed and interrupted, but are considerably faulted; from my mapping, I estimate that about 5,000 feet of lavas occur along the lakeshore between Silver Bay and the top of the succession at Tofte.

Northeast of Tofte and Lutsen, where the flows parallel the shoreline, about 12,000 feet of lavas have been measured above diabase of the Hovland complex (see cross section, fig. V-7). Below the diabase, there is a Lower Keweenawan succession of lavas about 9,000 feet thick, giving a total thickness in this area of about 21,000 feet.

Cross-section profiles give estimated thicknesses between 11,000 and 18,000 feet at Tofte above the Duluth Complex, assuming dips between 12° and 20°. Although the average dip at Tofte is about 12°, there is very little control on dips near the base of the sequence, for flow contacts are rarely exposed in the few inland outcrops. Farther northeast at the Cascade River, a thickness of about 15,000 feet of lavas is calculated, using an average dip of 12°, to occur above the Duluth Complex; possibly as much as 5,000 feet of lavas lie beneath the complex in this area, and farther east near the Brule River 10,000 feet of Lower Keweenawan lavas are inferred to underlie the southern prong of the Duluth Complex.

GENERAL CHARACTER

Petrography

The North Shore Volcanic Group resembles, both physically and chemically, plateau lava sequences of various geologic ages, but it is much more variable in composition than some, as for example, the Deccan, Coppermine, or Columbia River basalts. Similarities to the Tertiary plateau lavas of eastern Iceland are particularly striking. The lavas are almost entirely subaerial; they have highly vesicular (now amygdaloidal) upper portions, massive interiors, and various types of jointing, surface features, and textures, depending on their specific compositions.

Evidence for subaqueous extrusion is rare. At the base of the sequence west of Duluth the lowest flow is pillowed; on Grand Portage Island, the basal flow has spheroidal forms that could be pillows, and excellent, thick-rinded, vesicular pillows constitute a flow on the lakeward side of the island a short distance above the base of the succession (fig. V-8A). Unequivocal but less well formed pillows and pillow-breccia are rare higher in the succession, but were seen (1) in the west bank of the Brule River, sec. 27, T. 62 N., R. 3 E. (fig. V-8B); (2) in a creek about 2 miles to the west; and (3) along the Lake Superior shore near Little Marais. Also, Grogan (1940, *op. cit.*) described pillowed basalt from near Two Harbors. These local pillowed lavas could have formed in small lakes or streambeds on the lava surface.

The flows are generally tabular, and individual flows or flow groups can be traced along strike for distances of at least 20 miles (fig. V-6), giving the general impression of a broad, rather flat volcanic terrane. In contrast to eastern Iceland (Walker, 1964), however, definite evidence of volcanic centers, representing shield or composite volcanoes contemporaneous with the plateau volcanism, has not been found. White (1960) has drawn attention to the remarkable areal extent and volume of some Keweenawan flows in

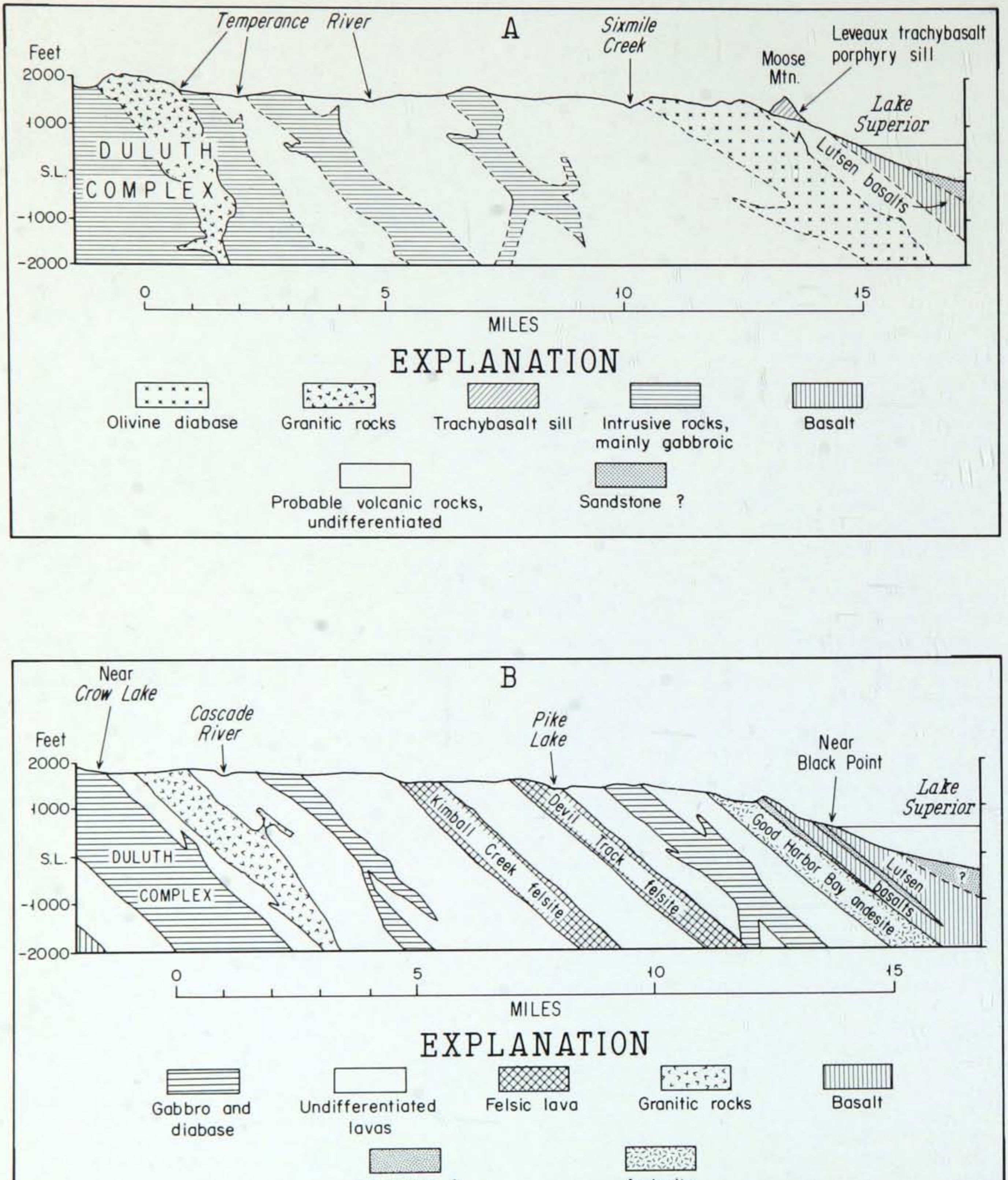

Figure V-7. Interpretive geologic sections across North Shore Volcanic Group, Cook County. A, from near Lutsen on Lake Superior shore to northwest; B, from near Cascade River on Lake Superior shore to north. Dips of formations not to scale. See Figure V-6 for lines of sections.

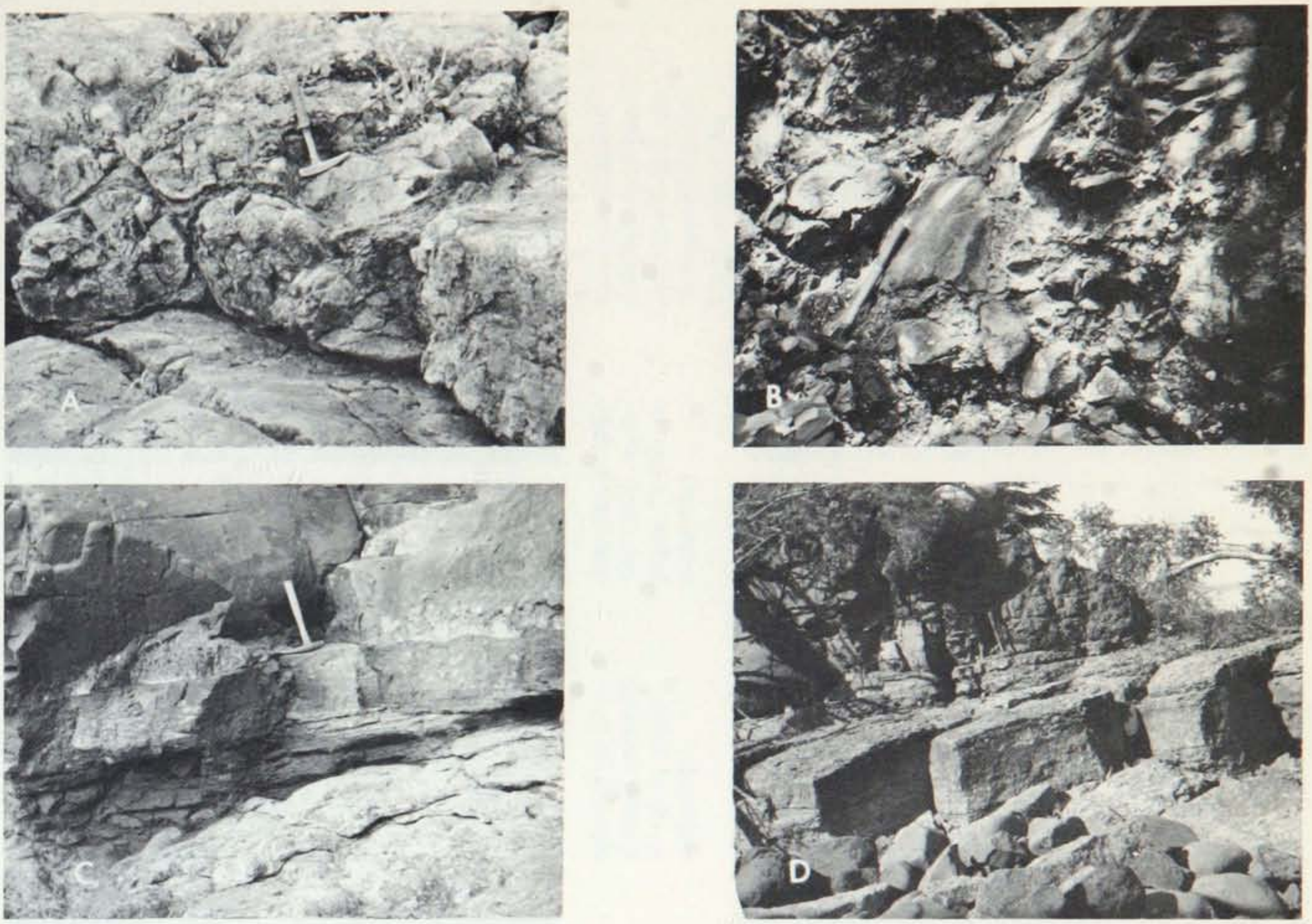

Figure V-8. Photographs of Keweenawan lavas and interbedded sedimentary rocks. A, pillow basalt, a few flows above base of section. SE. shore Grand Portage Island, Cook County; B, coarse basaltic pillow-breccia in Brule River lavas, W. bank Brule R., sec. 27, T. 62 N., R. 3 E., Cook County; C, two- to three-foot lens of red interflow sandstone, Lake Superior shore, northeast of Little Marais, Lake County. Uppermost foot of sandstone apparently has been baked by the overlying ophitic basalt. Note pipe amygdules in base of upper flow and non-eroded, lumpy surface of underlying lava; D, red interflow conglomerate bed overlain by basalt, Lake Superior shore near Lutsen, Cook County. Only Keweenawan rock types were identified in this conglomerate.

Michigan, and with ample justification called them flood basalts.

Interflow sediments make up a small part of the sequence. In a 21,000-foot succession of lavas measured at Duluth, Sandberg (1938) found 1.3 percent of interflow sandstones, and in the northeastern limb of the succession in Cook County, an estimated 18,000 feet of stratigraphic section contains 2.8 percent of interflow sediments (Grout and others, 1959). These sedimentary rocks are principally red, cross-bedded sandstones that occur sporadically as beds a few inches thick between flows (fig. V-8C). Conglomerate is rare (fig. V-8D). Some sand occurs in cavities in the upper parts of flows and forms a matrix for flow-top breccia in others (fig. V-9A). The sediments appear to have been deposited by temporary streams that meandered across the volcanic surfaces. There is little direct evidence for erosion between lava flows, but the sandstones evidently had a Keweenawan volcanic provenance. Pyroclastic deposits are extremely rare, but welded tuff and mixed sand and shards have been reported from the Cascade River in Cook County (Johnson and Foster, 1965) and basaltic or andesitic breccia, other than flow-top breccia, is present at a few localities.

Aside from the high potassium content of some mafic and intermediate members and the relative abundance of rhyolite, the lavas are very similar in composition to the plateau lava series in Iceland and elsewhere. Tables V-1 and V-2 show the general characteristics and abundance of the major types, and Table V-3 summarizes all chemical analyses of lavas of the North Shore Volcanic Group made since 1899. The chemical characteristics of the group are discussed in a later section.

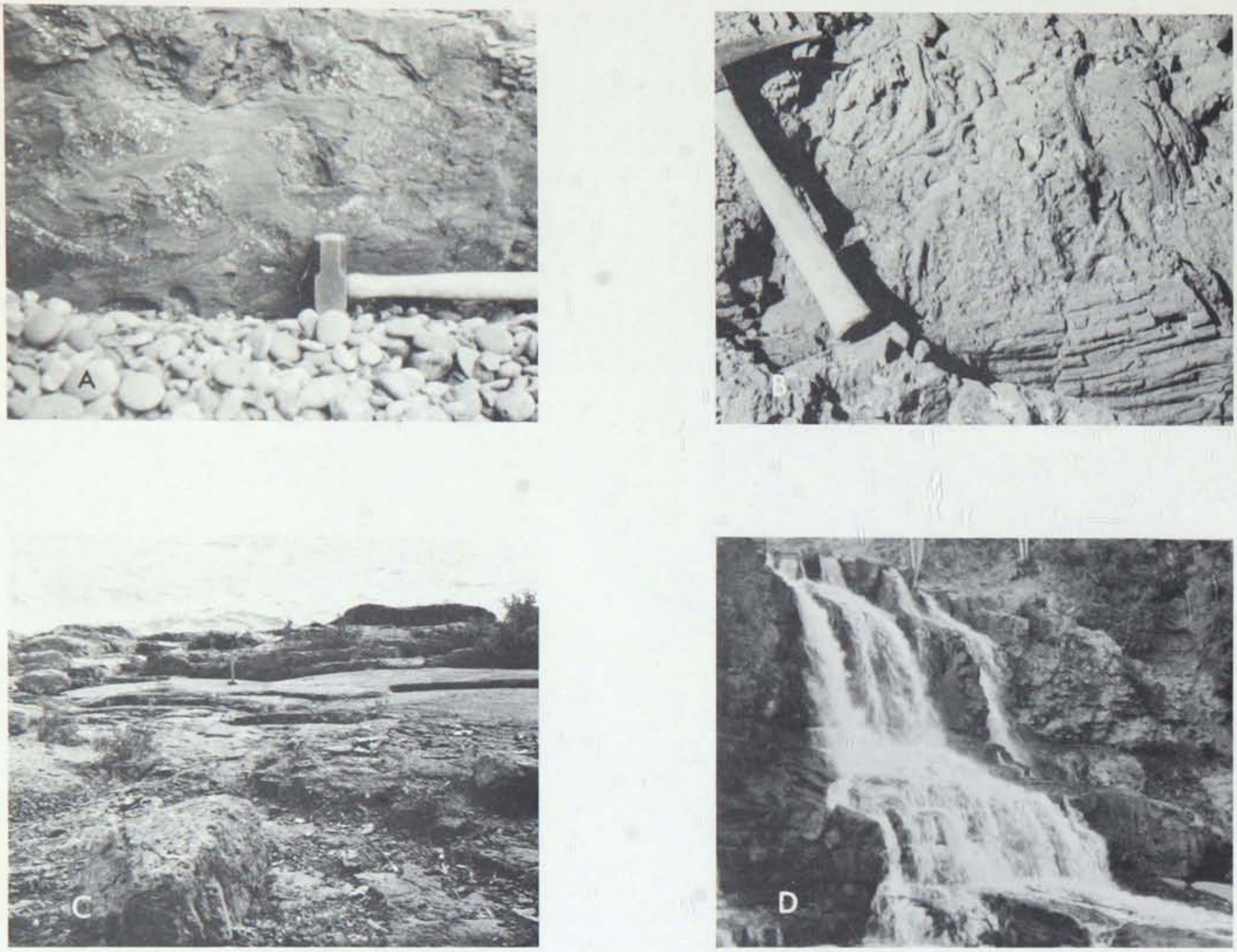

Figure V-9. Photographs of Keweenawan lavas. A, scoriaceous rubble of flow-top of Bell Harbor lavas in which interstices have been filled with red, laminated sandstone. Lake Superior shore southwest of Little Marais, Lake County; B, several layers of ropy crust on ophitic olivine basalt flow, Lake Superior shore near Little Marais, Lake County (Schroeder basalts); C, broad, smooth, billowy top surface on olivine basalt, Lake Superior shore southwest of Tofte, Cook County (Schroeder basalts). Overlying flow has been stripped off by wave action; D, two smooth, gently billowing flow tops of ophitic olivine basalt (middle and just above base of falls). Gooseberry River, just below U.S. Highway 61, Lake County.

The most abundant general type of lava is olivine basalt of several varieties. One widespread, important, and distinctive variety is mottled (ophitic), and is similar to what has been called ophite in earlier work on the Lake Superior district and olivine tholeiite in other areas. These lavas typically have ropy surfaces (fig. V-9B) and must have been very fluid. Rough columnar joints are common. Some flows lack ropy surfaces and instead have broad, smooth billows (figs. V-9C and V-9D). Other olivine basalts are relatively coarse grained, some with diabasic and some with other characteristic textures. Some of these rocks resemble the "feldspathic melaphyres" described from the Keweenaw Peninsula. In the Tofte-Lutsen area, a group of olivine basalts high in the sequence have abundant, small (1-3 mm) bytownite phenocrysts or crystal clots. These resemble some "porphyritic feldspathic melaphyres" and "glomeroporphyrites" of older reports on the Michigan Keweenawan. At the base of the sequence, both at Duluth and on Lucille Island east of Grand Portage, are distinctive basalts that contain abundant phenocrysts, 2-3 mm across, of augite and serpentinized olivine. These are particularly uncommon, for they have ferromagnesian rather than plagioclase

Table V-1. Characteristics of major lava types.

	Olivine tholeiite	Quartz tholeiite–trachyandesite		Andesite, trachyandesite	Intermediate quartz latite	Quartz latite–rhyolite
Thickness, in ft. (normal)	10-50	30-60		30-100		80-600
(max. thick., in ft.)	160	>100		240	200	1300
Jointing	rough columns in upper ½; sheeting at top	small joints, no good pattern		small joints, irregular and subhorizontal	irregular, blocky, platy near top	platy, small joints, large columns in some thick flows
Top zone	ropy, smooth; round amygdules	amygdular rubble-breccia; stretched amygdules		lobate, wrinkled crust; stretched amygdules	uneven, wrinkled crust or breccia	wrinkled, flow-banded crust, vesicular
Textures	obviously ophitic lower 2/3; amygdaloidal in upper 2/3	aphanitic to fine-grained, pilotaxitic to subophitic; aphyric		aphanitic to fine-grained; porphyritic (plag., augite, magnetite)	aphanitic, porphyritic (plag., ferroaugite, Fe-olivine, magnetite)	aphanitic to felsitic-poikilitic; most porphyritic (K-spar, qtz, ± plagioclase, ferroaugite, magnetite)
Mineralogy	olivine, augite, bytownite-labradorite, opaques, zeolites, etc.	augite, labradorite-andesine, opaques, some interstitial glass, K-spar, quartz		andesine, augite, magnetite, some pigeonite, interstitial K-spar	plagioclase, ferroaugite, Fe-olivine, magnetite, K-spar, quartz	as above (ferroaugite rarely preserved)
		Tholeiite	Trachyandesite			
SiO_2, wt.%	45.7 - 49.1	50.1-51.1	53.1-55.0	51.4-56.4	60.2-64.9	68.9-75.5
MgO, wt.%	5.3 - 8.1	4.7-5.9	2.9-4.3	2.3-4.5	0.9-1.9	Tr - 0.7
K_2O, wt.%	0.1 - 0.5	0.6-0.9	1.9-2.4	0.8-2.7	2.8-4.9	3.5-6.2
Other	pipe amygdules at base; thin flow units common	fine oxidation-banding common in lower ½; qtz, agate in amygdules		variable	red-brown; plagioclase commonly altered	red, pink, or gray

Table V-2. Comparative abundance of lava types.

<table>
<tr><th></th><th colspan="3">North Shore Volcanic Group, Keweenawan</th><th colspan="2">Eastern Iceland, Tertiary</th></tr>
<tr><th></th><th>NE limb (Tofte-Grand Portage)</th><th>Duluth-Two Harbors (Sandberg, 1938)</th><th>Two Harbors-Split Rock (Grogan, 1940, unpub. Ph.D. thesis, Univ. Minn.)</th><th>Reydar-fjordur plateau lavas (Walker, 1960)</th><th>Thingmuli volcano (Carmichael, 1964)</th></tr>
<tr><td>Olivine tholeiite ("ophite")</td><td>57</td><td>32.8</td><td>41</td><td>23</td><td>7*</td></tr>
<tr><td>Porphyritic basalt</td><td></td><td></td><td></td><td>12</td><td>1</td></tr>
<tr><td>(Quartz) tholeiite</td><td rowspan="2">7</td><td></td><td></td><td>48</td><td>50</td></tr>
<tr><td>Aphyric trachyandesite</td><td></td><td></td><td></td><td></td></tr>
<tr><td>"Melaphyre"</td><td></td><td>38.6</td><td>54</td><td></td><td></td></tr>
<tr><td>Andesite-porphyritic trachyandesite</td><td>7</td><td></td><td></td><td></td><td></td></tr>
<tr><td>Andesite</td><td></td><td></td><td></td><td>3</td><td>18</td></tr>
<tr><td>"Porphyrite"</td><td></td><td>10.4</td><td>2</td><td></td><td></td></tr>
<tr><td>Intermediate quartz latite</td><td>4</td><td></td><td></td><td></td><td></td></tr>
<tr><td>Rhyolite ("acid")</td><td>25</td><td>18.2</td><td>3</td><td>8</td><td>21</td></tr>
<tr><td>Other</td><td></td><td></td><td></td><td></td><td>3 pyroclastics</td></tr>
</table>

* Percentages in this column are "abundances"

phenocrysts; their flow tops are characterized by small lobes and coarse wrinkles (fig. V-10A). Another moderately abundant and distinctive basalt type is quartz tholeiite, which is aphanitic or very fine grained, nonporphyritic, and slightly more siliceous than the olivine basalts; it formed from more viscous lavas than did the olivine basalts. The quartz tholeiites characteristically have a rubbly or brecciated top, with the highly vesicular fragments set in a matrix of washed-in red sand (fig. V-9A) or, less commonly, hydrothermally deposited calcite and zeolites. Also, they commonly show narrow oxidation bands, 1-3 mm thick, along subhorizontal flowage-fracture planes. This quartz tholeiite grades into more potassium-rich varieties (trachybasalt, trachyandesite) that can be distinguished only by chemical analysis and microscope study; patches of interstitial K-feldspar are present in these rocks but are invisible in hand specimen. Both the quartz tholeiites and the fine-grained trachybasalts resemble the "fine-grained melaphyres" of older Lake Superior terminology.

The intermediate lavas are nearly all porphyritic, and have plagioclase, augite, magnetite and, in some specimens, iron-rich olivine phenocrysts. They have the compositions of andesites, trachyandesites, and intermediate quartz latites. Pigeonite is common in the groundmass of the more mafic types. Some of these flows resemble the "porphyrites" of older reports. Most are aphanitic, but one unusual flow, here called the Manitou trachybasalt, is at least 300 feet thick and granular, and can be traced for a distance of 5 miles (fig. V-10B); originally it extended an unknown distance farther in both directions. These flows are commonly brown or red, and are irregularly jointed or have platy subhorizontal joints. Lakeshore outcrops of the Manitou trachybasalt show distinctive liesegang-like weathering bands on joint surfaces (see fig. V-10C and Grout and Schwartz, 1939, fig. 12). Some of the trachyandesites and intermediate quartz latites resemble the "icelandites" of Carmichael (1964). Many of the flows near Hovland and on strike to the west-northwest are porphyritic trachybasalts and trachyandesites that have very conspicuous platy plagioclase phenocrysts (fig. V-10D).

The felsic lavas are anomalously abundant for a simple differentiation series from a basaltic parent magma. They are red, pink, or light gray, and have the composition of quartz latite and rhyolite. These flows tend to be much thicker than the other types; the thickest known flow, a few miles east of Grand Marais, is 1,300 feet thick, although it is possible that the Brule River rhyolite west of Hovland is locally as much as 3,500 feet thick. Their top surfaces are mostly strongly flow-banded, vesicular, and contorted, but not brecciated, and their bases are commonly flow-banded,

Figure V-10. Photographs of Keweenawan lavas. A, lobes and wrinkles in surface of augite-porphyritic basalt, SE. shore Lucille Island, east of Grand Portage, Cook County; B, Lake Superior shore exposure of Manitou trachybasalt, a thick flow in the Schroeder basalt sequence. View northeastward along strike from just northeast of Little Marais River, Lake County; C, banded weathering rings on Lake Superior shore outcrop of Manitou trachybasalt, northeast of Little Marais, Lake County; D, porphyritic trachyandesite flow showing large plagioclase phenocrysts just northwest of U.S. Highway 61, 0.9 miles SW. of Reservation River, Cook County.

folded, and locally brecciated. No evidence for ignimbrites has been seen, except for that reported by Johnson and Foster (1965) in the Cascade River. Spherulites are rare. Jointing ranges from large columns 5 or more feet across in the thickest flows (fig. V-11A) to small, subhorizontal platy or irregular joints; small tectonically produced parallel fracture sets a few mm apart commonly have broken the cooling-joint blocks into small pieces. Most of the felsites are porphyritic, with quartz and feldspar phenocrysts (sanidine-orthoclase with or without oligoclase-andesine), but some are only weakly porphyritic or aphyric. Poikilitic quartz surrounding stout alkali-feldspar laths ("snowflake texture") is a common microscopic texture in the thicker flows (Green, 1970a and b), and thin plates of quartz, probably pseudomorphic after tridymite, are also common. Even these siliceous lavas evidently flowed great distances, for one lava or flow group can be traced for at least 23 miles west from the Devil Track River, Grand Marais (see fig. V-6).

Alteration and Hydrothermal Deposition

The lavas have been strongly but irregularly affected by secondary solutions that deposited low-temperature minerals in vesicles, fractures, and other cavities, and altered some of the minerals in the lavas themselves. For example, all the olivine has been altered in the lavas, although fresh olivine is common in the intrusive diabases. A broad zonation of the alteration is apparent. At Duluth and at Grand Portage, in the lower parts of the lava sequence, much of the groundmass pyroxene has been converted to actinolite and some plagioclase has been saussuritized, indicating conditions approaching low greenschist facies. Here also, the amygdule minerals are characteristically quartz, prehnite, calcite, epidote, and chlorite, the same basic assemblage found in the Portage Lake Lava Series on the Keweenaw Peninsula (Stoiber and Davidson, 1959). J. A. Kilburg (1972, *op. cit.*) has found wairakite, as well, in some specimens near the Duluth Complex. In and northeast of Duluth,

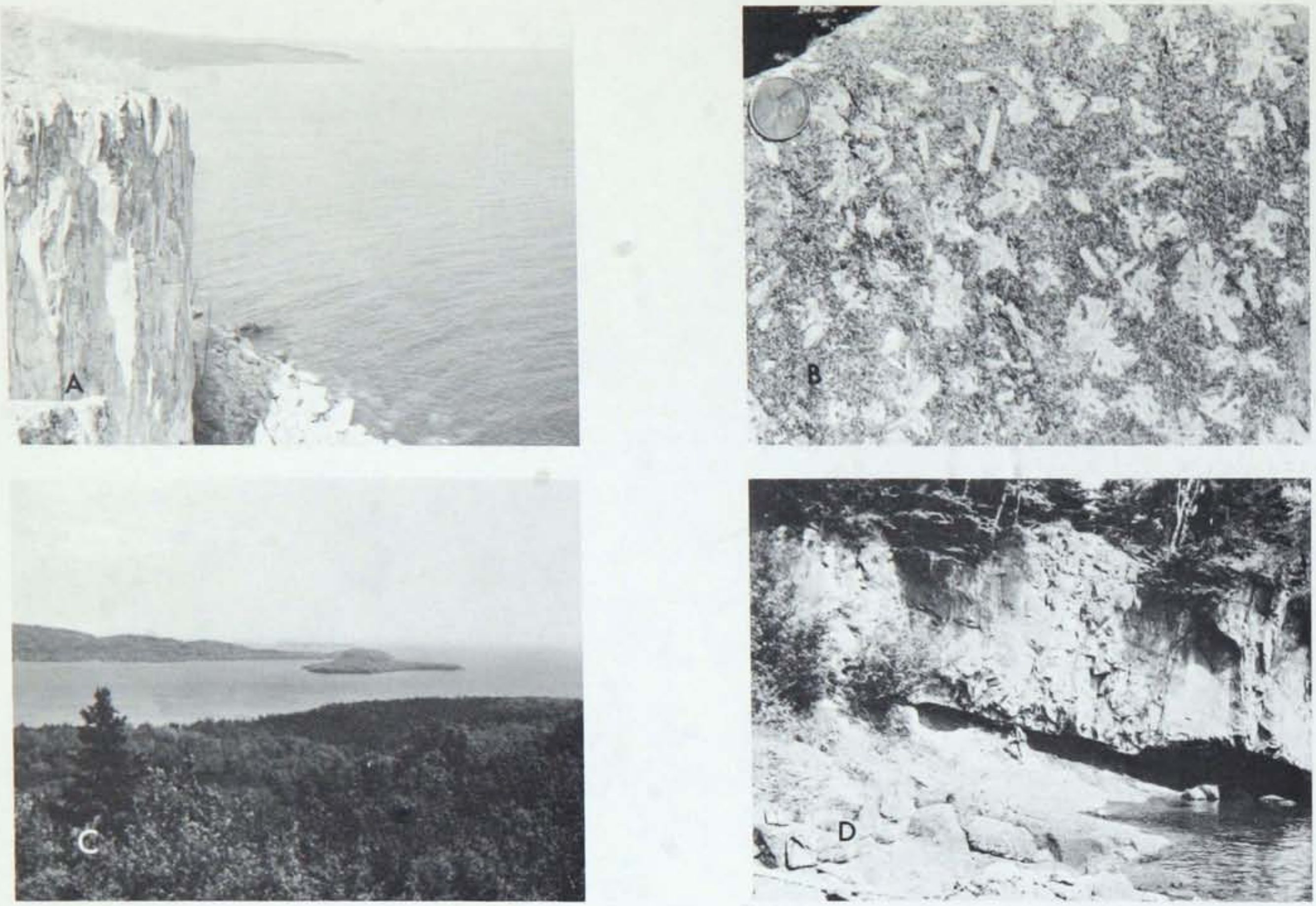

Figure V-11. Photographs of Keweenawan lavas. A, Palisade rhyolite flow, exposed in Palisade Head (foreground) and Shovel Point or Little Palisades, 2 miles away in distance. Cliffs are about 200 feet high. Northeast of Silver Bay, Lake County; B, plagioclase-porphyritic basalt flow, small islet just west of Lucille Island, east of Grand Portage, Cook County (Grand Portage lavas); C, view eastward along base of Keweenawan from hill 1.4 miles southwest of Grand Portage village. Basal lavas overlying Puckwunge Fm. are exposed in slope behind spruce tree in foreground, at base of hill on Grand Portage Island in middle and on left-hand end of Lucille Island (seen over low neck of Hat Point); Isle Royale, Michigan, composed of younger Keweenawan lavas on the northwest side of the Lake Superior syncline, is faintly visible to right; D, flow contact in ophitic olivine basalts of Schroeder basalt series. Massive base of a thick flow overlies softer, amygdaloidal top of flow beneath. See man at contact for scale. Lake Superior shore near Schroeder, Cook County.

K-feldspar also occurs and laumontite is abundant. Pumpellyite has not yet been identified in the North Shore Volcanic Group. Higher in the sequence, various zeolites, together with calcite, are dominant except in the quartz tholeiites and similar lavas where agate, crystalline quartz, and chlorite are more common. The most abundant zeolites are laumontite, stilbite, heulandite, thomsonite, scolecite, and mordenite; but analcite, natrolite, mesolite, and apophyllite also have been found. Saponite is common in olivine basalts at these higher stratigraphic levels (Whelan and Lepp, 1961). Andradite garnet has been discovered in amygdules and veins from lava types ranging in composition from basalts to rhyolites and occurring at different stratigraphic levels, and traces of native copper have been found at several localities. Thus, the secondary zonation in the North Shore Volcanic Group spans both the deeper level, hotter type found on the Keweenaw Peninsula and the higher level, cooler type characteristic of the lower parts of the Tertiary plateau lavas of eastern Iceland, as described by Walker (1960). Apparently, Walker's upper zeolite-free zone is not represented in Minnesota. According to estimates from Walker's data, the presently exposed top of the sequence on the Lake Superior shore was approximately 5,000 feet below the surface during the hydrothermal alteration. Although detailed work has not yet been done, clear crosscutting relations of zeolite zones to stratigraphy *within* the lavas have not been recognized; the apparent Upper Precambrian, post-volcanic unconformity that follows the north shore must have postdated the alteration, for it cuts across the zeolite zones.

It should be stressed, however, that almost none of the flows have been converted entirely to secondary minerals. In fact, although fresh olivine has not been seen, the plagioclase and augite are typically unaltered or only locally altered in most mafic and intermediate rocks. Typically, the opaque minerals are partly oxidized, and most pigeonite has oxidized rims. In many intermediate and felsic lavas, plagioclase phenocrysts are albitized and/or zeolitized; some of this alteration could be deuteric. In the more felsic lavas, ferromagnesian silicates (ferrous olivines, ferroaugites) are nearly everywhere destroyed by oxidation. Fresh, undevitrified volcanic glass is still present locally, notably in an andesite exposed about 2 miles west of the mouth of the Brule River (MI-35, table V-3B).

The alteration (low-grade metamorphism to greenschist and zeolite facies) must have occurred under a cover of several thousand feet of overlying deposits, subsequently eroded. As pointed out by White (1957, p. 11), extrapolation of the present geothermal gradient in the deep mines of the Keweenaw Peninsula (18°/km; Birch, 1954, p. 19) would give a temperature of 250° C, probably characteristic of the low greenschist facies, at a depth of about 40,000 feet, which is not unreasonable compared to current estimates of the total thickness of the Upper Precambrian lavas (21,000-29,000 feet) and intrusions (17,000-20,000 feet). With the higher geothermal gradients that probably existed in the Keweenawan, of course, such a temperature would have been reached at a considerably shallower depth; however, the entire thickness of cover must have been eroded before deposition of the Upper Keweenawan Fond du Lac sandstone unconformably on the greenschist-facies Ely's Peak basalts at Duluth.

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From personal observation, I have seen that flow thicknesses (table V-1) are generally similar to those of Tertiary Icelandic plateau lavas. Many flow contacts are well exposed along the shoreline and in streambeds, but determination of true flow thicknesses is dependent on distinguishing flow-unit contacts from flow contacts. Flow units, defined as overlapping tongues or sheets of a single eruption, might be distinguished from flows by the absence of sediments, weathered rock, or eroded material along the contacts between similar rock types. If a lithic change or evidence of surficial processes is present, a flow contact is implied, but the converse is not true; even with a time span of several years, a flow surface may show no evidence of weathering, erosion, or stream sedimentation. If a new eruption of the same lava type follows, the next contact will be indistinguishable from an internal, flow-unit contact that may represent a time span of only a few hours or days, as is commonly the case in Hawaii. Therefore, the numbers of flows given below are maximal, as units are counted as flows, and flow thicknesses are minimal. Another problem in determining flow thickness arises when a thick flow is intersected at a small angle by the lakeshore; the top and base may be separated along the outcrop by distances of a few miles, so that thickness of the flow cannot be measured directly. In such cases, geometric construction has been used to estimate thicknesses.

From the data of Sandberg (1938), the 179 flows measured and counted between Duluth and Two Harbors average 74 feet in thickness, whereas the 70 dominantly basaltic flows measured by Grogan (1940, *op. cit.*) northeast to Split Rock River, average only 37 feet in thickness. Near Tofte, most flows, all ophitic olivine tholeiites, are between 1 and 40 feet thick. According to Grout and others (1959), the succession from Tofte to Grand Portage contains 94 flows that are an average of 187 feet thick; my mapping, however, shows that the number of flows probably was underestimated considerably, partly because of the many gaps in outcrop east of Grand Marais. Between Lutsen, where the section starts to descend stratigraphically, and Grand Marais, about 25 flows are present, averaging 84 feet in thickness. These are dominantly basaltic and are uncommonly thick. East of Grand Marais, several thick intermediate and felsic flows are present, and four of these are more than 300 feet thick. In the poorly exposed, magnetically reversed succession at Grand Portage, the 38 basaltic flows that can be seen range in thickness from 6 to about 40 feet; the well exposed pyroxene-porphyritic basalts in Lucille and Magnet Islands are, on the average, only 16 feet thick.

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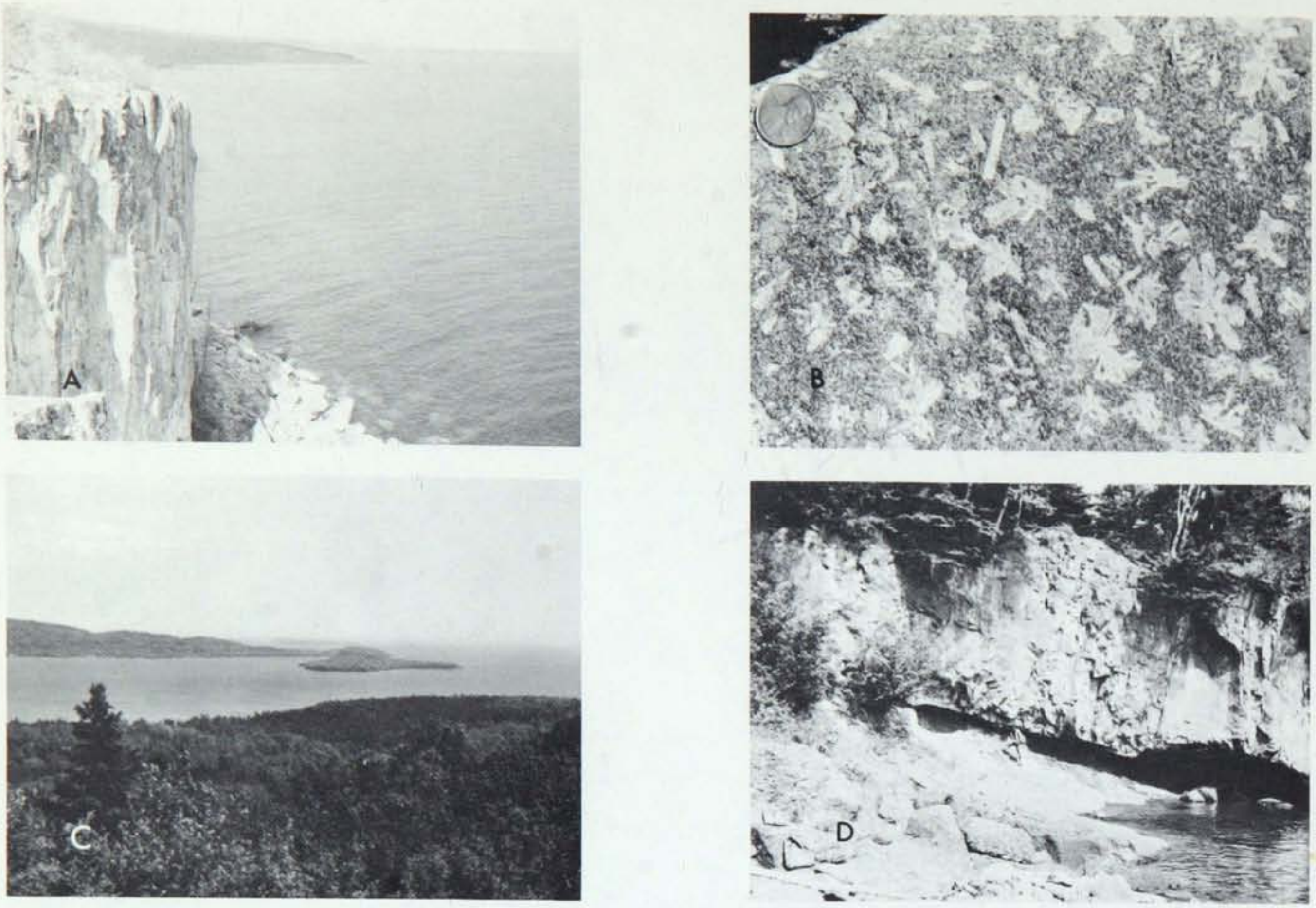

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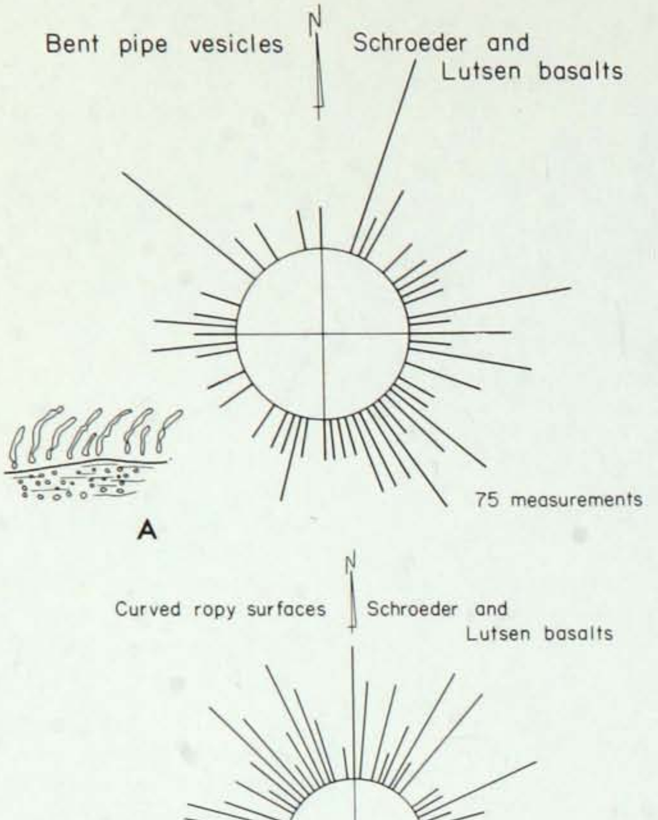

Figure V-12. Directions of lava movement for the Schroeder and Lutsen basalts (highest in North Shore Volcanic Group) as inferred from (A) bent pipe vesicles and (B) curved ropy surfaces. No clear preferred orientation is evident. Length of bar from circle is proportional to number of observations.

give a weak suggestion of flow toward the southeast. Crossbedding in the disconformably underlying Puckwunge Formation (Mattis, this chapter) at the latter locality indicates a southerly paleoslope.

The lack of a uniform or preferred flow direction, as expressed by the primary features in the North Shore Volcanic Group, can be interpreted in several ways. First, it is possible that structures such as bent pipe amygdules at the bases of flows, stretched amygdules, and looping ropy surfaces are not reliable indicators of the main flow direction; unless large numbers of measurements are made, the data may reflect many local deflections in direction of flow resulting from local irregularities on the underlying lava surfaces. However, in several areas, both in the Keweenawan province (Butler and Burbank, 1929; Sandberg, 1938) and elsewhere (Schmincke, 1967), a uniform orientation is given by similar directional indicators, which suggests that they are meaningful. Assuming that the primary features are valid directional indicators, the observed measurements on the north shore may mean that the North Shore Volcanic Group either was fed primarily from fissures or vents distributed randomly with respect to the shoreline, or that the vents were located in the vicinity of the shoreline, with the lavas spreading out in all directions. The observations on the north shore do not support a dominant source in fissures along the present axis of the Lake Superior basin, although statistically they may not disprove it. The zone of mantle rifting may well have been located elsewhere in pre-Portage Lake time.

Another indication of the possible locations of vents is the distribution of dikes that could have served as feeders for overlying lava flows. Several dike swarms occur in the area, and generally have trends that are roughly parallel to the axis of the Keweenawan basin. The known dikes cut either the lavas in the lower parts of the sequence—as at Duluth and Grand Portage—or pre-Keweenawan rocks beyond the limits of presently exposed lavas. In the Cloquet area (Wright and others, 1970), basaltic dikes occur as much as 9½ miles west of the lowermost Keweenawan lavas, and near Ely (Green and others, 1966) similar dikes occur as much as 4½ miles from the Duluth Complex. These indicate that many of the lavas had source fissures well beyond the present limits of Keweenawan outcrops.

Chemical Characteristics

The North Shore Volcanic Group ranges in composition from olivine melabasalt to rhyolite. A wide variety of basaltic types, as well as a nearly complete gradational series through andesites, trachyandesites, intermediate quartz latites, and felsic quartz latites, are present. The 46 modern chemical analyses are given in Table V-3.

Figure V-13 is a Harker variation diagram for major oxides. Sampling to date shows two small gaps in SiO_2 con-

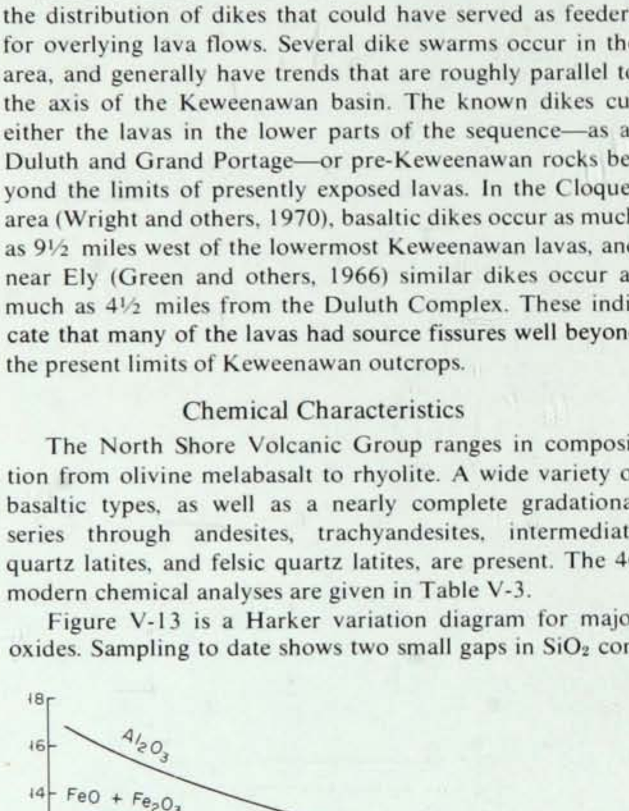

Figure V-13. Harker variation diagram, North Shore Volcanic Group.

Table V-3A. Chemical analyses of basaltic lavas, North Shore Volcanic Group.

	T-45	FFG-IV-2	T-56	KC-9	LW-10	LW-1	GH-25	DY-6b	PP-16	T-22	GP-44	T-59	PP-2	TH-2	F-96	GP-4a	F-108	DT-104
SiO_2	45.71	46.80	46.87	46.98	47.06	47.10	47.19	47.69	48.55	49.11	49.13	49.82	49.95	50.19	50.54	50.90	51.08	51.12
TiO_2	1.28		0.81	1.81	2.10	2.87	0.95	1.28	1.86	1.64	3.19	2.18	1.19	1.51	1.49	1.05	2.20	1.65
Al_2O_3	17.47	15.21	19.20	16.06	14.35	14.23	17.04	18.36	9.02	13.35	15.14	14.08	15.29	15.15	16.59	14.50	14.17	14.77
Fe_2O_3	7.80	13.13	5.42	5.28	9.13	9.50	2.63	6.47	2.63	9.16	5.55	6.82	1.12	5.51	5.65	1.28	10.25	5.03
FeO	3.41		2.97	6.63	3.71	4.93	7.69	4.74	10.20	3.47	8.30	7.15	9.05	5.82	5.47	9.63	3.69	7.76
MnO	0.15		0.12	0.17	0.17	0.20	0.14	0.17	0.24	0.14	0.18	0.26	0.22	0.15	0.16	0.18	0.19	0.17
MgO	6.80	8.13	5.86	6.32	5.70	3.86	8.11	5.27	11.69	6.23	4.24	4.57	8.38	5.91	4.51	6.49	4.73	5.63
CaO	10.53	11.11	12.41	9.59	8.58	6.53	10.76	11.21	11.29	10.09	6.17	7.88	6.29	9.13	10.06	8.41	8.08	7.79
Na_2O	2.61	1.95	2.27	2.64	3.65	3.70	2.23	2.35	1.62	2.91	4.06	3.56	3.28	2.71	3.23	2.80	2.98	3.27
K_2O	0.31	0.01	0.12	0.50	1.32	2.66	0.35	0.46	0.71	0.43	1.13	1.14	1.27	0.62	0.76	0.85	0.87	0.86
H_2O+	1.77	2.79	1.55	2.79	3.41	3.20	2.55	1.62	1.75	2.47	2.21	1.90	3.40	1.92	1.38	3.47	1.20	1.56
H_2O-	1.78		2.08	1.60	1.02	1.04				1.36				1.76	0.74		0.77	
P_2O_5	0.14		0.08	0.25	0.32	0.77	0.13	0.15	0.25	0.23	0.40	0.32	0.19	0.17	0.23	0.17	0.38	0.27
CO_2	0.12		0.10	0.16	0.09	0.01	0.07	0.00	0.02	0.07	0.13	0.01	0.00	0.13	0.02	0.08	0.18	0.00
Total	99.88	99.13	99.86	100.78	100.61	100.60	99.84	99.77	99.83	100.66	99.83	99.69	99.63	100.68	100.83	99.81	100.77	99.88

T-45 Ophitic black basalt, Cross R. at 1,090 ft.; NW¼NW¼ sec. 36, T. 59 N., R. 5 W., Cook Co.; anal. M. Kumanomido
FFG-IV-2 Thomsonite-bearing basalt, Good Harbor Bay; sec. 34, T. 61 N., R. 1 W., Cook Co. (Grout, 1910a, table IV, no. 2); anal. F. F. Grout
T-56 Ophitic olivine basalt with small plagioclase phenocrysts, L. Superior shore E of Onion R.; range line T. 59 N., R. 3/4 W., Cook Co.; anal. M. Kumanomido
KC-9 Ophitic basalt, Durfee Ck. at 750 ft.; SE¼NW¼ sec. 8, T. 61 N., R. 2 E., Cook Co.; anal. Tadashi Asari
LW-10 Olivine trachybasalt with small plagioclase phenocrysts, L. Superior shore NE of Lester R.; SE¼SW¼ sec. 34, T. 51 N., R. 13 W., St. Louis Co.; anal. Tadashi Asari
LW-1 Granular olivine trachybasalt with small plagioclase phenocrysts, L. Superior shore NE of Lakewood pumping station; SW¼SE¼ sec. 26, T. 51 N., R. 13 W., St. Louis Co.; anal. Tadashi Asari
GH-25 Fine-gr., black, ophitic basalt, Hwy. 61 cut at Good Harbor Bay; NW¼NW¼ sec. 34, T. 61 N., R. 1 W., Cook Co.; anal. K. Ohta
DY-6b Coarse, brown, ophitic olivine basalt, L. Superior shore W of Cascade R.; SE¼SE¼ sec. 2, T. 60 N., R. 2 W., Cook Co.; anal. K. Ohta
PP-16 Dark-gray basalt with augite and altered olivine phenocrysts, E end Lucille I.; SW¼ sec. 4, T. 63 N., R. 7 E., Cook Co.; anal. K. Ohta
T-22 Coarse, gray-brown olivine basalt (amygdular clay hand-picked out), 1,070 ft. el. W of Schroeder; SW¼NW¼ sec. 2, T. 58 N., R. 5 W., Cook Co.; anal. M. Kumanomido
GP-44 Fine-gr., subophitic basalt, Hollow Rock Ck. at 660 ft. el.; NW¼NE¼ sec. 35, T. 63 N., R. 5 E., Cook Co.; anal. K. Ohta
T-59 Fine-gr., brown basalt with small plagioclase and augite phenocrysts, L. Superior shore SW of Onion R.; NW¼NW¼ sec. 13, T. 59 N., R. 4 W., Cook Co.; anal. K. Ohta
PP-2 Red-brown, fine-granular basalt, W end Lucille I.; NW¼ sec. 8(?), T. 62 N., R. 7 E., Cook Co.; anal. K. Ohta
TH-2 Fine-gr., dk. gray basalt, L. Superior shore at Town Park, Burlington Bay, Two Harbors; SE¼NW¼ sec. 6, T. 52 N., R. 10 W., Lake Co.; anal. M. Kumanomido
F-96 Granular, reddish basalt, knob just N. of Hwy. 61 W. of Caribou R.; NE¼SW¼ sec. 36, T. 58 N., R. 6 W., Lake Co.; anal. Shiro Imai
GP-4a Diabasic gray basalt, 1st road cut Hwy. 61 SW of Grand Portage; NE¼NE¼ sec. 17, T. 63 N., R. 6 E., Cook Co.; anal. K. Ohta
F-108 Fine-gr. black basalt, L. Superior shore NE of Kennedy Ck.; SE¼SW¼ sec. 36, T. 57 N., R. 7 W., Lake Co.; anal. Shiro Imai
DT-4 Fine-gr., dk. gray basalt, N slope of hill 0.8 mi. W of Fall R.; S½NE¼ sec. 14, T. 61 N., R. 1 W., Cook Co.; anal. K. Ohta

Table V-3B. Chemical analyses of trachybasalts, andesites, and trachyandesites, North Shore Volcanic Group.

	NL-5	GM-9	S&G-5-9	T-36	F-54	H-5b	F-7a	GP-27	MI-4a	GH-2b	MI-35	B-T-111
SiO_2	51.37	52.63	52.70	52.82	53.08	53.22	53.60	54.88	55.01	55.42	56.38	57.55
TiO_2	2.10	2.04	1.76	2.15	1.99	2.05	1.70	1.69	1.65	2.15	1.69	2.13
Al_2O_3	16.65	13.12	14.47	13.67	13.25	17.55	15.33	15.36	14.13	12.29	13.64	10.84
Fe_2O_3	3.88	6.23	7.44	7.66	8.73	4.63	6.34	4.91	6.90	9.00	2.28	6.56
FeO	8.08	6.10	5.55	6.61	3.74	6.18	6.65	6.50	2.80	3.32	9.45	6.04
MnO	0.14	0.18	0.24	0.15	0.15	0.13	0.19	0.19	0.11	0.23	0.20	0.21
MgO	2.90	4.51	3.70	3.66	4.34	2.33	3.11	3.27	2.88	2.89	2.58	4.60
CaO	4.84	5.58	8.01	6.94	5.78	4.79	5.91	3.03	5.62	3.25	5.76	7.21
Na_2O	4.38	4.18	3.19	2.94	3.37	5.35	3.63	4.84	3.79	4.47	3.71	3.02
K_2O	2.51	1.54	1.14	1.19	1.91	1.76	1.26	2.40	2.07	2.72	0.79	0.14
H_2O+	2.20	2.70	0.68	1.67	2.11	1.32	1.66	1.97	3.58	2.86	2.40	0.94
H_2O-		1.35	0.48		1.98				1.84	1.71		
P_2O_5	0.57	0.37	0.25	0.42	0.38	0.51	0.28	0.65	0.39	0.42	0.76	0.35
CO_2	0.00	0.07		0.01	0.07	0.11	0.01	0.06	0.10	0.15	0.01	0.05
S			0.02									0.01
BaO			0.05									
Total	99.62	100.60	99.68	99.89	100.88	99.93	99.67	99.75	100.87	100.88	99.65	99.65

NL-5 Fine-gr., brown trachybasalt with platy plagioclase phenocrysts; USFS road 309 W of Stony Ck.; sec. line 4/5, T. 63 N., R. 2 E., Cook Co.; anal. K. Ohta

GM-9 Fine-gr. dark brown trachybasalt, L. Superior shore NE of Grand Marais; SW¼SW¼ sec. 14, T. 61 N., R. 1 E., Cook Co.; anal. Shiro Imai

S&G-5-9 Granular, brown trachybasalt of columnar flow, Grand Marais harbor; sec. 14, T. 61 N., R. 1 E., Cook Co.; anal. E. D. Burr (Sandell and Goldich, 1943, table 5, no. 9)

T-36 Granular, brown, magnetite-rich trachybasalt, W side of hill NE of Onion R.; SE¼NW¼ sec. 1, T. 59 N., R. 4 W., Cook Co.; anal. K. Ohta

F-54 Fine-gr., brown trachybasalt, Hwy. 61 cut N of Kennedy Landing; E½NE¼ sec. 36, T. 57 N., R. 7 W., Lake Co.; anal. Shiro Imai

H-5b Fine-gr. trachybasalt with large plagioclase phenocrysts, NW of Hwy. 61 W of Reservation R.; NE¼NW¼ sec. 12, T. 62 N., R. 4 E., Cook Co.; anal. K. Ohta

F-7a Fine-gr., red-brown trachyandesite with small plagioclase, augite, and olivine phenocrysts, L. Superior shore near Little Marais R.; NW¼NE¼ sec. 21, T. 57 N., R. 6 W., Lake Co.; anal. K. Ohta

GP-27 Fine-gr., black basalt, Hwy. 61 cut NE of Deronda Bay; SE¼NW¼ sec. 25, T. 63 N., R. 5 E., Cook Co.; anal. K. Ohta

MI-4a Fine-gr., brown-gray trachyandesite, L. Superior shore E of Cook Co. 14; SW¼SE¼ sec. 32, T. 62 N., R. 3 E., Cook Co.; anal. M. Kumanomido

GH-2b Fine-gr., red-brown trachyandesite with small plagioclase, augite, and magnetite phenocrysts, L. Superior shore ENE of Good Harbor Bay; NW¼SE¼ sec. 26, T. 61 N., R. 1 W., Cook Co.; anal. S. Imai

MI-35 Black andesite with abundant interstitial glass, logging road NW of Paradise Beach; NW¼NE¼ sec. 32, T. 62 N., R. 3 E., Cook Co.; anal. K. Ohta

B-T-111 Vesicular, magnetite-rich basalt, sec. 26, T. 59 N., R. 11 W. (Bonnichsen, this chapter, table V-30, no. 5)

Table V-3C. Chemical analyses of intermediate quartz latites, North Shore Volcanic Group.

	MC-3b	MI-2	F-21	T-40	D-28	KC-1	M-4600
SiO_2	60.21	62.60	63.22	63.37	63.45	64.32	64.95
TiO_2	0.96	1.09	1.20	1.10	0.69	1.11	0.82
Al_2O_3	14.64	12.01	10.61	12.18	12.80	12.64	12.58
Fe_2O_3	2.71	8.18	8.73	6.46	3.75	5.89	4.70
FeO	5.82	2.02	1.43	2.64	7.40	2.72	4.83
MnO	0.17	0.12	0.09	0.14	0.18	0.12	0.16
MgO	0.85	1.40	1.81	1.88	1.48	0.90	0.93
CaO	3.56	2.22	1.49	1.63	1.79	2.40	2.07
Na_2O	3.66	4.04	2.69	4.31	3.22	3.78	3.46
K_2O	3.92	4.15	4.95	2.76	3.94	4.12	4.21
H_2O+	1.24	1.72	2.80	2.45	0.81	1.74	0.54
H_2O-		0.93	1.44	0.75		0.75	0.32
P_2O_5	0.50	0.28	0.35	0.38	0.15	0.31	0.15
CO_2	1.36	0.04	0.03	0.17	0.00	0.01	0.03
S							0.03
BaO							0.12
Total	99.60	100.80	100.84	100.22	99.66	100.81	99.90

MC-3b Fine-gr., brown "andesite," L. Superior shore NE of Deronda Bay; cen. sec. 25, T. 63 N., R. 5 E., Cook Co.; anal. K. Ohta

MI-2 Aphanitic, red-brown quartz latite with small plagioclase, ferroaugite, olivine and magnetite phenocrysts; Hwy. 61 cut at W. edge of sec. 6, T. 61 N., R. 3 E., Cook Co.; anal. Tadashi Asari

F-21 Fine-gr., red-brown quartz latite with small phenocrysts of plagioclase and ferroaugite, low hill W. of road; NE¼ sec. 28, T. 57 N., R. 7 W., Lake Co.; anal. S. Imai

T-40 Fine-gr., red trachyandesite; low railroad cut at Cook Co. 1, NE¼SW¼ sec. 5, T. 58 N., R. 5 W., Cook Co.; anal. M. Kumanomido

D-28 Aphanitic, red, porphyritic quartz latite, W. end 8th St., Duluth; NW¼NW¼ sec. 27, T. 50 N., R. 14 W., St. Louis Co.; anal. K. Ohta (same flow, locality as M-4600)

KC-1 Fine-gr., red quartz latite with small plagioclase, subcalcic augite, and magnetite phenocrysts; low cuts of Hwy. 61 ½ mi. E. of Kadunce Ck., NW¼ sec. 1, T. 61 N., R. 2 E., Cook Co.; anal. Tadashi Asari

M-4600 Aphanitic, red-brown porphyritic quartz latite; 8th St. at 3rd Ave. W., Duluth, NW¼NW¼ sec. 27, T. 50 N., R. 14 W., St. Louis Co.; anal. Goldich and Smith (same locality, flow as D-28) (Taylor, 1964, table 17, p. 54)

Table V-3D. Chemical analyses of felsic lavas, North Shore Volcanic Group.

	KC-17	S&S-1-22	MC-7	GM-14	GM-10	MC-1	F-201	G&O-5-2	S&S-1-7
SiO_2	68.88	71.12	71.61	72.23	73.42	74.17	74.41	75.40	75.48
TiO_2	0.48	0.45	0.25	0.45	0.45	0.26	0.24	0.31	0.21
Al_2O_3	12.77	12.58	11.87	11.38	11.63	11.98	10.95	11.53	12.30
Fe_2O_3	4.32	5.20	2.48	4.08	3.18	3.60	1.64	4.06	2.54
FeO	2.66	0.15	3.59	0.24	0.57	0.18	2.91	0.22	0.36
MnO	0.09	0.06	0.05	0.04	0.04	0.01	0.05	0.03	0.02
MgO	0.67	0.08	0.21	0.44	0.27	0.26	0.30	0.16	Tr
CaO	0.94	0.58	0.70	1.07	0.99	0.39	0.50	0.16	0.14
Na_2O	4.53	2.85	3.81	2.62	3.50	4.87	1.93	3.33	3.43
K_2O	3.45	6.19	4.10	5.50	4.65	3.88	5.64	4.44	5.17
H_2O+	0.74	0.22	0.52	1.54	1.54	0.88	0.79	0.07	0.24
H_2O-		0.05		0.50	0.48	0.24		0.06	0.04
P_2O_5	0.11	0.03	0.05	0.03	0.08	0.03	0.04	0.04	0.02
CO_2	0.10	0.18	0.44	0.45	0.03	0.13	0.26		
Total	99.74	99.74	99.68	100.57	100.83	100.88	99.66	99.81	99.95

KC-17 Granular, pink felsite with quartz and plagioclase phenocrysts; Hwy. 61 ½ mi. NE of Kimball Ck., SW¼SE¼ sec. 3, T. 61 N., R. 2 E., Cook Co.; anal. K. Ohta

S&S-1-22 Aphanitic red rhyolite; Lake Superior shore, approx. 1.2 mi. SW of Lakewood Rd. SW¼SW¼ sec. 34, T. 50 N., R. 13 W., St. Louis Co.; anal. Perlich (Schwartz and Sandberg, 1940, table 1, no. 22)

MC-7 Aphanitic red rhyolite with quartz and orthoclase phenocrysts; Hwy. 61 cut ½ mi. W. of Deronda Bay, SW¼SE¼ sec. 26, T. 63 N., R. 5 E., Cook Co.; anal. K. Ohta

GM-14 Fine-gr., gray felsite; L. Superior shore approx. ½ mi. E. of Devil Track R., NW¼NW¼ sec. 18, T. 61 N., R. 1 E., Cook Co.; anal. Tadashi Asari

GM-10 Aphanitic, red, spherulitic rhyolite with oligoclase, ferroaugite, olivine, and magnetite phenocrysts; L. Superior shore, Croftville, S. center line sec. 14, T. 61 N., R. 1 E., Cook Co.; anal. S. Imai

MC-1 Fine-gr. whitish felsite with plagioclase, alkali feldspar, quartz, and magnetite phenocrysts; L. Superior shore 0.6 mi. W. of Reservation R., NE¼NE¼ sec. 12, T. 62 N., R. 4 E., Cook Co.; anal. Tadashi Asari

F-201 Aphanitic, pink felsite with plagioclase, orthoclase, quartz, and altered fayalite (?) phenocrysts; NE cliff of Palisade Head, NE¼NW¼ sec. 22, T. 56 N., R. 7 W., Lake Co.; anal. K. Ohta

G&O-5-2 Aphanitic, lt. gray felsite with alkali feldspar and quartz phenocrysts; Hwy. 61 cut ½ mi. W. of Grand Marais harbor, NE¼SW¼ sec. 20, T. 61 N., R. 1 E., Cook Co.; anal. S. Fruehling (Grout and others, 1959, table 5, no. 2)

S&S-1-7 Aphanitic, red, banded rhyolite above Endion sill, Tischer Ck. near 2nd St. E., Duluth; NE¼NW¼ sec. 13, T. 50 N., R. 14 W., St. Louis Co.; anal. S. S. Goldich (Schwartz and Sandberg, 1940, table 1, no. 7) tent, one from about 57 to 60 percent SiO_2 and another from 65 to 68 percent SiO_2. The diagram also shows a higher-than-normal K_2O content in the intermediate SiO_2 range. This suite gives an alkali-lime index of about 54; thus, it would be classed as alkali-calcic. These analyses differ from analyses from the Keweenaw Peninsula (Broderick, 1935) in two ways: (1) intermediate rocks are present; and (2) some of the felsic lavas have a lower potassium content.

The composition of the lavas is intermediate between typical calc-alkaline continental rocks and the strong iron-enrichment trend of the Skaergaard liquids, as can be seen in Figure V-14. The Keweenawan trend is very similar to that of Thingmuli volcano, Iceland (Carmichael, 1964), although there are certain minor differences which will be pointed out below.

The basaltic rocks of this sequence are rather varied, as has been mentioned above, and they straddle the chemical dividing lines proposed by several petrologists to distinguish major petrogenetic classes. For example, in comparing total alkalis with silica percentage, one-third of the basaltic rocks lie in MacDonald and Katsura's (1964) alkali-basalt field and two-thirds in their tholeiite field, although the tholeiites are all close to the boundary. Using Kuno's (1960) plot, twelve of the basaltic and andesitic rocks fall in the alkali-basalt field, nine fall in the tholeiite field, and the remaining four fall in the high-alumina basalt field. With one exception (LW-10, a trachybasalt) they are hypersthene-normative, and therefore tholeiitic according to Yoder and Tilley (1962). The olivine tholeiites are similar to those of many other areas (see table V-4).

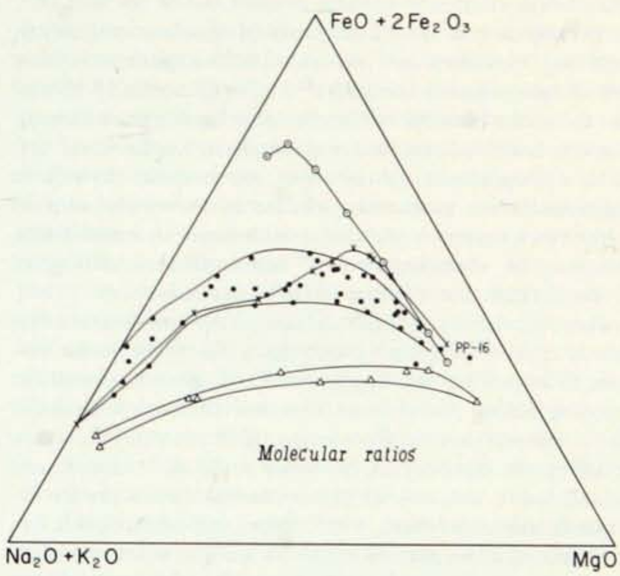

Figure V-14. Iron-enrichment (alkalis/MgO/total iron) diagram for ●, North Shore Volcanic Group, compared to △, Cascade Province calc-alkali trend (Turner and Verhoogen, 1960), ⊙, Skaergaard liquid trend (Wager, 1960), and ×, Thingmuli volcano trend (Carmichael, 1964). Sample PP-16 contains abundant augite and olivine phenocrysts and probably does not represent a liquid composition.

Figure V-15 shows that the North Shore suite contains a higher content of alkalis than typical tholeiitic suites, such as the Skaergaard liquids and Thingmuli volcano in the Tertiary of eastern Iceland. The more mafic rocks have less K_2O than typical alkali-olivine basalt series, and many of the least siliceous basalts, in fact, are rather typical tholeiites according to these two components. Except for four North Shore lavas, the Thingmuli series is consistently lower in K_2O in the mafic and intermediate range; the North Shore lavas, in contrast, include many trachybasalts, trachyandesites, and intermediate quartz latites, all containing significant to abundant K-feldspar. In the intermediate to felsic range, the North Shore lavas plot on this diagram with the Hawaiian alkalic suite, but are actually very different because of their higher silica content.

Origin of the Magmas

Table V-4 shows the strong similarity between the very abundant ophitic olivine tholeiite of the North Shore Volcanic Group (col. 1) and basalts assumed to have come more or less directly from the upper mantle in other areas. The North Shore olivine tholeiites bear a striking similarity to Kuno's (1960) high-alumina basalts and to Engel and others' (1965) oceanic tholeiites.

Green and others (1967) have shown experimentally that under an intermediate-pressure regime in the mantle (that is, depths of about 15-35 km), a primary partial melt such as the relatively olivine-rich average Hawaiian tholeiite (table V-4, col. 6) would separate clinopyroxene and some

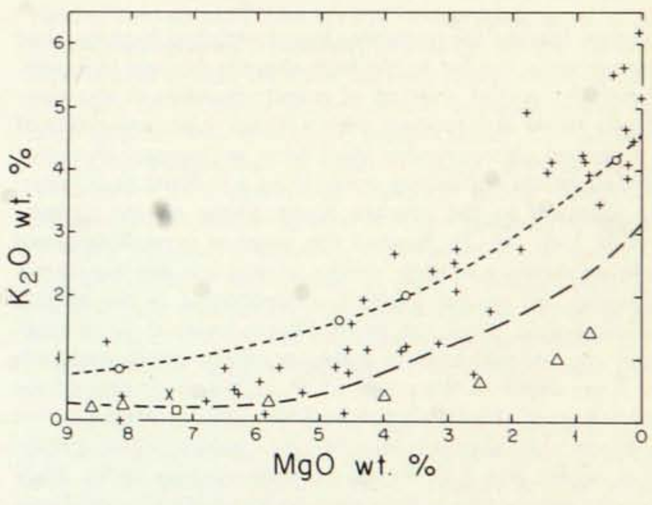

Figure V-15. K_2O/MgO diagram for North Shore Volcanic Group and other suites. Note general low K_2O content of the most magnesian rocks, but rapid rise in K_2O with falling MgO compared to other tholeiitic suites (Skaergaard, Thingmuli). △, Skaergaard liquids (Wager, 1960); ○, Hawaiian alkali trend (MacDonald and Katsura, 1964); +, North Shore Volcanic Group; - -, Thingmuli volcano (Carmichael, 1964); ⊡, average of ten oceanic tholeiites (Engel and others, 1965); and ×, average high-alumina basalt (Kuno, 1960).

Table V-4. Compositions of various tholeiitic lavas.

	1	2	3	4	5	6	7	8
SiO_2	48.53	48.94	49.94	50.19	50.3	49.36	51.69	50.20
TiO_2	1.38	0.85	1.51	0.75	1.8	2.50	2.20	3.19
Al_2O_3	17.83	20.05	16.69	17.58	17.9	13.94	15.19	12.90
Fe_2O_3	2.44	1.82	2.01	2.84	1.7	3.03	3.02	3.31
FeO	8.78	6.57	6.90	7.19	7.9	8.53	9.93	11.88
MnO	0.16	0.13	0.18	0.25	0.2	0.16	0.17	0.27
MgO	6.85	6.12	7.28	7.39	6.8	8.44	5.26	4.78
CaO	10.89	12.96	11.86	10.50	10.8	10.30	7.99	9.27
Na_2O	2.55	2.37	2.76	2.75	3.0	2.13	3.34	3.07
K_2O	0.41	0.13	0.16	0.40	0.2	0.38	0.89	0.57
P_2O_5	0.17	0.08	0.16	0.14		0.26	0.31	0.58
Total	99.99	100.02	99.45	99.98			99.99	100.01

1. Average of 4 olivine tholeiites, North Shore Volcanic Group (reduced to $Fe^{+++}/Fe^{++} = 1/4$; neglecting H_2O, CO_2)
2. T-56: plagioclase-porphyritic olivine tholeiite (reduced to $Fe^{+++}/Fe^{++} = 1/4$; neglecting H_2O, CO_2)
3. Average of 10 oceanic tholeiites (Engel and others, 1965)
4. Average high-alumina basalt of Japan (Kuno, 1960)
5. "Oceanic tholeiite" derivative, synthetic (Green and others, 1967, table 2, col. 9, p. 47)
6. Average of 181 Hawaiian tholeiites (MacDonald and Katsura, 1964, p. 124)
7. Average of 4 quartz tholeiites, North Shore Volcanic Group (reduced, etc.)
8. Average of 7 (quartz) tholeiites, Thingmuli Volcano (reduced, etc.) (Carmichael, 1964, p. 439)

olivine, leaving the remaining liquid enriched in plagioclase components, similar to the high-alumina oceanic tholeiites. Similarly, partial melting of mantle pyrolite at the same depth would also produce such a liquid. Also, separation of 10 percent clinopyroxene alone from an "oceanic tholeiite" under similar conditions would leave a residual liquid nearly identical to the average North Shore olivine tholeiite (table V-4, col. 5). Beyond this stage of crystallization of olivine tholeiite at these depths, plagioclase appears on the liquidus and further alumina concentration in the magma is impossible. Thus, the North Shore olivine tholeiite magmas are assumed to have segregated from mantle materials at some depth in the range 15 to 35 km. Estimates of the thickness of the Keweenawan lavas and intrusions in the western Lake Superior area (see for example White, 1966a; this paper) give 6 to 15 km of mafic igneous rocks. They overlie more granitic basement rocks of Middle and Early Precambrian age at the edges of the basin, but if the olivine tholeiites at the top of the Keweenawan (say, 13 km above the base) were derived from the mantle at a maximum depth of 35 km, this would allow a stable pre-Keweenawan crust of at the most only 22 km thickness. This may be interpreted as support for the idea of origin of the Keweenawan igneous province by rifting (Smith and others, 1966; King and Zietz, 1971). The sialic crust may actually have been pulled apart under part of the Lake Superior basin.

As most of the North Shore basalts are aphyric, there must have been a very efficient separation of crystals from residual liquid during fractionation. Those phenocrysts that are present are in nearly all cases plagioclase, indicating, according to Green and others' (1967) experiments, that crystallization had proceeded at shallower depths (5-15 km) than those at which the olivine tholeiite liquids were formed. In a few basaltic lavas, as for example in Onion River and Tofte, clinopyroxene phenocrysts are present as well as plagioclase, but magnesian olivine is extremely rare as phenocrysts except in the basal Keweenawan melabasalts. This may be a consequence of more efficient settling of olivine crystals than of plagioclase or pyroxene.

Saturated quartz tholeiites (table V-4, col. 7) are a less abundant yet distinctive basalt type in the North Shore Volcanic Group. They are fine grained and aphyric, except for rare plagioclase phenocrysts in a few flows. Very similar basalts are well known from many other plateau sequences, including the Tertiary of Northern Ireland, Scotland, and Iceland (table V-4, col. 8). Experimental studies (as for example Green and others, 1967) show that such liquids can be produced from mantle materials only at relatively shallow depths (0-15 km), by either partial melting of pyrolite or fractional crystallization of olivine tholeiite (or high-alumina tholeiite) that had been segregated at some greater depth. This relationship is controlled by the fact that only at these shallow depths does the liquidus field of olivine extend across the silica-saturation boundary into the oversaturated area (because of the incongruent melting of orthopyroxene), so that separation of olivine can drive the residual liquid across this boundary from olivine-normative to quartz-normative compositions.

As mentioned above, Green and others (1967) concluded from their experiments that "... segregation of liquid from partially melted pyrolite at (15-35 km) depths will yield high-alumina basalt liquids. . . . The liquid so formed may rise to the surface as high-alumina basalts or may undergo fractional crystallization at shallow levels (0-15 km) yielding quartz tholeiite derivative liquids complementary to residual stratified anorthosite-peridotite complexes" (p. 49).

Phinney (1970) has shown recently that the 14 basaltic and andesitic analyses then available for North Shore lavas fit well a model of derivation of lava liquids by fractional removal of the crystalline components of gabbroic anorthosite, similar to that which constitutes the bulk of the Duluth Complex, from a primary melt rich in Al_2O_3 and CaO. This deduced primary melt is close in composition to Sample T-56, an olivine basalt (table V-4, col. 2), but although this rock has a high MgO content and a low K_2O, P_2O_5, and SiO_2 content, this lava type is not abundant on the north shore and it is crowded with small bytownite phenocrysts, which indicate that it possibly may itself have been enriched physically in Al_2O_3 and CaO. However, these phenocrysts are small (1-3 mm) and evenly distributed, and do not appear to be xenocrysts. The next most primitive basalts of Phinney's calculations, T-45 and KC-9, are typical aphyric, ophitic olivine tholeiites that are the most widespread and abundant single lava type and might be more likely to have been primary mantle differentiates, as discussed above. On the other hand, it is possible that very efficient fractionation removed all phenocrysts from these liquids before extrusion and that the initial magma actually was more anorthositic. As large plagioclase phenocrysts are present or abundant in many minor intrusions and some basalt flows of the North Shore Volcanic Group, it is obvious that plagioclase fractionation played a major role in the differentiation.

The K_2O contents of a few of the basalts and several of the intermediate rocks, however, do not fit well with any simple process of removal of gabbroic or anorthositic crystals from a low-potassium primary magma. It should be noted that whereas the K_2O-rich basalts (for example trachybasalts LW-1, LW-10) could be called alkali-rich olivine basalts, they are unlike the oceanic alkali basalts in having a relatively high K/Na ratio and lacking feldspathoids or analcite. Thus Engel and others' (1965) conclusions as to derivation of oceanic alkali basalts from primary tholeiite probably are not directly applicable, and it is unlikely that these potassic mafic rocks were derived directly from an "ordinary" pyrolite source.

Another distinct possibility is that the K-rich character of these rocks, and the relatively large volume of intermediate and felsic flows as compared to basalts, may be the result of large-scale contamination or assimilation of crustal rocks by the enormous amount of mafic magma that penetrated the crust during the formation of the Keweenawan intrusive and extrusive rocks. Interstitial granophyric material is common in many rocks of the Duluth Complex, and minor intrusions, especially in the higher level Beaver Bay Complex and others along the Lake Superior shore, show a continuous range of compositions from olivine gabbro through granogabbro and syenogabbro to augite granodiorite to adamellite or granite, parallel to the trend of flow compositions. It is difficult to imagine the voluminous Keweenawan mafic magmas penetrating and intruding granitic basement rocks (as is clearly the case at the edges of the basin) without considerable melting and assimilation of granitic material and consequent enrichment of the differentiates at least locally in potassium. On the other hand, it would be difficult to enrich these liquids in K without at the same time enriching them in Si; such an assimilation hypothesis cannot explain the K-rich character of some basalts and intermediate rocks. Available initial Sr isotope ratios from mafic intrusions in the Duluth area (Faure and others, 1969; 0.7055 for the Duluth Complex and 0.7046 for the Endion sill) suggest slight crustal contamination of dominantly mantle-derived liquids. Further Rb-Sr and Sr isotope studies should help to determine the role of contamination in the development of this series.

A further possibility that warrants investigation is that the North Shore Volcanic Group actually represents a mixture of two differentiation suites, one having an olivine tholeiite parent and another characterized by a high K_2O content and a high Fe/Mg ratio. The more potassic group could have been derived from an anomalous mantle material that contained a large proportion of "KREEP," analogous to some of the lunar rocks. If so, the higher radioactivity of this material may have contributed to the heat necessary to produce the voluminous Keweenawan magmas.

STRATIGRAPHY

The lavas of the North Shore Volcanic Group can be divided conveniently into several stratigraphic units of coherent petrographic character, primarily on the basis of exposures at or near the Lake Superior shore. Many of these units can be traced for a considerable distance inland, but interruptions and structural complications by intrusive bodies, as well as lack of outcrop inland, prevent the reconstruction of a continuous sequence. Irving (1883) and Elftman (1898) described several major members or groups of flows, but in the following description a new and considerably different terminology is used. Considering the entire sequence, no clear trend of compositional change is evident; in fact, although the most ferromagnesian lavas occur at the base, the uppermost flows, in the Tofte-Lutsen area, are entirely olivine basalts.

Tables V-1 and V-2 show the general characteristics and abundance of the major lava types in the northeastern limb of the basin (Tofte to Grand Portage). In the southwestern limb, descriptions from previous work are insufficient to differentiate between some of the chemical types; nevertheless, Table V-2 shows the recorded abundances in the Duluth to Two Harbors and the Two Harbors to Beaver Bay Complex segments. Between the Beaver Bay Complex at Silver Bay and the top of the section at Tofte the structure is more complicated, and estimates of abundances are not given for this segment.

Description of Units

In this section, a generalized description of the stratigraphic sequence will be given, starting with the Lake Superior shoreline section, which is best exposed. Brief comments will then be made on inland areas where some volcanic rocks are exposed and known. Columnar sections of the North Shore Volcanic Group are shown in Tables V-5 and V-6; many of the units are represented by analyzed samples in Table V-3.

The northeastern limb, in Cook County, has been mapped most recently (figs. V-6 and V-16). The basal flows at Grand Portage directly overlie the Puckwunge Formation. The lowest 4,500 feet or so are basalts of various types; especially notable is a succession about 250 feet thick of porphyritic melabasalts, crowded with augite and serpentinized olivine phenocrysts, which lies at the very base on Lucille and Magnet Islands, a few miles east of Grand Portage (fig. V-10A). Directly above these pyroxene-porphyritic lavas on Lucille Island, and also in the basal exposures at Portage Brook, 22 miles to the west, are single flows of a remarkable plagioclase porphyry (fig. V-11B). The uppermost 460 feet or so of these magnetically reversed Grand Portage lavas, in the vicinity of Deronda Bay, consists of a thick trachyandesite-quartz latite and an overlying thick rhyolite flow (Red Rock rhyolite). The Grand Portage sequence is cut by a swarm of diabasic dikes, some of which are plagioclase-porphyritic, and which trend about N. 80° E. (see below). This succession of lavas, herein called the Grand Portage lava series, extends westward, bounded on the north side by a 100-300-foot escarpment, where it disconformably overlies the basal Upper Precambrian Puckwunge orthoquartzite and the underlying softer Middle Precambrian Rove Formation (fig. V-11C). It is intruded by large volumes of gabbroic and diabasic rocks of the "Logan" intrusions and, to the southwest, by the Reservation River diabase complex.

Table V-5. Stratigraphy of northeast limb (Tofte-Grand Portage), North Shore Volcanic Group. (Interflow sedimentary rocks not included.)

Approx. thickness (ft.)	Lithostratigraphic unit	Lithic character
Top	Middle Keweenawan	
1020	Lutsen basalts	Olivine basalts, olivine tholeiites
160	Terrace Point basalt flow	Thomsonite-bearing ophitic basalt
310	Good Harbor Bay andesites	Brown, porphyritic andesite, trachyandesite
360	Breakwater trachybasalt flow	Brown, columnar, granular trachybasalt
500	Grand Marais rhyolite flow	Pink, red, gray porphyritic rhyolite
600	Croftville basalts	Various fine-grained basalts
1020	Devil Track felsites	Aphyric and porphyritic rhyolite flows
400-900	Red Cliff basalts	Amygdaloidal, ophitic olivine basalts
1300	Kimball Creek felsite	Pink to tan, porphyritic felsite
1800	Marr Island lavas	Mixed tholeiitic basalt, intermediate, felsic lavas
1000	Brule River basalts	Granular-diabasic basalts
3500	Brule River rhyolite flow	Pink to gray porphyritic rhyolite
	Hovland diabase complex	
	Lower Keweenawan	
4000 (est.)	Hovland lavas	Mixed porphyritic basalt, trachybasalt, rhyolite
	Reservation River diabase complex (Middle Keweenawan)	
200	Red Rock rhyolite flow	Red, porphyritic rhyolite
260	Deronda Bay andesite flow	Gray-brown, aphyric andesite
4500	Grand Portage basalts	Mixed tholeiitic to diabasic basalts; porphyritic metabasalts locally at base
Base	Disconformity	
	Puckwunge Formation	Cross-bedded quartz sandstone
	Disconformity	
	Middle Precambrian	
	Rove Formation	Shale and graywacke

Table V-6. Generalized stratigraphy of southwest limb (Tofte-Nopeming), North Shore Volcanic Group. (Interflow sedimentary rocks not included.)

Approx. thickness (ft.)	Lithostratigraphic unit	Lithic character
Top	Middle Keweenawan	
4000	Schroeder basalts	Amygdaloidal ophitic olivine tholeiites
>300	Manitou trachybasalt flow	Red-brown granular trachybasalt to basalt
	(more of the Schroeder basalts)	
>280	Bell Harbor lavas	Mostly quartz tholeiites, other basalts
>300	Palisade rhyolite flow	Gray to pink, porphyritic rhyolite
few 100's	Baptism River lavas	Mixed lavas, mostly basalts
	Beaver Bay intrusive complex	
3200	Gooseberry River basalts	Mixed basalts, one felsite
	Lafayette Bluff, Silver Creek Cliff intrusions	
1025	Two Harbors fine-grained basalts	"Melaphyres," some quartz tholeiites
1615	Larsmont ophitic basalts	Amygdaloidal ophitic olivine basalts
	Knife River diabase intrusion	
4930	Sucker River basalts	Mixed basalts, mostly ophitic
4400	Lakewood basalts	Mixed basalts, mostly non-ophitic
	Lester River diabase sill	
3600	Lakeside lavas	Mixed basalts, andesites, felsites
	Endion diabase sill	
2560	Leif Ericson Park lavas	Mixed basalts, andesites
	Duluth Complex	
	Lower Keweenawan	
1200	Ely's Peak basalts	Porphyritic melabasalts, diabasic basalts
Base	"Nopeming sandstone"	Quartz sandstone
	~~Angular unconformity~~~~~~~	
	Middle Precambrian	
	Thomson Formation	Slate and graywacke

Just west of the Reservation River, lavas reappear along the lakeshore; they are herein called the Hovland lavas. These too have reversed magnetic polarity according to Palmer (1970) and Green and Books (1972). The Hovland lavas near the shore include rhyolite and a distinctive porphyritic trachybasalt that contains plagioclase phenocrysts as much as 10 cm across (fig. V-10D). Most of the low area near the lakeshore that is probably underlain by these lavas is covered, however, and the rock types comprising much of the unit are not known. They are probably similar to the varied basaltic, intermediate, and felsic flows exposed a few miles inland near Farquhar Peak. Similar plagioclase-porphyritic basaltic to trachybasaltic flows are common about 18 miles on strike to the west near Greenwood Lake. The total thickness of this group of lavas is not known as few dips have been measured, but is estimated to be 4,000 to 5,000 feet.

The various gabbroic and syenogabbroic rocks of the Hovland diabase complex form the lakeshore section southwestward to the Brule River.

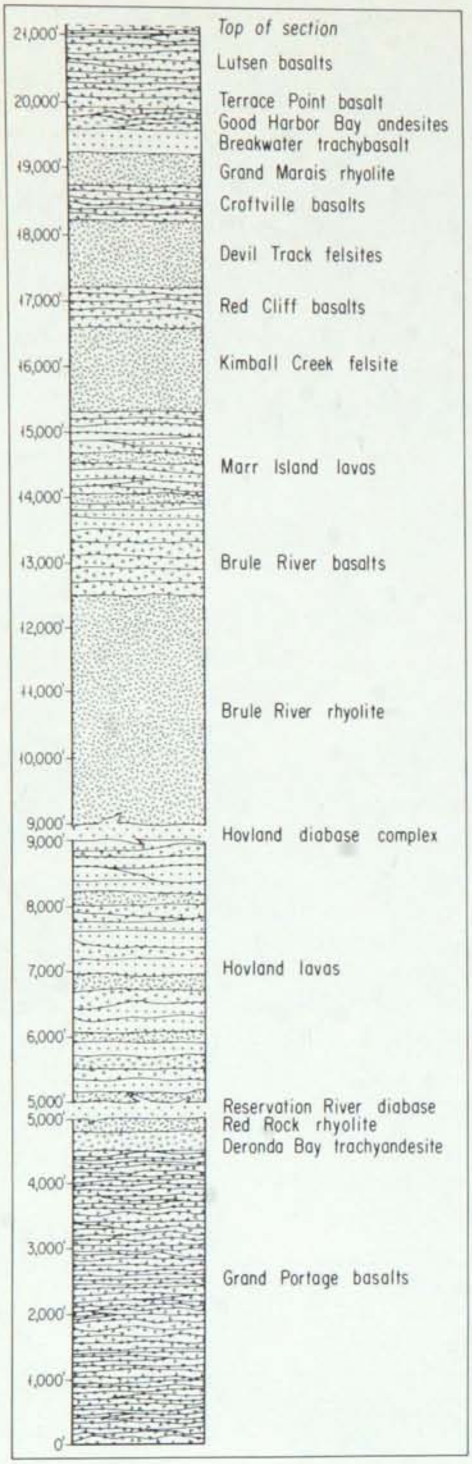

Figure V-16. Diagrammatic columnar section for northeast limb (Lutsen to Grand Portage) of the North Shore Volcanic Group. Extreme thickness of Brule River rhyolite may represent lava dome, but intrusions, other structural complications, and lack of outcrops prevent a clear interpretation.

From the Brule River southwestward to Duluth, the entire lava sequence has normal magnetic polarity. At the mouth of the river, a group of about five diabasic basalts (Brule River basalts) is exposed; these are underlain, upriver, by basalt breccia and pillow-breccia (fig. V-8B), which in turn overlie a very thick porphyritic rhyolite flow complex (Brule River rhyolite, as much as 3,500 feet thick). The basalt group is overlain by a thick series (about 1,800 feet) of interlayered basaltic, intermediate, and less common felsic flows (Marr Island lavas), westward nearly to Kadunce Creek. One andesitic flow from this series (MI-35, table V-3) contains abundant glass having minute, immiscible iron-rich glass spheres. Another thick rhyolite, the Kimball Creek felsite (1,300 feet thick) follows; this is traceable at least 16½ miles westward along strike to the Cascade River. It is overlain at Red Cliff by another series of about five ophitic basalt flows (the Red Cliff basalts, about 900 feet thick), some of which have a few plagioclase phenocrysts that have floated in one flow and sunk in another. Overlying these basalts are two felsite flows (Devil Track felsites) totaling about 1,020 feet in thickness; the deep canyon of the Devil Track River has been cut into the upper, thicker one. This flow has been traced west-southwestward inland for about 23 miles. This felsite is overlain by at least six basalt flows (the Croftville basalts, about 600 feet thick) just south of the Devil Track River east of Grand Marais, but a fault probably separates this group from the lakeshore outcrops west of the river mouth, where a variety of basalts (probably the same Croftville group) and a porphyritic rhyolite, much intruded and deformed by diabase, extend westward to Grand Marais.

The town of Grand Marais is mainly underlain by the porphyritic rhyolite, discussed above, which is locally intruded by diabase. The Grand Marais rhyolite is estimated to be 500 feet thick, and has been traced with certainty for about 8 miles. The harbor is protected by the massive base of a thick trachybasalt (the Breakwater trachybasalt) that directly overlies the rhyolite. This massive trachybasalt, 320-400 feet thick, is thought to be a flow because it has a zeolitized, amygdaloidal top zone (although the actual top surface is not exposed). It may, however, be a sill.

Two thick intermediate flows totaling 310 feet in thickness (the Good Harbor Bay andesites, equivalent to at least six flows on strike farther west) overlie the trachybasalt and are in turn overlain at Cutface Creek by a thick unit of sandstone that grades upward into interbedded fine sandy and shaly siltstone. This sandstone unit continues inland for at least 6 miles, and thins to the west. It is overlain by a major ridge-forming, fine-grained, ophitic basalt flow that is characterized by thomsonite in its amygdules as well as by its texture; it can be traced at least 16 miles west-southwestward to the Poplar River. Known as the Terrace Point basalt, it contains some complex units (some possibly pillowed) and is about 160 feet thick.

The Terrace Point basalt is followed stratigraphically by a thick succession (the Lutsen basalts) that starts just east of the Cascade River with rather thick, coarse-grained olivine basalts. This group includes two flows at the base, totaling at least 220 feet thick, an overlying 300-foot-thick wedge of red sandstone, and about six more olivine basalts which are together 700 feet thick. Overlying these are a group of seven or eight thinner olivine basalts, which mark the apparent top of the section at Lutsen village. Unfortunately, faulting in this area makes it difficult to ascertain the exact thickness of this group and precisely which segment of the continuous shoreline exposure contains the highest flow. The group is estimated to be at least 100 feet thick.

Southwest from Lutsen, the basalt succession descends stratigraphically to the Onion River; in this sequence many of the olivine basalts contain abundant small plagioclase phenocrysts, and the Terrace Point ophite appears to pinch out. At the Onion River, there are fine-grained, porphyritic andesitic and basaltic flows with rubbly tops that may be equivalent to the andesitic group seen to the northeast beneath the Terrace Point basalt at Good Harbor Bay. The high hills just inland from there (Leveaux, Onion, and Eagle Mountains) are cuestas held up by a thick porphyritic trachybasalt sill that crosscuts the flows at a small angle (see below).

Proceeding toward Tofte from the Onion River, progressively younger flows are met again, all basalts, but abundant faults (see fig. V-4C) make tracing of individual flows difficult. The basalts probably are the lateral equivalents of the Lutsen basalts, and include some similar plagioclase-porphyritic olivine basalts. The topmost flow is thought to be that exposed at the old boat dock at Tofte village. These basalts, here called informally the Schroeder basalts, are almost entirely of the ophitic olivine-tholeiite type, and include many thin flow units, although only a mile inland, on the flank of the intrusion-cored Carlton Peak, rhyolite and quartz tholeiite are exposed. At least 700 feet of basalts are exposed at Tofte at the top of the succession (figs. V-9B, 9C, and 11D).

The ophitic Schroeder basalts can be traced nearly continuously to the southwest past Schroeder and Taconite Harbor to the Cook-Lake County line, and apparently thicken in this direction; southwest of Taconite Harbor, about 4,000 feet of basalts are present (fig. V-6). The

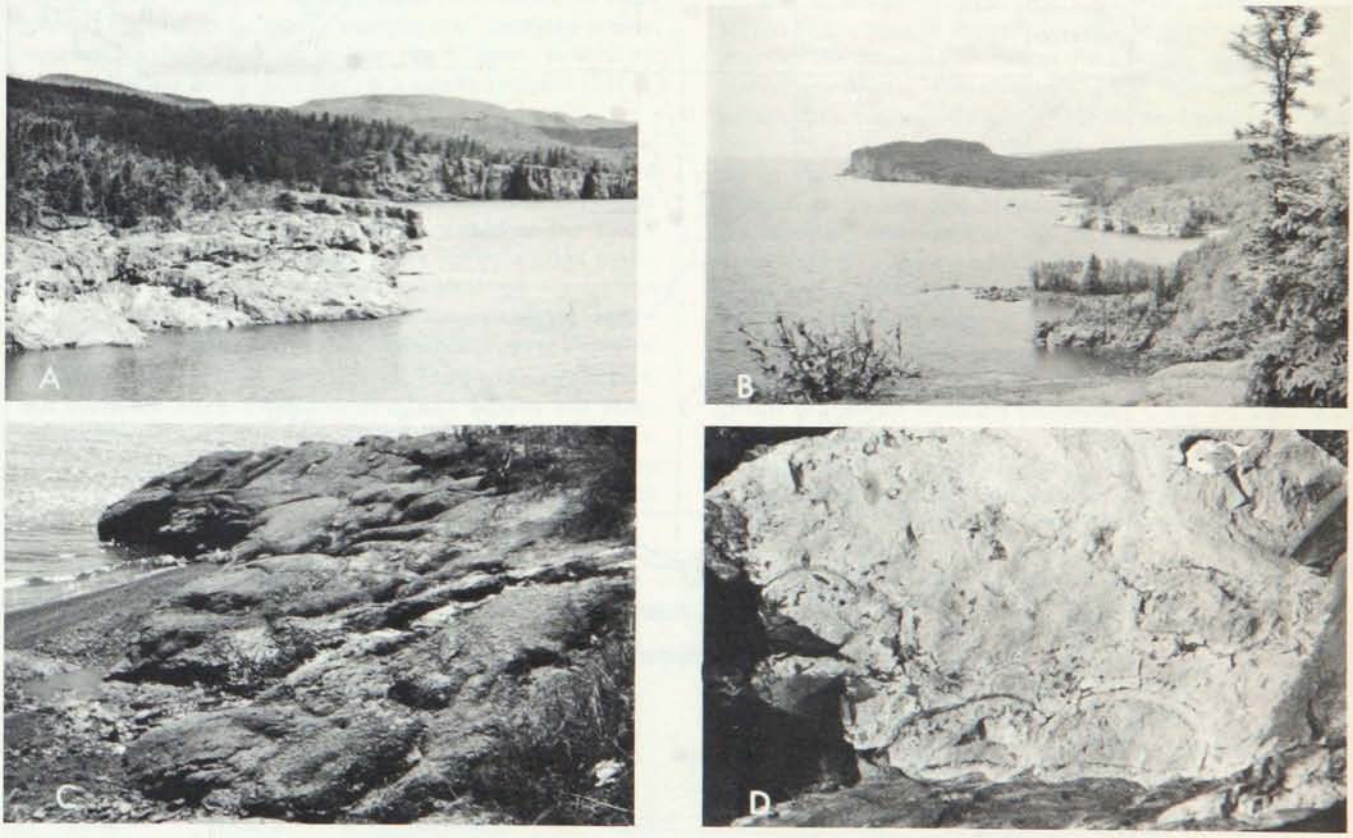

Figure V-17. Photographs of Keweenawan rocks. A, Bell Harbor lavas (dark, middle distance) overlying Palisade rhyolite (left foreground). View northeastward from Shovel Point, Illgen City, Lake County. Hills and deep road cut in distance are intrusive rocks of Beaver Bay Complex; B, view southwestward toward Palisade Head from top of Shovel Point, Illgen City, Lake County. Viewpoint and Palisade Head both held up by Palisade rhyolite flow; intervening shoreline is cut into underlying Baptism River lavas, except for fault-dropped section of Palisade rhyolite that forms light-colored cliffs in middle distance by Baptism River mouth; C, wave-stripped, coarsely wrinkled basalt flow surface in Lakewood basalts. Lake Superior shore at 7710 Congdon Blvd., Duluth; D, sections across several lava toes or lobes in diabasic basalt of Ely's Peak lavas: southwest slope of Ely's Peak on Duluth, Winnipeg, and Pacific Railroad. Some structures such as these in this unit bear strong resemblances to pillows, but they occur with good ropy surfaces and are thought to be subaerial rather than subaqueous.

Schroeder basalts continue southwestward past Little Marais, and thin in this direction; they are interrupted in the vicinity of the Manitou River by a thick, reddish trachybasalt flow that rests on a series of interbedded thin basalts and breccia-agglomerate, and locally, on felsite. The top of the Manitou trachybasalt is not exposed, but the flow is at least 300 feet thick (figs. V-10B and 10C).

Southwest of Little Marais, faulting again interferes with detailed accounting of the lava succession, but the ophitic Schroeder basalts are underlain by a series of fine-grained quartz tholeiites, trachybasalts, and other basaltic varieties (the Bell Harbor lavas, at least 280 feet thick), which in turn overlie a great rhyolite flow at Crystal Bay at Illgen City (fig. V-17A). This flow, at least 300 feet thick, forms Shovel Point and re-enters the shore 2 miles to the southwest to form Palisade Head (fig. V-11A). The Palisade rhyolite is in turn underlain by a few hundred feet of mixed basaltic flows exposed in and near the mouth of the Baptism River (Baptism River lavas; fig. V-17B). In this region, the flows are intruded by many large dikes and small plutons of the Beaver Bay Complex, and their attitudes are locally strongly disturbed from the regional north-northeasterly trend. Recent mapping of the flows has not yet been extended southwest of Palisade Head.

According to Grout and Schwartz (1939), both basalts and rhyolites form the volcanic remnants between the domi-

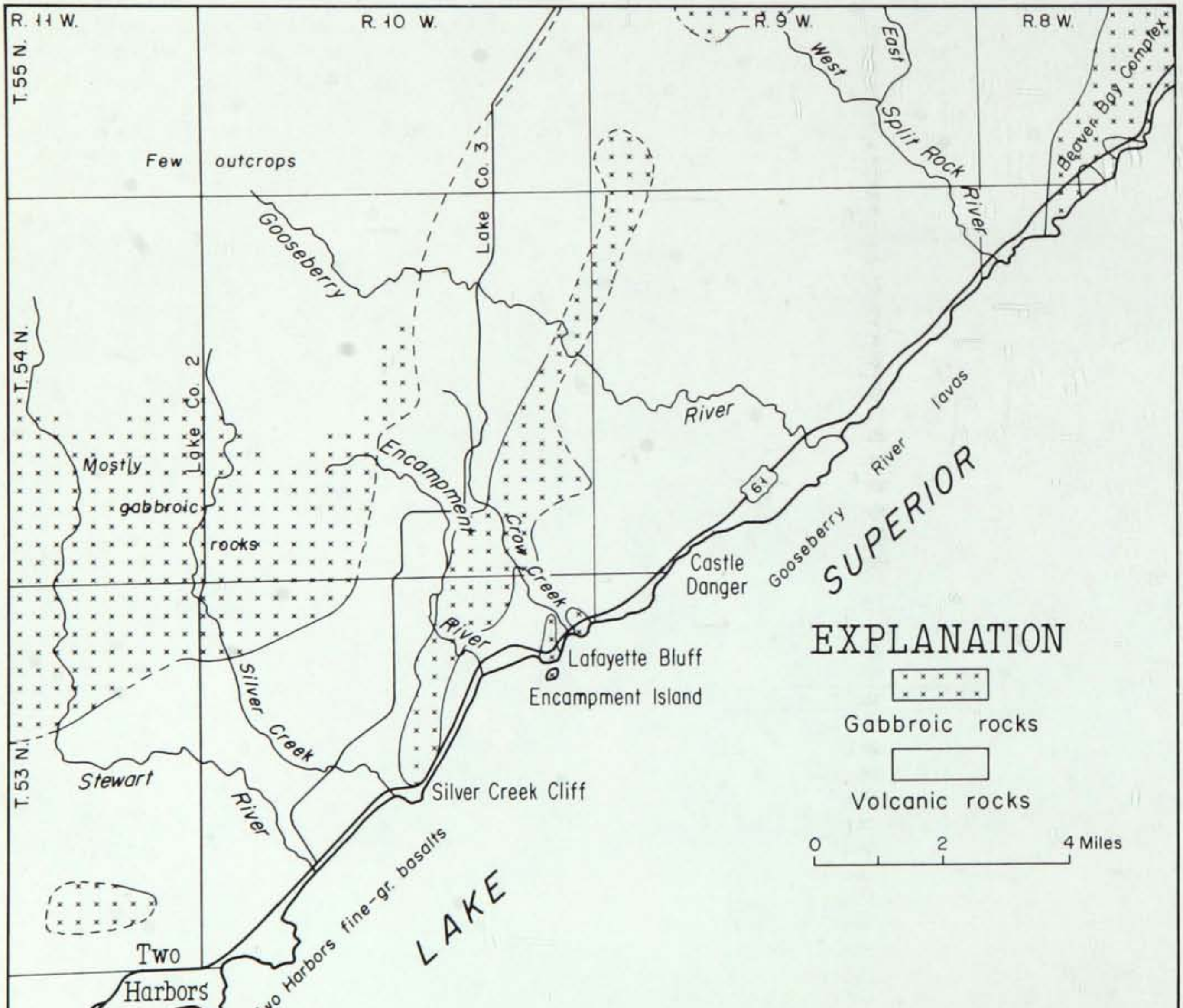

Figure V-18. Generalized bedrock geologic map of Lake Superior shore between Two Harbors and Split Rock Point, Lake County, and adjacent inland areas. Northwestern part of map area has few outcrops but they are nearly all of gabbroic rocks (after Bonnichsen, 1971).

nant intrusions of the Silver Bay-Beaver Bay area, but they cannot yet be tied stratigraphically to the succession farther removed from the intrusive complex.

To the southwest from the Beaver Bay Complex at Split Rock River, the stratigraphic succession descends, and the shoreline as far as Two Harbors has been mapped by R. M. Grogan (1940, *op. cit.*). From this work and that of Sandberg (1938), tentative groupings of flows into informal stratigraphic units (table V-6; fig. V-18) can be given. For a distance of 8 miles, a succession of generally basaltic rocks ("melaphyres and ophites"), including one glomeroporphyritic basalt and one porphyritic felsite, forms shoreline exposures to the vicinity of Crow Creek, where a thick diabase intrusion has deformed the flows and forms Lafayette Bluff and Encampment Island. This group is herein called the Gooseberry River basalts. Additional basaltic flows progressively underlie this succession southwestward to Silver Creek Cliff, where they again are intruded by a thick gabbroic sill that crosscuts the flows at a small angle and which probably once connected inland with the Lafayette Bluff intrusion. More basalts (Two Harbors fine-grained basalts), principally "melaphyres," continue southwestward to Two Harbors; at least one of these, at the city tourist park on Burlington Bay, is a quartz tholeiite (TH-2, table V-3).

Sandberg's mapping (1938), from Two Harbors to Duluth, shows the sequence of melaphyres at Two Harbors to be underlain by a thick group of ophitic basalts ("Larsmont ophitic basalts") that extends southwestward to the Knife River, where the flows once again are cut by a thick crosscutting diabase sheet that projects eastward into the lake to form Knife Island and extends southwestward to Stony Point (see fig. V-19). Quaternary lake clays cover much of the succession between Stony Point and the French River, but there are a few exposures of ophitic basalts and other ophitic basalts underlie the shoreline to the Talmadge River. This group of dominantly ophitic basalts between Stony Point and the Talmadge River is herein called the Sucker River basalts. These are succeeded down-section by dominantly thin "melaphyre" lavas ("Lakewood basalts") which extend past the Lakewood pumping station (fig. V-17C). These are underlain successively by additional melaphyres (some of which are porphyritic olivine trachybasalts), two felsite flows, and the thick Lester River diabase sill, characterized by its red, granitic top zone. Other melaphyres appear beneath the diabase at the Lester River, and are underlain in turn in the East End of Duluth by a mixed sequence of porphyrites, melaphyres, and felsites, including a thick, banded flow into which the Tischer Creek gorge has been cut. This latter group, beneath the Lester River sill, is here termed informally the Lakeside lavas.

Further southwestward in Duluth, the thick, differentiated Endion sill underlies the Tischer Creek felsite. It is in turn underlain in the vicinity of Leif Ericson Park by a mixed series of "melaphyres" and "porphyrites" (Leif Ericson Park lavas), which are truncated by the upper contact of the Duluth Complex.

Beneath the Duluth Complex at Nopeming, west of Duluth, a wedge of lavas (the Ely's Peak basalts) is preserved at the base of the volcanic sequence; it conformably overlies the basal Keweenawan Puckwunge sandstone (fig. V-4A; see Mattis, this chapter). Their reversed magnetic polarity (Green and Books, 1972) shows these Ely's Peak basalts to be Lower Keweenawan. They are strikingly similar to the basalts at the northeastern extremity of the sequence on Lucille and Magnet Islands at Grand Portage; identical pyroxene-and-olivine-porphyritic basalts at the base are overlain by massive, diabasic basalts (fig. V-17D; see Kilburg, 1972, *op. cit.*). The Ely's Peak basalts extend northward for a distance of 7 miles, where they wedge out beneath the Duluth Complex. Kilburg's mapping indicates that only about 1,200 feet of lavas are represented in this group, in contrast to earlier estimates.

In addition to the extensive areas underlain by volcanic rocks that are in direct continuity with the lakeshore outcrops, several isolated patches or inliers of flows have been

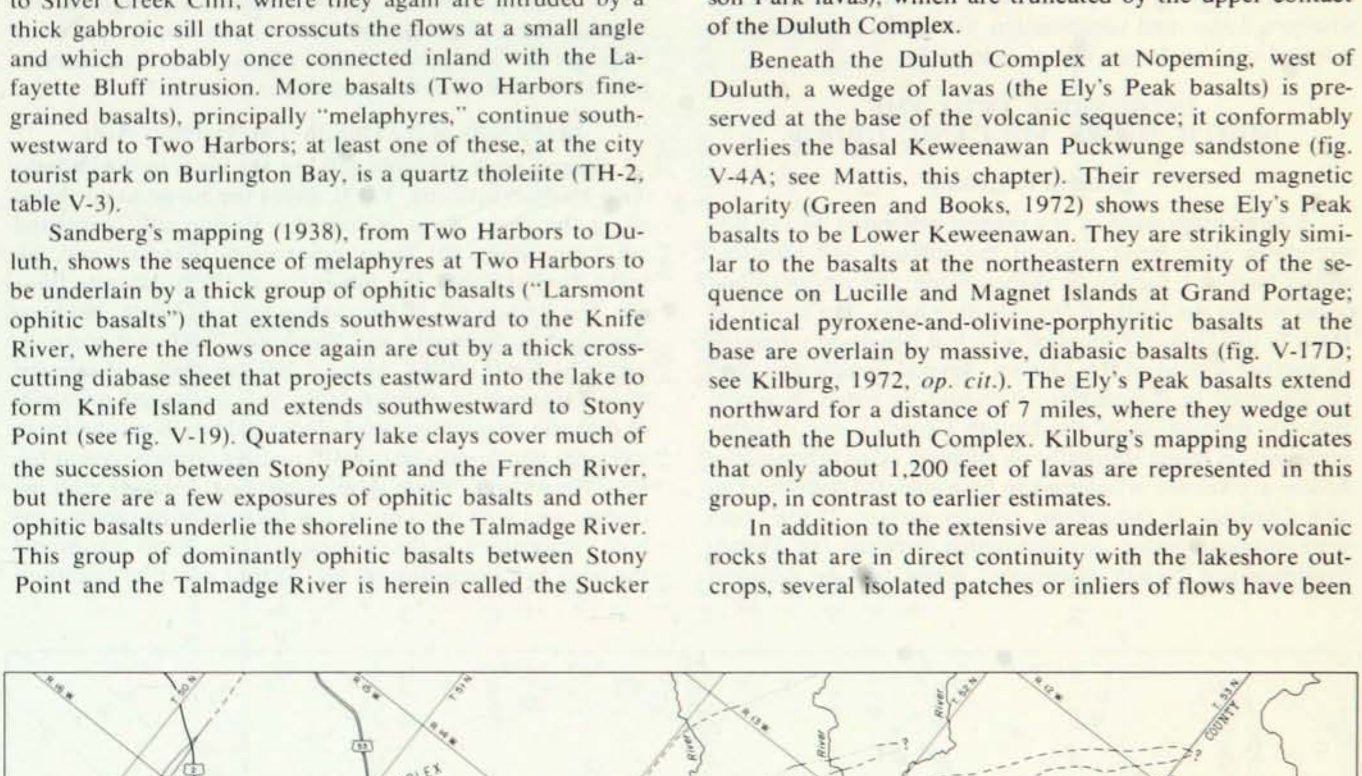

Figure V-19. Generalized bedrock geologic map of Lake Superior shore zone from Duluth to Two Harbors, with adjacent inland area, showing intrusive bodies and volcano-stratigraphic units.

found that are surrounded by intrusive bodies of the Duluth Complex. About 6 miles northeast of Hoyt Lakes, in sec. 19, T. 59 N., R. 13 W., Bonnichsen (this chapter) has recognized a thin zone of basalts immediately underlying the complex. These lavas have been thoroughly recrystallized by the overlying troctolite. In southwestern Lake County, between T. 56 N., R. 11 W. and T. 58 N., R. 10 W., southwest of Whyte, is a long, northeasterly-trending area of sparsely outcropping lavas of several varieties. They also have been partly recrystallized, and many of the basalts contain abundant magnetite. North and northeast of Dumbbell Lake, in east-central Lake County (T. 60 N., R. 7 W.) is another area of both basalts and rhyolites, which may or may not be isolated within the Duluth Complex. Although their boundaries are obscure because of poor outcrops, these lavas probably cover an area of at least 7 square miles.

In addition to the above, most of the extensive regions that contain little or no outcrop (in southwestern Cook, southern Lake, and southeastern St. Louis Counties), are inferred to be underlain by volcanic rocks.

INTRUSIONS INTO THE NORTH SHORE VOLCANIC GROUP

General Relations

A large proportion of the magma generated during Keweenawan time never reached the surface, and intrusive bodies having a variety of shapes, sizes, and compositions occur within the limits of the extrusive lavas. The largest of these is the Duluth Complex, which is discussed separately in another section of this chapter. Several smaller units will be discussed in this section; it should be noted, however, that the delineation of the Duluth Complex and of other intrusions is based primarily on whether or not the smaller bodies are known or inferred to be connected with the Duluth Complex at the present erosion surface. If the rocks were exposed better, the distinction between the Duluth Complex and "minor intrusions" might well be different from that made here. The distinction is particularly difficult to make for the Beaver Bay Complex, which merges to the north into the Duluth Complex, and for the intrusions in the area north and west of Two Harbors, where outcrops are sparse. With respect to the Beaver Bay Complex, its northern boundary is drawn arbitrarily at the middle of the Cramer 1:62,500 quadrangle (see fig. V-23, section on Duluth Complex). Bonnichsen (also in the section on the Duluth Complex) interprets the eastern boundary of the complex in the area north of Two Harbors to be the eastern limit of exposed troctolitic and anorthositic rocks that are clearly continuous with the remainder of the Duluth Complex; this gives a fairly smooth contact on the map (see fig. V-19). Bonnichsen discusses the minor intrusions that are east of, and stratigraphically above, this contact in another section in this chapter. The locations of the various intrusive bodies discussed below are given on Figures V-6, V-18, V-19, and V-20; many of them are designated informally. Several smaller intrusive bodies that are known to exist are not described in this report. Chemical analyses published since 1899 are compiled in Table V-7.

Mafic Sills of the Duluth-Two Harbors Area

Several large gabbroic sills cut the lavas in the Duluth-Two Harbors area (fig. V-19). From southwest to northeast along the shore these include the Endion sill, Northland sill, Lester River sill, Stony Point-Knife Island sill and, beyond Two Harbors, the Silver Creek Cliff-Lafayette Bluff sill. The Endion sill, in the eastern part of Duluth, has been described by Schwartz and Sandberg (1940) and more recently by Ernst (1960). About 1,500 feet thick and exposed over a distance of about 6 miles, it is composed primarily of diabasic, medium-grained olivine gabbro, but contains in its upper part approximately 40 percent intermediate to felsic differentiates. These show gradational contacts with the diabase and with each other, but because of the anomalous bulk composition of the exposed portions, Ernst interpreted the body as a composite sill that formed either by migration

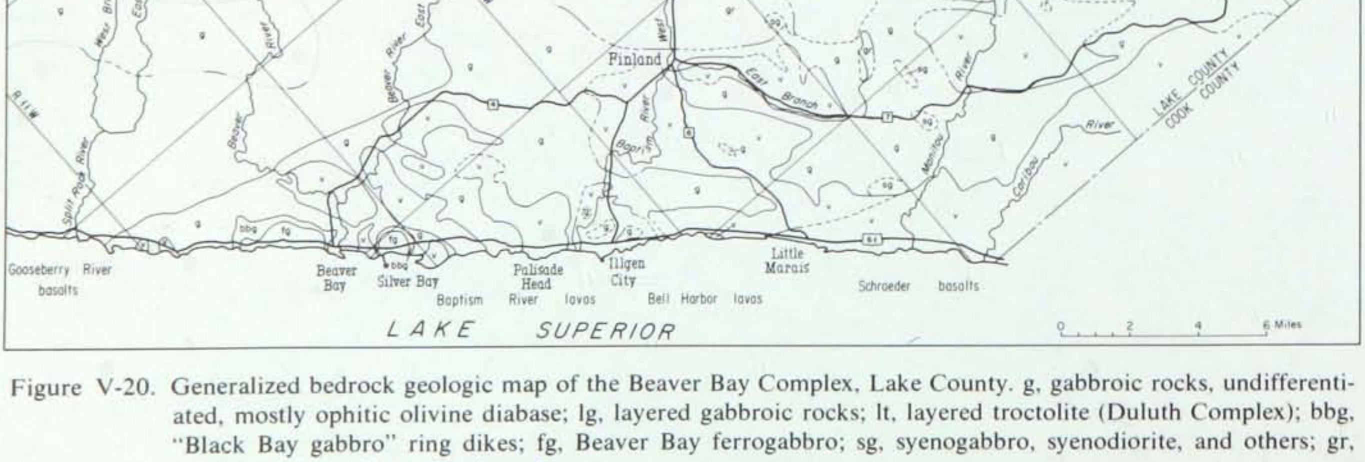

Figure V-20. Generalized bedrock geologic map of the Beaver Bay Complex, Lake County. g, gabbroic rocks, undifferentiated, mostly ophitic olivine diabase; lg, layered gabbroic rocks; lt, layered troctolite (Duluth Complex); bbg, "Black Bay gabbro" ring dikes; fg, Beaver Bay ferrogabbro; sg, syenogabbro, syenodiorite, and others; gr, granitic rocks; and v, volcanic rocks.

Table V-7. Chemical analyses* of Keweenawan intrusive rocks (exclusive of Duluth Complex).

	Beaver Bay Complex											Endion sill				
	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16
SiO_2	43.21	45.84	45.87	46.94	47.88	48.77	49.16	49.83	50.04	50.86	63.18	47.25	48.51	52.42	61.07	61.46
TiO_2	4.16	3.44	4.65	3.86	3.00	1.52	1.36	2.21	3.36		1.51	2.89		2.66	1.37	1.37
Al_2O_3	14.22	11.19	10.66	11.87	11.86	18.25	17.09	11.59	13.00	15.72	13.27	15.00	13.79	12.66	13.66	13.22
Fe_2O_3	5.11	4.02	4.11	4.66	4.05	2.25	2.28	4.51	4.20	9.77	4.03	2.64	19.34	3.90	3.04	3.08
FeO	13.46	16.30	15.59	11.33	15.56	8.44	7.99	17.19	10.71	2.48	4.42	11.09		9.55	5.54	5.42
MnO	0.20	0.27	0.29	0.24	0.30	0.15	0.19	0.37	0.22		0.12	0.21		0.25	0.15	0.18
MgO	4.53	3.66	3.50	5.99	2.47	4.94	7.43	1.06	3.61	3.55	1.62	6.52	4.81	3.74	2.48	2.00
CaO	9.31	8.39	8.28	10.88	7.22	10.81	10.89	7.34	7.59	10.52	2.74	8.40	8.34	5.16	2.36	2.96
Na_2O	2.66	2.68	2.84	2.26	2.89	2.77	2.35	2.93	2.66	3.89	3.47	2.52	1.67	3.01	3.40	3.33
K_2O	0.67	0.83	0.96	0.48	1.29	0.45	0.45	1.15	1.17	0.90	3.57	0.81	0.19	2.44	4.10	4.30
H_2O+	0.73	0.87	1.42	0.64	0.92	0.84	0.60	0.53	1.51	2.53	1.20	1.63		2.00	1.79	1.24
H_2O-	0.60	0.45	0.47	0.63	0.85	0.45	0.39	0.48	0.95		0.40	0.35		0.80	0.45	0.32
P_2O_5	0.78	1.74	1.70	0.17	1.41	0.24	0.13	0.54	0.69		0.31	0.56		1.14	0.48	0.40
CO_2	0.01				0.03		0.02	0.01	0.05						0.20	0.53
S		0.04														
BaO																
Total	99.65	99.72	100.34	99.95	99.73	99.88	100.33	99.74	99.76	100.22	99.84	99.87	96.65	99.73	100.09	99.81

* Published since 1899
See footnotes p. 326.

Table V-7 (cont'd). Chemical analyses of Keweenawan intrusive rocks (exclusive of Duluth Complex).

	Lester River sill						Northland sill					Pigeon Point sill		Logan sills	
	17	18	19	20	21	22	23	24	25	26	27	28	29	30	31
SiO_2	50.46	51.62	54.86	56.76	58.88	61.22	47.50	58.01	58.06	62.85	63.33	49.18	61.09	50.04	51.00
TiO_2	1.64	2.26	1.97	1.62	1.38	1.62	3.74	2.14	2.16	1.59	1.55	1.09		3.76	1.82
Al_2O_3	16.90	11.39	11.92	12.55	11.84	12.30	12.94	13.63	13.09	13.10	12.99	19.01	15.34	11.70	18.93
Fe_2O_3	4.68	4.94	5.04	4.70	6.82	9.51	3.94	4.59	3.34	2.71	2.87	0.89	5.74	2.28	1.19
FeO	6.47	12.94	10.32	8.04	5.63	2.93	11.52	5.19	7.22	5.56	5.12	7.79	3.69	13.51	9.54
MnO	0.17	0.30	0.26	0.33	0.20	0.28	0.22	0.19	0.17	0.14	0.16	0.51		0.15	
MgO	4.01	3.00	1.53	3.57	2.06	1.36	5.62	3.05	2.05	1.40	2.21	6.42	1.33	4.20	4.04
CaO	8.45	6.36	5.84	1.75	3.63	1.46	8.38	3.63	4.78	3.31	2.05	9.12	3.10	7.16	9.46
Na_2O	2.82	2.59	2.96	2.76	2.82	3.24	2.39	4.23	3.48	3.51	3.97	3.32	3.41	3.47	1.46
K_2O	1.21	2.03	2.25	3.09	4.07	3.84	1.07	3.30	3.02	4.02	3.52	0.82	3.65	1.03	1.20
H_2O+	1.63	1.32	1.68	3.06	1.18	1.05	1.31	1.02	1.48	1.08	1.25	2.06	1.80	1.28	1.18
H_2O-	1.28	0.78	0.53	1.16	0.58	0.32	0.68	0.52	0.39	0.16	0.58			0.07	0.16
P_2O_5	0.42	0.54	0.66	0.55	0.41	0.54	0.69	0.41	0.70	0.40	0.37			0.47	
CO_2					0.58	0.11								0.25	
S														0.11	0.02
BaO														0.02	0.15
Total	100.14	100.07	99.82	99.94	100.08	99.78	100.00	99.91	99.94	99.83	99.97	100.21	99.15	99.60	100.15

See footnotes p. 326.

Table V-7 (cont'd). Chemical analyses of Keweenawan intrusive rocks (exclusive of Duluth Complex).

	Anorthosites						Dikes					Miscellaneous			
	32	33	34	35	36	37	38	39	40	41	42	43	44	45	46
SiO_2	47.05	49.78	50.59	50.68	51.45	51.54	48.10	49.18	49.21	49.34	57.31	46.88	52.02	52.82	53.91
TiO_2		0.00	0.05	0.07	0.15	0.12	1.49	2.99	2.83	3.16	1.62	1.11	2.17	2.15	1.22
Al_2O_3	32.03	29.37	29.88	30.67	28.47	28.87	15.88	13.82	14.24	13.03	13.39	20.98	14.15	13.67	17.25
Fe_2O_3	2.01	0.34	0.35	0.21	0.80	1.06	2.23	2.46	2.36	2.50	4.23	3.32	6.07	7.66	2.50
FeO		0.60	0.49	0.29	1.00	0.22	10.16	10.99	10.59	13.74	8.05	5.56	7.76	6.61	4.87
MnO		0.08	0.01	0.01	0.02	0.01	0.18	0.20	0.19	0.51	0.17	0.12	0.17	0.15	0.09
MgO	0.15	1.01	1.18	0.42	1.07	0.39	7.54	5.44	5.73	3.64	1.89	4.74	4.40	3.66	2.38
CaO	15.85	11.86	13.08	13.79	12.08	12.70	8.82	9.16	9.14	7.40	3.61	11.15	7.29	6.94	3.38
Na_2O	1.00	4.39	3.48	3.40	3.90	3.72	3.26	2.72	2.72	4.55	3.88	2.49	2.92	2.94	6.08
K_2O	0.05	0.46	0.12	0.09	0.33	0.27	0.49	0.98	0.97	1.57	2.66	0.29	1.03	1.19	3.21
H_2O+	1.36		0.79	0.50	0.63	0.82	1.44	1.04	1.00	0.69	2.05	1.45	1.48	1.67	1.93
H_2O-		1.76	0.22	0.03	0.18	0.08	n.d.	0.20	0.19	0.24	0.61	1.59			0.33
P_2O_5			0.05	0.16	0.04	0.09	0.16	0.56	0.50	0.15	0.59	0.15	0.30	0.42	0.40
CO_2				0.03		0.03	0.14	0.04	0.02	0.96		0.12	0.00	0.01	2.12
S								0.09	0.07	0.12		0.02			0.04
BaO										0.03					0.10
Total	99.50	99.65	100.29	100.35	100.12	99.92	99.89	99.87	99.76	101.69	100.06	99.97	99.76	99.89	99.84

See footnotes p. 326.

1. Hortonolite-ferrogabbro; L. Superior shore SE¼NE¼ sec. 14, T. 55 N., R. 8 W., Lake Co.; anal. D. Thaemlitz (Gehman, 1957, unpub. Ph.D. thesis, Univ. Minn., table 6, no. 9)
2. Diabasic gabbro; Hwy. 61, SW¼SW¼ sec. 14, T. 55 N., R. 8 W., Lake Co.; anal. R. B. Ellestad (Grout and Schwartz, 1939, table 2, no. 2)
3. Iron-rich diabase (M 3174); Hwy. 61, ⅝ mile SW of settlement, Beaver Bay, Lake Co. (NE¼SW¼ sec. 14, T. 55 N., R. 8 W.); anal. J. W. Scoon (Muir, 1954, table 1, no. 2)
4. Diabase with olivine spots; L. Superior shore, SE¼SW¼ sec. 6, T. 55 N., R. 7 W., Lake Co.; anal. T. Kameda (Grout and Schwartz, 1939, table 2, no. 3)
5. Ferrohortonolite-ferrogabbro; L. Superior shore, NW¼NW¼ sec. 13, T. 55 N., R. 8 W., Lake Co.; anal. D. Thaemlitz (Gehman, 1957, *op. cit.*, table 6, no. 11)
6. Mottled diabase; Hwy. 61, SW¼SW¼ sec. 28, T. 55 N., R. 8 W., Lake Co.; anal. R. B. Ellestad (Grout and Schwartz, 1939, table 2, no. 1)
7. Fine-gr. border of Beaver River gabbro; Reserve Mining Co., Silver Bay, T. 55 N., R. 8 W., Lake Co.; anal. D. Thaemlitz (Gehman, 1957, *op. cit.*, table 1, no. 1)
8. Fayalite ferrogabbro; L. Superior shore, NW¼NW¼ sec. 13, T. 55 N., R. 8 W., Lake Co.; anal. C. O. Ingamells (Gehman, 1957, *op. cit.*, table 6, no. 12)
9. Black Bay gabbro; cut near diesel repair shop of Reserve Mining Co., Silver Bay, Lake Co., T. 55 N., R. 8 W.; anal. D. Thaemlitz (Gehman, 1957, *op. cit.*, table 11, no. 18)
10. Diabase; E. of Baptism R., Lake Co.; anal. J. A. Dodge and C. F. Sidener (Winchell and Grant, 1900, p. 226, no. 156)
11. "Intermediate rock;" near Finland, SE¼SE¼ sec. 8, T. 57 N., R. 7 W., Lake Co.; anal. R. B. Ellestad (Grout and Schwartz, 1939, table 2, no. 5)
12. Ophitic diabase; L. Superior shore at 18th Ave. E., Duluth, St. Louis Co.; anal. S. S. Goldich (Schwartz and Sandberg, 1940, table 1, no. 1)
13. "Traprock, cupriferous;" Tischer's Ck. quarry, Duluth, St. Louis Co.; anal. J. A. Dodge (Winchell and Grant, 1900, p. 149, no. 57)
14. Diabase; L. Superior shore at 22nd Ave. E., Duluth, St. Louis Co.; anal. R. W. Perlich (Schwartz and Sandberg, 1940, table 1, no. 3)
15. "Red rock" within a few feet of overlying rhyolite; Tischer Ck. near 2nd St. E., Duluth, St. Louis Co.; anal. R. W. Perlich (*ibid.*, table 1, no. 6)
16. "Intermediate red rock;" McLean quarry near L. Superior shore between 24th and 25th Ave. E., Duluth, St. Louis Co.; anal. R. W. Perlich (*ibid.*, table 1, no. 4)
17. Diabase; L. Superior shore 4,600 ft. E. of Lester R., SW¼NE¼ sec. 4, T. 50 N., R. 13 W., Duluth, St. Louis Co.; anal. S. S. Goldich (*ibid.*, table 1, no. 15)
18. Massive, dark diabase; Hwy. 61 (now St. Louis Co. 61) 2,000 ft. E. of Lester R., Duluth, St. Louis Co.; anal. L. A. Danielson (*ibid.*, table 1, no. 14)
19. Massive, dark, medium-gr. diabase; Hwy. 61, 1,000 ft. E. of Lester R., Duluth, St. Louis Co.; anal. S. S. Goldich (*ibid.*, table 1, no. 13)
20. Reddish diabase; L. Superior shore, NE¼NE¼ sec. 4, T. 50 N., R. 13 W., Duluth, St. Louis Co.; anal. S. S. Goldich (*ibid.*, table 1, no. 18)
21. "Red rock;" main facies, L. Superior shore, SE¼NE¼ sec. 4, T. 50 N., R. 13 W., Duluth, St. Louis Co.; anal. S. S. Goldich (*ibid.*, table 1, no. 17)
22. "Red rock;" along fractures in reddish diabase, L. Superior shore, NE¼NE¼ sec. 4, T. 50 N., R. 13 W., Duluth, St. Louis Co.; anal. S. S. Goldich (*ibid.*, table 1, no. 19)
23. Diabase; L. Superior shore at 30th Ave. E., Duluth, St. Louis Co.; anal. S. S. Goldich (*ibid.*, table 1, no. 8)
24. "Red rock;" Skyline Blvd. at 41st Ave. E. proj., Duluth, St. Louis Co.; anal. S. S. Goldich (*ibid.*, table 1, no. 10)
25. Reddish diabase; Amity Ck., NW¼SW¼ sec. 32, T. 51 N., R. 13 W., Duluth, St. Louis Co.; anal. S. S. Goldich (*ibid.*, table 1, no. 9)
26. "Intermediate red rock;" ¼ mi. W. of center sec. 33, T. 51 N., R. 13 W., Duluth, St. Louis Co.; anal. R. W. Perlich (*ibid.*, table 1, no. 11)
27. "Red rock;" Amity Ck. above bridge at Maxwell Rd., SW¼NE¼ sec. 32, T. 51 N., R. 13 W., Duluth, St. Louis Co.; anal. S. S. Goldich (Schwartz and Sandberg, 1940, table 1, no. 12)
28. Olivine diabase; Pigeon Point, Cook Co.; anal. W. F. Hillebrand (Winchell, 1900, p. 213, no. 1)
29. Med.gr., miarolitic red granite; N. side Pigeon Point, SE¼NW¼ sec. 26, T. 64 N., R. 7 E., Cook Co.; anal. J. A. Dodge and C. F. Sidener (Winchell and Grant, 1900, p. 303)
30. Chilled phase of diabase near top of sill, sec. 25, T. 65 N., R. 2 W., Cook Co.; anal. R. B. Ellestad (Grout and Schwartz, 1939, table 8, no. 1); total includes 0.10% Cl., 0.00% ZrO_2
31. Quartz diabase; SW¼NW¼ sec. 35, T. 65 N., R. 2 W., Cook Co.; anal. H. F. Kendall (*ibid.*, table 8, no. 3)
32. Anorthosite; foot of Caribou Peak, St. Louis Co. (?); anal. C. F. Sidener (Grout, 1918d, p. 650, no. 24)
33. "Plagioclasyte;" Carlton Peak, Tofte, T. 59 N., R. 4 W., Cook Co.; anal. A. N. Winchell (Winchell, 1900, p. 281, no. 1)
34. Very coarse anorthosite; cove in L. Superior shore, sec. 5, T. 56 N., R. 7 W., Lake Co.; anal. W. T. Kameda (Grout and Schwartz, 1939, table 4, no. 9)
35. Anorthosite; Split Rock quarry, sec. 5, T. 54 N., R. 8 W., Lake Co.; anal. R. B. Ellestad (*ibid.*, table 4, no. 7)
36. Black anorthosite; Beaver Bay, sec. 12, T. 55 N., R. 8 W., Lake Co.; anal. W. T. Kameda (*ibid.*, table 4, no. 8)
37. Brown anorthosite; Hwy. 61, sec. 1, T. 56 N., R. 7 W., Lake Co.; anal. R. B. Ellestad (*ibid.*, table 4, no. 6)
38. Dark green, massive, fine-gr. olivine diabase dike; NW shore of W. end Moose Lake, NE¼ sec. 36, T. 64 N., R. 10 W., Lake Co.; anal. K. Ramlal (Green, 1970a, table 2, M-7129); intrudes Knife Lake Gp.
39. Basaltic chilled zone of microgabbro dike (M 3744-1); near TV tower, sec. 28, T. 50 N., R. 14 W., Duluth, St. Louis Co.; anal. E. Oslund (Taylor, 1964, table 14, no. XXVI); intrudes Duluth Complex.
40. Microgabbro dike (M 3744-2); near TV tower, sec. 28, T. 50 N., R. 14 W., Duluth, St. Louis Co.; anal. E. Oslund (*ibid.*, table 14, no. XXVII); intrudes Duluth Complex
41. Olivine diabase dike; Carlton Co.; anal. W. H. Truesdale (Grout, 1910a, table 8, no. 1); intrudes Thomson Fm.
42. Aphanitic, brown trachyandesite dike (H-4); Hwy. 61 at old road jct., NE¼NW¼ sec. 12, T. 62 N., R. 4 E., Cook Co.; anal. T. Konda (cuts Lower Keweenawan Hovland lavas)
43. Diabase; Lafayette Bluff, sec. 12, T. 53 N., R. 10 W., Lake Co.; anal. S. S. Goldich (Grout and Schwartz, 1939, table 2, no. 4)
44. Fine- to med.-gr., brown trachybasalt sill; S. of Devil Track Lake, SE¼NE¼ sec. 8, T. 61 N., R. 1 W., Cook Co.; anal. K. Ohta (Bally Creek sill)
45. Fine- to med.-gr., brown, nonporphyritic lower part of large trachybasalt sill; W corner of Hill NE of Leveaux Mtn., SE¼NW¼ sec. 1, T. 59 N., R. 4 W.; anal. K. Ohta (Leveaux porphyry sill)
46. Coarse-gr., red syenite; (drill hole) NW¼NE¼ sec. 2, T. 58 N., R. 15 W., Aurora, St. Louis Co.; anal. E. Oslund (White, 1954, p. 65) ("Aurora sill")

and concentration of felsic differentiates up-dip within a single larger body, or by separate intrusion of the granophyric magma from some outside source into the still-hot diabasic sill.

The Northland and Lester River sills have been studied by Schwartz and Sandberg (1940). The Northland sill is at least 6 miles long and varies in thickness from 30 to approximately 1,030 feet. It forms a major scarp in the East End of Duluth (Lakeside district). It also contains intermediate and felsic differentiates, but they are less regularly distributed than in the Endion sill. The Lester River sill forms a conspicuous ridge still farther east, and can be traced for about 17 miles; it is approximately 1,000 feet thick. Its major phase is ophitic olivine gabbro or diabase, but it also contains local irregular zones of felsic and intermediate differentiates. Small amounts of interstitial granophyre are common in the diabase of both sills. Schwartz and Sandberg concluded that the "red rocks" of these sills are magmatic differentiates of their host gabbroic magmas.

At Stony Point (sec. 1, T. 51 N., R. 12 W.), another gabbroic sill intrudes the lavas; its thickness is about 1,600 feet, and its exposed length is about 3 miles (Sandberg, 1938). The sill appears to be conformable with the basalts beneath it at Stony Point, where it strikes slightly east of north, but it swings to the east and crosscuts the flows at Knife River, where it forms Granite Point and Knife Island as it passes beneath Lake Superior. The upper contact appears to be irregular, beneath ophitic basalts. The sill is an olivine diabase or gabbro that contains several local pegmatitic patches and an anorthosite xenolith, but lacks appreciable felsic differentiates.

Approximately 5 miles east of Two Harbors (fig. V-18), Silver Creek Cliff, which rises abruptly above the volcanic terrane, is held up by another large, slightly crosscutting olivine gabbro sill. This sill trends north by east for 2 miles, then swings to the east and then back south to form Lafayette Bluff and Encampment Island. Diabase bodies to the southwest (SW¼ sec. 21, T. 53 N., R. 10 W.) and to the northeast across Crow Creek (SE¼ sec. 1, T. 53 N., R. 10 W., and SW¼ sec. 6, T. 53 N., R. 9 W.) are probably outliers or are connected with the main sill. According to R. M. Grogan (1940, *op. cit.*), the sill at Silver Creek Cliff dips 10°-20° W., transgressing the easterly-dipping flows, whereas at Lafayette Bluff it dips eastward at a high angle. Just east of Crow Creek, it has gently-dipping layering. Its contacts are complex and some are faulted, and in some places the intrusion evidently has forcefully deformed the overlying lavas. The sill forms a scarp 200 feet high along much of its length; Grogan estimated it to be 600 feet thick at Lafayette Point, but it is probably thinner over most of its extent. It is an olivine gabbro that has little or no segregated felsic material. Near the base in Silver Creek Cliff some layering can be seen, the individual layers ranging in thickness from an inch to two feet. Evidently, the base has remelted some of the rhyolite flow immediately beneath it at this locality, and basaltic dikes project downward into the flow. The upper parts of the sill contain scattered amygdules in some outcrops. Plagioclase phenocrysts as much as 2 inches long are common in some parts of the sill, especially in the eastern exposures.

Beaver Bay Complex

For several miles on either side of Beaver Bay, in Lake County, a complex series of intrusive rocks is exposed that range in composition from troctolite to granite (fig. V-20). They have been studied by Grout and Schwartz (1939) and, more recently, the southwesternmost part was studied by H. M. Gehman (1957, unpub. Ph.D. thesis, Univ. Minn.). I have done reconnaissance in the eastern and northern parts of the complex. Pyroxenes from the complex have been analyzed by Muir (1954) and Konda (1970). Inliers of lavas appear at several places within the complex. Although its lower (western) margin is rather abrupt for about 11 miles inland from the Lake Superior shore, its relations with the underlying flows and Duluth Complex to the north are obscured by a thick cover of glacial drift. There is no question that a continuous series of mafic intrusive bodies connects the Duluth Complex proper, in the northern half of the Cramer 15-minute quadrangle, with the Beaver Bay Complex, as mapped near the Lake Superior shore, and an arbitrary contact tentatively is placed between the two (see fig. V-23). I here describe the intrusive rocks south of the middle of the Cramer quadrangle (S. edge of T. 60 N., approx. lat. 47°38′ N.), whereas Davidson (this chapter), discusses those to the north in his discussion of the eastern part of the Duluth Complex. From the vicinity of Silver Bay to Illgen City, the Beaver Bay Complex penetrates and deforms the lavas along the Lake Superior shore; north of Illgen City, it passes inland, forming highlands that extend north-northeast into the Cramer quadrangle.

The dominant rock type of the Beaver Bay Complex is medium- to coarse-grained, ophitic olivine gabbro; the smaller bodies are generally fine- to medium-grained diabase. This rock type, informally termed the "Beaver River gabbro" by Gehman (1957, *op. cit.*), underlies most of the southern part of the complex, and consists of thick sill-like and less regular bodies, which extend from Split Rock Point northeastward through Silver Bay, Finland, and Cramer. In many exposures it contains angular to rounded blocks of light-colored anorthosite that range in size from single xenocrysts to blocks more than a quarter of a mile across (Grout and Schwartz, 1939, p. 50). According to Grout and Schwartz (1939) the anorthosites contain at most a small percentage of olivine, augite, and/or interstitial zeolite (most commonly thomsonite). The plagioclase varies in composition from one inclusion to another, but is in all cases either labradorite or bytownite. The diabase that encloses the anorthosite commonly shows intrusive relations to it, and the blocks are thought to have been rafted upward from some deeper source, most likely from the uppermost mantle or lower crust, although their origin is still unclear (Phinney, 1968).

Many local variants of olivine gabbro occur in the complex. Plagioclase phenocrysts, approximately 1 cm across, are widespread but not abundant in many of the ophitic diabases. At Beaver Bay and Silver Bay, Gehman (1957, *op. cit.*) has described two round plugs of ferrogabbro ("Beaver Bay ferrogabbro"), one or two miles in diameter, that are each surrounded by an uneven-textured gabbroic ring dike and that cut the dominant ophitic olivine diabase. The ferrogabbros have a horizontal foliation and cryptic layering, with plagioclase ranging from An_{48} to An_{28} and olivine ranging from Fo_{34} to Fo_{1} (Gehman, 1957, *op. cit.*, p. 57-58). Discontinuous flow-banding can be seen in olivine gabbros in the great north-trending ridge east of Cramer (sec. 10, T. 58 N., R. 6 W.), in a large sill north of Illgen City (SE¼ sec. 26, T. 57 N., R. 7 W.), and in a small troctolite knob near Finland (NE¼ sec. 20, T. 47 N., R. 7 W.). A layered intrusion of olivine gabbro occurs near Sonju Lake, northeast of Finland (secs. 21, 22, 27, 28 and 29, of T. 58 N., R. 7 W.). Near Shoepack and Crooked Lakes, north of Cramer (secs. 4 to 9, T. 59 N., R. 6 W.), is a body of layered olivine gabbro or troctolite that appears to be continuous with the uppermost part of the Duluth Complex.

In some intrusions of the Beaver Bay Complex, interstitial granophyre is a significant component. In such bodies the olivine, if present, is generally altered, there is commonly some hornblende as well as augite, apatite is abundant, and the plagioclase is strongly zoned and has andesine or oligoclase rims. Intrusions of such augitic granodiorites, syenodiorites, or syenogabbros are found northeast of Illgen City (sec. 1, T. 56 N., R. 7 W., fig. V-21A), north of Little Marais (sec. 4, T. 57 N., R. 6 W.), near Blesner Lake and Blesner Creek (secs. 19 and 29, T. 58 N., R. 6 W.), and in the high ridge northwest of Illgen City (sec. 3, T. 56 N., R. 7 W.). The large (up to 20×20 cm), skeletal iron-rich pyroxenes and olivines in some of these bodies are common and distinctive.

Granitic or adamellitic intrusions also occur in the Beaver Bay Complex. Generally brick red from their hematite-clouded feldspars, these "red rocks" have granophyric combined with porphyritic and hypidiomorphic textures. Locally, such as just west of Finland, miarolitic cavities are common, whereas elsewhere (such as northeast of Finland) the rocks approach a rhyolitic texture; both textures suggest a relatively shallow depth of intrusion. The largest bodies occur north and west of Finland (secs. 4, 5, 6, 7, 8, 17, 18 and 19 of T. 57 N., R. 7 W.), and farther north (sec.

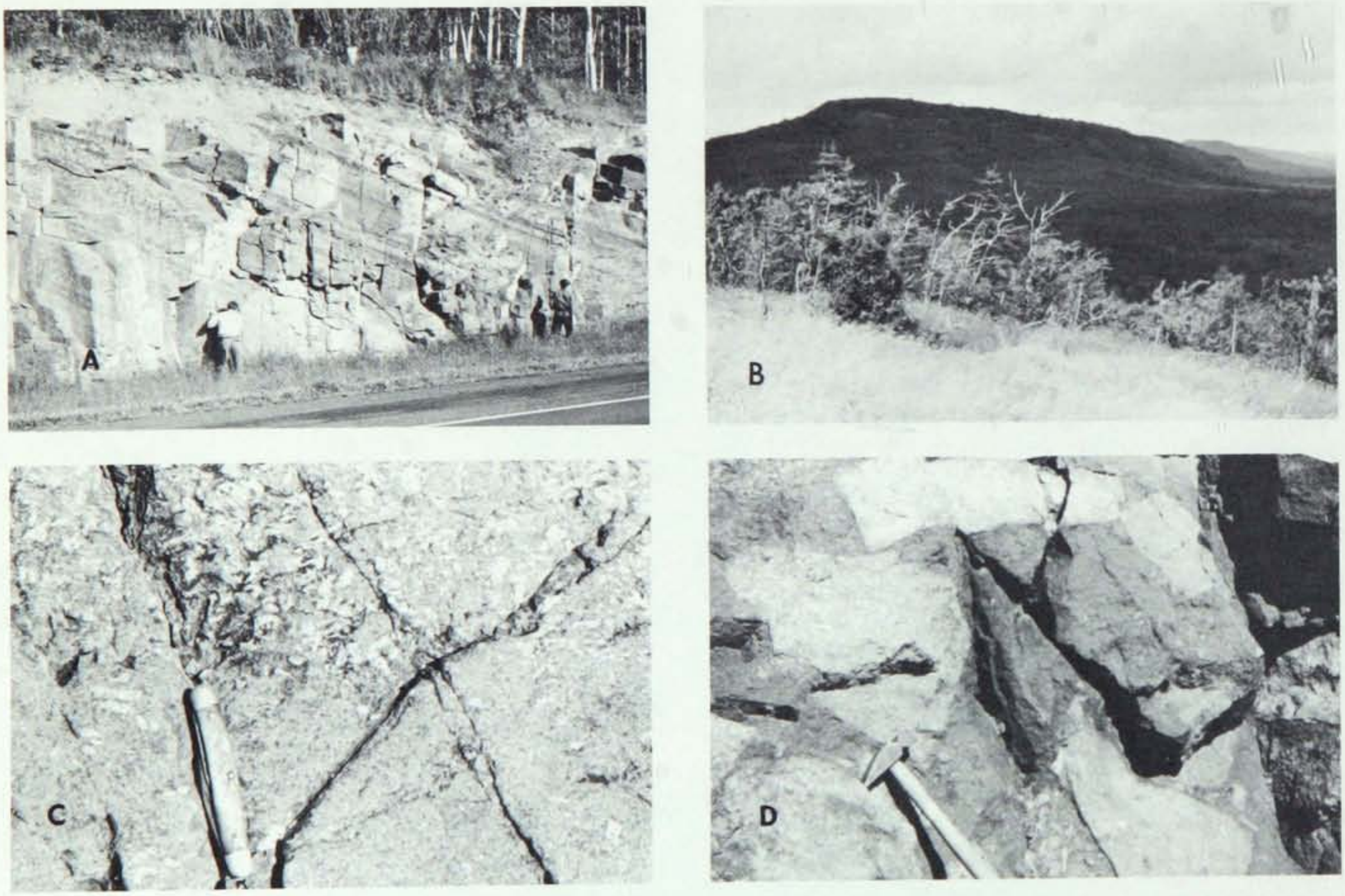

Figure V-21. Photographs of Keweenawan intrusive rocks. A, layered syenogabbro of Beaver Bay Complex, cut in U.S. Highway 61, 1.5 miles northeast of Illgen City, Lake County; B, Leveaux Mountain, north of Tofte, Cook County. Hill from which photo was taken (Leveaux Mtn.) and other hills in right distance are held up by large, dipping sill of Leveaux trachybasalt porphyry, in which plagioclase phenocrysts have floated to tops; C, contact between richly and sparsely porphyritic (upper and lower, respectively) parts of Leveaux trachybasalt sill, as seen on Bear Island, Taconite Harbor. There is no evidence of intrusion of one type by the other, and the plagioclase phenocrysts are inferred to have floated; D, diabase choked with inclusions of anorthosite and gabbro, Carlton Peak quarry, Tofte, Cook County.

16, T. 58 N., R. 7 W.). Smaller red granitic dikes and irregular bodies that cut the other rocks of the complex are widely distributed.

Leveaux Porphyry

A large differentiated sill of porphyritic trachybasalt, informally named the Leveaux porphyry, slightly transgresses the basalts in the area between Taconite Harbor and the Cascade River in southwestern Cook County (figs. V-6 and V-7). It forms large cuesta-like ridges (Leveaux and Eagle Mountains) behind Tofte and Lutsen (fig. V-21B) as well as two hills near the Tofte airstrip (secs. 10, 11, 14 and 15, T. 59 N., R. 4 W.); its southwestern extension has been eroded away. Outliers of the sill form Bear and Gull Islands at Taconite Harbor. Another small outlier is present 4½ miles north of Tofte in sec. 33, T. 60 N., R. 4 W. Its total known length is 24 miles. However, if a very similar sill outlier which forms a knob known as Pincushion Mountain just east of Grand Marais (SE¼ sec. 11, T. 61 N., R. 1 E.) is part of the same intrusion, its minimum length is 36 miles. Neither its top nor its base is exposed, but it has a minimum thickness of 150 feet at Leveaux Mountain and a possible thickness of 350 feet at Eagle Mountain. It is a brown, granular, iron-rich trachybasalt containing pigeonite and augite, large magnetite crystals and, in the upper half of the sill, abundant (about 40 percent), blocky, 1-2 cm labradorite phenocrysts that are assumed to have floated. The contact between the porphyritic and non-porphyritic parts of the sill—which can be readily examined on Bear Island and on the southwest end of the hill just northeast of Leveaux Mountain and the Onion River—is gradational over a distance of one to three feet and lacks any evidence of chilling or intrusive relations. The nonporphyritic, lower part of the sill (no. T-36, table V-7) is the same rock as the groundmass of the upper porphyry (fig. V-21C). On the hilltop on the peninsula in Caribou Lake (NE¼ sec. 2, T. 60 N., R. 3 W.), the contact is gradational over 3 to 10 feet and is vertical. Here, it may have been disturbed by the intrusion of the nearby diabase.

Diabases of the Tofte-Lutsen Area

A few miles inland from the shore, behind Tofte and Lutsen, is a group of irregular hills underlain by gabbroic and diabasic intrusions. These extend from just west of the Temperance River (sec. 30, T. 59 N., R. 4 W.) northward for 6 miles along the east side of the river to Six Mile Creek, and thence northeastward behind the Leveaux porphyry sill past the Poplar River. The complex continues past Caribou Lake and probably crosses the Cascade River to merge with diabase sills of the Grand Marais area, mentioned below.

Most exposures appear to be sills (or one large sill) and consist of massive, ophitic olivine diabase that has rare plagioclase phenocrysts. Locally, banded gabbros occur, such as in the road cut just west of Temperance River in the NW¼ sec. 30, T. 59 N., R. 4 W. and a cut southwest of Caribou Lake, in the NE¼ sec. 11, T. 60 N., R. 3 W. One of the most notable parts of this complex is Carlton Peak, at Tofte, which is held up largely by massive inclusions of anorthosite similar to those in the Beaver Bay Complex (see Grout and Schwartz, 1939, p. 64-67). Rising 924 feet above the level of Lake Superior, this hill has recently been quarried for riprap, and the relations between the anorthosite blocks and host diabase and gabbro are excellently exposed. At one place in the quarry, the diabase is choked with angular blocks of a variety of coarser anorthositic gabbros and anorthosites (fig. V-21D).

Grand Marais Intrusions

At Grand Marais, the lavas, especially the thick rhyolite flow that underlies much of the town, are intruded in a complex manner by ophitic olivine diabase, which is exposed in many places along the Lake Superior shore just east of the harbor. Five Mile Rock, five miles east of the harbor, is also made up of diabase, and probably is an extension of the complex. The Grand Marais diabase passes westward beneath the Grand Marais rhyolite to form a major cuesta (including Murphy Mountain) that extends to the Cascade River, which it crosses in sec. 24, T. 61 N., R. 2 W. From there it merges into a branch of the Tofte-Lutsen diabase complex just described.

North of Grand Marais, a long, low strike-ridge is underlain by another tabular mafic body that probably is a sill at least 100 feet thick. It forms the 9-mile-long ridge just south of Little Devil Track River and Bally Creek between the Gunflint Trail (Cook Co. Hwy. 12) and the Cascade River. The contacts of the sill (?) are not exposed. The rock has a hypidiomorphic to diabasic texture but is not ophitic like the normal diabases, and it contains abundant iron-rich pigeonite and granophyre. It is here informally called the Bally Creek trachybasalt.

The knob called Pincushion Mountain, 3 miles northeast of Grand Marais in the SE¼ sec. 11, T. 61 N., R. 1 E., also is held up by a nonophitic trachybasalt sill, which may be an outlier of the Bally Creek sill. However, it contains very abundant blocky plagioclase phenocrysts and it may be an outlier of the very similar Leveaux porphyry.

Northeast of Grand Marais, several isolated hills in secs. 24 and 25, T. 62 N., R. 1 E. are underlain by diabase that appears to trend in a west-northwesterly direction, and may constitute another minor sill.

Diabases of Central Cook County

Several large sills and dikes, each a few hundred feet thick, intrude the lavas in the area between the upper Cascade River (T. 62 N., R. 2 W.), Crescent Lake (T. 62 N., R. 4 W.), and White Pine Lake (T. 61 N., R. 3 W.). One large diabase intrusion has an arcuate map pattern, and forms the western and southern shore of Crescent Lake. Another, longer diabase trends northward from a point west of Mistletoe Lake (W. side of T. 61 N., R. 3 W.) to just southeast of Crescent and Lichen Lakes, and then swings east to pass north of Little Cascade Lake and Swamp Lake to the Cascade River; its inferred length is at least 14 miles. A third major dike-like unit trends northward to northeastward from east of White Pine Lake to east of Tait Lake (T. 62 N., R. 3 W.). Where studied, all the rocks are ophitic olivine gabbros.

Hovland Diabase Complex

In the vicinity of Hovland (Ts. 62 and 63 N., Rs. 3 and 4 E.), nearly all exposures consist of mafic intrusive rocks. Earlier published works (for example Grout and others, 1959) considered some of these an extension of the Duluth Complex eastward to Lake Superior, but in a more recent study Jones (1963, *op. cit.*) concluded that the Duluth Complex does not extend east of the Brule River and that the intrusive rocks near Hovland belong to five separable units. To the east, the Hovland complex appears to merge with rocks assigned to the Reservation River diabase complex and with the large sills and dikes of the Logan intrusions. The Hovland complex (Middle Keweenawan according to its normal magnetic polarity) intrudes the reversed-polarity Hovland lavas of Early Keweenawan age (Green and Books, 1972).

The Hovland sill is the southernmost of the units mapped by Jones; it forms the shoreline ledges and upland outcrops between the Brule River and Chicago Bay. An underlying smaller unit of similar lithology that may be connected to it continues eastward across the northern parts of sections 19 and 20 (T. 62 N., R. 4 E.). The sill strikes about N. 80° E. and dips 10° S. and, although its top is eroded away, it is at least 500 feet thick. It is a remarkable, brown- or red-weathering ferrogabbro, the lower part of which has a pronounced igneous lamination and the upper part of which is a coarser grained, unlaminated fayalite-syenodiorite. The sill crosscuts lavas at the Brule River in sec. 27, T. 62 N., R. 3 E.

A smaller, sill-like ferrogabbro underlies the Hovland lookout hill in sec. 6, T. 62 N., R. 4 E. and sec. 1, T. 62 N., R. 3 E. From the topography and igneous lamination, the "Lookout sill" appears to strike about N. 80° W. and dip 15° S. and must be at least 200 feet thick. It is similar to the Hovland sill in that, in most samples studied, it contains abundant interstitial granophyre. A possible lower extension of this unit, south of Moosehorn Lake on the south edge of sec. 36, T. 63 N., R. 3 E., includes some highly granophyric red monzonite.

A larger, irregular area of intrusive rocks north of the sills described above has been called the "Tom Lake-Swamp River unit" by Jones (1963, *op. cit.* p. 70). It underlies the low ridge along the south side of Tom Lake, passes eastward through Moosehorn Lake, and thence northeastward across the Swamp River to large highlands in secs. 14 and 15, T. 63 N., R. 4 E. The eastern part appears to be sill-like, but the western parts probably are more discordant, and sharp, crosscutting contacts have been observed. Nearly all samples are ophitic olivine gabbro or diabase.

Two high, northeasterly-trending ridges northeast of Hovland are held up by thick gabbroic dikes, which are part of what Jones (1963, *op. cit.*) called the "Moose Valley-Farquhar unit." The northwesternmost dike (in secs. 33 and 34, T. 63 N., R. 4 E. and secs. 4 and 5 of T. 62 N., R. 4 E.) is a medium- to coarse-grained, fresh, ophitic olivine gabbro which appears to range in thickness from 1,000 to 3,000 feet. Some samples have poikilitic inverted pigeonite as well as augite. The other large dike—to the southeast across Moose Valley (in sec. 35, T. 63 N., R. 4 E. and secs. 2, 3, 9 and 10, T. 62 N., R. 4 E.), including Farquhar Peak—is more feldspathic, contains more interstitial granophyre and/or alkali feldspar, and is more highly altered than the northwestern dike. It shows strong shearing and retrograde alteration in Carlson Creek, in section 10. Both of these large dikes appear to pass eastward into the Reservation River diabase complex, but relative ages have not been determined.

The relationships of the other exposed mafic intrusive rocks, particularly between Hovland and Big Bay, are obscure. Most are olivine gabbros.

Reservation River Diabase Complex

A large area of gabbroic rocks underlies the Lake Superior shore zone at and east of the Reservation River, mostly in secs. 27, 28, 29, 31, 32, 33 and 34, T. 63 N., R. 5 E. and secs. 4, 5 and 6, T. 62 N., R. 5 E. Along their western and northwestern sides the gabbroic rocks cut the Hovland lavas and apparently merge with the Hovland diabase complex. On the east, they intrude the uppermost part of the Grand Portage lavas and may pass into the great northeasterly-trending dikes of the Logan intrusions. Although crosscutting, the rocks appear to be generally sill-like in character; foliated gabbro with low dips can be seen in some outcrops. The dominant rock type is ophitic olivine gabbro, which has a conspicuous luster-mottling. A larger area in the northern parts of sections 28 and 29, separated apparently from the main gabbro mass, is composed of a reddish-brown-weathering syenogabbro or granogabbro, which contains distinctive long, curved augite crystals as well as about 20 percent granophyre. A coarse granogabbro dike, apparently connected with this complex, is crossed by U.S. Highway 61, and forms a point on the lakeshore at the west side of sec. 12, T. 62 N., R. 4 E., about 1.3 miles southwest of the Reservation River. It is 100 to 200 feet thick and contains 20 percent interstitial quartz and alkali feldspar. At the road cut it apparently cuts an unrelated, aphanitic trachyandesite dike.

The Reservation River diabase complex has normal magnetic polarity, and thus is Middle Keweenawan, whereas the Grand Portage lavas which it cuts have reversed polarity and are Lower Keweenawan.

Grand Portage Dike Swarm

Basaltic to trachybasaltic dikes are common in the Lower Keweenawan lavas of the Grand Portage area, and similar dikes have been seen on strike to the west, north of Hovland. Generally more resistant than the enclosing lavas, they tend to form low ridges on land and small points projecting into the lake. The great majority of the dikes strike eastward or slightly north of east (between N. 65° E. and N. 70° W.) and dip 68° or more to the north; a few strike about N. 10-20° E. and dip steeply southeast, and others strike about N. 30° W. and dip steeply northeast. Although outcrops are sparse inland from the lakeshore, one large dike has been traced for 1½ miles and others for distances of about a mile. The thicker dikes are 30 to 200 feet thick, and are gray or dark-brown, granular, plagioclase-porphyritic trachybasalts and basalts, whereas the thinner ones (10 feet or less thick) are generally black, fine-grained or aphanitic basalts. Two dikes on Lucille and Brick Islands, east of Grand Portage, are red granophyric-porphyritic rhyolites. All the dikes have chilled contacts. The Grand Portage dike swarm also has reversed magnetic polarization, as do the Hovland lavas (Green and Books, 1972).

Other Dikes

About 40 basaltic dikes were noted by Sandberg (1938) in the succession of lavas between Duluth and Two Harbors, mainly in the lower part; generally, they strike a little east of north and dip steeply west. Taylor (1964) has also described some dikes that cut the Duluth Complex in Duluth. Generally, the dikes in both areas have diabasic texture, chilled margins, and normal basaltic composition. The dikes are the youngest igneous rocks of the area.

Basaltic dikes thought to be of Keweenawan age have been found in many areas beyond the limits of the North Shore Volcanic Group and the Duluth Complex. Although Hanson (1968) and Hanson and Malhotra (1971) have found that some basaltic dikes, particularly in areas other than the northeastern part of the state, are Middle Precambrian in age, there are several dikes that cut older rocks near the main mass of Keweenawan igneous rocks which, on structural and petrographic evidence, are probably Late Precambrian in age. One such dike set was mapped by Green (Green and others, 1966; Green, 1970a) in the Gabbro Lake quadrangle, east of Ely in Lake County. Here, the dikes are unmetamorphosed aphanitic basalts or fine-grained, massive olivine diabases. They range in thickness from 3 inches to 50 feet, and show chilled borders. Most strike about N. 65° E., parallel to the regional trend of the base of the Duluth Complex, but a few trend in other directions. The dikes cut all the Lower Precambrian formations in the area, and are as much as 4½ miles from the Duluth Complex. Whole-rock K-Ar ages ranging from 955 to 1,100 m.y. have been obtained by Hanson and Malhotra (1971) on similar dikes farther east in the Saganaga Lake area.

Near Carlton, in northeastern Pine County, a major swarm of dikes is exposed in the St. Louis River valley (Wright and others, 1970). The dikes trend about N. 30° E., are nearly vertical, and are generally about 30 feet thick, but some are as much as 220 feet thick. They are fine-grained ophitic diabases, in which some of the augite has been retrograded to amphibole, as is the case in the Ely's Peak lavas 5½ miles to the east. Hanson and Malhotra (1971) reported a whole-rock K-Ar age of 1,050 m.y. for one of these dikes. The dikes cut the Middle Precambrian Thomson Formation and at least one cuts the Ely's Peak basalts. Both of these dike sets may have acted as feeders for now-eroded overlying Keweenawan lavas.

ECONOMIC GEOLOGY

Despite the general similarities between the North Shore Volcanic Group and the richly endowed Portage Lake Lava Series of Michigan, no economically profitable mineral deposits have yet been discovered in the Keweenawan lavas and minor intrusive bodies of Minnesota. The important copper-nickel deposits in the Duluth Complex are discussed in another section.

After the Treaty of Fond du Lac in 1826 opened up all the Chippewa country to mineral exploration, and especially after the discovery of the immense copper deposits of the Keweenaw Peninsula soon afterward, the north shore area was the scene of intense prospecting activity, which lasted throughout most of the remainder of the 19th century. Only a small number of mineral occurrences have been discovered, however, and none of these has as yet shown promise for development. The localities and character of many of these occurrences are obscure; this brief report covers the approximately 20 occurrences known to me from a wide variety of sources, including Hall (1889), and Foster (1962, unpub. M.S. thesis, Univ. Missouri; 1963). The aid of the St. Louis County Historical Society is hereby acknowledged, with appreciation.

Most of the known mineralized outcrops are found in two areas: (1) between Duluth and Two Harbors; and (2) between Tofte and Grand Marais (see fig. V-22). In each, the principal metal of value is copper, which occurs either in the metallic state or as sulfides.

The majority of occurrences contain native copper. In these, the copper occurs in much the same way as it does in the deposits on the Keweenaw Peninsula: in veinlets and amygdule fillings in amygdaloidal basalt; as disseminated specks in massive basalt; as flecks in amygdaloidal prehnite; in veins with calcite; in laminae and veinlets in interflow sands; and as thin sheets along joints in mafic intrusions. It is nearly everywhere accompanied by calcite and/or prehnite; other minerals also associated locally with native copper are quartz, chlorite, epidote, laumontite, heulandite, datolite, stilbite, and natrolite. In two prospects northeast of Duluth (sec. 17, T. 51 N., R. 12 W.; sec. 25, T. 52 N., R. 12 W.), masses of copper weighing as much as 15 pounds had been recovered by 1866 (Hall, 1889), but more recent (1929) intensive exploration in the area by the Mining Corporation of Canada (see Schwartz, 1949, p. 130-133) proved unsuccessful. Similar disappointing results were obtained from diamond drilling in 1969 by the New Jersey Zinc Company in similar rocks east of Grand Marais (sec. 8, T. 61 N., R. 2 E.).

A few occurrences of copper sulfides are known from the north shore area. At the trap-rock quarry at Ely's Peak, west of Duluth, small calcite veins in fractures carry rare, small crystals and stringers of pyrite, chalcopyrite, chalcocite, and bornite. Native copper is present elsewhere in the same quarry, in amygdules and veinlets. On a small hill just north of Finland (sec. 17, T. 57 N., R. 7 W.), sulfide minerals occur as disseminations in a red, medium- to coarse-grained syenodiorite. Trace-element analyses of a mineralized sample by the U.S. Geological Survey show 89 ppm copper, 120 ppm zinc, 5 ppm nickel, and 0.6 ppm silver. Near the mouth of the Cascade River, an old prospect worked by H. Mayhew in 1868-1869 is reported to have consisted of stringers of bornite in a four-foot-wide (calcite?) vein (Hall, 1889). Another of Mayhew's prospects farther up the Cascade River now shows only secondary copper stains (Foster, 1962, *op. cit.*, p. 119). One-fourth mile up the canyon of Cutface Creek, 5 miles west of Grand Marais, Foster (1962, *op. cit.*; 1963) reported chalcocite