THE STRATIGRAPHY AND STRUCTURE OF THE MESABI RANGE, MINNESOTA

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THE STRATIGRAPHY AND STRUCTURE OF THE Mesabi Range, Minnesota

BY

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FOREWORD

The Mesabi range of Minnesota, because of its great economic importance, has received comprehensive geologic study in past years. Although there have been several publications on the geology of the Mesabi district, only recently has development of the range proceeded far enough to permit reasonably accurate determination of its detailed structure. In addition, changed concepts of the principles of stratigraphy indicated the desirability of a new study of the Mesabi iron-formation; such a study would use modern concepts in an attempt to explain the reasons for variations in the nature of the formation along its explored length. This fact was first brought to my attention by Professor F. M. Swain of the department of geology at the University of Minnesota.

The problem was suggested to Mr. White as a suitable subject for a thesis for the degree of Doctor of Philosophy. In large measure, his report is built upon the foundation of knowledge established by previous workers, and its aim has been to contribute and interpret additional detailed geologic information of both practical and theoretical value. His field work was done in the summer of 1951 and in six months in 1952 under the auspices of the Minnesota Geological Survey, a department of the University of Minnesota.

The result of this intensive work by Mr. White has contributed greatly to our understanding of the stratigraphy and structure of the greatest iron ore producing district in the world, and it is hoped that the structural maps will prove helpful in continued development of the iron-ore deposits. The Minnesota Geological Survey is indebted to Mr. White for his devotion to this task.

G. M. Schwartz

ACKNOWLEDGMENTS

The research on which this report is based was made possible by successive fellowship grants from Dartmouth College, the Taconite Fellowship Committee of the University of Minnesota, and the National Science Foundation, and by the support of the Minnesota Geological Survey.

Many courtesies, including permission to examine exploration records, drill cores, and mining properties, were extended by the following companies: The Cleveland-Cliffs Iron Co., the Eveleth Fee Office, Great Northern Iron Ore Properties, The M. A. Hanna Co., Inter-State Iron Co., Meriden Iron Co., W. S. Moore Co., the Oliver Iron Mining Division of the U.S. Steel Corporation, Pickands Mather & Co., and the Division of Lands and Minerals, Department of Conservation, State of Minnesota. Great Northern Iron Ore Properties and the Oliver Iron Mining Division also supplied base maps, and Pickands Mather & Co. furnished some partial chemical analyses.

So many of the company officials, mining engineers, and professional geologists gave aid that individual acknowledgments cannot be made here, but Mr. J. F. Wolff, formerly General Mining Engineer of the Oliver Iron Mining Division, must be singled out for his invaluable review of the structure maps. The following members of the University of Minnesota's department of geology also gave aid of various kinds: Professor J. W. Gruner has given much advice and constructive criticism, and suggestions have been made by Professors W. C. Bell, S. S. Goldich, F. F. Grout, G. M. Schwartz, F. M. Swain, G. A. Thiel, and H. E. Wright, Jr. The drafting of maps and figures was done by Dr. N. P. Prokopovich of the Minnesota Geological Survey, who also gave advice on some of the details of the map layouts. Two new chemical analyses were provided by the Mines Experiment Station, University of Minnesota, and one was provided by the Minnesota Geological Survey. To my wife goes the credit for typing the manuscript and for doing some of the other tedious work entailed in the preparation of this report.

D. A. W.

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ABSTRACT

The later Precambrian Animikie group in northeastern Minnesota consists of three sedimentary units: the Pokegama (quartzite), Biwabik (iron-rich rock), and Virginia (argillite) formations. "Mesabi range" designates the preglacial outcrop belt, 1/4 to 3 miles wide and 120 miles long, of the Biwabik formation.

Varieties of iron-rich rock ("taconite") are either granular or slaty and consist dominantly of chert, iron silicates, magnetite, and siderite.

The Lower Cherty, Lower Slaty, Upper Cherty, and Upper Slaty members of the Biwabik formation, which averages 600 feet in thickness, can be further subdivided as shown on a detailed longitudinal stratigraphic section. These members are fairly uniform along most of the range, but only one cherty and one slaty member exist on the Westernmost Mesabi, where the lithic units are intertongued. The areal distribution of rock units on the Westernmost Mesabi is shown on a geologic map. The Biwabik, Pokegama, and Virginia formations are considered conformable. Mesabi rocks probably correlate with those of the Emily district 30 miles away.

Chert, greenalite, minnesotaite, stilpnomelane, magnetite, some hematite, and siderite probably formed either during deposition or diagenesis. The rocks are essentially unmetamorphosed. The Pokegama and Biwabik formations were probably produced by the migration of a series of coexisting environments of deposition during an advance, a retreat, and a second advance of the Animikie sea. The deposits formed, during the retreat, in successive environments seaward from shore, were clastic material, carbonaceous-pyritic mud, chert-siderite, chert-magnetite, and iron silicate. Fine clastics of the Virginia formation, perhaps furnished by an outburst of volcanic activity, spread across the former environments of chemical sedimentation. Possible conditions of iron sedimentation were as follows: derivation of iron and silica by weathering of a low-lying land mass, perhaps under an atmosphere rich in carbon dioxide, and a seasonal climate; tectonic stability; and deposition in a shallow, quiescent epicontinental sea.

The Animikie beds strike about N. 75° E. and commonly dip 6–12° SE. A structure contour map on the base of the Biwabik formation shows numerous small anticlines, synclines, monoclines, and faults. Three major joint sets are present. The few rocks intrusive into the Biwabik formation include diabase sills, the Duluth gabbro, and the Aurora syenite sill. Contact metamorphism by the soda-rich Aurora sill has produced crocidolite in adjacent taconite. Minor internal folding of Animikie beds

seems to be more prevalent where the underlying rocks are volcanic or sedimentary rather than granitic.

The Mesabi range is covered by glacial drift which thickens southward, commonly from 20 to 200 feet, away from a ridge known as the Giants Range. Drift is as much as 500 feet thick over the Westernmost Mesabi. A map of the thickness of drift shows many drift-buried preglacial bedrock valleys that extend from notches in the Giants Range southward across the Mesabi range. Cretaceous iron-ore conglomerates, which at places overlie the Biwabik formation, occur as erosional remnants on bedrock ridges.

The scattered soft iron-ore bodies in the Biwabik formation are residual concentrates of oxidized iron minerals formed by the leaching of silica from the chert and iron silicates in taconite. Conditions favoring ore concentration are thought to be as follows: accentuated fracturing at folds and faults allowing ready circulation of leaching solutions; a high iron content in taconite; reducing rather than oxidizing conditions of deposition of the original taconite; the fine size of the grains in taconite and the intimate intermixing of different minerals; a lack of metamorphism, which coarsens the grain; and the availability of large amounts of solutions. The soft ores may have been concentrated by downward-circulating ground waters.

THE STRATIGRAPHY AND STRUCTURE OF THE MESABI RANGE, MINNESOTA

I. INTRODUCTION

LOCATION

The Mesabi range of Minnesota lies northwest of Lake Superior (Fig. 1). "Mesabi range" designates the preglacial outcrop area, now mostly buried under drift, of the Biwabik iron-bearing formation. This belt, ¼ to 3 miles wide, extends east-northeast for 120 miles along strike from eastern Cass County through Itasca County to Birch Lake in St. Louis County (Fig. 2). Although the Biwabik formation has only half a dozen natural outcrops west of Mesaba, its north and south boundaries beneath the drift have been accurately determined by thousands of drill holes.

The Mesabi range may be divided for convenience into four parts: the "East Mesabi" is between Birch Lake and Mesaba; the "Main Mesabi" extends from Mesaba westward to Nashwauk; the "West Mesabi" extends from Nashwauk to Pokegama Lake; and the "Westernmost Mesabi" extends from Pokegama Lake to a point in eastern Cass County, which is the western limit of exploration.

TOPOGRAPHY AND DRAINAGE

The Mesabi range occupies the middle slopes of the south flank of the Giants Range,* which is a chain of low hills extending from a point north of Grand Rapids to Birch Lake at the St. Louis County-Lake County line. In the east, the summits are as high as 1900 feet above sea level† and stand some 400 feet above the general level of the surrounding country. The elevation of the crest diminishes westward, and beyond Grand Rapids the entire ridge is buried by thick glacial deposits which form a lowland at an elevation of about 1400 feet. The Giants Range has gentle slopes to the south and somewhat steeper slopes to the north, but on both sides these slopes grade into lowlands characterized by kettle holes, lakes, and swamps. The upper slopes and crest are notched by many drainage channels. The "Virginia Horn" in R. 17 W. is a Z-shaped bend in the crest, which parallels the fold pattern of the Mesabi range.

The Giants Range forms part of the Laurentian drainage divide. The western sector of the south flank is drained south by the Mississippi River and its tributaries; the central sector is drained south by the St. Louis River, which discharges into Lake Superior; but the eastern sector is drained north by the Dunka River.

* "Range" will be capitalized when it refers to this topographic feature. The definition of the Mesabi range and the distinction between it and the Giants Range are important to the understanding of parts of this report.

† All elevations are referred to the datum of mean sea level, which is 602 feet below the Lake Superior datum.

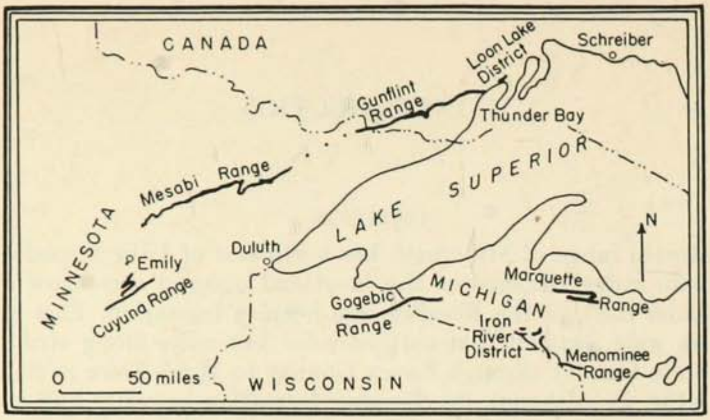


FIGURE 1. — Index map of the Lake Superior region.

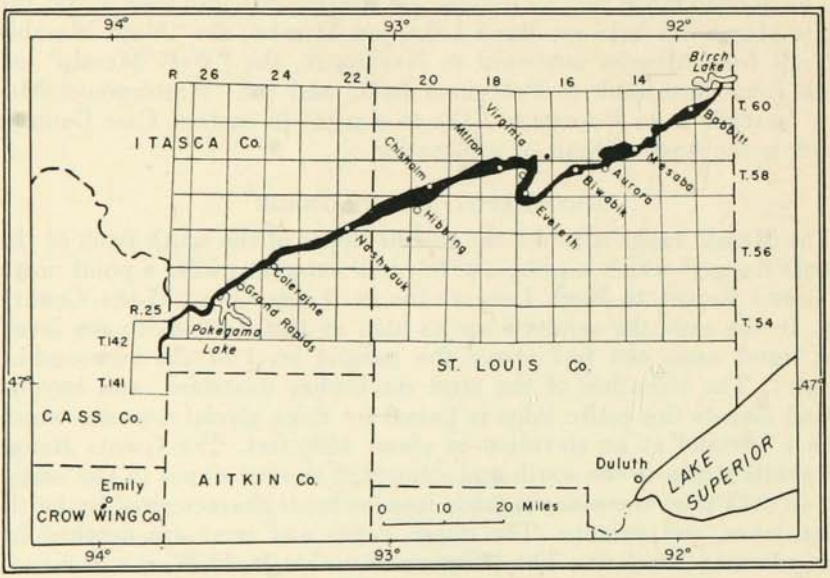


FIGURE 2. — Index map of the Mesabi district, Minnesota.

The northern slopes east of Hibbing are within the Hudson Bay drainage basin except where the south-flowing Embarrass River crosses the Range, but west of Hibbing the north flank is drained south by the Mississippi and Prairie rivers, which transect the Range. A point north of Hibbing, near the south line of Sec. 26:58–21* (E. Peterson and C. R.

* For the sake of convenience in the many references to public-land divisions required, a form such as "Sec. 25, T. 59 N., R. 15 W.," indicating section-township-range, will be shortened to "Sec. 25:59–15."

Knowles, personal communication), marks a junction of drainage divides. Water flowing northeast from this point into a tributary of the Sturgeon River reaches Hudson Bay by the Nelson River; water flowing northwest into a tributary of the Prairie River reaches the Gulf of Mexico by the Mississippi River; and water flowing south into tributaries of the St. Louis River reaches the Atlantic Ocean by the St. Lawrence River.

TABLE 1. CHRONOLOGIC AND STRATIGRAPHIC SEQUENCE IN NORTHERN MINNESOTA

(Pre-Cambrian column modified from Grout *et al.*, 1951)

PHANEROZOIC EON

CENOZOIC

Pleistocene (Wisconsin glacial drift)

.................unconformity.................

MESOZOIC

Upper Cretaceous
  Coleraine formation (shale, conglomerate)

.................unconformity.................

PRE-CAMBRIAN (OR CRYPTOZOIC) EON

LATER PRE-CAMBRIAN

Keweenawan group
- Upper
  - Hinckley formation* (sandstone)
  - Fond du Lac formation* (red shale, sandstone)
- Middle
  - Scattered granites
  - Duluth gabbro
  - Beaver Bay complex* and Logan intrusives
  - Keweenaw Point volcanics* (flows, tuffs)
- Lower
  - Puckwunge formation* (conglomerate, sandstone)

.................unconformity.................

Animikie group
- Virginia formation (argillite) = Rove
- Biwabik formation (iron-bearing) = Gunflint
- Pokegama formation (quartzite, argillite, conglomerate) = Kakabeka

.................unconformity.................

MEDIAL PRE-CAMBRIAN

Algoman batholiths: Giants Range and Vermilion* granites
Knife Lake group (slate, graywacke, iron-bearing beds, conglomerate, with tuffs, flows, intrusives)

.................unconformity.................

EARLIER PRE-CAMBRIAN

Batholithic intrusives: Saganaga granite
Keewatin group
- Soudan formation* (iron-bearing)
- Ely greenstone (flows, tuffs)

*Not present locally at Mesabi or Gunflint ranges.

REGIONAL GEOLOGIC SETTING

The Precambrian rocks of northern Minnesota (Table 1) form the southern margin of the Canadian Shield. The Giants Range, which parallels the north boundary of the Mesabi range, is made up chiefly of the Giants Range granite, which is intrusive into the Ely greenstone and into rocks of the Knife Lake group that crop out at places on the southern slopes of the ridge.

Parts of all these older rocks are overlain unconformably by the Animikie group, which consists of three conformable sedimentary units: the Pokegama, Biwabik, and Virginia formations. The Animikie beds, being on the north limb of the great Lake Superior syncline, generally dip 5° to 15° south-southeast, although abrupt local changes in magnitude or direction of dip are not uncommon. The narrow buried outcrop belt of the Pokegama formation lies north of the Biwabik formation, and the Virginia argillites underlie the drift for an unknown distance south of the Mesabi range.

Though most of the Animikie rocks are essentially unmetamorphosed, the intrusion of the Duluth gabbro has recrystallized the iron-bearing rocks of the East Mesabi. The gabbro cuts diagonally across the Animikie group in the east, obliterating all the sediments except a few scattered inclusions for a distance of fifty miles, between Birch Lake and a point ten miles west of Gunflint Lake, near the Canadian border. In the Gunflint district, the three units of the Animikie group corresponding to the Pokegama, Biwabik, and Virginia formations of the Mesabi district are called respectively the Kakabeka, Gunflint, and Rove formations; the group is intruded by the Logan sills and rests unconformably on Saganaga granite, Ely greenstone, and rocks of the Knife Lake group. The strike and dip of Animikie beds at the Gunflint and Mesabi ranges are much alike. The Gunflint range extends for a distance of one hundred miles from the International Boundary, through Port Arthur to the Loon Lake district in Ontario, on Thunder Bay. The only reported Animikie rocks farther east are on the shore of Lake Superior at Schreiber, seventy miles from Loon Lake (Harcourt, 1939, pp. 13–15).*

The Westernmost Mesabi has been explored only as far as eastern Cass County because the iron-formation there and westward is lean and the drift is very thick. Rocks much like those of the Biwabik formation occur thirty miles farther south near Emily, and a number of geologists believe that the rocks of the Cuyuna range are an extension of the Animikie group.

Scattered deposits of thin Cretaceous shale and conglomerate lie with angular unconformity upon the eroded edges of the Animikie beds of the Mesabi range, and most of the range is blanketed by glacial deposits.

The rock (taconite) of the Biwabik formation is locally oxidized and leached of silica, which results in bodies of soft ore; but no leached ores are found on the metamorphosed East Mesabi. Because material was

* See references at the end of this volume.

removed and pore space was formed during the leaching process, the residual ore layers show structures resulting from slump. The entire iron-formation is partly oxidized and leached on the West Mesabi. The unoxidized taconite of the Main and East parts of the range constitutes a vast reserve of material from which magnetite concentrates can be extracted.

HISTORY OF PRODUCTION

On November 16, 1890, the first rich iron ore on the Mesabi range was discovered near what is now the Mountain Iron Mine, and the first shipment of 4245 tons was made in 1892. The 1952 shipments amounted to 59,461,866 tons, and this brought the total cumulative production of the Mesabi district to more than 1,800,000,000 tons (Wade and Alm, 1953, p. 238), which is two thirds of the total production of the Lake Superior district and about 55 per cent of the nation's entire production since 1891.

II. ROCKS OF THE ANIMIKIE GROUP

POKEGAMA FORMATION

Rock Types

Quartzite. Much of the Pokegama formation is vitreous quartzite. Its color is white or gray, green, brown or yellow, orange or red, depending on whether the matrix consists of quartz overgrowths, chlorite, or, more rarely, of limonite or hematite. The grains, mostly of quartz and a few of microcline and plagioclase, range in size from fine to coarse sand, and small lenses are conglomeratic. Larger grains are well rounded, whereas smaller ones are subangular. Accessory minerals are anatase, apatite, epidote, hornblende, leucoxene, magnetite, muscovite, pyrite, sphene, tourmaline, tremolite, and zircon (Tyler *et al.*, 1940, p. 1495). The quartzite commonly is massive but at places has finer-grained, darker beds and laminations.

Quartz-mica argillite. Micaceous quartzite may grade into thin-bedded, gray quartz-mica argillite composed of silt-size quartz grains and muscovite flakes in an argillaceous matrix.

Conglomerate. The basal conglomerate of the Pokegama formation consists largely of pebbles and boulders derived from the underlying rocks (graywacke, slate, granite, vein quartz, volcanic rocks, and chert).

BIWABIK FORMATION

Nomenclature

H. V. Winchell (1893, p. 124) called the banded siliceous rocks of the lower part of the iron-formation "taconyte," because he correlated the Animikie group with the "Taconic group." Spurr (1894, p. 248) used this name "as a designation of the iron-bearing rock in general" and spoke of "taconyte slates" and "taconyte cherts." Leith (1903, p. 101) preferred the term "ferruginous cherts," but "taconite" was so firmly established that it has been used in almost all reports on the district. Gruner (1946, p. 5) refers to all rocks of the Biwabik formation, other than oxidized ores, as taconite. In the present report, all rocks of the Biwabik formation other than oxidized ores and partly detrital rocks (quartz-mica argillite, graphitic argillite, and iron-bearing sandstone) are referred to as taconite.

The name "slate" has been previously applied to all fine-grained, thin-bedded rocks throughout the Animikie group, but nearly every writer has pointed out that this term is a misnomer, because these rocks do not have the cleavage of a true slate but merely a parting parallel to the bedding. Slate as a rock name is therefore not used in this report, and the term "argillite" is substituted as a name for the hardened rocks derived from clastic shales and siltstones. But the adjective "slaty," meaning thin-bedded, is retained, and "slaty taconites" are the non-clastic fine-textured rocks of the Biwabik formation.

Classification of Taconite

Taconite consists of admixtures of minerals which belong to four principal chemical groups, as listed here in order of decreasing abundance: silica (chert), iron silicates (minnesotaite, stilpnomelane, greenalite), iron oxides (magnetite, hematite), and carbonates (siderite, calcite, dolomite). Pyrite and graphite are accessory minerals. A detailed description of these minerals is given by Gruner (1946, pp. 7–22). The present discussion does not include the metamorphosed rocks of the East Mesabi.

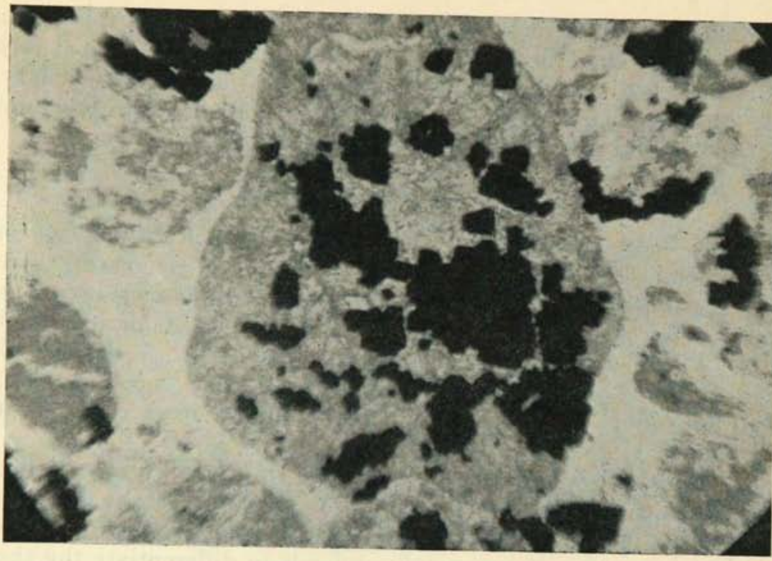
Most of the chert is microcrystalline quartz, although a few particles are as much as .2 mm. in diameter. The iron silicates occur as individual microscopic plates and needles, which are commonly either matted or in radiating aggregates. It is generally difficult to differentiate the three iron silicates because of intimate intermixing and fine grain size; therefore they are treated as a group. Magnetite forms tiny octahedra that are rarely more than .1 mm. in diameter, and siderite forms small rhombs or irregularly round grains, whereas the other carbonates commonly occur as larger rhombs and grains.

On the basis of textures that can be seen in hand specimen, taconites can be classed as seemingly coarse-grained or "granular" types and fine-grained or "slaty" types. Actually, nearly all the mineral particles in taconites are very fine-grained (Fig. 3), but the organization of particles into spherical or ellipsoidal clumps averaging about .5 mm. in diameter gives some rocks a coarse-textured appearance. Spurr (1894, p. 49) first called these rounded bodies "granules." Although even the densest slaty taconite commonly has a few granules, they are visible only under the microscope.

Granules may be formed by any mineral or combination of minerals embedded in a matrix of the same or other minerals. For example, some granules of fine chert in a coarser chert matrix can be recognized only under a microscope, but others may be visible to the naked eye because of rims and inclusions of magnetite, silicate needles, graphitic dust, or hematitic dust. Granules are rarely greater than 1 mm. in diameter, although aggregates may make larger blobs. Some granules have cracks filled with the material of the matrix, and a few have small "tails" or comma-like appendages. Granules have no regular internal structure, but some taconite contains rounded bodies that are true oölites, having concentric layers of jasper or hematite around a quartz or iron oxide core.

In general, taconite composed dominantly of chert with iron silicates

A



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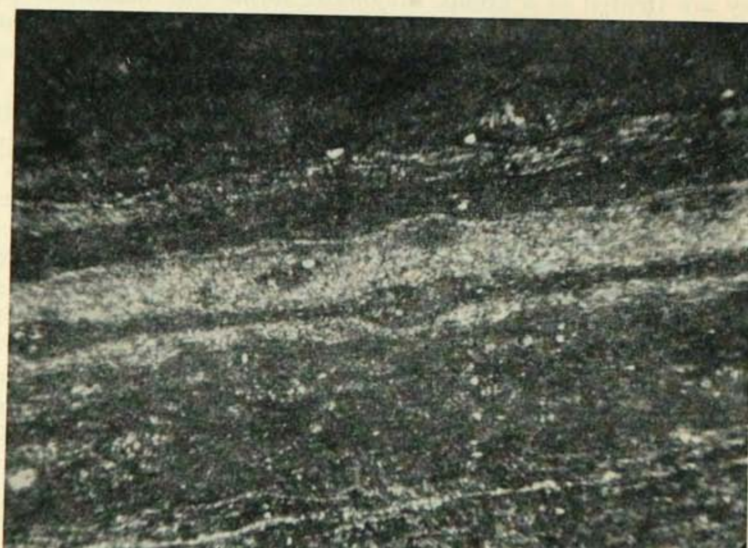

FIGURE 3. — Granular and slaty taconites as seen under the microscope. (A) Magnetic silicate taconite. Granules are of very fine-grained minnesotaite (gray) with some chert in a matrix of chert (white). Black areas are euhedral magnetite. Without analyzer, ×45. (After Gruner, 1946.) (B) Green slaty taconite. The dark portion consists mostly of stilpnomelane and siderite. Light bands are largely chert with some minnesotaite needles. Without analyzer, ×20. (After Gruner, 1946.)

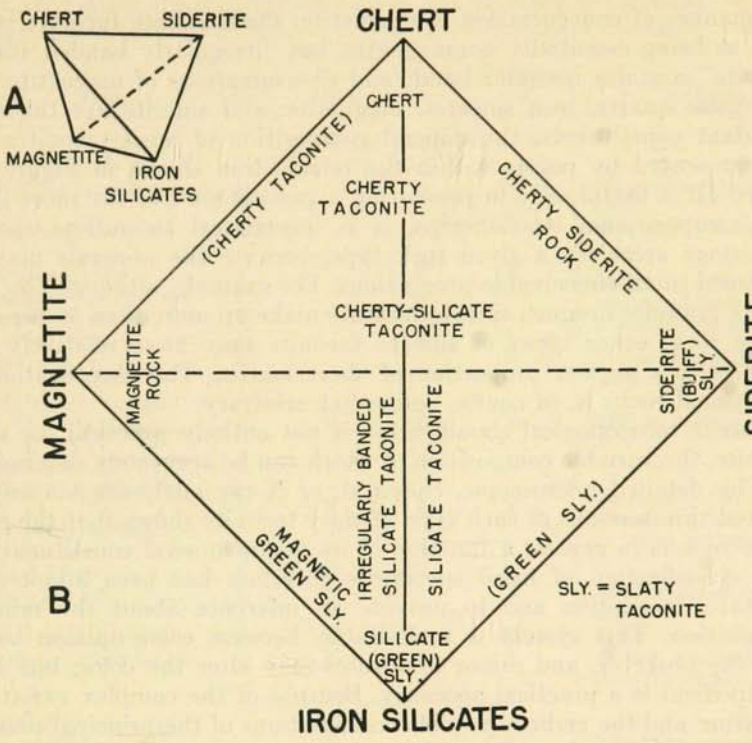

FIGURE 4. — Diagram showing generalized mineral compositions of taconites. (A) The tetrahedron by which the mineral composition of most taconites can be represented. (B) The same tetrahedron with the edge between chert and iron silicates facing up. Rock names are inside the tetrahedron, generally near the upper surfaces, and the locations of the names roughly show the more common compositions and interrelationships of rock types. Taconite composed dominantly of magnetite and siderite is rare and is therefore not shown.

or magnetite is apt to be granular, whereas rock that is made up mostly of siderite and/or iron silicates is apt to be slaty. Ideally, slaty taconite can be classified according to its major mineral constituents. The matrix of granular taconite is ordinarily chert; the rock is classified according to the composition of its dominant granules and given such names as cherty taconite, silicate taconite, and cherty-silicate taconite. "Carbonate taconite" and "magnetite taconite" are not included in this classification according to dominant granules because the carbonates are generally inconspicuous and in lesser amount, and the distribution of magnetite is described in other ways.

By definition, cherty and cherty-silicate taconites contain magnetite as disseminated octahedra and granules, and commonly also as irregularly banded, regularly banded, laminated, or mottled concentrations. These textural adjectives preceding a rock type indicate the presence and manner of concentration of magnetite. Plain silicate taconite is defined as being essentially nonmagnetic, but "irregularly banded silicate taconite" contains irregular bands and disseminations of magnetite.

Because quartz, iron silicates, magnetite, and siderite are the most abundant constituents, the mineral composition of most taconites can be represented by points within the tetrahedron shown in Figure 4A. Figure 4B is useful only in presenting a general view of the more common compositional relationships; it is impractical to outline specific percentage areas for a given rock type, because the minerals may be combined in all conceivable proportions. For example, although the iron silicate granules in much silicate taconite make up more than 50 per cent of the rock, other types of silicate taconite may have relatively few granules in a greater proportion of chert matrix. The classification of gradational rocks is, of course, somewhat arbitrary.

A strict mineralogical classification is not entirely practical for slaty taconite, the variable composition of which can be accurately determined only by detailed microscopic, chemical, or X-ray analyses. A study of selected thin sections of each type of slaty taconite shows that the color of the rock is in general a function of its major mineral constituents. A color classification of hand specimens therefore has been adopted to facilitate description and to provide an inference about the mineral composition. This system is undesirable, because color opinion varies with the observer, and minor impurities may alter the color, but such an expedient is a practical necessity. Because of the complex variations in texture and the endless possible combinations of the principal mineral constituents, only the most common types of taconite are named. The characteristics implicit in these names are summarized in Table 2.

Granular Taconites

Chert. Few of the more highly siliceous rocks in the Biwabik formation are pure cherts, and those containing minor amounts of magnetite are called lean cherty taconite. The gray cherts of the Westernmost Mesabi are both granular and oölitic and contain a small proportion of siderite but practically no magnetite or iron silicates. The bluish or greenish cherts in the upper part of the formation commonly contain a very few minute silicate fibers or granules.

Cherty taconite. Cherty taconite has granules of chert, magnetite, carbonate (siderite and calcite), and minor amounts of iron silicates set in a white or gray chert matrix, which gives a "salt and pepper" effect. (Jaspery taconite, which is oölitic in places, is a variety of cherty taconite in which the chert granules are colored red by minute hematite inclusions.) The magnetite in plain cherty taconite is all disseminated, but there are four textural varieties (Fig. 5) in which about 25 per cent of the total magnetite is disseminated and the remainder is in larger aggregates. Although these varieties differ in appearance, probably they all have about the same bulk mineral composition.

TABLE 2. GENERALIZED COMPOSITIONS AND TEXTURES OF TACONITES*

Name	Chert	Iron Silicates	Magnetite	Carbonate	Remarks
		GRANULAR TACONITES			
Chert	E				
Jaspery taconite (jsp.)	E		E	X	Magnetite disseminated only.
Cherty taconite (chy.)	E		E	X	Magnetite disseminated only.
Laminated chy.	E		E	X	Magnetite partly disseminated but mostly in larger aggregates.
Regularly banded chy.	E		E	X	
Irregularly banded chy.	E		E	X	
Mottled chy.	E		E	X	
Lean chy.	E			X	Minor magnetite present.
Cherty-silicate taconite (chy.-sil.)	E	E	E	X	Magnetite disseminated only.
Laminated chy.-sil.	E	E	E	X	Magnetite partly disseminated but mostly in larger aggregates.
Regularly banded chy.-sil.	E	E	E	X	
Irregularly banded chy.-sil.	E	E	E	X	
Mottled chy.-sil.	E	E	E	X	
Lean chy.-sil.	E	E		X	Minor magnetite present.
Silicate taconite (sil.)	E	E		X	Little or no magnetite.
Irregularly banded sil.	E	E	E	X	(Other magnetite textures uncommon.)
		SLATY TACONITES (SLY.)			
Silicate slaty taconite		E			
Green slaty taconite		E		X	Minerals uncertain.
Magnetic green slaty taconite		E	E	X	Minerals uncertain except magnetite.
Black slaty taconite		E			Minerals uncertain, except graphite.
Cherty siderite rock	E			E	
Brown slaty taconite	E			E	Minerals uncertain.
Buff slaty taconite				E	Carbonate is siderite.
Limestone				E	Carbonate is calcite.
Red slaty taconite	X	X		X	E, Hematite.
Hematitic slaty taconite	X			X	E, Hematite.

*E = Essential constituent; X = Constituent may be present in important amounts. Blank spaces indicate that a constituent is generally present in insignificant amounts, although it may be lacking entirely; rarely, the constituent may be abundant.

1. Laminated cherty taconite is a fine-grained slaty-looking rock in which part of the magnetite is in the form of closely spaced straight laminations. The dark color commonly has a reddish tone caused by small amounts of hematite. Granules are rare, small, and poorly formed, and iron silicates may be somewhat more abundant than in the other textural varieties.

2. Regularly banded cherty taconite has relatively straight, parallel,

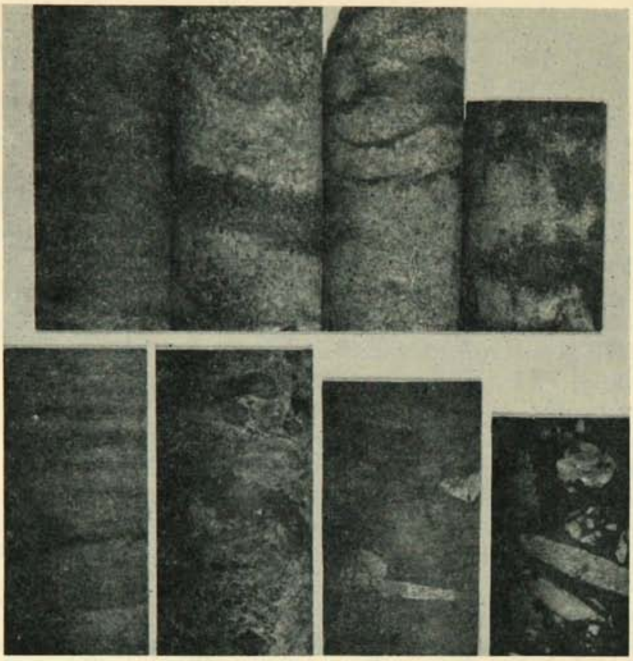

FIGURE 5.—Drill cores of taconites. Top, left to right: four textural varieties of cherty taconite — laminated, regularly banded, irregularly banded, and mottled. Bottom, left to right: green slaty taconite; conglomerate of jasper and magnetite pebbles in a cherty matrix; cherty siderite rock, showing chert lenses; conglomerate of chert pebbles in a matrix of black slaty taconite. (All natural size; photographs by R. D. Taylor.)

magnetite-rich bands alternating with cherty beds. These bands contain coarser granules of magnetite and have a more cherty groundmass than the irregular bands, and their boundaries are commonly gradational. They may be as much as 2 or 3 inches thick, and they generally comprise 10 to 40 per cent of the rock's volume.

3. In irregularly banded cherty taconite, light-colored cherty beds alternate with brown or black bands composed mostly of magnetite with interstitial chert, iron silicates, and carbonate. The wavy bands, which generally make up 5 to 20 per cent of the volume of the rock, are roughly parallel although they do swell and pinch out, split and reunite, or connect with adjacent bands. They range from minute stringers to bands 1½ inches thick and commonly have sharp boundaries, each lined with a thin layer of siderite or dolomite.

4. Mottled cherty taconite contains scattered patchy concentrations of magnetite, a half inch or so in diameter, which generally have much cherty groundmass and irregular gradational boundaries. Some asso-

ciated mottles are magnetite-deficient areas, and others are composed of yellow, orange, or pink carbonate.

Cherty-silicate taconite. As its name implies, this rock is gradational between cherty taconite and silicate taconite. It contains abundant granules of chert, iron silicates, and disseminated magnetite, and the matrix commonly contains iron silicates as well as chert. Carbonate may be abundant, and the magnetite may be concentrated in regular and irregular bands, mottles, or laminations. Lean cherty-silicate taconite contains very little magnetite.

Silicate taconite. Silicate taconite has green silicate granules in a matrix of white or gray chert; the matrix may also contain carbonate and needles or radiating sheaves of iron silicates, particularly of minnesotaite. Plain silicate taconite contains essentially no magnetite, but irregularly banded silicate taconite contains magnetite disseminations and bands. The other textural aggregates of magnetite in a groundmass of silicate taconite are uncommon.

Slaty Taconites

Silicate slaty taconite. As its granules become more abundant, indistinct, and merged, silicate taconite grades into green silicate slaty taconite in which the small amount of cherty matrix is not visible and only matted silicate fibers can be seen. Carbonate may be present at the expense of chert.

Green slaty taconite. Probably the dark-green, thin-bedded, nonmagnetic slaty taconites are largely silicate slaty taconites, but they are so fine-grained that the constituents cannot be identified without X rays. Chert is generally present, but iron silicates and siderite are undoubtedly the dominant constituents of most varieties. The name "magnetic green slaty taconite" indicates the presence of magnetite.

Black slaty taconite. Graphite and some finely divided pyrite account for the color of this very fine-grained nonmagnetic rock, which grades into green slaty taconite and is probably composed mostly of iron silicates, including much stilpnomelane, with minor amounts of chert and carbonate. Silicate granules and thin beds of pure white chert may occur.

Cherty siderite rock. This gray-brown or green-brown, nonmagnetic, somewhat irregularly laminated rock is composed of intimately intermixed siderite and chert. Two thirds of the rock generally is siderite, although chert may predominate; white chert lenses about ½ inch in thickness, some of which contain calcite rhombs around the edges, are common. A few strata are rich in calcite, and a small amount of fine graphite is peppered through the rock. Chert grains are rarely more than .01 mm. in diameter, and much of the siderite is also microcrystalline, although some particles are as large as .05 mm. Occasionally there is a small amount of presumably clastic material — chlorite and a few rounded quartz grains. According to the analyses reported by Gruner (1946, p. 50), cherty siderite rock contains an average of 15 to

20 per cent iron. The specific gravity of the rock ranges from 2.95 to 3.25, averaging about 3.05.

Brown slaty taconite. Only the rock of certain uniform lithic units west of Nashwauk can with certainty be called cherty siderite rock. For the sake of convenience, taconite of similar appearance along the rest of the range is called "brown slaty taconite," although the color is more truly a brown-green or olive drab; this rock commonly contains some iron silicates, and the proportions of chert and siderite are variable. Specific gravity may be as high as 3.60. Small amounts of buff slaty taconite, composed mostly of fine-grained dusty siderite, are scattered throughout the iron-formation.

Limestone. The gray or green limestone of the Biwabik formation consists of interlocking, irregular grains of calcite. Thin beds of granular chert or gray argillite may be included, and at places the limestone contains some iron silicates. Specific gravity ranges from 2.75 to 2.90 and it averages about 2.80.

Red and green slaty taconite. This rock has alternating straight bands and laminations of dull reddish and greenish slaty taconite. The green bands are like silicate slaty taconite but commonly contain a high proportion of siderite. The major constituent of the slightly magnetic red bands is hematite, which forms closely interwoven stringers and laminations set in a groundmass of iron silicates, carbonate, and chert.

Hematitic slaty taconite. Red hematitic slaty taconite grades from nearly pure hematite into very fine-grained hematitic jaspery taconite. Some interbedded layers are rich in siderite, calcite, and iron silicates, or contain a few octahedra of magnetite. Where this rock rests directly on top of the Pokegama formation, minor chlorite and clastic white quartz grains that form the nuclei of oölites may be incorporated, which gives the rock a spotted appearance.

Conglomerates and Algal Structures

Conglomerates. Beds of conglomerate, generally only a few inches thick — but in one case as much as 25 feet thick — occur throughout the Biwabik formation. All are intraformational conglomerates, for they are composed entirely of fragments of taconite cemented in a taconite matrix. The only possible exception to this generalization is the red basal conglomerate, which commonly contains clastic quartz grains and some chlorite. But even here (except on the East Mesabi) all the particles are derived from adjoining sediments which were probably formed almost contemporaneously with the red basal conglomerate.

Most of the pebbles are small, flat, and rounded. Some beds in which the fragments are very angular are breccias, and others in which the pebbles are variously oriented or standing on end are "edgewise conglomerates." Many of the conglomerates in cherty rocks have pebbles of cherty taconite, chert, hematitic slaty taconite, or solid magnetite in a matrix of cherty taconite, silicate taconite, or, most commonly, granu-

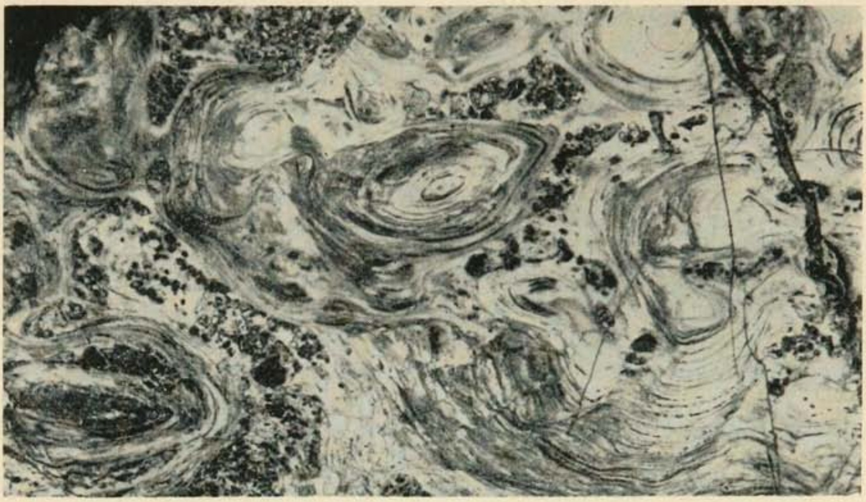

FIGURE 6. — Algal structures. Laminae are pink in color. Sample taken from Corsica Mine, Sec. 18:58-16, and reproduced at 1½ times its natural size. (After Gruner, 1946.)

lar and oölitic jaspery taconite. A few pebbles are surrounded by a shell of nearly pure magnetite which grades outward into the matrix. Some of the conglomerates in slaty rocks have slaty pebbles in a matrix of calcite or chert, whereas others consist of flat white chert pebbles in a slaty matrix. These chert pebbles are commonly coated with pyrite.

Algal structures. Algal structures, generally red and white but at places green, form either the matrix of some of the conglomerates or the adjoining massive, resistant beds which are as much as several feet thick. At places these structures are in separate concretionary mounds which are two or three feet in diameter and a foot high. Algal structures are generally composed of uniformly fine-grained, gray chert containing red laminations of jasper. The laminations form gnarled patterns or whorls and upwardly convex arches, or columnar structures resembling piles of inverted bowls (Fig. 6). At places granules of magnetite are concentrated in the areas between the columns, but iron silicates and carbonates are very rare. Associated with these larger structures are small oölites of similar composition.

Clastic and Partly Clastic Rocks

Quartz-mica argillite. Dark-gray, thin-bedded, quartz-mica argillite is found in the Biwabik formation only in the Westernmost Mesabi. It is typically composed of about 25 per cent quartz, 25 per cent biotite and muscovite, and 50 per cent sericite, chlorite, and graphite. Feldspar and pyrite are accessory minerals. The subangular or angular quartz grains are generally the size of silt but may be as large as .1 mm. in diameter; where the coarser grains predominate, the rock might be called a subgraywacke (Pettijohn, 1949, p. 255). Bedding planes, on which the larger flakes of mica can be seen with the naked eye, are due to alternating quartz-rich and matrix-rich layers. At places the rock is very cherty, and it also grades locally into a calcareous phase in which rhombs of calcite and scattered grains of detrital quartz and mica are imbedded in a graphitic matrix. Specific gravity ranges from 2.55 to 2.85, averaging about 2.70. Probably the iron content is never more than 10 per cent.

Graphitic argillite. In thin section, black graphitic argillite is mostly opaque, and the clay minerals that are probably in the matrix are obscured by abundant graphite. Subangular clastic grains of quartz and a few of feldspar, all smaller than .05 mm. in diameter, generally make up less than 5 per cent of the rock. Tiny disseminated crystals of pyrite, which can be seen in polished sections, may comprise very little or as much as 10 per cent of the rock. Laminations, barely visible in hand specimen, are due to thin layers of biotite or iron sulfide.

An incrustation of white iron sulfate forms on the surface of some of the drill core in storage, cracking the argillite into thin discs. This alteration suggests that marcasite is present as well as pyrite, and that the original content of iron sulfide was higher than that shown in the analyses (Table 3). Specific gravity ranges from 2.25 to 2.75, and averages about 2.50; the higher values probably indicate a higher pyrite content. The radioactivity of samples 1 and 2 (Table 3) is negligible.

The black argillites of the Virginia formation look much like graphitic argillite, but the latter can generally be recognized in hand specimen by one or more of the following characteristics: a darker streak, a lower specific gravity, a surface alteration of white iron sulfate, a lack of gray interbeds, and a lesser brittleness when scratched with a knife point.

Iron-bearing sandstone. This rock is well cemented, but it breaks around grains and not through them. Generally, more than half of it is composed of fairly well sorted subangular grains of quartz, and a few of feldspar, most of which are .1 to .2 mm. in diameter. The rock is red or green, depending on whether the matrix is hematite or chlorite. Although most of the hematite matrix appears to be primary, at places a red staining follows joint planes. Both siderite and calcite are commonly found in this rock, and chert may be a minor constituent. The rock is associated with hematitic slaty taconite.

Partly Oxidized and Leached Taconites

Where large areas of taconite have been partly oxidized and leached, it is difficult to infer the original mineralogy of the rock. Correlations of this rock with the few fresher rocks in these areas give reasonable indications, however, of the nature of the original constituents.

Oxidized cherty taconite. Oxidized cherty taconite consists mostly of pitted white chert which contains magnetite or martite (hematite pseudomorphous after magnetite) in any of the typical textural concentrations. This rock may be altered cherty, cherty-silicate, or silicate taconite in which the carbonates and iron silicates have been largely leached or oxidized. In general, larger amounts of limonite in the altered rock indicate a greater proportion of iron silicates in the original rock.

Oxidized white cherty taconite. This rock, composed mostly of white pitted chert containing some limonite granules, is found in a single stratigraphic unit that, where fresh, consists of a silicate taconite containing relatively few silicate granules in a very cherty matrix.

Paint rock. "Paint rock" is the local name applied to yellow, white, or in most cases red soft rocks that are altered silicate, green, or black slaty taconite. Paint rock gives X-ray patterns of quartz, hematite, and kaolinite (Gruner, 1946, p. 22). Whereas hematitic slaty taconite is believed to be colored by primary hematite, paint rock owes its color to secondary oxidation. It is not always easy to tell the two rock types apart without a microscope, but hematitic slaty taconite commonly has a higher specific gravity. Cherty siderite rock, limestone, and brown

TABLE 3. ANALYSES OF GRAPHITIC ARGILLITE

Constituent	1	2	3
SiO_2	47.29	53.42	36.67
Al_2O_3	6.92	13.43	6.90
Fe_2O_3	11.46*	.23*	**
FeO	4.46*	4.46*	2.35
MnO	.77	.19	.002
CaO	none	none	.13
MgO	1.92	2.08	.65
Na_2O	.99	1.62	.26
K_2O	3.86	5.77	1.81
TiO_2	.83	1.17	.39
P_2O_5	.09	.13	.20
V_2O_5	n.d.	n.d.	.15
CO_2	1.18	.34	
SO_3	2.87†	1.77†	2.60
C	4.25	6.83	7.28
Organic matter	n.d.	n.d.	.32
FeS_2	5.66	2.70	38.70
H_2O-	....‡	....‡	.55
H_2O+	5.72	5.02	1.25
	98.27	99.16	100.21
Total Fe	14.13	4.89	19.89

1. Representative sample of 35 feet of drill core, Sec. 4:54-26. V. Bye, analyst, Mines Experiment Station, University of Minnesota.
2. Representative sample of 175 feet of drill core, Sec. 13:142-25. V. Bye, analyst, Mines Experiment Station, University of Minnesota.
3. Sample from 10th level, Buck Mine, Iron River district, Michigan (James, 1951a, p. 255).

* The relative proportions of Fe_2O_3 and FeO are uncertain, owing to the presence of pyrite and the possible presence of organic matter.
† SO_3 occurs in iron sulfate, an alteration product of FeS_2.
‡ Samples dried at 100° C.
** Blank spaces indicate "not reported."

slaty taconite that does not contain iron silicates, generally decompose with little color change.

VIRGINIA FORMATION

ROCK TYPES

Argillite. Most of the Virginia formation consists of very fine-grained, thin-bedded black and gray argillites. A few clastic grains of quartz, muscovite, biotite, and rarely of feldspar, which are generally less than .02 mm. in diameter, lie in a matrix of clay minerals, chlorite, and graphite. A higher graphite content accounts for the many black beds and laminations, which commonly contain a few pyrite crystals. The rock is generally tough, having only a poor parting parallel to the bedding planes.

The argillite may grade into thin beds of somewhat coarser-grained, lighter-colored argillaceous limestone, or it may be hard and cherty, breaking with a conchoidal fracture. The grain size may differ somewhat in alternate beds, and Leith (1903, p. 169) states that some of the rock is coarse enough to be called graywacke. However, the Virginia formation is very uniformly fine-grained throughout thousands of feet of drill core all along the range, with the exception of an occasional thin intraformational conglomerate consisting of flat argillite pebbles in a matrix of argillite or argillaceous limestone. The specific gravity of the argillite ranges from 2.50 to 2.90, and it averages 2.70.

Limestone. The argillites of the lower part of the Virginia formation contain thin beds and concretions of gray limestone and argillaceous limestone similar in appearance and specific gravity to the limestone of the Biwabik formation. The massive rock contains some graphite, a few flakes of muscovite, and some small grains of quartz, but it has no chert or iron silicates. The limestone concretions may be as large as 3 feet in diameter by 1 foot in thickness.

Cherty siderite rock. The cherty siderite rock of the Virginia formation is like that of the Biwabik formation, except that it commonly contains interbeds of argillite.

III. DESCRIPTIVE STRATIGRAPHY

HISTORY OF NOMENCLATURE

Hunt (1873, p. 339) wrote: "This older series of Thunder Bay and its vicinity, which may be named the Animikie group from the Indian name of the bay, is the lower division of the upper copper-bearing series of Logan." (The upper division of the copper-bearing series was what is now known as the Keweenawan group.) Logan (1863, p. 67) believed that the rocks unconformably underlying the Animikie could be correlated with Huronian rocks (named by Murray, 1857, p. 168) of the type locality on the north shore of Lake Huron. Logan (1863, p. 87) also thought that the Animikie itself was the probable equivalent of the Cambrian Potsdam sandstone of New York. N. H. Winchell (1882, p. 135) correlated the Animikie group with the Lower Cambrian "Taconic group" underlying the Potsdam sandstone.

Bell (1873, p. 93) observed that the rocks at Gunflint Lake were similar to those at Thunder Bay, and Irving (1883, pp. 381–390) correlated the Mesabi range with the Gunflint range and also with the Gogebic range of Wisconsin and Michigan. Irving, however, linked the Animikie itself, rather than the unconformably underlying rocks, with the original Huronian of Murray.

Alexander Winchell (1890, p. 370) recognized an unconformity within the Huronian at the type locality, which divided the sequence in two. Van Hise (1892, p. 193) accordingly called the Animikie "Upper Huronian" and the unconformably underlying rocks "Lower Huronian." This usage was approved by a committee of Canadian and American geologists (Adams *et al.*, 1905, p. 103). Allen (1920, p. 191) discovered an unconformity at the Gogebic range and split the "Upper Huronian" there into two divisions. By analogy with this condition in Michigan, Leith *et al.* (1935, p. 15) split the Animikie of Minnesota, classing the Virginia formation as "Upper Huronian" and the Biwabik and Pokegama formations as "Middle Huronian," even though the local evidence of unconformity was negligible.

Lawson (1929, p. 383) concluded that the Animikie of Thunder Bay was not Huronian in any sense, and Miller and Knight (1913, p. 125) recommended dropping the term "Huronian" entirely, because of the confusion involved. Others (Allen, 1920, p. 189; A. Leith, 1935, p. 325) advocated doing away with the term "Animikie," but Grout *et al.* (1951, p. 1042) state: "We retain the name because Minnesota outcrops are well correlated with the original Animikie locality and not well correlated with the type locality of the Huronian." In the Mesabi district, therefore, "Animikie" appropriately designates the group of three formations — the Pokegama, the Biwabik, and the Virginia.

METHOD OF REPRESENTATION

The longitudinal stratigraphic section of the Biwabik formation (Plate 1) was constructed by projecting logged diamond drill holes up or down the dip onto a vertical plane following the map contact between the Biwabik and Virginia formations. Some of the projected drill holes actually lie to the north and some lie to the south of this contact, but few are more than one mile away from it. The holes so projected fall into their proper stratigraphic places relative to the arbitrarily straightened-out base line chosen for plotting — a line representing the base of the Biwabik formation. (No structural relations are shown.) Correlation of the rock types logged in adjacent holes results in the outlines of lithic units. About two thirds of the total number of holes examined are plotted. The vertical exaggeration is necessarily very large; a distance equal to 100 feet on the vertical scale is equal to about 6400 feet on the horizontal scale.

The uppermost part of each hole, representing unconsolidated surficial deposits and any Cretaceous material, has been omitted on Plate 1. The percentages by volume of the different interbedded rock types, shown within some lithic units, are averages of the varying estimated percentages found from hole to hole. The mining companies generally saved only about 2 feet of core from each 5-foot run of the bit, but errors introduced by possibly unrepresentative samples are probably averaged out for the most part. The holes at the Emily district, where the beds are tightly folded, have been shortened to compensate for an assumed average 30° dip, so that the thicknesses shown are more nearly accurate than if the footages had been plotted directly.

No correction for the gentle dips along the Mesabi range has been applied, however. The bedding generally makes an 80° to 90° angle with the core axis, and many of the holes seem to have been slightly deflected at depth to a position more nearly perpendicular to the bedding. A thickness correction for decomposed, slumped rocks is so difficult to evaluate that it has not been attempted; however, the holes selected, other than those on the West Mesabi, are generally in fresh rocks.

The specific gravities shown on Plate 1 are averages of determinations that were made ordinarily every 10 feet with a simple beam balance on representative, naturally dried, fresh cores of the slaty rocks in lithic units. Where units consist of interbedded slaty rocks and cherty or silicate taconite, specific gravities of only the slaty rocks were determined.

This method of correlation of slaty taconites by specific gravity, originated by Gruner (1946, p. 51), is particularly useful in distinguishing cherty siderite rock having a high iron content and a high specific gravity from iron-poor Virginia argillite having a lower specific gravity. However, graphitic argillite containing as much as 15 per cent total iron generally has a lower specific gravity than either quartz-mica argillite or Virginia argillite, both of which have a low iron content. The specific gravity of graphitic argillite is low, presumably because the low specific gravity of the graphitic matter and some of the other constituents more than offsets the high specific gravity of the pyrite. Gruner (1946, p. 74) reports that the average specific gravity of the whole iron-formation is 3.28.

Gruner (1946), in his longitudinal section of the Biwabik formation, gives estimates of the average percentage of iron occurring as magnetite in the drill cores.

On Plate 1, the symbols indicating lateral gradational changes of lithology within a unit are placed between the drill holes near which the major part of the change occurs. Many local variations are too abrupt to be shown, and the rock types are therefore generalized to some extent. It should be emphasized that most contacts between units are gradational and are therefore placed somewhat arbitrarily. A few of the minor subdivisions, particularly in the Lower Cherty member, are based on relatively slight differences in texture or on gradational changes in the content of iron silicates within rocks having much the same mineral composition. In addition, the great vertical exaggeration of Plate 1 makes these changes appear more abrupt and more significant than they actually are.

The stratigraphy of the Westernmost Mesabi is based on a piecing together of the structural and stratigraphic information provided by the few drill holes available. The succession of rock types is reasonably well established, but the thicknesses of some units, particularly those in the Virginia formation, are, at places, not definitely known. Much more information would be required before the areal distribution of these rocks could be determined with any degree of exactness, and the geologic map of the district (Plate 2) represents only the best approximation possible at present.

POKEGAMA FORMATION

The Pokegama formation was so named because it crops out at Pokegama Falls on the Mississippi River near Grand Rapids (H. V. Winchell, 1893, p. 123; N. H. Winchell, 1900, p. 45).

Very few holes (Table 4) have ever been drilled through the entire Pokegama formation, but its thickness probably does not exceed 150 feet along most of the Main and West parts of the range. On the East Mesabi its maximum thickness is 30 feet (Grout and Broderick, 1919b, p. 7), and at places this formation is missing entirely. Judging from meager evidence concerning dip and the width of the buried outcrop belt on the Westernmost Mesabi, the formation thickens westward to perhaps as much as 350 feet.

The stratigraphic sections of the formation presented by Tyler *et al.* (1940, p. 1494) and the descriptions by Gruner (1924, pp. 5–6) indicate that the basal conglomerate is a few inches to 8 feet thick. It fills small

TABLE 4. TOTAL THICKNESSES OF THE POKEGAMA FORMATION
(Taken from exploration records)

Location of Drill Hole	Thickness in Feet	Location of Drill Hole	Thickness in Feet
SW¼NW¼ Sec. 1:59-14	14	SE¼SE¼ Sec. 17:58-19	52
SW¼NE¼ Sec. 12:59-14	23	NE¼SW¼ Sec. 13:58-20	48
SE¼NW¼ Sec. 3:58-16	63	SW¼NE¼ Sec. 35:58-21	140*
NE¼SW¼ Sec. 9:58-17	78	? Sec. 16:56-23†	87
NW¼SE¼ Sec. 35:59-18	40		

* This figure is questionable.
† The specific forty-acre tract in Sec. 16:56-23 is unknown.

channels in the uneven pre-Animikie erosion surface and is commonly overlain by micaceous quartzite or quartz-mica argillite, which is in turn overlain by massive quartzite. The top of the formation on the Westernmost Mesabi is at places argillaceous, however, and it may contain chert lenses and limonitic or hematitic stains. In some areas, the entire formation consists of massive micaceous quartzite.

BIWABIK FORMATION

GENERAL FEATURES

The name of the iron-formation was given by Van Hise and Leith (1901, p. 356), "because the word *Biwabik* is the Chippewa word for 'a piece or fragment of iron' and the Biwabik mine is one of the earliest and larger of the mines located upon the formation." The total thickness of the formation ranges from about 340 feet at Birch Lake and in R. 26 W. to about 800 feet at Eveleth.

The general stratigraphy of the Biwabik formation on the Main and West parts of the range is summarized in Table 5, which is based on Plate 1. In a broad sense, the iron-bearing* formation is defined by its high iron content (Table 6) and by its characteristic granular texture. The base of the formation is easy to recognize because of the abrupt change from iron-poor quartzite to obviously iron-bearing or granular rocks. East of R. 21 W. the top of the formation is the top of a limestone unit that does not contain much iron but does contain some iron silicates and a few interbeds of granular chert. Spurr (1894, p. 9) considered this layer the basal unit of the Virginia formation, but it was properly placed in the Biwabik formation on the charts of Wolff (1917a) and Gruner (1924, 1946). The limestone layer pinches out in R. 21 W., and the top of a unit of cherty siderite rock must be considered the top of the iron-formation to the west of R. 21 W. This unit is absent in R. 27 W., however, and the top of the formation is there the top of a unit of graphitic argillite, which is commonly but not invariably an iron-bearing rock.
The four subdivisions of the formation (Lower Cherty, Lower Slaty,

* According to Pettijohn (1949, p. 333), "Only those sedimentary rocks which contain 10 or more percent of iron (15 or more percent ferric oxide or its equivalent) are considered iron-bearing."

TABLE 5. GENERALIZED LITHIC SEQUENCE WITHIN THE BIWABIK FORMATION BETWEEN MESABA AND COLERAINE*

	Range of Thickness in Feet
Upper Slaty member	20–295
Cherty siderite rock (top of formation west of R. 21 W.)	0–165
Limestone, with some chert and gray argillite (top of formation east of R. 21 W.)	0–55
Brown slaty taconite and silicate taconite, with some cherty-silicate taconite or chert	0–130
Silicate or cherty-silicate taconite and magnetic green slaty taconite	0–250
Upper Cherty member	20–235
Regularly banded, mottled, and laminated cherty or cherty-silicate taconite, with some brown, or red and green slaty taconite (magnetite-bearing unit)	0–175
Algal structures and conglomerate (This unit at places lies under the unit listed below or within the unit listed above.)	0–25
Cherty-silicate taconite and red and green, brown, black (rarely), green or silicate slaty taconite	0–135
Laminated, mottled, regularly and irregularly banded cherty taconite or at places cherty-silicate taconite (magnetite-bearing unit)	10–205
Green slaty taconite and cherty-silicate taconite	0–35
Laminated, regularly or irregularly banded cherty-silicate taconite	0–65
Lower Slaty member	5–225
Green (in part magnetic), brown or silicate slaty taconite, with some cherty or cherty-silicate taconite	0–65
Silicate taconite, with some green slaty taconite	0–130
Green and silicate slaty taconite with some interbedded silicate taconite (paint rock where oxidized)	0–125
Black slaty taconite (paint rock where oxidized)	0–35
Lower Cherty member	125–375
Rocks deficient in magnetite	7–75
Silicate taconite (white cherty taconite where oxidized)	......
Lean cherty-silicate taconite	......
Rocks rich in magnetite (the following sequence is typical, but omissions and reversals in order are common)	70–340
Irregularly banded or mottled silicate taconite	......
Mottled cherty-silicate or cherty taconite	......
Irregularly banded cherty or cherty-silicate taconite	......
Regularly banded cherty taconite	......
Brown and/or red and green slaty taconite, commonly interbedded with cherty taconite	0–80
Regularly banded cherty taconite	0–55
Rocks rich in hematite ("red basal taconite")	......
Jaspery taconite and hematitic slaty taconite	0–35
Iron-bearing sandstone (rarely, this unit underlies the unit listed below)	0–45
Conglomerates and algal structures, or chert	0–25
Total Biwabik formation between Mesaba and Coleraine	510–800

* The complete sequence is never found at any one place.

Upper Cherty, Upper Slaty) were first defined by Wolff (1917a, p. 148), who stated: "They are named from the predominant physical characteristic of the rock in them." However, the texture of taconite is in general related to its mineral composition, and Wolff apparently recognized this when he chose the term "cherty" rather than the term "granular."

TABLE 6. AVERAGE CHEMICAL COMPOSITION OF THE BIWABIK FORMATION, EXCLUDING THE HEMATITE-LIMONITE ORES

(Estimated by Gruner, 1946, p. 65)

Constituent	Per Cent	Constituent	Per Cent
SiO_2	51.0	CaO	1.1
Al_2O_3	1.0	CO_2	5.0
Total Fe*	27.0	P	.035
Mn	.4	Combined H_2O	2.2
MgO	2.8		

*About three fifths of the iron is in the ferrous state.

Because of lateral changes of lithology within units, the division of the formation into members on this basis is at places arbitrary, but it is nevertheless useful from both genetic and economic standpoints. In general, lithic units included in the slaty members contain 40 per cent or more of slaty taconite, although locally within a given unit the proportion of thin-bedded rocks may be less. The rocks of the slaty members characteristically contain little iron oxide, and the associated granular rock is silicate taconite. The cherty members, on the other hand, are rich in magnetite and contain cherty or cherty-silicate taconite in abundance. Locally, however, thin slaty units are included in the cherty members, and rocks more typical of the cherty members are included in the slaty members.

LOWER CHERTY MEMBER

The relatively uniform Lower Cherty member is thinnest on the East Mesabi and thickest near Calumet. The algal structures at the base of the Lower Cherty occur in separate concretionary mounds associated with conglomerate; therefore, many drill holes have failed to penetrate them. The unit of jaspery taconite and hematitic slaty taconite contains much calcite, and the lower beds are commonly of almost pure hematite. The lower three rock units of the member (Table 5), in which most of the iron is in the ferric state, are collectively called the "red basal taconite." The overlying unit of regularly banded cherty taconite, which contains both magnetite and hematite, is at many places separated from the succeeding strata, which are rich in magnetite, by a thin unit containing slaty rocks. In general, the amount of iron silicates in the thick magnetite-bearing sequence increases upwards. The layers of intraformational conglomerate scattered throughout the member have not been shown on Plate 1 because they are very thin and generally cannot be correlated from hole to hole.

The persistent unit at the top of the member is composed of nonmagnetic coarse-grained silicate taconite or, where partly oxidized and leached, of white cherty taconite. Although the contact between the Lower Cherty and Lower Slaty members is normally well defined, this uppermost unit contains interbeds of slaty taconite at a few places.

Lower Slaty Member

The black, green, or silicate slaty taconites (paint rock where oxidized) in the lower part of this member form the so-called "Intermediate Slate," the base of which is a good horizon marker along most of the range. At a few places in Rs. 16, 17, 18, and 22 W., however, the lowermost unit of the member consists of slaty taconite interbedded with cherty-silicate or silicate taconite, and the exact contact with the Lower Cherty member is somewhat difficult to locate. West of Nashwauk, the entire Lower Slaty member consists of paint rock and is generally less than 20 feet thick; in two of the deeper, less oxidized holes the fresh rock is dark green slaty taconite. Along the rest of the range, the lower few feet of the "Intermediate Slate" are at places oxidized to paint rock.

Gruner (1946, p. 45) includes "a lens of cherty, banded and irregularly banded taconite rich in magnetite, and having a maximum thickness of 140 feet," in the Lower Slaty member between Mountain Iron and Eveleth. In the present report, these rocks are considered as part of the Upper Cherty member instead of the Lower Slaty member.

One minor layer of algal structures and conglomerate occurs near the top of the Lower Slaty member in R. 15 W., and a few other thin conglomerate beds are locally present. On the Main Mesabi, the contact between the Lower Slaty and Upper Cherty members is gradational and must be located arbitrarily.

Upper Cherty Member

The Upper Cherty member is not so uniform as the Lower Cherty member. It is very thin near Calumet, and at Keewatin the abnormally great thickness of the overlying slaty rocks and the presence of some cherty rocks within the Upper Slaty member make it difficult to determine the boundaries of the members with any degree of accuracy. The unit containing much red and green slaty taconite is put in the Upper Cherty member because of the interbeds of magnetite-bearing cherty-silicate taconite, the high hematite content of the red slaty bands, and the location of the unit between thick units of cherty taconite at many places.

Between Mountain Iron and Biwabik, the textures of the main magnetite-bearing unit beneath the algal structures are complexly variable. The textural names listed in small print by segments of the drill holes in Plate 1 show that the textures of this unit in holes more than one quarter of a mile apart generally cannot be correlated. This suggests that the four textural varieties of cherty taconite — regularly and irregularly banded, mottled, and laminated — probably have much the same mineral composition.

The Upper Cherty member contains many intraformational conglomerates, the most prominent of which is associated with the nearly continuous algal structures near the middle of the member. These beds, which cannot be recognized west of Hibbing, form blocky and resistant layers in some of the open pits to the east of Hibbing. In Rs. 18, 19, and 20 W. the algal beds cross and recross the unit of cherty-silicate taconite and red and green slaty taconite. The contact between the Upper Cherty and Upper Slaty members is generally gradational.

Upper Slaty Member

Gruner (1946, p. 50) made the important discovery that in the West Mesabi some of the slaty rocks that had long been classed as argillites of the Virginia formation have in reality a high iron content and belong more properly to the Upper Slaty member of the Biwabik formation. Eastward toward Keewatin, where this cherty siderite rock pinches out, the rock grades into argillite, becoming grayer and having a lower specific gravity indicative of a lower iron content. Aside from iron content, further strong evidence that this unit actually belongs to the Biwabik formation is the gradual transition in Rs. 22 and 23 W. from cherty-silicate taconite, upwards through interbedded silicate taconite and cherty siderite rock, to cherty siderite rock having no associated granular taconite.

Minor algal zones occur in the lower part of the member near Eveleth and Buhl. A spherical mass of almost pure chert about 15 feet in diameter occurring at this same horizon is exposed in the Miller-Mohawk pit near Aurora. The chert is chalcedonic (fibrous) but contains none of the laminations typical of algal structures. Adjacent beds wrap around the mass concordantly.

Rocks of the Westernmost Mesabi Range

The Lower Slaty member pinches out near Grand Rapids, and west of this town the Lower and Upper Cherty members are joined as a single member; thus the normal fourfold subdivision of the Biwabik formation does not exist in the Westernmost Mesabi. On the basis of texture and composition the lower part of the formation can be referred to as a cherty member and the upper part as a slaty member (Table 7).

Both the thickness and the iron content of the cherty rocks diminish westward, and in eastern Cass County these rocks are only 20 feet thick and contain almost no iron-bearing minerals. The basal conglomerate in this area contains algal structures.

Partial analyses of the sequence of slaty rocks are given in Table 8. The lowermost unit of cherty siderite rock commonly contains more chert and less siderite than the other units of this rock. The lower unit of graphitic argillite generally seems to have a higher content of iron than the upper unit of this rock. (In the analyses shown in Table 3, the lower unit has an average of 14.13 and the upper unit has an average of 4.89 per cent total iron.) Some gray argillite is associated with the cherty siderite rock of unit 7 (Table 7) in Sec. 4:54-26.

TABLE 7. GENERALIZED LITHIC SEQUENCE WITHIN THE BIWABIK FORMATION IN THE WESTERNMOST MESABI RANGE*

(Rs. 26 and 27 W., Itasca Co., and R. 25 W., Cass Co.)

	Range of Thickness in Feet
Slaty member	100–405
10. Cherty siderite rock	0–35
9. Graphitic argillite	15–200
8. Quartz-mica argillite, with minor chert at top	0–100
7. Cherty siderite rock	0–15
6. Quartz-mica argillite, becoming cherty eastward	0–15
5. Graphitic argillite	10–50
4. Cherty siderite rock	0–30
Cherty member	20–450
3. Chert and lean cherty taconite	15–150
2. Relatively iron-rich rocks (in R. 26 W. only)	0–300
Oxidized cherty taconite	.....
Paint rock	.....
Regularly banded cherty taconite	.....
Jaspery taconite and hematitic slaty taconite	.....
1. Conglomerate and algal structures, or chert	0–10
Range of thickness of total Biwabik formation in the Westernmost Mesabi range	340–560

* Rock units are numbered to show correspondence with units in Table 8.

TABLE 8. PARTIAL ANALYSES OF SLATY ROCKS OF THE WESTERNMOST MESABI RANGE*

Rock†	Depth in Feet	SG‡	Total Fe	SiO_2	Al_2O_3	Mn	P
			VIRGINIA FORMATION				
Cherty siderite rock	217–265	3.19	16.74	50.27	2.97	1.36	.098
Argillite, with some cherty siderite rock**	265–315	2.58	7.44	64.97	4.28	.54	.219
			BIWABIK FORMATION (SLATY MEMBER)				
10. Cherty siderite rock	315–340	3.23	20.23	38.94	2.11	1.05	.405
9. Graphitic argillite	340–390	2.60	11.99	....	...	...	...
8. Quartz-mica argillite	390–440	2.67	7.50	....	...	...	...
7. Cherty siderite rock	440–455	3.07	16.89	....	...	...	...
5. Graphitic argillite	455–480	2.66	14.59	....	...	...	...
4. Cherty siderite rock	480–485	3.20	15.14	....	...	...	...

* Analyses are averages of combined core and sludge analyses made on 5-foot samples from a drill hole in the SE¼SW¼ Sec. 4: 54-26. Analyses furnished by Pickands Mather & Co.

† Numbers refer to the rock unit numbers shown in Table 7.

‡ Specific gravity.

** Analyses and specific gravity represent the argillite portion only.

VIRGINIA FORMATION

Van Hise and Leith (1901, p. 356) named the formation from its typical occurrence in drill holes and test pits near the town of Virginia. The thickness of the formation is unknown, but one drill hole in the SE¼ NE¼ Sec. 30: 58-15 penetrated 2000 feet of Virginia argillite, and the total thickness of the formation is probably much greater.

In the Mesabi district the formation consists mostly of argillite, generally having only a few beds and concretions of limestone near the base. Grout and Schwartz (1933, p. 11), however, found considerable graywacke and quartzite in addition to argillite in the Rove formation, which is the equivalent of the Virginia formation in northeastern Minnesota. On the Westernmost Mesabi range, an iron-bearing member, which at places is perhaps as much as 200 feet thick, is composed of cherty siderite rock with some argillite and is separated from the base of the formation by 50 to 140 feet of argillite containing some cherty siderite rock. This iron-bearing member thins eastward and pinches out at Calumet.

RELATIONS BETWEEN FORMATIONS

The lithologic difference between the Pokegama and Biwabik formations is very marked and the contact is generally sharp. Leith (1903, p. 167) concludes that a "minor erosional interval" occurred after deposition of the Pokegama formation, although he recognizes that the two formations are everywhere structurally concordant. He attaches great significance to the basal conglomerate of the Biwabik formation. Where the Pokegama formation is present, however, this conglomerate consists dominantly of taconite pebbles in a taconite matrix; a few clastic quartz grains but no pebbles of older rocks occur in it. Grout and Broderick (1919b, p. 7) state, "The conglomerate pebbles in most cases are not fragments of the Pokegama, such as would indicate its erosion; but they are chert fragments, probably indicating a breaking up of newly formed sediments above the sand." This conglomerate seems to have no greater significance than the other prominent intraformational conglomerates within the Biwabik formation. Where the Pokegama formation is missing, pebbles and boulders of granite and other older rock are imbedded in an iron-bearing matrix.

The occurrence of iron-bearing sandstone above the basal conglomerate of the Biwabik formation between Virginia and Gilbert suggests that sand deposition and iron deposition occurred at the same time. Apparently the Pokegama and Biwabik formations are perfectly conformable. Further evidence that they are genetically related is presented in the next chapter.

The relation between the Biwabik and Virginia formations is a subject of controversy. Most workers have considered these two formations conformable, but Wolff (1917a, p. 165) was one of the first to raise doubt. He stated: "Prior to the deposition of the Virginia slate, the iron-formation may have been raised above water, and its upper surface somewhat eroded." Leith *et al.* (1935, p. 15) regard an unconformity between these formations as "probable but still unproved."

The contact between the two formations is exposed in the Embarrass and Miller-Mohawk pits near Aurora and seems at these places to be conformable (Fig. 7). Over the rest of the range the contact area can be studied only in drill cores. As Wolff points out (1917a, p. 165), east of

FIGURE 7. — Contact between the Biwabik and Virginia formations at the Miller-Mohawk Mine near Aurora. Length shown is about 6 feet.

Nashwauk the two formations are interbedded to only a minor extent quantitatively. However, argillite and limestone beds occur to either side of the contact, and the limestone in particular seems significant even though its bulk is small. The limestone in both formations is continuous along nearly all of the Main Mesabi range and shows no evidence of erosion where covered by a thick sequence of Virginia argillite; where the limestone is locally decomposed close to its outcrop beneath the drift, the adjacent argillite beds are altered also. Likewise, the association of bleached Virginia argillite with some ore bodies shows that leaching and oxidation took place on the Main Mesabi after deposition of the Virginia formation.

Wolff (personal communication) believes that much of the extensive oxidation and leaching of the granular taconites of the West Mesabi range took place during an erosional interval prior to deposition of the Virginia formation. West of Nashwauk the contact between the two formations is sharp at some places but is definitely gradational at others. In R. 23 W., for example, the cherty siderite rock at the top of the Biwabik formation grades downwards and laterally into silicate taconite and grades upwards into argillite. The hybrid rock in the base of the Virginia formation has an appearance and specific gravity intermediate between cherty siderite rock and argillite, and lenses of more nearly pure cherty siderite rock are included in it. These gradations suggest continuous sedimentation.

Any variations in thickness of rock units near the top of the Biwabik formation can be more readily explained by variations in sedimentation than by erosion. Many of the thin conglomerates, which are located near the top of the Biwabik formation in R. 23 W., are shown on Plate 1. These beds are all intraformational in character, consisting of pebbles of adjacent rock types; they are scattered at various levels in both the Biwabik and Virginia formations and therefore seem to have little or no bearing on the question of unconformity.

The distribution of the oxidation and leaching on the West Mesabi does not suggest the presence of an unconformity. Oxidation and leaching are related to the erosion surface beneath the glacial drift, even though minor oxidation may extend as deep as 800 feet beneath the bedrock surface. Where the Biwabik formation lies directly beneath the drift, its rocks are generally much decomposed. But where the formation is covered by two or three hundred feet of Virginia argillite, the rocks of the Upper Cherty member are far less decomposed and the rocks of the Lower Cherty member are mostly fresh. Both argillite and cherty siderite rock are commonly somewhat decomposed at the bedrock surface but are fresher at depth.

If the iron-formation had been oxidized before deposition of the Virginia formation, the evidences of oxidation would not diminish so greatly down the dip of the beds. Unfortunately, however, no really deep hole has ever been drilled south of the West Mesabi, and absolute proof that the cherty rocks are completely unaltered at greater depth is lacking on this part of the range, although such proof is abundant on the Main Mesabi.

The contact between the Gunflint and Rove formations is not exposed at the type locality of the Animikie group at Thunder Bay (Tanton, 1931, p. 34), but Broderick (1920, p. 442), Gill (1924, p. 39), and Grout and Schwartz (1933, p. 5) regard the two formations as being conformable in the part of the Gunflint range near the International Boundary. The existence of an unconformity anywhere in the part of the Animikie group studied is very improbable. The Pokegama, Biwabik, and Virginia formations are here considered to be conformable, and a splitting of the group into "Middle Huronian" and "Upper Huronian" seems untenable.

ANIMIKIE EXTENSIONS IN MINNESOTA

The correlation between the Mesabi and Gunflint ranges is well established, but the relation between the Mesabi and Cuyuna ranges is a matter of lively controversy. The geology of the Cuyuna range is complex, and generalizations are not apt to be sound. For this reason the Cuyuna range itself is not discussed here, but it is perhaps worthwhile to point out the relations between the rocks of the Mesabi range and those of the Emily district, which lies between the Mesabi and Cuyuna ranges. (See Fig. 1 for location.)

The rocks of the Emily district are tightly folded, and the dips ascertained in the drill cores range from 0° to 90°, although most dips seem to be between 30° and 40°. It is difficult to determine the sequence of rock units except where the drill holes are long, so only the lower part of the column shown on Plate 1 is known with any accuracy.

Characteristics that the rocks of the Emily district share with those of the Mesabi range are as follows (in ascending stratigraphic order):

1. The lowermost formation contains quartzite. However, the formation contains much white, gray, green, or red soft sericitic argillite. In one hole that penetrated 235 feet (measured perpendicular to the bedding) into this unit, the lower half of the footage is sericitic argillite and the upper half is interbedded sericitic argillite and dark red quartzite. Mica flakes can be seen on the bedding planes of the quartzite.

2. Overlying the quartzite is an iron-formation containing ferruginous cherts and slaty rocks which are here classed as varieties of taconite for easier comparison with the rocks of the Biwabik formation, which they very much resemble. Most of these rocks are partly oxidized and leached, and the martite in them commonly forms rude granules. A few beds, however, contain clastic quartz grains.

a) The lower member of this iron-formation is cherty and contains algal structures at the base; these are overlain at one place by jaspery and hematitic slaty taconite. The rocks above these are regularly banded cherty taconite (50 to 80 feet thick), paint rock (5 to 10 feet thick), regularly banded cherty taconite (60 to 95 feet thick), and mottled cherty taconite (105 to 155 feet thick). The total thickness of this cherty member ranges from 255 to 300 feet. The thin paint rock layer near the base of the member is reminiscent of the slaty layer near the base of the Lower Cherty member on the Main Mesabi.

b) Overlying the cherty member is a slaty member, consisting of paint rock (25 to 90 feet thick).

c) Overlying this slaty member is a second cherty member of unknown thickness.

d) Units of quartz-mica argillite and graphitic argillite, lithologically similar to those in the upper part of the Biwabik formation on the Westernmost Mesabi, presumably overlie these other rocks, but the relationship cannot be proved.

3. In one drill hole, argillite like that of the Virginia formation overlies graphitic argillite.

The rocks of the two districts are only 30 miles apart and are geologically "in strike." The three formations at Emily appear to be similar in lithology and sequence to the three Animikie formations of the Mesabi district. The possible fourfold subdivision of the iron-formation at Emily is good evidence for correlation, but it should not be overemphasized. Perhaps the most important factor to bear in mind in this connection is the enormous lateral change in lithology and thickness which takes place within the Biwabik formation over a distance of 6 miles in R. 26 W. Profound changes might easily take place in the 30 miles between the explored end of the Mesabi range and the Emily district, and the correlation of individual members seems a more precarious undertaking than the correlation of formations. On the other hand, this same possibility of change can readily be used to explain the differences between the rocks of the two areas.

No conclusive evidence has been presented as yet to refute this correlation. The magnetic belt of the Mesabi range does not connect with the magnetic belt at Emily because the rocks of the Westernmost Mesabi are all nonmagnetic. There is no geologic reason why there may not be gentle folding of beds at one place and sharp folding of an extension of these beds 30 miles away. Indeed, a transition to more complex structure may well take place on the Westernmost Mesabi, for here (Plate 2) the Animikie beds are somewhat more deformed than they are farther east.

IV. ORIGIN OF THE ANIMIKIE ROCKS

INTRODUCTION

The origin of these ancient rocks, in particular the iron-bearing varieties that have no exact counterpart in the more recent sections of the geologic column, involves vexing questions, for many of which there are no ready answers. Yet the Mesabi district is an ideal area in which to investigate the origin of these rocks, because of the area's general freedom from structural and metamorphic complications, the abundance of data, and the completeness of the rock sequence and its instructive nature.

One aim is that of distinguishing "original" features — those formed during deposition or diagenesis — from features produced by later metamorphism or other alteration. "Diagenesis" is the change produced in sediments after deposition that leads to consolidation without the influence of unduly increased temperature or, of lesser importance, pressure. No sharp division can be recognized between diagenesis and metamorphism, but those changes which can be directly correlated with features produced by sedimentation seem more properly related to diagenesis than to metamorphism, especially where definite evidence of metamorphism is lacking. Rocks altered by oxidation and leaching are not considered here.

A number of hypotheses developed by other workers at various localities are examined in the course of this and later discussions. These hypotheses are evaluated only for their applicability to the Mesabi range, largely on the basis of evidence available at the Mesabi itself. The conclusions apply specifically to the Mesabi and may or may not have application in other localities.

ORIGIN OF THE MINERALS

Chert

Evidence suggesting both deposition of silica and replacement by silica is present. At the Gogebic range of Wisconsin, Aldrich (1929, p. 151) reports rhomb-shaped grains of chert in which the rhombohedral cleavage of a carbonate is preserved. Because the distribution of oxides in some chert granules and in some carbonate granules is similar, he believes that some granules of carbonate were replaced by chert. Aldrich concludes that "silicification of carbonate is evidenced on a tremendous scale," although he also believes (1929, p. 162) that much silica was originally precipitated as a gel.

Some oölites in the taconites of the Biwabik formation consist of concentric layers of chert or jasper, and at places large grains of quartz cut across several of the layers, suggesting either replacement or recrystallization. Pettijohn (1949, p. 75) points out that all oölites formed in recent times are calcareous and that older noncalcareous oölites generally show evidence of replacement. Nevertheless, oölites are far less abundant than the structureless granules in taconite, and the granules show no certain evidence of replacement.

Most geologists agree that the algal structures of the Biwabik formation are of organic origin, because the forms so closely resemble some modern growths of algae. Gruner (1922, p. 420) reports microscopic bacilli and algae in the algal beds, but some observers (Tanton, 1931, p. 46) consider these to be of inorganic origin. Because most modern algae other than diatoms are calcareous, the siliceous algal structures in the Biwabik formation may be replacements. But Grout and Broderick (1919a, p. 202) state: "There is no reason to expect early forms to be so uniformly calcareous as recent forms." Clark (1916, p. 604) contends that siliceous organisms would be likely to thrive in waters rich in silica. Moreover, even if the algal structures were replaced by silica, it does not necessarily follow that all the cherts of the formation are replacements, since calcareous organic structures throughout the geologic column commonly seem to have a greater affinity for silicification than their calcareous host rocks.

Many of the arguments favoring deposition of chert are negative in character, for direct evidence is elusive, and the arguments must therefore be based on the general lack of evidence for replacement. None of these arguments conclusively eliminates the possibility of a wholesale replacement that left almost no record of the pre-existing material, but they perhaps combine to show that such a possibility is unreasonable. Certainly, clear-cut evidence of replacement is scarce. Most of the grains and rhombs of carbonate embedded in chert show no evidence of replacement. Angular pebbles of chert in intraformational conglomerates strongly suggest (but do not prove) that this chert existed as such at the period in which the layers were broken up during sedimentation. Laminations wrap over and around these pebbles and other chert nodules and do not pass through them. Stylolites, which presumably formed in the soft horizontal sediments not long after deposition, cut through chert beds.

Bedded chert is abundant over great areas at definite stratigraphic levels, and the distribution of this chert is uniform and bears no relation to folding and faulting. The limestone layer at the top of the Biwabik formation contains chert beds at places, but these beds are granular whereas the limestone is nongranular, and the rocks therefore seem to be related by sedimentary interbedding rather than by replacement.

If silicification occurred after burial of the sediments, the source of so much silica and the thorough permeation of silica throughout the formation would be difficult to explain. However, much silicification could have taken place almost contemporaneously with deposition. The problem of the origin of the chert remains unsolved, but the following assumption is made as a basis for further discussion: the chert of the Biwabik formation is believed to be a product of deposition, and the few textures suggesting possible replacement are thought to result from the diagenetic interaction of part of this silica with the other constituents of the sediment.

Iron Silicates

Greenalite is generally considered to be an original mineral (Spurr, 1894, p. 227; Leith, 1903, p. 259; Jolliffe, 1935, p. 405; Gruner, 1936, p. 453; Tyler, 1949, p. 1102) and no evidence to the contrary has yet been produced.

The origins of stilpnomelane and minnesotaite are, however, not generally agreed upon. Gruner (1946, p. 12) believes them to be original minerals: "As stilpnomelane forms from, let us assume, a colloidal gel, it will take the ions in its neighborhood which are most convenient and of the necessary charge. If it cannot find any more it will stop growing. The leftover gel material, then, may be of the proper composition to form minnesotaite or greenalite, or quartz and siderite, if CO_2 is available in considerable concentration."

Tyler (1949, p. 1103), on the other hand, regards stilpnomelane as a product of low-grade metamorphism, because some of the stilpnomelane in the iron-formations of the Lake Superior district is related to dikes and veins, and because the mineral occurs in metamorphic rocks in other parts of the world. He assigns the same metamorphic origin to minnesotaite, because of its "association with stilpnomelane and magnetite." Tyler contends that regional metamorphism and local metamorphism next to dikes, sills, and veins has caused minnesotaite and stilpnomelane to form at the expense of chert, siderite, and greenalite. Underlying this contention is the premise that most varieties of rock now found in the iron-formations are the metamorphosed and/or weathered derivatives of a single phase of "original rock" composed chiefly of chert, siderite, and greenalite. This concept of the "original rock" was advanced by Irving and Van Hise (1892, p. 246) and Spurr (1894, p. 227), and it has been supported by Leith (1903, p. 101), Mann (1953, p. 259), Tyler and Twenhofel (1952, p. 128), and others.

Tyler (1949, p. 1108) further holds that "stratigraphic horizons with a high iron content, a granular texture, thin alternating beds of chert and siderite, or zones located adjacent to sills may be altered almost completely to iron silicate rock." He calls this process "stratigraphic alteration" and believes that the formation of new minerals tends to destroy banding and to produce a slaty appearance.

The problem at hand is to determine whether the present distribution of these iron silicates can be more easily related to sedimentation or to subsequent local and regional metamorphism. The taconites of the East Mesabi range, which were metamorphosed by the Duluth gabbro, contain a rather high-grade metamorphic mineral assemblage consisting of relatively coarse quartz,* grunerite, carbonate, garnet, fayalite, magnetite, and minor pyroxene. Near Mesaba, these rocks grade rather abruptly into taconites composed dominantly of fine quartz, greenalite, minnesotaite, stilpnomelane, magnetite, and siderite. Such taconites are characteristic of the entire Main and West parts of the range, and no further lateral metamorphic gradient is evident.

The rocks of the Westernmost Mesabi contain practically no magnetite or iron silicates, but the disappearance of these minerals coincides with a profound lateral change in lithic type, sequence, composition, thickness, and texture, a change which can be explained only by reference to sedimentation. Furthermore, the grain size of the quartz of the Westernmost Mesabi is no different from that of the West Mesabi, nor is there any textural or mineral change in the units of cherty siderite rock which extend through both areas. Again, no metamorphic gradient is evident.

About eight small dikes and one larger sill are known to cut the iron-formation between Gilbert and Mesaba. The contact effects of the sill have been studied carefully and are reported in more detail in the next chapter; the effects are minor, however, and the distribution of minnesotaite and stilpnomelane shows no relation to the sill's contacts, which at places are discordant to the bedding. A few veins of what is probably hydrothermal quartz occur in the taconites and ores east of Hibbing, but none of those observed in fresh drill core produced any noticeable effect on the taconite more than an inch or two from the contact. However, some coarse stilpnomelane occurs within a few of the veins (Grout and Thiel, 1924, p. 228). Obvious sources for local metamorphism other than the Duluth gabbro seem inconsequential.

A study of Plate 1 shows that the distribution of the iron silicates is stratigraphically controlled on a dominantly regional rather than a local scale. The greatest concentration of iron silicates is in the slaty members. There is no reason to believe that any one thin but laterally extensive unit of the Biwabik formation should have been any more or less regionally metamorphosed than any of the other units above or below. The grain size of the chert in cherty siderite rock is no different from that in silicate-bearing taconites. An extensive unit of cherty siderite rock on the West Mesabi contains little or no iron silicate, although these minerals may be abundant in adjacent units. As Gruner (1946, p. 31) notes, a nearly pure thin bed of each of the minerals composing taconite is found interlayered, at one place or another, with an equally

* Tyler (1949, p. 1104) uses the grain size of quartz as a metamorphic indicator, considering coarser grains the products of stronger metamorphism. This interpretation seems generally reliable insofar as the Biwabik formation is concerned, but it should be pointed out that the grain size commonly varies widely in a single thin section and may therefore be subject to broader variations not related to metamorphism.

pure thin bed of each of the other minerals. It is difficult to see how such a state of inequilibrium could survive any metamorphism credited with producing some of these minerals from the others. Differences in chemical composition — for example, the distinctively higher Al_2O_3 content of the slaty silicate-bearing units — can be far more readily explained by sedimentary differentiation than by metamorphic differentiation.

Any control, by texture, of "stratigraphic alteration" of other minerals to iron silicates, as proposed by Tyler, seems doubtful. The silicate-bearing slaty rocks are thin-bedded, whereas most of the granular rocks are either thick-bedded (banded) or contain little or no bedding whatsoever. Metamorphic production of thin beds from thick beds, or in places where no beds existed, is not likely. The bedding characteristics of units of the Main Mesabi can be traced into the highly metamorphosed East Mesabi, where, as Grout and Broderick (1919b, p. 9) state, "white quartz bands are much more conspicuous than the original gray chert bands." The thin beds of the siderite-cherts are commonly somewhat wavy and nodular, whereas the thin beds of silicate slaty taconite are generally very even and straight. Because each rock has its own distinctive bedding, it is not likely that one could be derived from the other.

Gruner (1944, p. 371) states: "Many of the green granules which look exactly like greenalite . . . have been identified as minnesotaite by X-rays." He also notes (1946, p. 15): "Occasionally the observer is not certain whether he is examining greenalite or very fine-grained stilpnomelane, since both of them occur as granules." Inasmuch as granules are generally regarded as original features, and greenalite is commonly assumed to be an original mineral, and the grain size of all three iron silicates in many granules is equally fine, it is not unreasonable to assume that minnesotaite and stilpnomelane in this form are original constituents.

However, the fibers of the abundant minnesotaite and stilpnomelane in the matrix of some taconites generally penetrate grains of quartz and, at a few places, grains of siderite and granules of silicate. Some dark green, nearly opaque silicate granules are encompassed by a sheath of minnesotaite fibers cutting the interstitial quartz. Replacement is possibly suggested, but in view of the sedimentary control of silicate distribution, any such replacement is here believed to have taken place during diagenesis by reaction between small, adjacent, but chemically different portions of the sediment. James (1951a, p. 261) reports similar relations, such as replacement of clastic quartz grains by a chamosite matrix, occurring in the younger unmetamorphosed ironstones of the world.

The possibility (James, 1954) that some of the iron silicates were derived by mild metamorphism from some pre-existing silicate material cannot be eliminated, but direct evidence for such an intermediate stage is lacking.

Although it is believed that almost all stilpnomelane and minnesotaite in the Biwabik formation are original constituents, the production of minor amounts by local metamorphism, in the manner envisioned by Tyler, is neither impossible nor necessarily inconsistent with this belief.

Magnetite

It has been generally assumed that the magnetite in Precambrian iron-formations is a metamorphic mineral, and Tyler (1949, p. 1104) believes that magnetite was formed from siderite and greenalite where oxygen was available during metamorphism. However, the fact that the distribution of magnetite, like that of the iron silicates, has a definite relation to sedimentation suggests that it too is an original constituent. The lines of evidence supporting this contention are mustered below.

It is believed, as noted above under the discussion of the iron silicates, that the Biwabik formation west of Mesaba is essentially unmetamorphosed, since the chert is generally fine-grained, the minerals give evidence of no metamorphic gradient, and obvious sources of local metamorphism are negligible. Any effect of possible deep burial was insufficient to produce changes that can clearly be recognized as metamorphic rather than diagenetic. Certainly, different stratigraphic levels were not subjected to any different grades of metamorphism; yet bedded magnetite occurs consistently at specific stratigraphic levels within the Biwabik formation. The disappearance of magnetite on the Westernmost Mesabi can readily be correlated with changes in sedimentation in that area.

Moreover, the rather highly metamorphosed taconites of the East Mesabi contain little more magnetite, bed for bed, than the rocks of the Main Mesabi. Gruner (1946, p. 73) states: "The changes in the iron minerals brought about by the metamorphism on the East Mesabi range did not produce additional Fe_3O_4 but iron amphiboles, olivines, and pyroxenes. The chief reason was undoubtedly the unavailability of additional oxygen." If oxygen were unavailable on the East Mesabi, it would probably likewise be unavailable on the Main Mesabi during any supposed metamorphism there. The grain size of magnetite on the East Mesabi is small, like that on the Main range, although aggregates of grains are commonly larger and more compact on the East Mesabi than on the Main Mesabi (Gruner, 1946, p. 72). However, some magnetite on the East Mesabi has been derived from hematite (Schwartz, 1923, p. 412), and Grout and Broderick (1919a, p. 201) observe that the normal red color of the algal beds is changed to gray by the alteration of hematite to magnetite. The amount of magnetite so produced must be relatively small, because the amount of hematite in the fresh taconites of the rest of the range is relatively small.

Although any mineral in the taconites west of Mesaba may at places be associated with any other mineral, the association of magnetite with siderite or iron silicates is not marked. Magnetite is concentrated in the cherty members and is characteristically rare in the silicate-bearing slaty members. The magnetite in magnetic green (silicate) slaty taconite is not nearly as abundant as it is in the common varieties of cherty and cherty-silicate taconite. Silicate taconite was defined as being nonmagnetic simply because most of it contains no magnetite. Magnetite-siderite rocks are among the rarest of all types of taconite, and the extensive units of cherty siderite rock on the West and Westernmost parts of the range are completely nonmagnetic throughout.

As Gruner (1946, p. 72) notes, the mineral most commonly associated with magnetite is quartz. In addition, magnetite has unique bedding characteristics not typical of any other constituent: irregular bands and, to a lesser extent, mottles and regular bands having gradational boundaries. Sharply defined flat pebbles of nearly pure magnetite in intraformational conglomerates were, presumably, first consolidated as beds of magnetite which were later broken up and incorporated in the next-deposited layer of sediment. Coatings of magnetite, which are generally about one-quarter inch thick and grade outwards into the matrix, surround some pebbles of other composition; it is likely that this magnetite accumulated as the pebbles were rolled around during sedimentation. In view of these mineral and textural associations, it is very difficult to see how magnetite could have been derived from any of the other minerals now contained in taconite.

Brown (1943, pp. 138–141) describes several occurrences of original magnetite in the younger unmetamorphosed ironstones of France, England, and Russia. Weiser (1935, pp. 88–89) reviews the conditions under which magnetite has been synthesized at room temperature and pressure. Spiroff (1938, pp. 818–828) cites reports of original magnetite in modern sediments. James (1954) states that much of the magnetite in the iron-formations of the Lake Superior district is original, and he summarizes additional literature on the occurrence of sedimentary magnetite. Gruner (personal communication) has believed for several years that magnetite is an original constituent of the Biwabik formation. Spencer and Percival (1952, p. 378) regard the magnetite in the Precambrian iron-formation at Singhbhum, India, as primary. Another example of sedimentary magnetite is found in the unmetamorphosed conglomerates of Cretaceous age which overlie the Biwabik formation. The ferruginous cements of many of these ore conglomerates contain unabraded crystals of magnetite.

As seen under the microscope, octahedra of magnetite are generally imbedded in grains, granules, or a matrix of the other minerals. Such a relation is often used as a criterion of replacement, but the interpretation that it results at many places from the earlier crystallization of magnetite during either deposition or diagenesis is highly probable. In discussing the general case of such textures, Bastin (1950, p. 60) states: "The replacement alternative should be avoided unless evidence of replacement is definite, for . . . replacement usually betrays itself by spatial relation to fractures, by selecting certain minerals as host, or in other ways."

Most of the few obvious replacement relations, such as magnetite replacing oölites of hematite, can logically be ascribed to diagenetic reactions caused by the usual increase in reducing tendencies after burial of newly formed sediments (Krumbein and Garrels, 1952, p. 16). A few of the more pronounced irregularities of magnetite bands and aggregates in irregularly banded or mottled taconite are interpreted as replacement textures by some observers; probably most of these can just as well be related to local irregularities of sedimentation, and the rare instances of cross-cutting stringers and veinlets of magnetite are here ascribed to modifications produced during diagenesis.

Other Minerals

Carbonates. The bulk of the carbonate in the Biwabik formation is generally regarded as original, although Leith (1903, pp. 101, 102, 120, 150), who believes the "original rock" was composed mostly of greenalite and chert, assigns a replacement origin to much of the carbonate. Some larger rhombs are imbedded in other minerals, but in most cases this texture could have resulted from the earlier crystallization of the carbonate; an example indicating a sequence in crystallization is the cockscomb structure of quartz which has grown outward from the faces of calcite rhombs contained in some of the chert lenses in cherty siderite rock. In the absence of evidence of replacement except on a small scale, carbonate is believed to be largely a product of deposition, and the remainder is assumed to have been produced during diagenetic reactions. Likewise, pyrite and carbonaceous material are considered to be original constituents.

Hematite. The most conspicuous concentration of hematite in the Biwabik formation, other than the soft ores, occurs in the "red basal taconite." This stratigraphic unit, which can be traced along the entire Main and West parts of the range, contains a remarkably consistent succession of rock types. No other than a sedimentary origin can be logically assigned to the hematite in these beds. Gruner (1922, p. 8) has summarized the arguments showing that these rocks cannot be products of secondary oxidation and leaching: the distribution of the beds has no relation to ore bodies; this hematite is found in the deepest holes ever drilled south of the range, where the entire formation is completely unaffected by secondary oxidation; the hematite is not pseudomorphous after magnetite, and it is associated with considerable carbonate, which is one of the first minerals removed by leaching.

Original hematite in taconite occurs as laminations, as small scattered plates, as layers in oölites, or as dust. Such hematite is in the stratigraphic units of interbedded red and green slaty taconite and also in the jaspery taconite that is commonly associated with the intraformational conglomerates in the cherty members. Gill (1924, p. 42; 1927, p. 702) regards hematite having these relations in the Gunflint district as primary.

Wolff was among the first to question the concept of the "original rock" and to suggest the sedimentary origin of both magnetite and hematite. He states (1917b, p. 234): "The great mass of iron was laid down as original oxide . . . cemented together in an amorphous silica matrix." James (1951a, p. 259) has also doubted that one rock type could be the predecessor of all others. Because of the individual sedimentary characteristics and distribution of each rock type in the Biwabik formation, the concept of an "original rock" is here considered untenable.

THE PATTERN OF SEDIMENTATION

The Basis for Interpretation

Although many observers have regarded iron sedimentation as a very special circumstance unrelated to "normal" sedimentary processes, Twenhofel (1950, p. 429) states that chemical sediments could have been precipitated in a sea at the same time that clastic sediments were being deposited closer to the shore line. Twenhofel adds: "Rises of water level would have led to migration of the areas of iron deposition over those of clastic sediments, whereas falls of sea level would have produced the opposite effect." This recognition of the possible contemporaneity and the migratory nature of associated environments of deposition is a sound interpretation based on modern knowledge of the origin of sedimentary rocks. Tyler and Twenhofel (1952, p. 1) have shown that the application of these concepts to Precambrian rocks too often has been overlooked. Fortunately, direct evidence concerning the pattern of sedimentation is present in part of the Mesabi section.

As McKee (1949, pp. 36–40) has pointed out, lateral changes of lithology within a genetically related body of sediment are brought about in two ways:

1. In a section parallel to a shore line, lateral gradation of one lithology into another may result from local variations in the source area, in the configuration of the sea bottom, in the transporting power of adjacent streams, or in other such factors. A deposit made directly offshore from a river mouth may differ somewhat from one made at the same time in the area between river mouths, even though the two deposits are equally distant from the shore. In general, however, the aspect of sediments formed simultaneously at a given distance from shore, as observed in such a section, is characterized by an over-all uniformity which at most places outweighs the changes produced by lateral gradation.

2. In a section perpendicular to a shore line, greatly differing sediments may be deposited simultaneously at different distances from shore. In such a section, the effects of advances (transgressions) and retreats (regressions) of the sea can best be observed. In the most simple case, each environment of deposition is in the form of a belt parallel to the

shore at a given distance from it. As the sea advances, each of the coexisting series of environments advances with it, and the lithology produced in each environment is formed continuously at its particular distance from the migrating shore. Thus one type of deposit is laid down over and covers the other deposits that previously had been formed closer to shore, and a sequence of differing rock types is built up. The shoreward extension of each lithic unit becomes progressively younger with the continued advance of the sea. If the sea then retreats, the different environments of deposition are carried back, each maintaining its relation with the others and with the shore. Near-shore deposits come to be laid down where offshore deposits had previously formed. Such movements of the sea produce wedge-shaped, intertongued lithic units, as shown in Figure 8.

The Mesabi range represents a section of rock somewhat arbitrarily exposed to view by earth forces which were not related exactly to the configuration of the original basin of deposition. This section is a long, narrow, relatively straight belt which, for the most part, gives only a two-dimensional picture of the distribution of the rocks. Along the entire Main and West parts of the range, the Biwabik formation can be divided into four members, each characterized more by lateral uniformity than by sharp differences in lithology; this part of the section, then, is similar to the general case of a section roughly parallel to the old shore line. On the Westernmost Mesabi, however, the rock types differ from

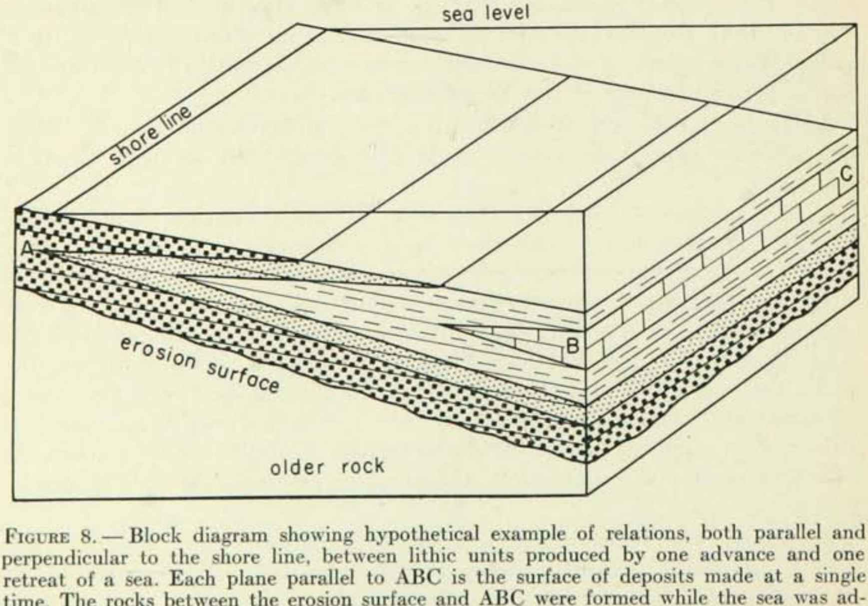

FIGURE 8. — Block diagram showing hypothetical example of relations, both parallel and perpendicular to the shore line, between lithic units produced by one advance and one retreat of a sea. Each plane parallel to ABC is the surface of deposits made at a single time. The rocks between the erosion surface and ABC were formed while the sea was advancing toward the left; the rocks above ABC were formed while the sea was retreating toward the right. Patterns show the lithic units produced by these migrations of the environments of deposition.

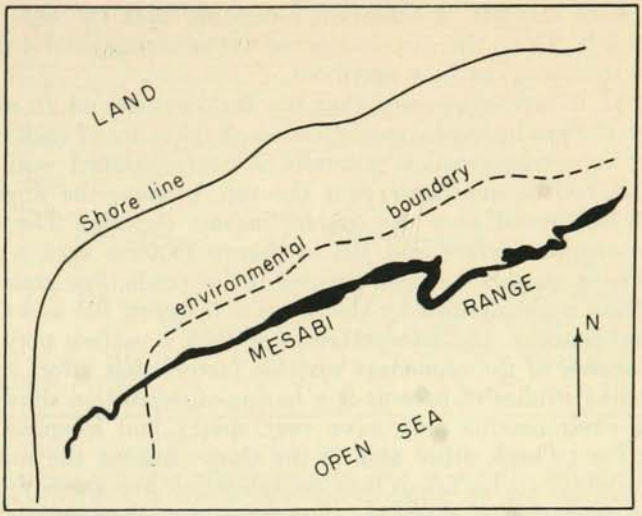

FIGURE 9. — Position of the presently exposed Mesabi section relative to one probable position of the old shore line of the Animikie sea. Only one of several parallel environmental boundaries is shown.

those of the rest of the range, and the lithic units are intertongued; these relations are those of a section essentially perpendicular to the old shore line.

Because the exposed Mesabi section is relatively straight, this situation could only have been produced by a major bend in the old shore line occurring somewhere near the western part of the district. The shore line migrated, but it must have been situated generally either north or south of the present Mesabi range while the iron-bearing rocks were being deposited. Because rocks correlated with the Animikie group lie to the south, at the Emily district and also in Wisconsin, the assumption is made that the closest shore of the basin of deposition was generally north of the Mesabi range (Figure 9).

The history of advances and retreats of a sea can be read more clearly in a rock section perpendicular to the old shore line, and therefore the rocks of the Westernmost Mesabi are the key to an understanding of the conditions of sedimentation. The actual and assumed relations between lithic units in this area are generalized in Figure 10, which is based on sheet 4 of Plate 1. The intertonguing of lithic units shown in the figure suggests that the sequence was built up by the migration of a series of coexisting environments of deposition. Line BB of Figure 10 is like line AB of Figure 8. Each of these lines is part of a surface of deposits that were made at a single time, and the different lithic units crossed by each line are therefore products of contemporaneous environments of deposition. The rock types crossed by BB of Figure 10 are progressively coarser toward the left, a condition indicating that the old shore line was to the left. Thus, the distribution of the environments of deposition relative to the shore can be worked out.

In general, it can be assumed that the first deposits on an old erosion surface are the products of a transgression. A sequence of rock types produced during a transgression generally is coarse-grained near the base and fine-grained or nonclastic near the top, because the finer offshore deposits were carried over the coarser inshore deposits. The rocks between the erosion surface and BB of Figure 10 form such a sequence. In a regressive sequence, coarse-grained rocks overlie fine-grained rocks; this condition is exemplified by the deposits between BB and CC.

In actual practice, the interpretation of such a pattern may be made difficult because of the numerous variable factors that affect sedimentation. Detailed studies of present-day basins of deposition show that the coexisting environments may have very spotty and irregular distributions relative to each other and to the shore. Among the many interrelated variables to be taken into consideration are earth movements, rate of sea migration, character of the source area, chemical and physical nature of the mediums of transportation and deposition, climatic factors, biologic factors, sea-bottom configuration, and the character of the sedimentary material. Obviously, the reconstruction of ancient environments requires many generalizations and some outright guesses.

The generalized basis for the interpretation that follows is shown in Figure 10, but Plate 1 must be consulted for details.

The Interpretation

Origin of Pokegama and Biwabik formations west of Nashwauk. The pre-Animikie erosion surface contained a few irregularities, but it was probably for the most part a broad, relatively flat plain. That the land mass was deeply eroded is attested to by the pre-Animikie exposure of the Giants Range granite. The present topographic ridge, the Giants Range, could not have existed in a form that might block an advance of the sea, for the distribution of rocks relative to the shore (Fig. 9) indicates that the old shore line must have been at least several miles north of the present Mesabi range at its farthest advance. The rocks making up the surface of this land mass were probably those of the Keewatin and Knife Lake groups, as well as granite; if any post-Algoman but pre-Animikie sediments were deposited, they were eroded away, for no record of them remains.

Subsidence presumably caused the Animikie sea to spread slowly across this low-lying land mass. The beach and inshore deposits of clastic material were spread as a blanket (the Pokegama formation) during the sea's continued advance. Local irregularities in the thickness of the Pokegama formation probably are to some extent a reflection of the local irregularities of the surface crossed by the advancing sea; it is likely that sands filled depressions but were deposited thinly over hills. The

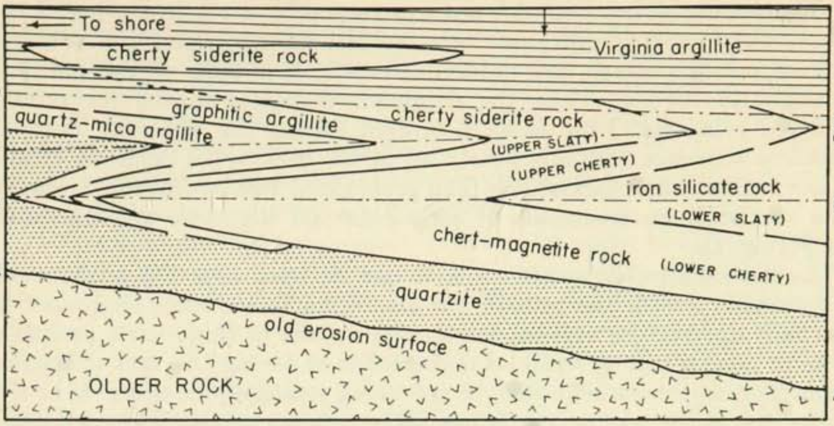

FIGURE 10. — Section perpendicular to the old shore line in the Westernmost Mesabi district showing the generalized relations between lithic units of the Animikie group (not to scale). Rocks between A and B were formed during the first transgression; rocks between B and C were formed during the first regression; rocks between C and D were formed during the second transgression; the nature of sea movements during which the rocks above D were formed is uncertain. The vertical arrow shows the general position of the West Mesabi section, which is essentially parallel to the old shore.

Pokegama formation may not everywhere be the product of a single advance of the sea, but for the purposes of this discussion the entire Pokegama formation in the Mesabi area is assumed to be a product of the first advance.

During this first advance much nonclastic iron-bearing sediment was being deposited seaward from the Pokegama environment. The contemporaneous deposits being formed, in order of increasing distance from shore, were (1) clean sand; (2) silt and clay (missing at many places); (3) lean chert, with algal structures; (4) hematitic rocks; (5) minor chert-siderite rocks; (6) chert and magnetite, with more iron silicates seaward; and (7) iron silicates, with some siderite. As the environments migrated, their deposits were built up into a succession of lithic units which, listed in ascending stratigraphic order, are now as follows: the Pokegama formation, consisting of (1) quartzite, having (2) quartz-mica argillite at the top (locally on the Westernmost Mesabi); the Lower Cherty member of the Biwabik formation, containing (3) lean chert and algal structures, (4) jaspery and hematitic taconite, (5) minor cherty siderite rock, and (6) extensive chert-magnetite taconite having more iron silicates at the top; and the lower part of the Lower Slaty member of the Biwabik formation, consisting chiefly of (7) silicate slaty taconite.

The sea did not advance far enough to carry the offshore deposits across the entire Westernmost Mesabi, and each of these deposits pinches out in turn toward the west, leaving only the inshore deposits farthest west. The reason for this, according to this interpretation of the record, is that here the sea began its first retreat.

The retreat of the Animikie sea was caused by a relative rising of the land. The increased difference between the levels of land and sea produced a difference in the nature and rate of supply of the material brought to the sea. Thus, the sediments deposited during the retreat differed somewhat from those deposited during the advance. This regressive sequence contains nearly all of the rock types common to the Lake Superior iron-formations. The generalized spatial relations between the coexisting environments of deposition and the shore are illustrated in Figure 11.

As the inshore environments were carried back over the offshore environments during the retreat, the stratigraphic sequence continued to be built up, but the vertical stratigraphic order of lithic units produced was in general the reversal of that produced during the first transgression. The regressive sequence, in ascending stratigraphic order, is as follows: the upper part of the Lower Slaty member of the Biwabik formation, consisting chiefly of silicate slaty taconite; the Upper Cherty member of the Biwabik formation, consisting dominantly of chert-magnetite rocks having more iron silicates at the base and lean chert at the top; cherty siderite rock; graphitic-pyritic argillite; quartz-mica argillite having local cherty siderite rock at the top; and quartzite (assumed).

The quartz-mica argillite now found intertongued with iron-rich sediments on the Westernmost Mesabi is thought to be a later product of the same migrating environment in which the upper part of the Po-

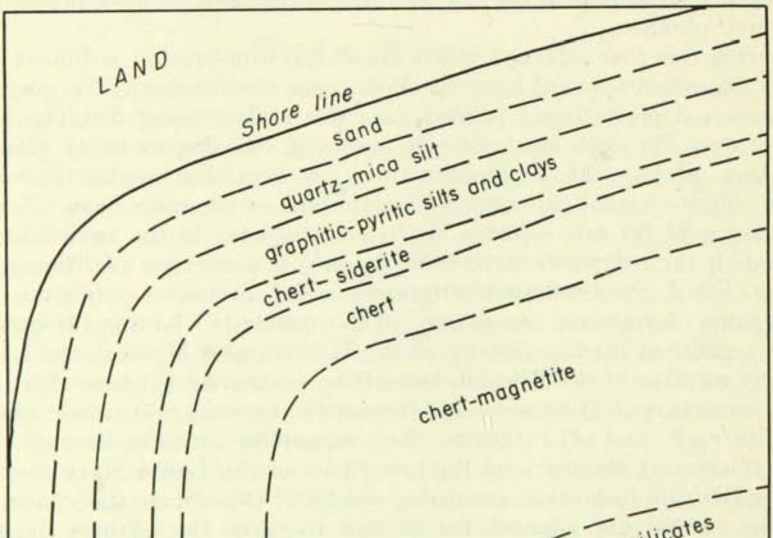

FIGURE 11. — Generalized relations between contemporaneous environments of deposition existing during the first regression of the Animikie sea.

kegama formation was produced. It is also assumed that a tongue of sandstone was included within this quartz-mica argillite, but that the western end of the exposed section was not close enough to shore to contain this sand; however, such a near-shore deposit is apt to be reworked or destroyed during a regression. The rocks of the Upper Cherty and Lower Cherty members are likewise considered to be the products of a single migrating environment of deposition.

The Animikie sea then stopped its retreat and began to advance again, presumably because of subsidence of the fringe of the land mass. The record of this second advance is more difficult to interpret because the advance was accompanied by changes in conditions which ultimately altered the entire pattern of sedimentation. The record on the Westernmost Mesabi indicates the following near-shore part of the series of deposits being made at the same time during the advance, listed in order of increasing distance from shore: sands (assumed); quartz-mica silts with minor near-shore chert-siderite; graphitic silts and clays; and chert-siderite. The offshore deposits, such as chert-magnetite and iron silicates, were probably still being formed but were not carried across the site of the Westernmost Mesabi.

Origin of Biwabik formation east of Nashwauk. The Main Mesabi range is roughly parallel to the old shore line; it is not exactly so, however, and variations in lithology are due not only to lateral gradation parallel to the shore but also to the varying distances of parts of the exposed range from the mean position of the shore line. East of Virginia, the present Mesabi section is farther from the old shore, because the folds associated with the "Virginia Horn" have caused rocks that were originally farther south to be exposed (Fig. 9). The evidence is largely the increasing thickness of the silicate-bearing Lower Slaty member toward the east, inasmuch as greater thicknesses of these offshore deposits could be built up only in areas generally more distant from shore.

The interpretation concerning the first transgression and regression can be readily applied to the rocks of the Main Mesabi range. The determination of the effects of the second transgression in the Main Mesabi sequence is, however, subject to much uncertainty. Near Keewatin the Upper Slaty member apparently grades laterally from cherty siderite rock eastward into silicate taconite, brown and green slaty taconites, and limestone; a few cores from the Keewatin area indicate that the relations between lithic units are complex, but the exact pattern could not be worked out from the information obtained. Because the Upper Slaty member east of Keewatin consists of a vertical sequence not having any repetition of lithic units, and the rocks themselves are a mixture of what seem to be both near-shore and offshore types, it is difficult to decide whether the movement producing this member on the Main Mesabi was transgressive, regressive, or both. None of these three interpretations is entirely satisfactory; the solution must await further detailed work in the Keewatin area.

The iron-bearing sandstone near the base of the Biwabik formation in the Virginia-Eveleth area is probably the product of a minor fluctuation in sea level that was rapid enough to cause intermingling of chemical and clastic deposits.

The crossing of a lithic unit by the algal structures in the Upper Cherty member in Rs. 18 and 19 W. (Plate 1, sheet 2) might be due to the growth of algae on top of two adjacent environments having a somewhat irregular contact. The minor slaty layers in the Upper Cherty member might be due either to minor fluctuations in sea level or to local irregularities in bottom configuration, in which lenses of differing deposits were formed.

Origin of the Virginia formation: The relation between the sediments of the Biwabik and Virginia formations is a major problem. Although some of the very fine-grained chloritic argillite of the Virginia formation was probably being formed contemporaneously with iron-bearing sediments, the change from chemical sedimentation to clastic sedimentation was apparently rather abrupt. It is likely that such a change in sedimentation would be a result of an abrupt change in the source area, if the sedimentary sequence is considered conformable. One explanation of the advent of clastic deposition would be that a period of vulcanism in the source area began. With an abundance of fine-grained volcanic debris available at the surface, streams could become heavily loaded with fine clastics. Van Hise and Leith (1911, p. 614) state: "Contemporaneous basic igneous extrusions . . . doubtless furnished an unusual source for mud." Such a vast bulk of this material might be contributed to the sea that deposition would necessarily spread across a number of pre-existing environments of chemical sedimentation.

The evidence is largely on the Gunflint range. Both Gill (1924, p. 64) and Tanton (1931, p. 35) describe ellipsoidal lavas included within the Gunflint formation near the upper algal beds. These flows perhaps represent the beginnings of a period of vulcanism which reached its climax at a somewhat later time. Gill (1924, p. 67) reports tuffaceous material, such as angular fragments of orthoclase and plagioclase, in some beds of the Upper Slaty member of the Gunflint formation. Locally, the beds of this member are folded and brecciated, and Gill (1924, p. 40) concludes that "violent volcanic disturbances occurred nearby toward the end of the iron formation period and before deposition of the Rove slates." Regarding the origin of the Rove formation, Grout and Schwartz (1933, p. 25) state: "Volcanic tuffs, more or less worked over, may have contributed parts of the series."

The iron-bearing member of the Virginia formation may represent a local recurrence of iron sedimentation in an area which was not overwhelmed by clastic materials because of its remoteness from the main area of vulcanism, a local lapse in volcanic activity, or some other factor. On the other hand, this unit might represent a product of an additional regression and transgression of the same environment in which the uppermost layer of cherty siderite rock in the Biwabik formation was produced. This interpretation, however, would require the contemporaneous deposition of chloritic argillite farther from the shore than cherty siderite rock; and this, although not inconceivable, is perhaps unlikely.

Regional interpretation: A knowledge of the general pattern of sedimentation (Fig. 10) makes it possible, to some extent, to predict changes in lithology that might occur in unexplored areas of Animikie rocks, although the many complicating variables of sedimentation must always be considered. For example, it is probable that relatively near-shore deposits prevail in the buried outcrop belt of the Biwabik formation for at least a few miles south from the present western end of exploration of the Mesabi range. These rock types are of little or no economic value at the present time. Somewhere between eastern Cass County and the Emily district, however, the section cuts back into rock types of economic importance that were originally deposited relatively farther from shore. It is not possible to predict exactly where this transition takes place.

Many geologists believe that the Animikie sediments are continuous under the Lake Superior syncline because of the reappearance of correlative rocks at the Gogebic range. It may be that the Gogebic range bears a relation to the southern shore line of the Animikie sea similar to the relation of the Mesabi range to the northern shore line. Probably the first transgression spread its products uniformly over a great area, but the later movements were seemingly of lesser magnitude. Therefore, it is possible that the upper part of the iron-bearing sequence under Lake Superior is relatively rich in offshore iron silicate rocks as well as in other deposits of unknown nature.

CONDITIONS OF SEDIMENTATION

Throughout the foregoing discussion, it has been assumed that the Animikie sediments were deposited in a sea rather than, as suggested by Tyler and Twenhofel (1952, p. 137), in a fresh-water basin having little or no connection with the sea. They base their conclusion largely on the relative lack of magnesia and lime in the iron-formations of Michigan. Three factors, however, should be considered: the content of magnesia and lime in the Biwabik formation is not insignificant (see Table 6); lime may have been concentrated in deposits which are not fully represented in the presently exposed section, as suggested by Gruner (1922, p. 456); and pre-Cambrian marine sediments in general seem to contain much less lime than Paleozoic or later sediments (Nanz, 1953, p. 59). In view of the great areal extent and thickness of the Animikie group, and the variety and nature of the sediments deposited simultaneously with the iron-bearing rocks, an epicontinental sea seems the most likely medium of deposition.

Van Hise and Leith (1911, p. 516) contend that a large part of the iron and silica which went into the iron-formation was contributed directly by magmatic springs and, to a lesser extent, by the reaction of sea water with hot submarine lava flows. Gruner (1922, p. 459), on the other hand, concludes that weathering of a land mass of older volcanic rocks in the presence of abundant vegetation of a low form could adequately supply the necessary iron and silica to streams emptying into the sea. Several geologists (Gruner, 1924, p. 55; James, 1951a, p. 264; Tyler and Twenhofel, 1952, p. 136) have pointed out such inadequacies of the theory of direct volcanic contribution as the lack of direct evidence, the lack of association of iron-bearing sediments with contemporaneous vulcanism in later geologic times, the enormity of the volcanic source necessary, and the improbability that local magmatic sources could produce relatively uniform iron-formations of broad extent.

The evidence concerning the pattern of sedimentation in the Mesabi district indicates that the iron rocks were closely related to a shore line, and that they were closely associated with contemporaneous clastic sediments of a "normal" character. The general pattern of sedimentation is like that of marine deposits produced throughout the ages, and an appeal to a magmatic source of material seems neither necessary nor justified.

Moreover, there is some reason to believe that vulcanism contemporaneous with Animikie sedimentation was detrimental rather than conducive to iron concentration. Structural disturbances attended by vulcanism may have brought on deposition of the Virginia argillites by making available so much fine clastic material that the concentration of iron in the streams was much diluted and deposition became too rapid to permit effective separation of iron from the clastics. Such a condition does not seem applicable to the Keewatin iron-formations, some of which are intimately associated with graywacke and contemporaneous volcanic rocks; but, as Bruce (1945, p. 590) points out, differences between some iron-formations are marked enough to indicate different conditions of origin. The fact that there is graywacke with the Keewatin iron-formations and quartzite with the Biwabik iron-formation implies diverse tectonic frameworks of sedimentation.

The pre-Cambrian iron-formations, because of their great bulk and thick alternating iron-rich and silica-rich bands, differ from the later sedimentary iron deposits. This difference suggests that some environmental factor in the pre-Cambrian was unlike that in later periods. One school of thought maintains that what was unique to the pre-Cambrian was a certain combination, intensity, or duration of conditions rather than the essential nature of the conditions themselves (Bruce, 1945, p. 601). Thus James (1954) concludes that the pre-Cambrian iron-formations resulted from the coexistence of unusually prolonged chemical weathering and restricted basins in which iron deposits could be built up. Another school of thought holds that the essential nature of some specific condition was not the same then as it is today. Thus Lane (1917, p. 47) proposes that the pre-Cambrian atmosphere contained somewhat less oxygen and more carbon dioxide than today's atmosphere, a factor which might facilitate weathering and transportation of iron. This proposal is based on the widely accepted hypotheses that the oxygen in the present atmosphere has been derived largely from the photosynthesis of plants, and that plant life was less abundant in pre-Cambrian than in later times.

The work of Nanz (1953, p. 60) indicates that the iron content of pre-Cambrian shales and slates is higher than that of Paleozoic and Mesozoic shales and slates. Although complicating factors make any interpretation tenuous, it is possible that this trend in iron content results partly from a specific condition favoring mobilization and transportation of iron in the pre-Cambrian, a condition not limited solely to periods of prolonged chemical weathering. If the essential nature of conditions has not changed since pre-Cambrian times, it seems unlikely that a given combination of these conditions could leave such an abundant record in the later pre-Cambrian and yet leave little record in subsequent times. The entire matter is yet open to question, but a difference in the pre-Cambrian atmosphere seems to be a reasonable postulate.

A likely setting for the iron sedimentation that produced the Biwabik formation would have been a quiescent, shallow sea bordering a low-lying, relatively stable land mass made up of igneous and metamorphic rocks undergoing deep chemical weathering in the presence of organic matter, and a seasonal climate. An analysis of the broad relation between iron sedimentation and earth forces, somewhat like that outlined by James (1951b, p. 1452), can be inferred from the Animikie record. Stable conditions during the first sea transgression favored the contemporaneous deposition of clean sands and iron rocks. The first regression is taken as a sign of decreasing stability which increased the rate of deposition, thereby producing more deposits of silts and clays and more irregularities in the iron sediments. Examples of such irregularities are the marked changes in thickness and the bewildering variety of textures in parts of the Upper Cherty member. During the second transgression, tectonic instability and vulcanism increased to such an extent that clastic deposition ultimately overwhelmed all chemical sedimentation.

Thus, the most favorable tectonic condition for the concentration of iron seems in this case to be one of great stability. Under quiescent conditions of relatively slow subsidence and slow deposition, the clastic and nonclastic materials supplied by streams could be effectively sorted in the sea according to their physical and chemical properties. That this sedimentary differentiation could be remarkably complete is attested to by the sharp contacts between some of the deposits which were formed contemporaneously. This separation took place in a single body of water, and the auxiliary basin acting as a "clastic trap" suggested by Huber and Garrels (1953, p. 356) was not necessary. Nor is it certain that a restricted (barred) basin was a prerequisite for the concentration of iron. Circulation of water was probably slow, and the site of deposition may at times have had limited access to the open sea; yet no evidence at hand rules out a shallow, relatively tideless epicontinental sea as a possible medium of deposition. Any other restrictions in circulation would probably have been the fortuitous result of broad irregularities in the topography crossed by the advancing sea rather than the result of structural disturbances.

Generally shallow water is suggested by the proximity of iron sedimentation to the shore line, the abundance of intraformational conglomerates, the algal structures, and the scattered oölitic beds of primary hematite.

When deposition took place under optimum conditions of quiescence, the textural relations between magnetite and chert tended to become more irregular as distance from shore increased. This relation is shown by the general sequence of textures in the Lower Cherty member (Plate 1) and by the greater regularity of the textures in the cherty rocks of the Westernmost Mesabi, where near-shore conditions prevailed. Increasing distance from shore can normally be assumed to correlate with increasing depth of water. Therefore it appears that regular bands of magnetite formed in relatively shallow water, irregular bands in somewhat deeper water, and mottles in still deeper water.

This relation between irregularity of beds and depth of settling (or distance of transportation) of material is further evidence that these features originated by sedimentation and not by replacement. The banding of the chert-magnetite rocks was apparently forming at the same time that laminations and thin beds were forming in the more uniform deposits both to the landward and seaward. Van Hise and Leith (1911, p. 525) contend that the globular form of the granules in taconite resulted from surface tension between sea water and a precipitate having a low crystallizing tendency.

In much of the recent literature about the environmental conditions under which different iron minerals form, there has been a tendency to explain a succession of different iron-bearing units in terms of a succession of extensive physical and chemical changes occurring over a whole basin of deposition. Probably, however, many of these successions of different units were built up by the migration of a series of coexisting environments. Although the succession in the Biwabik formation may not be entirely typical, it would appear to be significant because of its completeness. Departures from this pattern should be expected in other areas as a result of nondeposition, or of other variations in the many factors of sedimentation.

According to Krumbein and Garrels (1952, pp. 13–16), the chemical environment of deposition — in particular its oxidizing or reducing nature — largely controls the type of iron mineral (sulfide, carbonate, oxide, silicate) that is formed. The oxidizing or reducing nature of the various environments of deposition that produced the Biwabik formation is not a simple function of depth, or of distance from shore. The preservation of organic matter is a key factor that is related partly to the rate of sedimentation.

Although organic matter was scattered throughout the contemporaneous environments and some was concentrated in deeper waters, most of it was deposited rather close to shore, if conditions permitted. During the first advance of the sea, deposition was apparently slow enough to allow the destruction of most of the organic matter, and the well oxygenated inshore environment of chemical deposition produced hematite. Seaward, conditions were progressively of a relatively more reducing nature. But during the first retreat of the sea, deposition was apparently somewhat more rapid, and a near-shore concentration of organic matter (now represented by graphitic argillite) was preserved. Up to a point, at least, conditions seaward from this deposit were of a relatively more oxidizing nature; locally even hematitic sediments were produced. Still farther seaward, conditions probably were again more reducing, as evidenced by the iron carbonate associated with the silicate rocks.

The general sequence of lithic types from shore to sea during the regression was sulfide, carbonate, oxide, silicate. Near-shore hematite and near-shore sulfide were mutually exclusive, as a result of other conditions that affected the preservation of organic matter. The common association of primary hematite with intraformational conglomerates is logically the result of aeration accompanying the storms or other minor disturbances that broke up the beds. Graphitic-pyritic sediments have been interpreted by some investigators as an indication of deep water and stagnancy in an entire basin of deposition and by others as an indication of a lagoonal environment; neither interpretation seems applicable to the specific case of the graphitic argillite in the Biwabik formation.

The clastic rocks and the nearest-to-shore environment of chert deposition tended to be rather lean in iron, which suggests that iron was normally carried at least a short distance out to sea before being precipitated. Alumina was concentrated in the clastic rocks and, to a lesser extent, in iron silicate rocks. Aside from these variations, the chemical compositions of the nonclastic iron-bearing rock types are very similar.

Most features of the Animikie group can be far more readily explained by reference to "normal" sedimentary processes than by an appeal to more unusual mechanisms. James (1951a, p. 263) has summarized some of the arguments against postconsolidation introduction of iron by replacement as follows: replacement producing iron minerals that are different in each of several successive stratigraphic units is unlikely; within each unit, the uniformity of chemical composition, the preservation of minor sedimentary features, and the occurrence of pebbles composed of the various minerals in intraformational conglomerates all indicate nonreplacement. Such replacement, magmatic activity, and extensive metamorphic changes, in addition to being unnecessary factors, are not supported by direct evidence.

V. STRUCTURE

METHOD OF REPRESENTATION

Because the base of the Biwabik formation is a widespread, gently dipping, and easily recognized horizon, and because of the abundance of drilling information, the Mesabi range presents an exceptional opportunity for a study of minor structures by a contour method. The structure contour map of the district (Plate 3) is based on more than 3000 vertical drill holes, on field observations in almost all open pits, and on a wealth of information supplied by geologists and mining engineers who have spent years working on the range.

Most of the drill holes on which interpretations are based are shown on Plate 3, which gives the reader a chance to evaluate the degree of certainty of the interpretation in each area. Not all drill holes actually used could be shown on the map, but those omitted are mostly in areas where other information is especially plentiful. In some areas where deep drill holes to the base of the Biwabik formation are few, elevations on the base were estimated from shallower holes that penetrated some other horizon marker, such as the base of the Lower Slaty member. The local stratigraphic thickness between such a horizon marker and the base of the Biwabik formation was subtracted from the elevation of the horizon marker in order to determine the elevation of the base at the point vertically beneath the hole. This local stratigraphic thickness was taken from the log of the nearest deep hole. But, owing to the possible variations in stratigraphic thickness over short distances, elevations derived in this manner are subject to some error.

Determinations of the strike and dip of the undecomposed beds exposed in many open pits furnish a useful guide to the general trend and spacing of contours on the base of the formation, although this usefulness decreases somewhat as the distance of the particular bed from the base increases.

The Lake Superior datum has been employed for much of the surveying on the Mesabi range, but there are several variations of this datum. The collar elevations of some holes drilled years ago were never determined. Fortunately, however, there is available a series of recent topographic maps of the district, made by the U.S. Geological Survey and having a contour interval of 10 feet. Wherever possible, the collar elevation of each hole has been checked with the aid of these maps, and any remaining errors are probably minimized. Much of the drilling information was obtained from exploration records, inasmuch as the drill core from each hole could not be personally examined. A few holes that did not reach the base of the iron-formation were erroneously reported as having done so, but elevations derived from such holes generally could be disregarded because of their unreasonable relations to neighboring elevations. The locations of some of the old holes, which were never precisely surveyed, are possibly somewhat inaccurate on these maps.

On Plate 3, the attempt has been made to draw contours as nearly parallel to one another as the data permit. As a result, many folds appear to become less accentuated both up and down the dip. This procedure seems justified by the known relations in areas where information is abundant. If folding is gentle, the beds slip along one another instead of flowing internally; the beds remain essentially parallel, and folds also die out vertically (Billings, 1942, p. 53). However, all but the very small structures shown on the base of the Biwabik formation carry through to the bedrock surface at the base of the glacial drift, because the distance is not very great.

Where information is not abundant, the simplest interpretation has been sought. If considerable doubt exists as to whether a feature is a fold or a fault, it is shown as a fold; actually, however, contours depicting a fold and those depicting a fault of the same magnitude are not much different. It may be argued that structures are detected only where drilling is abundant, especially at ore bodies, and that other areas of supposed uniform dip may be equally deformed. To some extent this argument is true, but enough information is generally available to show that the dips are actually rather uniform at many places on the Mesabi range. A very few drill holes are needed for detecting the presence of a large structural feature, although many holes may be required before its exact nature may be defined. Lahee (1941, p. 649) states: "In any event, a contour map must be read and interpreted with an open mind, for new data may necessitate altering the minor details of the contour lines. Usually the major features will persist through various modifications, provided correlations have been correct."

It is important to remember that a fault is mapped where it intersects the base of the Biwabik formation and not where it appears in the pit. Because the fault plane and the base are sloping surfaces, the direction of this intersection is not necessarily the true strike of the fault plane.

Most of the deformation is the result of tectonic forces, although a small part of the southward dip of the beds and a few of the minor irregularities may be initial dips produced during sedimentation.

DESCRIPTIVE STRUCTURAL GEOLOGY

General Features

The Animikie beds of the Mesabi district strike generally east–northeast; the only major change in direction of strike takes place between Virginia and Eveleth, where the beds strike somewhat west of south. The major folds in this area are the Virginia syncline and the Eveleth anticline. East of Virginia the average regional dip is about 12° southeast, and west of Virginia the average dip is about 6° southeast. Slippage along bedding planes, which has produced slickensides oriented down the dip, is a common feature east of Virginia. The axes of many minor folds trend at right angles to the general strike of the Mesabi range, although the axes of some folds are nearly parallel to the regional strike.

The fault pattern is rather indefinite, but most faults strike either about N. 75° W. or about N. 20° W. Faults are commonly difficult to recognize in the open pits because of decomposition and slumping of the adjacent beds, but the faults whose dips can be observed are all of the gravity type in which the hanging wall is downthrown. The fault planes dip steeply, and the maximum displacements, which are all less than 200 feet, are commonly less than 50 feet.

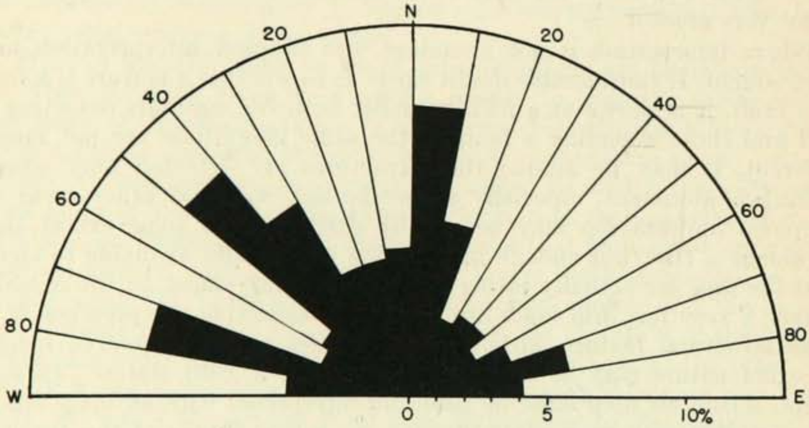

FIGURE 12. — Diagram showing strike of joints in the Biwabik formation between Mesaba and Coleraine, based on 290 determinations, most of which were made by Gruner. Length of rays is proportional to the number of determinations falling within each 10° sector.

The Animikie rocks are strongly jointed. The general strike of joints all along the Main and West parts of the range is shown in Figure 12. As noted by Gruner (1924, p. 25), the three main joint sets strike about N. 10° E., N. 45° W., and N. 80° W. Joint planes are commonly straight, long, and essentially vertical, the dips being steeper than 70°. At any one place, two of the joint sets are generally more prominent than the third. The set striking N. 45° W. is prominent at various places all along the range; the set striking N. 10° E. is more apt to be prominent east of Hibbing; and the set striking N. 80° E. is more apt to be prominent west of Virginia. But these relations are not marked.

At the north boundary of the Biwabik formation, the elevation of the base decreases toward the west, from as much as 1750 feet near Mesaba to about 1100 feet west of Pokegama Lake. Many local differences in the elevation of the north boundary are due to the effect of bedrock valleys and hills that are partly buried by glacial drift. The degree of deformation is relatively high east of Virginia, moderate between Virginia and Nashwauk, low between Nashwauk and Pokegama Lake, and moderate between Pokegama Lake and eastern Cass County.

Specific Features

Certain structural features are discussed below, in the order they occur from east to west along the range. No attempt is made to describe all structures, inasmuch as most are represented adequately on Plate 3, but alternate interpretations or additional data are pointed out. Mine exposures described are as of the 1952 field season.

East of R. 14 W. Although the structure along the Mesabi range east of R. 14 W. has not been mapped by the contour method, the very detailed outcrop maps by Grout and Broderick (1919b) give a good picture of the structure of this area. For 3 miles west of Birch Lake, where the Biwabik formation is obliterated by the Duluth gabbro, the Animikie beds dip 40°–70° southeast. Farther west, the dips flatten to about 10°. Two fairly prominent folds, each a few hundred feet wide, are in the SW¼ Sec. 26:60-13 and the NE¼ Sec. 25:60-13. Elsewhere, the dips are fairly uniform.

R. 14 W. Because of the relatively strong deformation and the scarcity of drilling in this area, the exact nature of the structural features is not known. The prominent structure near the east side of R. 14 W. is interpreted as a sharp fold, although the beds may well be faulted in part as Wolff (personal communication) believes. The results of two traverses with a dip needle and studies of outcrops and drill records tend to confirm the interpretation by Grout and Broderick (1919b, Plate 13) of the outcrop pattern in this area, which seems somewhat more suggestive of folding than of faulting. Very small amounts of ore occur along this structure in the NE¼ Sec. 14. Most of the strikes and dips in R. 14 W., shown on Plate 3, were determined by Grout and Broderick.

The structure near Mesaba is probably more complicated than the map shows it to be. Wolff (1915, p. 139) mentions the fold at the Wentworth mine in Sec. 21. This fold is probably faulted at places, for in one drill hole in the SE¼SW¼ Sec. 21 about 50 feet of the basal sequence of the Biwabik formation are missing. The representation of a syncline at the Knox and Adriatic mines in Secs. 19 and 30 may also be too simple.

R. 15 W. Wolff believes (personal communication) that there may be a fault striking southeast through the Donora reserve in Sec. 27. When the configuration of the base of the Biwabik formation was compared with the bedrock topography, it was possible to determine the outcrop pattern in Secs. 29, 32, 33, and 34. In the SE¼ Sec. 33, the dip of the base of the iron-formation is nearly the same as the topographic slope, and only a veneer of taconite covers the top of the Pokegama formation.

V. R. Campbell (personal communication) interprets the structure at the Meadow and St. James mines in Sec. 3 as a fault striking north, with the west side downthrown about 50 feet. A large syncline is exposed in the Miller-Mohawk pit in the SE$\frac{1}{4}$ Sec. 4, but part of this feature is probably due to slump. A structure in the Hudson Mine, which is also interpreted as a fault, is no longer exposed, but its displacement of about 50 feet is fairly well substantiated by drilling information. This fault might strike more nearly north and thus account for some of the downdropping shown in the supposed structural basin in Secs. 29 and 32. A quartz vein and brecciated beds can be seen along the strike of the Biwabik fault in the Embarrass pit, and the 200-foot displacement of the fault is well established by drilling information. Several minor faults and folds are exposed in this pit (Fig. 13).

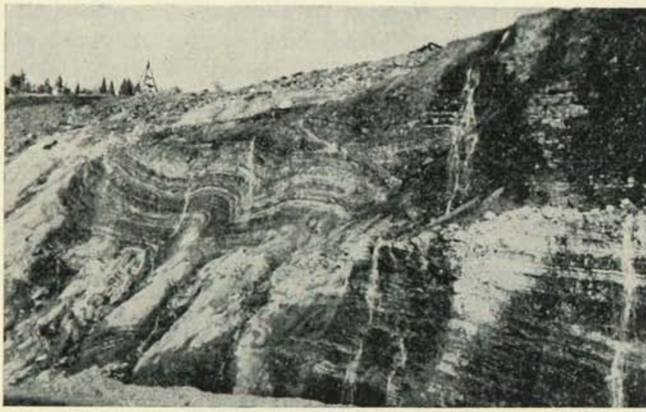

FIGURE 13. — Minor folding and faulting at the Embarrass pit near Aurora (looking southeast). The banded rocks are part of the Biwabik formation and the dark rocks in the upper right are part of the Virginia formation.

R. 16 W. The diversion channel of the Embarrass River crosses the Biwabik fault in Sec. 1 (see cross section F–F′, Plate 3, sheet 2), and here the rocks of the Lower Slaty member are in contact with greenstone, which suggests that there has been a minimum displacement of 175 feet (Gruner, 1946, p. 70). Several zones of oxidized taconite associated with quartz veins and slickensided surfaces are exposed in the diversion channel a short distance south of the main fault. The Biwabik fault is well exposed along the north wall of the Ajax-Hector Mine, where the brecciated zone is 20 feet wide, and along the north wall of the Biwabik pit. The displacement of the Biwabik fault apparently decreases toward the west (Gruner, 1946, p. 68). Most slickensides plunge directly down the dip of the fault plane.

Along the south wall of the Biwabik pit there is some indication of another fault parallel to the main one. The sharp southward projection of the south contact of the Biwabik formation in Sec. 11 is due to a deep valley now buried by glacial drift.

Spurr (1894, p. 22) mentions the anticline near the Belgrade Mine in Secs. 4 and 9. An examination of drill cores from the Belgrade Mine shows that some holes did not penetrate to the base of the Biwabik formation as reported on the exploration records. Some of these holes were used on Plate 3 (sheet 3), but inasmuch as they probably reach within 10 or 15 feet of the base, the error in picturing the anticline is presumably not great. F. S. Bergstrom and J. F. Wolff (personal communication) report a possible fault having a displacement of about 30 feet which strikes N. 30° W. along the east side of the Belgrade Mine.

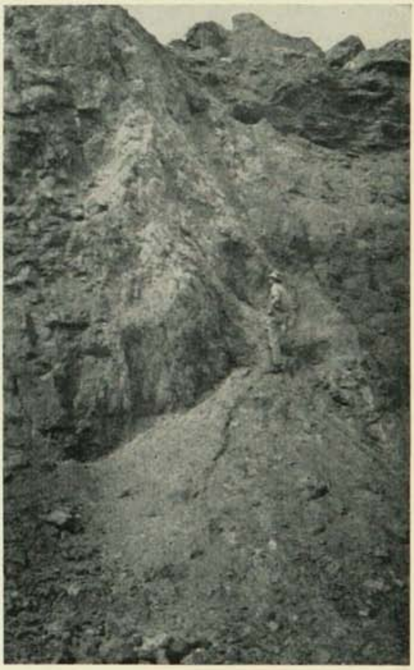

FIGURE 14. — The Corsica fault at the Corsica Mine near McKinley (looking north). The fault plane dips steeply toward the east. Most faults are not easily seen because the adjacent rocks are decomposed.

The wrinkles at the McKinley Mine in Sec. 8 are evident only because the drilling is very closely spaced; it is possible that similar structures are present elsewhere but remain undetected. At the south side of the Corsica Mine in Sec. 18, a fault brings the algal beds of the Upper Cherty member on the west into contact with bleached rocks of the Upper Slaty member on the east, which suggests a displacement of about 100 feet. Figure 14 shows this fault where it is exposed at the north side of the Corsica pit.

R. 17 W. The structure in the SE¼ Sec. 13:58-17 is probably more complicated than shown. The dips in the Gilbert area are generally steeper near the south contact of the Biwabik formation, as Gruner

(1946, p. 67) has pointed out. Several quartz veins strike N. 40° W. along the length of the Gilbert and Hobart ore bodies, which are elongated parallel to the major jointing and are in line with the strike of the Minnewas fault at Virginia. Bergstrom (personal communication) reports a small fault striking N. 60° E. across the south end of the Gilbert pit. Such a zone of disturbance might extend eastward toward the Sparta-Malta pit.

The Eveleth anticlinal axis, which has been traced by magnetic methods for some 15 miles (Schwartz, 1943, Plate 5), strikes about S. 40° W. from Eveleth. Three deep holes have been drilled near this axis. One, in the NW¼NW¼ Sec. 19:57-17, penetrates the base of the Biwabik formation at an elevation of 298 feet below sea level; the second, in the SW¼SE¼ Sec. 23:57-18, penetrates this horizon at 804 feet below sea level; and the third, in the NW¼NW¼ Sec. 26:57-18, penetrates this horizon at 838 feet below sea level. These elevations indicate that the average plunge of the anticlinal axis for the first 4 miles southwest of Eveleth is 5½°, whereas the average plunge for the next two miles is only 3°.

Several quartz veins, one of them as much as 15 feet thick, strike N. 20° E. through the Troy and Fayal ore bodies south of Eveleth. The small dislocations along joint planes and the minor deformation of beds, which occur in this zone, can be seen in these pits. A quartz vein in Sec. 31:58-17, striking N. 80° W., has been traced for nearly half a mile (Gruner, 1946, p. 109). Another quartz vein, associated with brecciated beds, parallels the length of the Virginia Mine in Sec. 30:58-17. Wolff (1917a, p. 159) believes that the Dorr fault, which has a maximum displacement of about 50 feet and which was once exposed in the NE¼ Sec. 31:58-17, continues northeastward along the base of a cliff of taconite which crops out a mile north of Eveleth. This interpretation was used for Plate 3 (sheet 4), because the dense drilling near the center of Sec. 20:58-17 defines a small break which is most likely a continuation of the Dorr fault.

The Minnewas fault in Secs. 16 and 17:58-17 is in a zone of vein quartz and breccia and has a displacement of 50 feet. Wolff (personal communication) reports two additional faults of small displacement which strike north-northwest along the west wall of the Rouchleau-Minnewas-Moose ore body. The Missabe Mountain fault is exposed in the SE¼NE¼ Sec. 8:58-17, where it has a 25-foot displacement, but it dies out both to the east and west. The syncline in the N½ Sec. 4:58-17 may extend to the outlier of iron-formation, reportedly synclinal (M. P. Walle, personal communication), which is in Secs. 35 and 36:59-17. The outcrop pattern in Sec. 33:59-17 is inferred from the assumed contour lines, the topography, and the one large outcrop in Sec. 28:59-17. The steeply dipping beds at the Alpena Mine in Sec. 5:58-17 are on the east flank of an asymmetric syncline. A detailed cross section of this fold has been prepared by Gruner (1946, p. 69).

R. 18 W. Several quartz veins occur in the Mountain Iron pit in Sec. 3. Wolff (personal communication) reports that two north-trending faults cut through the north end of this pit and that the block between these faults is dropped down about 40 feet. He describes two other faults; one, in the south end of the pit, strikes east, the rocks on the south being downthrown about 20 feet; the second, which strikes northwest, follows the northeast wall of the main pit.

R. 19 W. The Buhl monocline, located near the south boundary of the Mesabi range, extends across R. 19 W. and may at places be faulted (C. J. Calvin, personal communication). Large quartz veins occur in the Dean and Wanless mines in Sec. 15 (Gruner, 1946, p. 111). Beds dipping as steeply as 60° are exposed along the north side of the Grant pit in Sec. 20.

R. 20 W. According to Wolff (personal communication), the Buhl monocline is continuous between the Grant Mine and the Forster Mine in Sec. 24:58-20. The magnitude of the sharp monocline exposed in the SW¼SW¼ Sec. 13, which may be faulted in part, is approximate. Some of the beds here are vertical. The fault at the Sherman Mine in Sec. 23 is fairly well substantiated by drilling information but is not yet exposed in the pit; Wolff (personal communication) believes that this fault may continue westward almost to Chisholm. He also reports that two east-striking step faults pass near the center of Sec. 28; they dip north, their north sides are downthrown, and their combined displacement is about 20 feet at the Glen Mine and about 75 feet at the Monroe Mine.

R. 21 W. The structure in Sec. 36, which is a good example of parallel folding, has been substantiated by closely spaced drilling. The sharp roll here may be faulted toward the north, but no sign of a break is evident in the Burt pit to the south. Indications of a possible gravity fault striking N. 65° W. through the north side of the SE¼NE¼ Sec. 1 can be seen in the Hull-Rust pit. Apparently, this fault dips 70° southwest, and may have a displacement of about 75 feet (Wolff, personal communication). The steeply dipping beds exposed in the Hull-Rust Mine just north of the South Agnew Mine are shown in Figure 15. Wolff (1915, p. 139) regards the structure at the South Agnew Mine in Sec. 11 as a faulted fold, and he also believes that there may be a fault which trends northwest through the North Uno and Kerr mines in Secs. 2 and 3. Such a fault would be parallel to the probable Morton fault, which mining has not yet exposed, in Sec. 10. The displacement of the Morton fault seems to die out toward the northwest, but the zone of disturbance probably extends to the Lamberton Mine in Sec. 9.

A check of the collar elevations of drill holes in the W½ R. 21 W. indicates that all of them may be represented on Plate 3 (sheet 7) as being about 15 feet too high. Hence, all the contour lines in this area should perhaps have been drawn about 200 feet farther north than as indicated on Plate 3. Minor irregularities in the bedrock surface produce the irregularity of the map contact between the Biwabik and Virginia forma-

FIGURE 15. — Steeply dipping beds in the Hull-Rust pit, SW¼SE¼ Sec. 2: 57-21 (looking north).

tions in Sec. 17, because the dip of this contact is so nearly equal to the slope of the bedrock surface beneath the drift (see cross section T–T′, Plate 3, sheet 7). Leith (1903, p. 178) describes dips as steep as 60° at the Stevenson Mine in Secs. 7 and 8, and Wolff (personal communication) thinks there may be a fault parallel to the length of the ore body.

R. 22 W. The steeply dipping beds at the Galbraith and La Rue mines in Secs. 28 and 29 are well exposed; the structure appears to be generally monoclinal, although the beds are locally broken in the SW¼NW¼ Sec. 28 and in the SW¼SE¼ Sec. 29. Van Hise and Leith (1911, p. 175) note the sharp, steep fold at the Hawkins Mine in Sec. 31. The cross structure at this mine could be either a fold or a fault. The local steepening of the dip at the Harrison Mine in Sec. 6 does not appear to be faulted.

West of R. 22 W. The dips between Nashwauk and Pokegama Lake (Plate 3, sheets 10 and 11) are very uniform. West of Pokegama Lake, drilling is too scattered to allow for defining any but broad structures. The major feature is the Sugar Lake anticline, the axis of which strikes about S. 20° E. across Sec. 8: 54-26. The steeper dip of about 12° on the west flank of this anticline (shown on cross section Ii–Ii′, Plate 3, sheet 11) is suggested by one deep drill hole in Sec. 18: 56-26. The general structural pattern of the Westernmost Mesabi shown on Plate 2 indicates that folding in this area is sharper and dips are steeper than they are between Nashwauk and Pokegama Lake.

INTRUSIVE ROCKS

DIABASE SILLS

Grout and Broderick (1919b, Plate 1) have mapped several diabase sills, 10 to 20 feet thick, which intruded the Biwabik formation and the lower part of the Virginia formation on the East Mesabi. There are three sills near Birch Lake, two in Sec. 20:60-12, four in Sec. 26:60-13, one in Sec. 33:60-13, and one in Sec. 15:59-14. Several small dikes are also present. Grout and Broderick (1919b, p. 8) conclude that these rocks were intruded before the main mass of the gabbro was intruded, because they have been metamorphosed by the contact action of the gabbro. Grout and Broderick tentatively assign the dikes and sills to the Keweenawan group. The only other diabase in the Animikie rocks of the Mesabi range is reportedly in the NE¼SE¼ Sec. 13:57-22, where "drilling has penetrated 20 feet of diabase with iron-bearing formation both above and below" (Van Hise and Leith, 1911, p. 178).

Duluth Gabbro

According to Grout and Broderick (1919b, p. 8), "along the eastern Mesabi practically all of the gabbro is a coarse olivine gabbro with a somewhat variable composition, in alternating bands. The bands dip eastward at a steep, but variable angle."

The Aurora Intrusives

Intrusive Relations. The Aurora sill intruded the Upper Cherty member of the Biwabik formation near Aurora. The sill does not crop out, but it has been traced by recent drilling for 3½ miles, from the west side of Sec. 4:58-15 to the center of Sec. 36:59-15. The sill increases in thickness southward. To the north the sill diminishes; in some places it feathers out within the iron-formation, generally to two or more thin sheets; in others, its edge has been exposed and beveled by erosion. Several of the drill holes near the sill's north limit penetrate less than 8 feet of sill rock enclosed in taconite. To the south, successive drill holes penetrate sill thicknesses of 20, 24, 26, 45, 80, and 121 feet.

The sill is essentially concordant, and the structure contours indicating its top are nearly parallel with those which represent the base of the Biwabik formation (Plate 3, sheet 2). The top of the main mass of the sill transgresses somewhat across the bedding, however, from immediately beneath the algal structures to as much as 110 feet below them (Plate 1, sheet 1). The distribution of the sill is important to any future taconite production in this area, owing to the considerable volume of the iron-poor sill rock.

The Aurora intrusives were first recognized by mining engineers; later, Van Hise and Leith (1911, p. 178) wrote of intrusions "principally parallel to the bedding but partly across it" near Aurora, and Gruner (1924, p. 25) mentioned "kaolin dikes, originally igneous intrusives" as far west as Sec. 18:58-16. V. R. Campbell (personal communication) reports a vertical dike, 10 to 20 feet thick, altered to red clay, striking N. 60° E. through the Knox ore body in Sec. 19:59-14. Gruner (1946, p. 108) describes a dike trending northeast through the nearby Stephens Mine. F. S. Bergstrom (personal communication) has mapped three vertical dikes near Aurora; one, at the Miller-Mohawk Mine, is 8 feet thick and strikes N. 30° E.; the second, at the St. James pit, is 10 feet thick and strikes N. 10° E.; and the third, at the Hudson Mine, is 4 feet thick and strikes N. 25° W. A completely altered dike, 6 inches thick, is exposed in the west end of the Mary Ellen pit in Sec. 9:58-16; and Bergstrom and Wolff (personal communication) report another dike, 7 feet thick, which strikes N. 30° W. through the nearby Belgrade Mine.

Composition and origin. In its thicker parts the Aurora sill consists principally of dark-red coarse-grained rock. Closer to the contact, the rock is medium-grained for as much as 10 feet and grades into a red-orange layer, 2 to 8 feet thick, that contains phenocrysts of feldspar. The fine-grained marginal zone, which is commonly only 1 or 2 inches thick, is dark gray-green; and a number of thin separate sheets, some of which are as thick as 8 feet, are entirely green.

The southward thickening of the sill suggests that it was intruded from the south. The irregular north boundary of the sill is the result primarily of the irregular northward advance of magma. Most of the feldspar phenocrysts in the chilled margins are oriented with their long dimensions parallel to the contact. This flow structure suggests that the sill was injected as a mush of crystals and liquid. A detailed study of several drill holes in Sec. 36:59-15, where the sill is thin, indicated that none of the thin stratigraphic units has been eliminated and that therefore the sill must have been forcibly intruded. The injection of the sill simply increased the distance between the Pokegama and Virginia formations (cross section D–D′, Plate 3, sheet 2).

The Aurora sill is composed of a syenitic rock resembling some of the red rock types of the Duluth region. Its unusual composition is revealed in mineral and chemical analyses (Tables 9 and 10). Thin sections of three phases of the rock show 76 to 80 per cent silicic feldspar, listed as albite in Table 8. Only minor amounts (1 to 7 per cent) of potash feldspar, orthoclase, could be identified. In addition to albite, a feldspar having a very fine twinning is present, but it could not be positively identified. The appreciable amounts of potash shown in the chemical analysis suggest that this feldspar is anorthoclase. The average composition of the sodic feldspars is about 75 per cent $NaAlSi_3O_8$ and 25 per cent $KAlSi_3O_8$. The abundant chlorite and calcite are alteration products, and it is likely that the sill rock owes its unusual composition, in part at least, to late magmatic or hydrothermal activity.

The Aurora intrusives presumably are Keweenawan, but as they contain little or no zircon, their age could not be checked by the zircon method of Tyler *et al.* (1940).

Contact metamorphic effects. The intrusion of the Aurora sill had little effect on the adjacent taconite other than the formation of clumps and mats of a fibrous blue mineral, crocidolite. The X-ray pattern of this

TABLE 9. MINERALOGY OF THE AURORA SILL ROCK, IN VOLUME PERCENTAGES*

	1	2	3
Albite	78	76	80
Orthoclase	2	7	1
Calcite	6	16	6
Chlorite	11	X	X
Biotite		X	12
Quartz	X		
Magnetite	1	X	X
Sphene	1	X	X
Apatite	X	X	X
Muscovite	X	X	
Sericite	X	X	X
Hematite	X	X	X
Pyrite		X	X

1. Dark-red coarse-grained sill rock; exact location is unknown but probably from deep within the thick part of the Aurora sill.
2. Medium-grained red-orange porphyritic sill rock; 2 feet from contact of sill with taconite, in a drill hole in the SE¼SW¼ Sec. 36:59-15.
3. Medium-grained green porphyritic sill rock; from center of intrusive sheet 2 feet thick, in a drill hole in the NE¼SW¼ Sec. 36:59-15.

* X = present in quantities less than 1 per cent.

TABLE 10. CHEMICAL ANALYSIS OF THE AURORA SILL ROCK IN WEIGHT PERCENTAGES*

Constituent	Per Cent	Constituent	Per Cent
SiO_2	53.91	CO_2	2.12
Al_2O_3	17.25	TiO_2	1.22
Fe_2O_3	2.50	P_2O_5	.40
FeO	4.87	Cl	.03
MgO	2.38	S	.04
CaO	3.38	MnO	.09
Na_2O	6.08	BaO	.10
K_2O	3.21		99.84
H_2O+	1.93	Less O≡S and Cl	.02
H_2O-	.33		99.82

* Sample from near center of sill, about 60 feet from the sill's margins, in a drill hole in the NW¼NE¼ Sec. 2:58-15. Eileen Oslund, analyst, Laboratory for Rock Analysis, Department of Geology, University of Minnesota.

mineral is that of an amphibole, and the optical properties check closely with those given by Peacock (1928, p. 253) for fibrous crocidolite from South Africa. Crocidolite commonly occurs not farther than a few inches from the sill's contacts, although small amounts occur at a few places as much as 50 feet away. Thin inclusions of taconite in the sill are completely blue. The amount of crocidolite is greater where the sill is thicker and where the rock at the contact is cherty taconite. Green slaty taconite and irregular magnetite bands contain little or no crocidolite, whereas the mineral may be abundant in the adjacent cherty bands. The crocidolite, peppered throughout the rock, typically forms irregularly radiating clumps less than 1 mm. in diameter, although larger aggregrates occur. As seen with the microscope, the fibers seem most commonly to have replaced granules of microcrystalline quartz which are enclosed in coarser quartz. Quartz grains are laced with fibers, but siderite and iron-silicate grains are but little penetrated.

According to Miles (1942, p. 34) crocidolite (iron-amphibole asbestos) is found only in banded ferruginous cherts. In Africa and Western Australia, cross-fiber seams of blue asbestos as much as 3 inches thick follow the bedding of virtually unmetamorphosed Pre-Cambrian iron-formations, but the asbestos reportedly is not associated with intrusives (Peacock, 1928, p. 243; Miles, 1942, p. 9).

The composition of crocidolite is essentially $3H_2O \cdot 2Na_2O \cdot 6(Fe, Mg)O \cdot 2Fe_2O_3 \cdot 17SiO_2$ (Peacock, 1928, p. 281). All these constituents except soda are present in comparable proportions in the iron-bearing sediments of Africa, Australia, and the Mesabi range. Peacock (1928, p. 283) concludes that the African crocidolite results from "a mild, static, nonadditive metamorphic process, resulting in chemical union, along soda-rich bedding planes, of the necessary constituents already *in situ*." Turner (1948, p. 289) states that these crocidolite rocks are "products of low-grade, purely geothermal metamorphism of an assemblage of chemical substances specially sensitive to increased temperature." The difficulty remains, however, that soda-rich beds have not been found in the unaltered iron-bearing sediments.

The source of soda, although obscure in the African and Australian deposits, is obvious in the case of the Mesabi crocidolite. The Aurora sill rock is soda-rich, and crocidolite is known only in the contact area of the sill. The soda, derived from the cooling sill, combined with the other constituents already present in the adjacent taconite, probably at fairly low temperatures.

STRUCTURAL INTERPRETATION

Van Hise and Leith (1911, p. 623), noting the general parallelism between Keweenawan folds and folds in the earlier Precambrian rocks, conclude: "Therefore it is reasonable to assume that the dominant trend in the Lake Superior folds . . . was probably established before Keweenawan time." The outcrops at several places just north of the Mesabi range suggest that the present outcrop belt of the Biwabik formation coincides very closely with the south contact of the Giants Range batholith, and the north shore line of the Animikie sea was apparently also parallel to these features. Hotchkiss (1923, p. 671) believes that the present strike of the rocks of the Gogebic range is nearly parallel to the axis of the basin in which they were deposited. The established northeast-trending "grain" of the Canadian Shield in northeastern Minnesota, therefore, apparently affected the configuration of the Animikie basin of deposition as well as the subsequent deformation of the sediments.

Leith (1903, pp. 197–198) contends that the Animikie sediments were gently folded both before and at the end of Keweenawan time. Both

Leith and Grout (1918, p. 516) believe that in the Mesabi district the Duluth gabbro was intruded along the unconformity at the base of the Keweenawan. Such a relation, which would suggest some pre-Keweenawan tilting of the Animikie group, cannot be conclusively demonstrated, but this supposition is not unreasonable. Mann (1953, p. 275) reports that Keweenawan flows unconformably overlie the iron-formation on the eastern Gogebic range. It is likely that deformation continued intermittently throughout the Keweenawan and that the pattern of earth forces remained much the same, owing to the control by the established "grain" of the earlier Precambrian rocks. The main compressive forces probably acted northwest–southeast, perpendicular to the axis of the Lake Superior syncline and to the axes of such folds as the Eveleth anticline. An interpretation of the joint pattern of the Mesabi district (Fig. 12) is consistent with this regional structural picture; if a northwest–southeast compression is assumed, the joints striking N. 10° E. and N. 80° W. would be shear fractures, whereas those striking N. 45° W. would be tension fractures.

The fault pattern is too indefinite to be satisfactorily interpreted, but some faults at least seem to be younger than folds. The Morton fault in Sec. 10:57-21 apparently offsets folds. Dikes associated with the Aurora sill follow the Hudson fault and the folds (faults?) at the St. James and Belgrade mines, but the sill itself is seemingly cut by the Biwabik fault, although the evidence for this supposition is meager. According to Billings (1942, p. 210), "field observations show that gravity faulting often follows folding, which suggests that tensional stresses follow the relaxation of compressional forces."

The identities of the pre-Animikie rocks shown in the cross sections of Plate 3 are inferred from the geologic maps of the district (Grout and Broderick, 1919b; Gruner, 1924; and Plate 2, this report). The locations of contacts between these older rocks are therefore speculative, but a relation between the degree of folding of the Animikie rocks and the nature of the underlying older rock nevertheless seems evident. The Animikie rocks generally are folded only where they are underlain by older sediments or volcanic rocks. Rocks of the Knife Lake and Keewatin groups form the cores of the Eveleth and Sugar Lake anticlines and underlie the more deformed areas near Nashwauk and Hibbing, and between Virginia and Mesaba.

The Giants Range granite, on the other hand, was apparently an unyielding mass against which the Animikie beds could only be tilted without much internal folding. This factor explains the uniform regional dip of the West Mesabi, which is underlain by granite, better than the supposition that the intensity of earth forces was originally less toward the west; the degree of deformation is again greater on the Westernmost Mesabi, which, partially at least, is underlain by greenstone. A feature like the sharp linear fold at the Hill-Trumbull Mine in Sec. 17:56-23 (Plate 3, sheet 9) might be due to later movement along a pre-Animikie fault in the underlying granite. Structures in the Animikie rocks that may be related to zones of weakness along contacts in the "basement" rocks are the fold in the E½ R. 14 W., the Alpena monocline, the folds (faults?) at Mountain Iron, and the Buhl monocline.

VI. CRETACEOUS AND PLEISTOCENE GEOLOGY

PLEISTOCENE GEOLOGY

Lithology of the Drift

In general, the glacial drift of the Mesabi district consists of two till sheets which are at places separated by a layer of stratified sands and gravels. The lower, coarse-textured till is gray or gray-brown and contains abundant pebbles and boulders of rocks found in the source area to the north—granite, greenstone, slate, and graywacke. The upper till is a thin red ground moraine, confined to the south flank of the Giants Range east of Hibbing, that is much more clayey and less stony than the lower till. The drift of the Westernmost Mesabi includes much silty lake clay.

Figure 16. — Recumbent fold in a vertical section of stratified drift at the Embarrass Mine near Aurora.

The intense deformation of some of the beds of stratified drift (Fig. 16) is perhaps due to the pressure exerted by the ice that deposited the overlying till. Other features are the fingers of stony till and the sand "veins" that at places extend down into Precambrian soft ore; Spurr (1894, p. 204) reasoned that the till was forcibly packed into openings in the rigid ice-cemented ore and that the fine sand was carried in by glacial waters.

Thickness of the Drift

The map of the thickness of glacial drift on the Mesabi range (Plate 4) is based on about 1700 thickness measurements obtained from the drilling records of the mining companies. Because the drill holes had been selected primarily for bedrock information, their distribution is spotty, even though more complete data are available in the records. The density of the drill holes in Figure 17 is about average for the areas of Plate 4 containing unbroken thickness lines; drill holes are absent where lines are dashed. The local variations in thickness are so intricate that much more information would be required before the possibility of misinterpretation could be eliminated. No information about texture or composition was recorded, and the term "drift" as used here embraces all unconsolidated surface deposits; it probably includes preglacial stream gravels and perhaps even part of the Cretaceous shale, if the drillers mistook the choppings for drift. However, most of the Cretaceous sediments are easily recognized, and they are considered in this report as part of the bedrock.

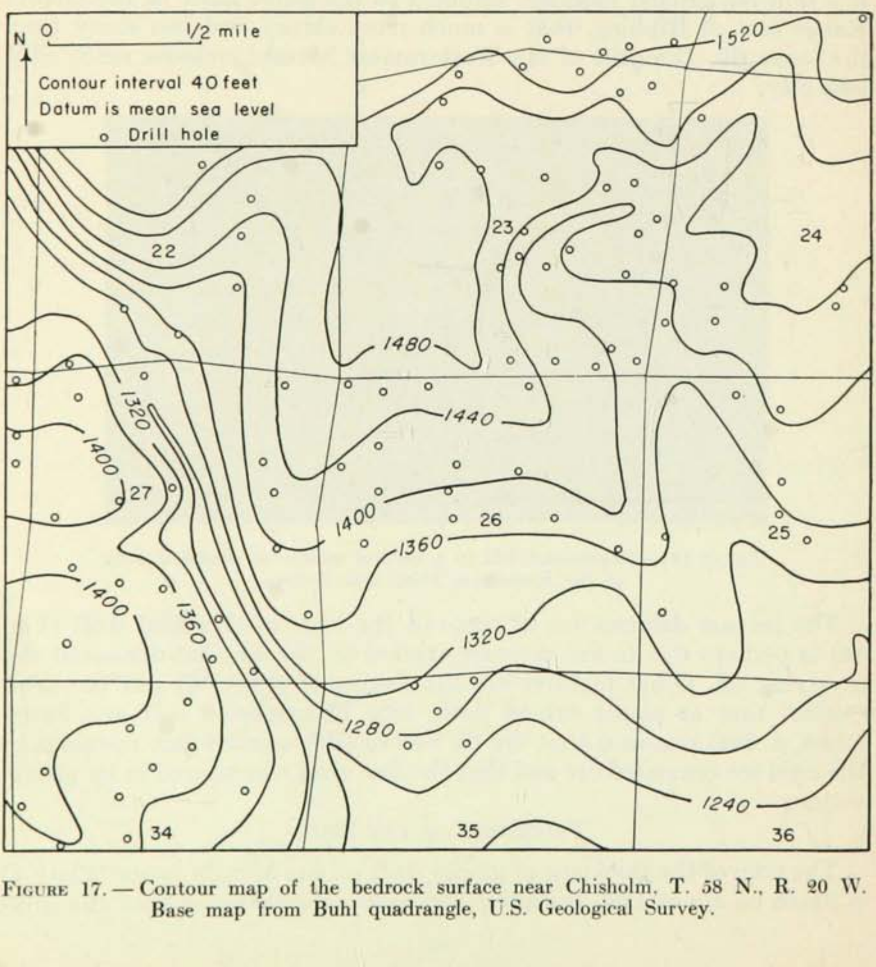

FIGURE 17. — Contour map of the bedrock surface near Chisholm, T. 58 N., R. 20 W. Base map from Buhl quadrangle, U.S. Geological Survey.

Drift thickness over the Biwabik formation increases toward the south away from the Giants Range, and it also increases toward the west. The higher elevations of the resistant, metamorphosed taconites of the East Mesabi account for the numerous natural outcrops and the thin drift, for hills were generally areas of glacial erosion whereas lowlands were areas of glacial deposition. Many outcrops occur along the summits and upper slopes of the Giants Range. The north contact of the Biwabik formation commonly is covered by less than 20 feet of drift; the south contact is generally buried by 80 to 100 feet of drift east of Chisholm but by nearly twice this amount to the west. Drift is very thick over the Westernmost Mesabi. The average depth of the mantle along the north margin of the range in Sec. 34:55-26, for example, is 120 feet. In R. 27 W. drift thickness is 250 to more than 500 feet, and it averages nearly 400 feet.

Bedrock Topography

Because the land surface over the Mesabi range is relatively flat and because there are only a few built-up glacial moraines, local increases of drift thickness generally represent filled depressions in the bedrock. The lines of equal drift thickness on Plate 4 are therefore a guide to the general trend of contours on the top of the bedrock. A comparison of the bedrock contours of Figure 17 with the thickness lines of the same area on Plate 4 illustrates this point. The bedrock contours were constructed by subtracting the thickness of the underlying drift from the elevations of surface contours shown on a recent topographic map and then connecting the resulting points of equal elevation on top of the bedrock. For a few small areas it was necessary to refer to the old topographic map of Van Hise and Leith (1911), in order to determine the surface configuration prior to mining operations.

The following descriptions of some of the principal topographic features of the bedrock (listed from east to west) are based on a visual comparison of drift thickness (Plate 4) with surface topography. Information about present-day land forms is taken from the series of topographic maps started in 1949 by the U.S. Geological Survey. Filled bedrock valleys generally trend at right angles to the strike of the Mesabi range. Because the flow of underground waters is partly controlled by these features, their locations are important to the future development of water resources in the area.

R. 15 W. The drift thickens beneath the gentle surface valley occupied by a small creek which flows from Old Mesaba Lake in SW¼ Sec. 27 south through Sec. 3. The underlying bedrock valley, like many others along the Mesabi range, therefore has a more accentuated relief than the surface valley above it.

The Embarrass River channel is one of the deepest erosional cuts through the Giants Range. The valley is now occupied by Sabin and Wine lakes. The elevation of these lakes is 1365 feet above sea level, and they lie some 300 feet below the summits to either side. Where the Embarrass River crosses the Mesabi range, the elevations of the flanking hills are only about 1500 feet, but the old valley is filled with as much as 300 feet of glacial debris. In Sec. 5, the axis of the bedrock valley is about a quarter of a mile east of the axis of the surface valley.

R. 16 W. A buried tributary to the Embarrass channel extends from Sec. 3 into Sec. 11. Although its bedrock relief is more than 100 feet, it has almost no present topographic expression other than the streamless notch in the Giants Range in Sec. 33. The Biwabik fault strikes toward this notch. A similar buried gorge trends north-northeast through the center of Sec. 9 into an exposed notch (50 feet deep) in Sec. 4. The gentle surface valley above the bedrock valley contains a small intermittent stream.

An old erosion channel trends northwest along the strike of the Corsica fault, from Sec. 18 to a streamless notch in the Giants Range in Sec. 7. The drift in Sec. 19 may be thicker only over the ore body there; the adjacent part of the Giants Range is not notched.

R. 17 W. The local thickening of drift to 40 feet in the N½ Sec. 24: 58-17 outlines a small hill and therefore represents a built-up moraine. Several bedrock valleys occur between Gilbert and Eveleth. A pronounced but completely buried erosion channel in Sec. 4:57-17, near Eveleth, probably connects with the steep-walled streamless ravine, which is 80 feet deep, in the S½ Sec. 28:58-17. The bedrock depression (perhaps partly a sink) in Sec. 5:57-17 follows the length of the Fayal ore body. A poorly defined area of thicker drift just north of Eveleth follows the axis of a well defined surface valley which has a relief of about 100 feet and begins in a swampy notch in the NW¼ Sec. 28:58-17. Because the drift is rather uniformly thin in the Eveleth area, the surface and bedrock topographies are closely similar.

A partially filled valley extends east-northeast from Virginia. The small area of very thick drift in the NW¼ Sec. 5:58-17 may represent a local sink along another bedrock valley that branches northward into several cuts in the Giants Range.

R. 18 W. Glacial debris, generally thin near Mountain Iron is somewhat thicker under a wide surface valley that fans into several tributary notches north of the town. The long valley which strikes north through the centers of Secs. 18, 7, 6, 31, and 30 is also only partially filled. Its relief diminishes southward, from 200 feet where it cuts through the ridge in Sec. 30, to 50 feet in Sec. 18. This decrease is due to the greater amount of fill southward as well as to normal reduction of valley relief downstream.

R. 19 W. Small streams occupy the several partially filled north-trending valleys in R. 19 W. Each valley terminates in a gap in the crest of the Giants Range.

R. 20 W. A shallow bedrock channel curves past the center of Sec. 23 and probably continues into a defile in Sec. 15. A very large trough through the Giants Range is now almost completely buried where it crosses the southern slopes at Chisholm. As much as 200 feet of fill occur under Longyear Lake at the center of the former valley, and extensive swampland covers the low area in the ridge to the north. A tributary channel enters the main valley near the center of Sec. 27.

A shallow depression in the bedrock follows the length of the Morris ore body in the E½ Sec. 31, and the two northward forks probably terminate in the nearby nicks in the Giants Range.

R. 21 W. A long hollow in the top of the bedrock may follow the length of the Hull-Rust-Mahoning pit in Secs. 1 and 2. Another possible channel may extend from Carson Lake, near the center of Sec. 10, northward to a notch in the W½ Sec. 34. The present surface along this route contains a stream and swampy ground. A small bedrock valley occurs in the W½ Sec. 9, and in the NE¼ Sec. 8 the drift is thicker under a stream which originates in a swampy notch in the Giants Range in Sec. 5.

R. 22 W. Bedrock valleys occur in each of Secs. 24, 22, and 29. All the valley heads are buried except for a notch cut 100 feet below the crest of the Giants Range in Sec. 21.

Origin of Modern Topography

Martin (in Van Hise and Leith, 1911, p. 104), in considering the origin of the long ridge of Giants Range granite, rejects folding and faulting as possible factors and suggests that the Pokegama quartzite has until recently capped and protected the granite. Allison (1925, p. 494) has proposed a theory relating the hills to little-eroded upward projections in the original roof of the granite batholith, and he suggests that the transverse valleys were cut in the downward projections of less resistant greenstone, slate, and graywacke. Allison considers that the plain north of the Giants Range is flat because it is underlain directly by the original roof of the batholith, which is flat in that area.

Leith (1903, p. 193) concludes that the major channels transecting the Giants Range were formed prior to glaciation, whereas the numerous smaller, higher notches were cut by glacial meltwater which, ponded north of the Range when the major channels were blocked, overflowed the crest at low points. When erosion again opened up the larger, lower routes, the higher gorges were abandoned. Martin (in Van Hise and Leith, 1911, p. 104), however, writes that "it seems possible that these gaps or cols were already in existence when the glacial streams found and modified them in the manner described by Leith." Almost every notch in the crest of the Giants Range is continuous with a bedrock valley which is buried under glacial debris, and the notches are therefore older than the drift contained in them. The U-shapes of many valleys are probably due both to modification by ice scouring and to partial filling.

In the eastern half of the district the present topography is a subdued impression of the preglacial topography. West of Chisholm even the heads of many former valleys are completely buried, and in the Westernmost Mesabi, where the entire Giants Range is buried, present land forms have little or no relation to the bedrock topography. The location of many surface valleys directly over bedrock valleys is due at places to a cover of drift that is too thin to neutralize the former topography. Where valleys were completely buried, the gentle overlying surface depressions may have been produced by increased compaction of the thicker drift over the valley centers, or by kettle holes formed by the melting of ice blocks buried in the valleys.

Most existing notches are streamless or contain streams that are too small for the size of their valleys. Late-glacial streams, which had a greater available discharge than their modern descendants, may have cut large valleys into the drift. The inheritance of partially filled mature valleys and excessive underflow into the loose drift might also account for some of the undersized young streams.

CRETACEOUS ROCKS

Lithology

Most of the rocks lying unconformably on top of the Animikie group are known to be Cretaceous by their fossil content. A few unfossiliferous, harder deposits may be older than Cretaceous, and Wolff (1915, p. 90) believes these to be possibly of Paleozoic age. For the purposes of the present discussion, all such deposits are classed as Cretaceous. Other conglomerates, like one in the Chisholm buried channel (Wolff, personal communication), contain reworked Cretaceous rocks and are probably Tertiary stream gravels. The following description of typical rocks of the Upper Cretaceous Coleraine formation is based on the report of Bergquist (1944, pp. 2–6).

The basal conglomerate consists of generally angular pebbles and boulders of iron ore, taconite, paint rock, and porphyry, all embedded in a ferruginous cement. Some limonite and hematite pebbles are highly polished. The conglomerate grades upward into grit, glauconitic sandstone, or green siltstone and shale. The Cretaceous deposits are generally less than 50 feet thick, and they contain shark teeth, fish vertebrae and scales, carbonized wood (lignite), and invertebrates that are dominantly Mollusca. Pyrite is commonly associated with the lignite, and octahedra of magnetite are present in the matrix of some of the conglomerates.

Location

The locations of Cretaceous deposits of the Main Mesabi mapped on Plate 4 were obtained from J. F. Wolff, C. H. Scheuer, J. V. Everett, and Josiah Royce. Van Hise and Leith (1911, p. 178) report an additional deposit near the Embarrass River. Cretaceous rocks are relatively plentiful near Calumet and Coleraine, and they form a veneer over much of the Westernmost Mesabi, but none have been found north of the Giants Range.

Many of the conglomerates make iron ores and some are useful for mixing with taconite concentrates. In the past, discovery of Cretaceous deposits was more or less a by-product of exploration for Precambrian ores; the need has arisen, however, for a more specific approach to the exploration for Cretaceous ores in their own right.

The shore line of the Cretaceous sea at its farthest advance probably extended along the south flank of the Giants Range. One conglomerate (now mined out) in the SW¼ Sec. 9:58-17 was at an elevation of 1580 feet, but it appears likely that it was deposited upslope from the shore. Bergquist (1944, p. 2) states: "The rocks are probably largely of continental origin in the east but grade westward into marine materials." As the sea advanced, a blanket of sediments was deposited across much of the Mesabi range, partially filling the pre-existing valleys and veneering the parts of ridges that were submerged. When the sea withdrew, the unconsolidated sediments were exposed to erosion, which continued to glacial times. During this period stream action completely stripped the sediments from the main valley floors, and erosional remnants were left only higher on the valley sides, on the ridge tops, or in depressions that did not contain large streams. Cretaceous rocks were also removed from some of the ridges.

Most of the known Cretaceous rocks shown on Plate 4 form part of the tops of bedrock ridges beneath the drift, so the eroded tops of the deposits are commonly convex up. The bottom surfaces are generally concave up, however, and the deposits are therefore commonly lenticular in shape. This condition, exemplified by the Judson deposit in Sec. 20:58-19, indicates that the sediments most likely to be preserved were originally channel fillings that formed ridges as a result of later erosion. Some sediments were deposited in depressions overlying Precambrian ore bodies, as at the Scranton pit near Hibbing. Bergquist (1944, p. 3) reports that here the Cretaceous conglomerates, 100 feet thick over the center of the ore body, thin to 10 feet at the sides, the top beds being flat-lying. Apparently, no drainage channel was developed over this deposit in post-Cretaceous time.

Most of the Coleraine formation has been stripped from the northern part of the Mesabi range, owing to more active erosion on the steeper slopes and to the original thinning of the sediments in that direction. Consequently, most known deposits (except on the West Mesabi) are near the south margin of the range, and good conglomerate ore overlies the Virginia formation at several places. Commercial deposits might exist a mile or two farther south along favorable ridges, but the thicker drift and probable dilution of ore pebbles by argillite pebbles would be detrimental to profitable mining.

Although the relation between Cretaceous deposits and bedrock topography may be useful for finding new deposits and extensions of known deposits, the target area of exploration is not much reduced, because the valley floors comprise only a small part of the total favorable area. The ore at the Elbern Mine in Sec. 25:58-20, for example, extends well down into a shallow bedrock valley. Gravimetric, electrical, and seismic methods of prospecting would be useful in determining topography beneath the drift, and actual contour lines of the bedrock surface should be drawn where drilling information is sufficient. Magnetic methods have already proved useful in detecting some of the Cretaceous deposits themselves.

VII. ORIGIN OF THE SOFT ORES

CHARACTER OF THE ORE

The soft ores are residual iron-rich concentrates formed by the leaching of silica from taconite. They are porous and friable and consist chiefly of fine-grained martite and goethite (limonite), with small amounts of magnetite and specularite. The only abundant gangue mineral is quartz. Gruner (1946, p. 100) summarizes the changes involved in the formation of ore: "Mineralogically stated, we may say that, in the decomposition of taconite, magnetite yields martite. Hematite remains unchanged. Minnesotaite and greenalite, also a part of the stilpnomelane, alter to goethite, which is brownish-yellow in the ore. Siderite at some places becomes hematite, in others, goethite. A part of the stilpnomelane, particularly in the Intermediate Slate, alters to hematite and kaolinite." Although most iron minerals in the ore are residual, minor amounts have been precipitated from moving solutions. A very detailed discussion of the physical and chemical properties of the ore is given by Gruner (1946, pp. 89–103).

Wolff (1915, p. 219) classifies the ore bodies by shape into fissure, trough, and flat-layered bodies. Gruner (1946, p. 104) estimates that, where conversion to ore is complete, 100 feet of taconite produce about 65 feet of ore. As a result of this shrinkage, the layers in an ore body are commonly slumped into a synclinal attitude (Fig. 18). Many ore bodies become narrower at depth. Bottoms of deposits are generally irregular, and walls are commonly abrupt as well as steep. All known ore bodies reach or closely approach the erosion surface beneath the glacial drift.

HYPOTHESES OF ORE CONCENTRATION

The Weathering Hypothesis

Irving and Van Hise (1892, pp. 283–290) outlined a weathering hypothesis for the origin of the soft ores of the Lake Superior region, and Leith (1903, pp. 260–272) applied this concept to the Mesabi range. According to this view, cold ground waters percolated down through the Biwabik formation for a long period of time, carrying away the silica, derived from chert and iron silicates, and leaving behind an oxidized, enriched ore. These ground waters, because of an alkaline content derived from the Giants Range granite, are considered to be effective solvents of silica (Van Hise and Leith, 1911, p. 539).

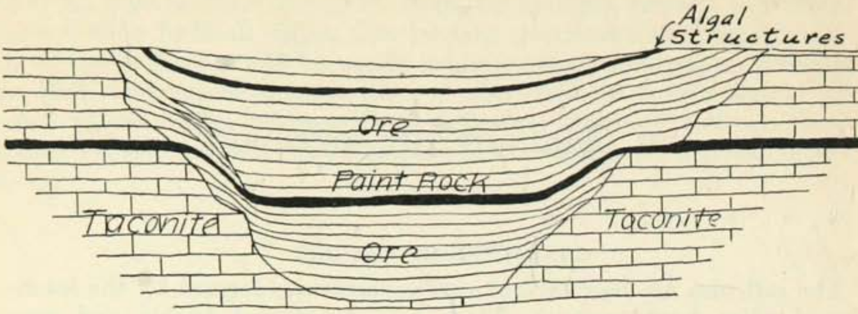

FIGURE 18. — Idealized cross section through a trough ore body, showing slumping of ore (after Gruner, 1924).

THE HYDROTHERMAL HYPOTHESIS

Newland (1922, p. 301) suggested that martite ores might have been oxidized by hydrothermal activity, and Gruner (1926a, p. 644) first presented a comprehensive hypothesis involving hydrothermal solutions. As most recently stated (Gruner, 1937, p. 130), this hypothesis holds that silica, derived from chert and iron silicates, was dissolved and carried away by upward-circulating ground waters which were made hot for a relatively short period of time by gaseous emanations rising from large basic intrusives. These hot ground waters are considered to be effective solvents of silica because the solubility of silica in water increases with increased temperature (Gruner, 1930, p. 698).

FACTORS IN ORE LOCALIZATION

Most factors in ore localization other than those involving the potency and availability of the leaching solutions are applicable under both weathering and hydrothermal conditions. The major controls of ore localization are considered to be structural and lithologic on the Main Mesabi and dominantly lithologic on the rest of the range.

STRUCTURE

Two years after the first Mesabi ore was shipped, Spurr (1894, p. 181) published a remarkably excellent analysis of the role of structure in the localization of ore bodies. He concluded that the circulation of solutions was favored by zones of accentuated fracturing at faults, incipient faults, anticlinal crests, and synclinal troughs. Wolff (1917a, p. 158) reaffirmed this conclusion when more detailed information about the ore bodies was available. It is believed that the present study again substantiates this conclusion, although many other factors must be considered. The following discussion of the effect of structure concerns only the Main Mesabi range.

The basic structural factor in ore localization is the regional dip of the beds of the Biwabik formation. Uniformly flat dips expose a wide belt of the Biwabik formation to ground waters, but this condition alone would not have been sufficient to produce ore; smaller structural irregularities within this belt are necessary. The outcrop belt is wide in the W½ R. 18 W., for example, but the dips are uniform and no soft ores are present.

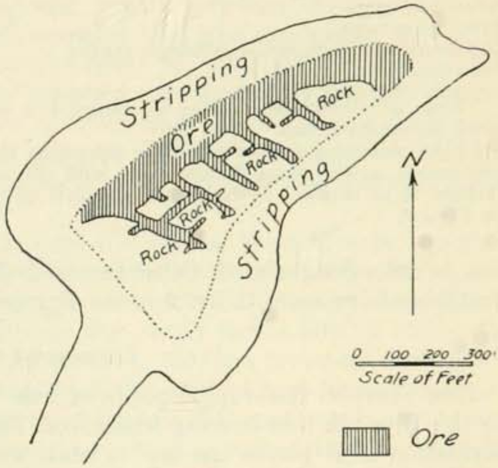

FIGURE 19. — Plan of fissure ore bodies at the Sparta-Malta pit near Gilbert (after Gruner, 1924).

As Spurr states, the most important structural features which are instrumental in localizing ore seem to be zones of accentuated jointing associated with flexures and breaks. Some ore bodies, like the Gilbert Mine (Plate 3, sheet 3), may be in zones of intensified fracturing where the bedding was not displaced. An excellent example of small fissure ore bodies following intersecting joints can still be seen at the Sparta-Malta Mine near Gilbert (Fig. 19). Many large trough ore bodies of the Main Mesabi trend parallel to one of the major joint sets.

The probable relative importance of the various structural features in determining the location of ores is summarized in Table 11. Accentuated jointing produced at folds may account for about 80 per cent of the ore concentration on the Main Mesabi. The proportion of ore produced along faults may be greater than indicated, however, for some of the features shown on Plate 3 may prove to be faults rather than folds. The importance of fracturing at crests and troughs of folds has probably been overemphasized in the past; a great deal of ore — for example, the vast concentration in the Hibbing-Chisholm area — has formed in all parts of minor folds. About three fourths of the ore that can be related definitely to anticlinal crests is associated with the Eveleth anticline, the flanks of which are largely unproductive. Leith's supposition (1903, p. 271) that the circulation of solutions was primarily controlled by the bedding in the troughs of broad, shallow synclines, rather than by joint-

TABLE 11. PROBABLE CONTROLS OF ORE LOCALIZATION ON THE MAIN MESABI RANGE BETWEEN MESABA AND NASHWAUK

Type of control*	Percentage of Total Ore Tonnage†
Folds (all parts of small anticlines and synclines)	50
Faults	15
Monoclines and flanks of folds	10
Anticlinal crests	10
Synclinal troughs	5
Dominantly stratigraphic or lithologic control	5
Unknown	5

* Zones of accentuated jointing associated with each type of structure are considered to be important factors.

† These percentages are obtained by comparing the total ore tonnage (mined and not yet mined) of the Main Mesabi range with the combined tonnages of the ore bodies thought to be localized by each type of control on the basis of the interpretations shown on Plate 3.

ing, is not substantiated. Other factors being equal, the greatest concentrations of ore seem to be in areas of greatest structural deformation.

LITHOLOGY

Iron content. Residual deposits of iron ore are found, of course, only in the Biwabik iron-bearing formation. The original iron content of the formation is at places too low to make ore under normal conditions of oxidation and leaching. For example, it is doubtful that any commercial deposit of soft ore exists between Sugar Lake and eastern Cass County, because all lithic units there have a relatively low iron content.

Guiding of solutions by less pervious layers. Some of the compact slaty layers of the Biwabik formation apparently guided solutions; nevertheless, owing to jointing, none of these layers is completely impervious. The Godfrey-Burt ore body near Chisholm (Plate 3, sheet 6) is an example of an extensive layer of ore (20 to 30 feet thick) which was formed from the layer of silicate taconite directly beneath the "Intermediate slate." The mineralogy of this particular layer of taconite is an additional significant factor in the formation of such ore bodies (Tyler, 1949, p. 1119), however, and solution-guiding by a less pervious layer may be relatively unimportant.

Minerals and textures. The very fine grain size and the intimate intermixing of different minerals are important factors in the great sensitivity of taconite to oxidation and leaching. In general, oxidation appears to have preceded leaching. It has long been recognized (Gruner, 1924, p. 47) that iron silicates are more soluble than chert. Examination of taconite in all stages of decomposition shows that carbonates and iron silicates were altered first; because of the intermixing of fine-grained particles of all the different minerals, it is likely that the pore space formed during this first alteration exposed enormous surface areas of fine chert and magnetite to the later solutions. In his experiments on the oxidation of magnetite by heating in air, Gruner (1926b, p. 379) noted that Mesabi magnetite could be altered to hematite more rapidly than any other of the various samples of magnetite studied, and this was probably because of its small grain size.

Several attempts have been made to explain the extensive oxidation but incomplete leaching of the Cherty members of the West Mesabi. The wash ores of this area consist mainly of geothite-martite bands alternating with partly decomposed, friable, powdery chert bands. Van Hise and Leith (1911, p. 186) contend that the diminished hydraulic head resulting from the lower elevations of the Giants Range in the West Mesabi has caused less vigorous circulation of ground waters. Gruner (1937, p. 123) holds that hydrothermal activity decreased toward the west. Tyler (1949, p. 1119) suggests that the decomposition of iron-silicate grains between chert grains has destroyed the bonding and made the chert friable.

The fact that oxidation is so extensive on the West Mesabi indicates that a decrease in the activity of the solutions is probably not an important factor. The incompleteness of the leaching and the resultant lower grade of the ore probably are due partly to the lack of structural deformation in this area, as Wolff (1915, p. 223) has proposed, and partly to the decrease in the iron content of the original iron-formation westward from Nashwauk. Yet the great extent of the alteration in the absence of much structural deformation suggests some lithologic control. The few fresh cores from the deeper, less oxidized holes in the Cherty members along the West Mesabi consist mostly of varieties of cherty-silicate or silicate taconite (Plate 1, sheet 3). As seen in a few thin sections, the iron silicates in these rocks seem to be abundantly scattered as individual needles rather than concentrated in granules or mats as they are, for the most part, on the Main Mesabi; such a scattering would presumably have favored a widespread partial decomposition and disintegration of the rock. Although this observation is inconclusive, because the available samples are few and may not be representative, Tyler's explanation seems at present to be possible.

A permeability that would promote effective leaching apparently cannot be readily developed in chert beds having little or no admixture of the other minerals, even where the beds are well jointed. Cherty rocks are far less decomposed on the Westernmost Mesabi than on the West Mesabi. Most cores from the area west of R. 26 W. are completely fresh. The reason for this lack of decomposition of the cherty rocks is probably the almost complete lack of iron silicates and other easily altered minerals in the chert. The algal beds of the Main Mesabi offer another example of this condition, for these highly cherty rocks remain as undecomposed blocky beds in many ore bodies.

Most minerals in taconite were formed under reducing conditions, and the mineral assemblage is therefore particularly unstable under the oxidizing conditions of the ore-forming process. On the other hand, the few beds that were originally deposited under oxidizing conditions resist concentration by secondary oxidation and leaching. The hematitic "red basal taconite" rarely is altered to ore, and the overlying unit of regularly banded cherty taconite, which contains both original hematite and magnetite but few iron silicates, is also resistant to decomposition at many places. On the Main Mesabi range the principal examples of structural features having no associated ore are near the north boundary of the Biwabik formation between Chisholm and Keewatin in Secs. 21 and 30:58-20, 36:58-21, and 3 and 5:57-21 (Plate 3, sheets 6 and 7). Nearby ore bodies do not extend northward into the basal units of the Biwabik formation, but most of these ore bodies do extend farther north where the rocks are more deformed.

Effect of Metamorphism. Most observers (Spurr, 1894, p. 199; Leith, 1903, p. 273; Grout and Broderick, 1919b, p. 46; Royce, 1942, p. 62) believe that the effects of high-temperature metamorphism on iron-formations are adverse to the secondary concentration of soft iron ores. No soft ores occur in the metamorphosed taconites of the East Mesabi range. The reason given for this lack of secondary concentration is that the great coarsening of the grain by metamorphism prevents effective solution of silica (Leith, 1903, p. 273).

Tyler (1949, p. 1102), however, contends that metamorphism has produced grunerite, minnesotaite, and stilpnomelane at the expense of primary chert, siderite, and greenalite; he believes that the soft ores resulted from the preferential oxidation and leaching of a metamorphic phase of the iron-formation "in which the chert has been largely or completely combined with iron to form iron silicates." He states: "Chert has not been removed as such in the quantities hitherto considered necessary for the formation of ore." Because of the following considerations, it is believed that Tyler's contention is not applicable to the Mesabi range:

1. For reasons already outlined above (see pp. 35–37), it is believed that minnesotaite and stilpnomelane are original constituents of the Biwabik formation rather than products of metamorphism. The evidence indicates that neither regional nor local metamorphism can reasonably account for the distribution of minnesotaite and stilpnomelane.

2. The grunerite-bearing metamorphosed taconites of the East Mesabi are almost completely unoxidized and unleached, yet these rocks are considered moderately favorable for secondary concentration by Tyler. He states (1949, p. 1118) that "the lack of extensive oxidation and leaching of the magnetite-grunerite phase cannot be correlated with the mineral composition or the grade of metamorphism attained, but appears to be dependent upon other factors, such as the source of the oxidizing solutions or the lack of adequate channelways." Under either the hydrothermal hypothesis or the weathering hypothesis, however, oxidizing solutions would have been especially active on the East Mesabi. The relatively strong structural deformation in parts of the East Mesabi

(see p. 57) indicates that adequate channelways are probably present. Only one small deposit of commercial soft ore occurs along the major fold in the E½ R. 14 W. (Plate 3, sheet 1). This deposit at the Siphon Mine is in the northernmost end of the fold, where the taconites are farthest from the Duluth gabbro and are therefore the least metamorphosed of any rocks in the fold. Although the deformation is greater toward the south, soft ores are there concentrated to a lesser extent or not at all, and the lack of concentration probably can be directly related to the higher grade of metamorphism toward the south.

3. The very detailed studies by Wolff (1917a, p. 156), Gruner (1946, pp. 99–101), and others show that each type of ore can be correlated with the type of taconite from which it was derived. Almost any mineral assemblage having a high enough iron content can be concentrated to ore. The Cherty members, where oxidized and leached, generally make the high-grade blue ore rich in martite, whereas the silicate-rich Slaty members generally make paint rock or the yellow ore that is rich in limonite. The mineral most abundantly associated with magnetite in fresh taconite is quartz rather than any iron silicate, and many ores consist mostly of martite. Ore layers composed dominantly of martite and containing little limonite grade into cherty taconite composed dominantly of chert and magnetite without much iron silicate. Although it appears true that silica in the form of iron silicates can be more readily leached than silica in the form of chert, it seems inescapable that much chert as such was removed from taconite during ore formation.

4. Tyler (1949, p. 1119) explains the localization of some Mesabi ore bodies by contact metamorphism next to quartz veins, and he suggests that the "formation of secondary iron silicates may be the factor which relates the quartz veins to the ore bodies." These quartz veins, which are probably of hydrothermal origin (Stephenson, 1952, p. 745), are slumped with the ore, and therefore were formed earlier than ore. Quartz veins are very rare west of Buhl. Metamorphic effects observed where such veins cut fresh taconite are inconsequential. It seems probable that the factor which commonly localized both quartz veins and ore bodies is accentuated jointing.

5. Regarding the contacts of ore bodies, Tyler (1949, p. 1118) states: "These abrupt contacts reflect the sharp limits of the essentially chert-free iron silicate areas which are observed in the metamorphic facies of the iron formation." This statement implies the existence of local contact metamorphic zones for which no obvious causes are apparent; a better explanation would be that the abrupt vertical contacts of many ore bodies are related to the boundaries of zones of accentuated vertical jointing.

CONCLUSION

None of the available field, experimental, and theoretical evidence bearing on the manner of origin of the soft ores can be considered at present as absolute proof of either weathering or hydrothermal activity. In the interpretation of available evidence, the conclusions of each observer depend upon his personal emphasis, for either hypothesis is capable of explaining the existing conditions. My personal feeling is that the weathering hypothesis can probably explain the existing conditions better than the hydrothermal hypothesis insofar as the Mesabi range is concerned.

The greatest single factor leading to this conclusion is the close relation of oxidation and leaching to the erosion surface beneath the glacial drift, a factor noted by Spurr (1894, p. 158) and all subsequent observers. All ore bodies that have been sufficiently explored are known to connect with this surface, either vertically or up-dip. Ore layers under the "Intermediate Slate" commonly extend down the dip from trough ore bodies which reach this surface. Oxidation and leaching of the iron formation die out southward under the Virginia formation in all known instances.

An inspection of nearly all the records of exploration on the range revealed the most striking feature in the distribution of oxidation and leaching to be that taconite characteristically is most altered directly beneath the surface at the base of the drift and is fresher at depth. This surficial alteration of taconite was noted in 112 out of 140 drill holes distributed all along the range; yet these holes were selected for study because their cores were the least altered of any available. The depth of this alteration commonly is 5 to 10 feet, although it is considerably greater at many places, particularly near ore bodies; each ore body can be considered as a locally more extensive development of this zone of alteration. Even the argillite of the Virginia formation is bleached and softened at places directly beneath the preglacial erosion surface, the maximum observed depth of alteration being 40 feet. Such relations suggest downward-circulating solutions.

Under the hydrothermal hypothesis, the highest temperatures and hence the most effective solvent powers would probably be developed in ground waters under greatest pressure at greatest depth. Under such conditions, the extent of decomposition of taconite would be expected to diminish upward rather than downward.

Many factors in the distribution of the superficial alteration cannot be fully evaluated. Glacial erosion probably stripped off much of the decomposed rock at the surface. Some types of taconite seem to decompose in a few years after exposure to the present climate, but others are completely unaffected. The few natural outcrops on the Main range are little altered, but most of them consist of the more resistant rocks of the basal part of the Lower Cherty member.

It may be argued that weathering is capable of causing minor alteration at the surface but not capable of causing oxidation and leaching to a depth of 800 feet or more. The quantity of alteration below 500 feet is generally minor, however, and the areas of greatest decomposition are all closer to the surface. Nevertheless, the alteration is deeper than that normally produced by the weathering of other types of rock.

If such alteration is ascribed to weathering, it must be assumed that the mineral assemblage in taconite is especially sensitive to surface oxidation and leaching for the reasons already outlined under *Minerals and textures* (see pp. 80–82). Such weathering would presumably be most effective under a humid tropical climate. The alteration could be carried to relatively great depth by the rapid flow of waters, perhaps under artesian pressure, down through a continuously advancing zone of highly permeable ore. The process would tend to reinforce itself, because the establishment of a permeable zone would localize the flow of much additional water. There would be no reason to expect a deposit of the previously leached silica at the base of ore bodies, because the silica would most likely remain in the solutions as they moved back toward the surface along joints and thence flowed toward the sea (Wolff, 1918, p. 1525).

Under the hydrothermal hypothesis relatively rapid solution of silica could have taken place during periods of hydrothermal activity, probably in the Keweenawan (Gruner, 1930, p. 864). Under the weathering hypothesis, relatively slow solution of silica could have taken place during most of the later Keweenawan, the Paleozoic, Mesozoic, and Cenozoic. The unslumped Cretaceous beds that overlie some ore bodies suggest that most enrichment was complete before the Cretaceous.

If ores were concentrated by cold ground waters, there might be a genetic relation between some small ore bodies and the preglacial bedrock drainage system. This drainage system probably has not changed much for a long time, for the scattered remnants of originally thin Cretaceous deposits indicate little post-Cretaceous erosion. The slumping that accompanied the formation of ore probably made surface depressions that may have localized some streams, but, on the other hand, seepage and stream underflow may have propagated ore. Because it is obviously difficult to distinguish cause from effect, any supposed genetic relation between streams and ore must be considered speculative. Yet it is possible that small ore bodies may have been localized where streams crossed minor structural irregularities of a type which would otherwise have remained barren without the larger amounts of leaching water supplied by the streams.

Many of the bedrock valleys discussed in Chapter 6 are associated with ore bodies, and several of the small ore bodies in R. 19 W. (Plate 3, sheets 5 and 6) are possible significant examples of the relation between streams and ore. The Atkins, Forsythe, and Midway mines are in an area of fairly uniform dip, but all lie along segments of partly buried former valleys that head in notches in the Giants Range. The ore bodies near the south contact of the Biwabik formation in this area follow the Buhl monocline; but the Woodbridge and Wabigon ore bodies (Secs. 16

and 17:58-19), which are the only two that extend northward into the area of uniform dip, both lie along the axes of buried valleys.

Gruner (1946, p. 106) states: "As it is quite certain that there has been erosion, perhaps 200 feet, probably more, since the formation of the ore bodies, the covering by Virginia slate must have been more widespread on some ore bodies at one time." If it is considered, however, that ore bodies were formed, for the most part, in the exposed iron-formation concurrently with erosional downcutting over a long period of time, there is no necessity for postulating the formation of ore under a thick cover of Virginia argillite (Leith, 1903, p. 270). Soft ores would not necessarily be eroded away much more rapidly than taconite; if slumping in an ore body made a depression in a stream course, for example, the stream would probably deposit rather than erode at that point.

A major obstacle to the acceptance of the hydrothermal hypothesis in the Mesabi district is the scarcity of direct evidence of igneous activity. It seems unlikely that igneous activity represented only by a few small dikes, sills, and veins could furnish the enormous quantities of emanations required to heat ground waters (Leith, 1931, p. 282). Inasmuch as the few igneous rocks are coarse-grained and were probably intruded at considerable depth, it is likely that any ores formed at the surface at the time of intrusion would now be eroded away. No special relation exists between the largest intrusive body, the Aurora sill, and ore. Some decomposed Aurora dikes are in ore bodies, but the dikes were probably localized along the same structural features that later localized ore. The slumped veins of quartz in some ore bodies indicate that hydrothermal solutions, where present, were depositing rather than removing silica at the sites of later ore concentration.

Circulation of cold ground water down through the Biwabik formation sufficient to account for the observed oxidation and leaching is a possible but unproved explanation. Both Leith (1903, p. 177) and Wolff (1915, p. 222) report that some holes drilled through the base of the Virginia formation encountered water under pressure. Grout and Broderick (1919b, p. 49) mention a flow of water under artesian pressure from a drill hole in Sec. 14:59-14.

For a fuller statement of the evidence for the hydrothermal hypothesis in the several districts of the Lake Superior region, the reader is referred to the papers of Gruner (1926a, 1930, 1932, 1937).

Many conditions at the Mesabi range are difficult to explain under any concept yet proposed; consequently, no present opinion concerning the manner of origin of the soft ores can be considered final. Conditions at some of the other iron ranges in the Lake Superior region differ markedly from the conditions at the Mesabi range, and these differences imply variations in the controls of ore genesis.

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