Geology of the Menominee Iron-Bearing District Dickinson County, Michigan and Florence and Marinette Counties, Wisconsin

GEOLOGICAL SURVEY PROFESSIONAL PAPER 513

Prepared in cooperation with the Geological Survey Division, Michigan Department of Conservation

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By R. W. BAYLEY, C. E. DUTTON, *and* C. A. LAMEY

With a section on THE CARNEY LAKE GNEISS

By S. B. TREVES

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1941 Lamey, C. A., and Dutton, C. E., Geology of the Menominee range vicinity of Iron Mountain [Michigan]: Michigan Dept. Conserv., Geol. Survey Div. Prog. Rept. 6, 14 p.

A preliminary geologic map is presented for the area extending from Quinnesec west along the south iron range to the Menominee River.

Stratigraphy, structure, and exploration possibilities are discussed.

1941 Pettijohn, F. J., and Hildebrand, F. A., 1941, Archean-Huronian unconformity of the Menominee iron range, Michigan [abs.]: Geol. Soc. American Bull., v. 52, no. 12, pt. 2, p. 1927.

"The pre-Lower Huronian age of the granite and granite gneiss complex northeast of the Menominee iron range is reaffirmed by discovery of a hitherto undescribed exposure of basal conglomerate near Fern Creek in Dickinson County, Michigan. * * *. The conglomerate lies between the granite gneiss and the Sturgeon quartzite."

1942 Dutton, C. E., Economic geology of a part of the Menominee range. Dickinson County [Michigan]: Michigan Dept. Conserv., Geol. Survey Div. Prog. Rept. 9, 27 p.

Several areas believed by the writer to warrant exploration for iron ore are described. Reference is made to beneficiation tests run on some specimens of iron-formation from the Menominee range.

1942 Lamey, C. A., and Dutton, C. E., Geology of Menominee Range, Norway to Waucedah [Michigan]: Michigan Dept. Conserv., Geol. Div. Survey Prog. 8, 20 p

A preliminary geologic map is presented of the area between Norway and Waucedah, Mich. Details of structure and stratigraphy are discussed. A major fault is said to separate the north and south iron ranges.

1943 Pettijohn, F. J., Basal Huronian conglomerates of the Menominee and Calumet districts, Michigan: Jour. Geology, v. 51, no. 6, p. 387–397.

The author describes an unconformable relationship between the Fern Creek formation and the pre-Huronian gneiss complex in Dickinson County, Mich.

1947 Higgins, J. W., Structural petrology of the Pine Creek area, Dickinson County, Michigan: Jour. Geology, v. 55, no. 6, p. 476–489.

Higgins describes the results of petrofabric investigation done on the belt of Sturgeon quartzite and adjacent gneiss that extends from sec. 20, T. 41 N., R. 29 W. (in central Dickinson County), south and southeast for a distance of 15 miles.

1948 Trow, J. W., The sturgeon quartzite of the Menominee district, Michigan: Chicago Univ. Ph. D. thesis, 60 p.

Includes outcrop map of quartzite belt and also detailed maps of selected areas.

1949 Richardson, E. J., Jr., Some Lower Huronian stromatolites of northern Michigan: Chicago Nat. History Mus., Fieldiana—Geology, v. 10, no. 8, p. 47–62.

A description and a classification are given of algal structures in the Lower Huronian dolomites of Michigan.

1950 Dutton, C. E., Progress of geologic work in Iron and Dickinson Counties, Michigan: U.S. Geol. Survey Circ. 84, 7 p.

Dutton gives a brief review of current work of the U.S. Geological Survey in the region.

1952 Prinz, W. C., The geology of a portion of the Quinnesec igneous complex, south of Iron Mountain, Michigan: Columbus, Ohio State Univ. M.S. thesis, 90 p., 19 pls.

1952 Shapiro, Norman, The geology of a part of the [Precambrian] Quinnesec greenstone complex, Dickinson County, Michigan: Columbus, Ohio State Univ. M.S. thesis, 78 p., 17 pls.

1953 Froelich, A. J., The geology of a part of the Wisconsin granite-Quinnesec greenstone complex, Florence County, Wisconsin: Columbus, Ohio State Univ. M.S. thesis, 120 p., 22 pls.

1953 Wier, K. L., Balsley, J. R., Jr., and Pratt, W. P. Aeromagnetic survey in part of Dickinson County, Michigan, with preliminary geologic interpretation: U. S. Geol. Survey Geophys. Inv. Map GP–115.

Dickinson County was surveyed with an airborne magnetometer, and traverses were flown at 500 feet above the ground along north-south lines one-third of a mile apart. The magnetic profiles and a map showing the crests of magnetic anomalies accompany the report.

1955 James, H. L., Zones of regional metamorphism in the Precambrian of northern Michigan: Geol. Soc. America Bull., v. 66, no. 12, pt. 1, p. 1455–1487.

Zones of regional metamorphism are delineated for northern Michigan, including the Menominee district. The isograd map shows the west half of the Menominee district of Dickinson County in the biotite zone of regional metamorphism, and the east half in the chlorite zone. Most of the granite area to the north of the district is shown in the garnet zone, as in the granite area to the south of the district.

1955 Thompson, G. L., A study of part of the Pre-Cambrian granite-greenstone complex in southeastern Florence County and northwestern Marinette County, Wisconsin: Columbus, Ohio State Univ. M.S. thesis, 86 p., 8 pls.

1957 Fulweiler, R. E., The geology of a part of the igneous and metamorphic complex of southeastern Florence County, Wisconsin: Columbus, Ohio State Univ. M.S. thesis, 68 p., 14 pls.

1958 James, H. L., Stratigraphy of pre-Keweenawan rocks in parts of northern Michigan: U.S. Geol. Survey Prof. Paper 314–C, p. 27–44.

A revised stratigraphic column is presented for the pre-Keweenawan rocks of Iron and Dickinson Counties, Mich. The Precambrian rocks are placed in three categories: lower Precambrian, middle Precambrian, and upper Precambrian. A sequence of lower Precambrian rocks—Archean of older reports—is established on the basis of detailed mapping in central Dickinson County. Middle Precambrian rocks are said to consist of the Animikie series—"Huronian" of older reports—and post-Animikie but pre-Keweenawan igneous rocks. The term "Huronian," used in the district for nearly 60 years, is not used because of the uncertainty of correlation with the type Huronian of Canada. The revised stratigraphic column follows (table 5).

1959 Prinz, W. C., Geology of the southern part of the Menominee district, Michigan and Wisconsin: U. S. Geol. Survey open-file report, 221 p.

GEOLOGY OF THE MENOMINEE IRON-BEARING DISTRICT, DICKINSON COUNTY, MICHIGAN AND FLORENCE AND MARINETTE COUNTIES, WISCONSIN

By R. W. Bayley, C. E. Dutton, and C. A. Lamey

ABSTRACT

The part of the Menominee district covered by this report is mainly in southern Dickinson County, in the northern Peninsula of Michigan, but partly in Florence and Marinette Counties, Wis. It is on the upland plateau that lies between Lake Superior and Lake Michigan about 1,000 feet above the lakes. The topography is diverse but gentle; local relief does not exceed about 500 feet, and commonly it is about 100 feet. Large parts of the district are poorly drained and swampy; the rest is made up of low ridges, rock hills, and sand plains. The drainage ways are young and poorly integrated. The Menominee River, which flows to Lake Michigan, is the main stream, and it is fed by several important tributaries, among them the Sturgeon and Pine Rivers and Pine Creek.

The district is chiefly underlain by lower and middle Precambrian rocks, formerly designated Archean and Algonkian rocks, which are extensively covered by Pleistocene glacial deposits, and locally by lower Paleozoic sandstone and dolomite.

The lower Precambrian rocks occur in two separate areas, one in the northeast quarter composed of granitic gneiss, the Carney Lake Gneiss, and one in the southwest quarter composed chiefly of basaltic and felsic metavolcanic rocks, the Quinnesec Formation. Between these areas of older rocks is a northwest-trending structural trough within which are middle Precambrian formations of the Animikie Series. They are, in ascending order: the Fern Creek Formation, the Sturgeon Quartzite, and the Randville Dolomite, which combined make up the Chocolay Group; the Felch Formation, and the Vulcan Iron-Formation, which make up the Menominee Group; and the Michigamme Slate and Badwater Greenstone, which make up the Baraga Group. The Chocolay Group is separated from the Carney Lake Gneiss by a profound unconformity. The groups are separated by slight unconformities or disconformities.

These rocks are cut by late-Animikie and post-Animikie igneous rocks of various kinds. They are cut at many places by metagabbro dikes and sills, some of the latter being several thousands of feet thick, and locally by granite plutons (Hoskin Lake Granite) and quartz-diorite (Marinette Quartz-Diorite), and by numerous pegmatite dikes.

The lower Precambrian rocks were folded, intruded by granitic plutons, and metamorphosed in pre-Animikie time; periods of uplift and erosion separated groups within the Animikie Series, and orogeny and regional metamorphism closed the Animikie. Of the igneous rocks noted, the metagabbro was intruded prior to or during the post-Animikie orogenic event. The times of intrusion of the Hoskin Lake Granite and Marinette Quartz Diorite have not been determined. The mineral ages of the intrusives suggest that they are of post-Animikie age. Fresh and undeformed diabase dikes, probably of late Precambrian (Keweenawan) age, are found at a few places.

Three main structural elements are recognized: the Carney Lake anticline, the Quinnesec or Wisconsin anticline, and the Menominee trough. The anticlines lie on either side of the trough, and they began moving upward intermittently in Randville time. The Menominee trough is homoclinal—a syncline whose south half has been removed by faulting. The homocline contains abundant secondary folds and is repeated near its middle by a major longitudinal fault; thus, it is separated into two homoclinal parts, each of which is marked by topographic ridges of Vulcan Iron-Formation which are referred to locally as the north and south iron ranges.

The principal economic asset of the district is the Vulcan Iron-Formation; materials of lesser importance are dolomite, sand, gravel, and quartzite. About 80 million tons of direct-shipping iron ore was produced, mainly from underground mines, in the period 1877–1945. At present nearly all iron mining has been abandoned. A large reserve of iron remains, but it must be won from low-grade iron-formation by modern beneficiation methods.

INTRODUCTION

Early geological investigations of the Menominee iron-bearing district of Michigan were conducted by the U.S. Geological Survey from 1896 to 1910 in connection with a study of the geology of the Lake Superior region. The results of this work were included in Van Hise and Bayley (1900), Bayley (1904), and in abbreviated form in Van Hise and Leith (1911).

In the early 1900's the Menominee Range matured and declined as a major producer of iron ore. By 1945, after nearly 70 years of continuous production of iron ore, the Menominee district became almost dormant, as the last of the underground mines was abandoned. In that 70-year period about 81 million long tons of iron ore was mined, constituting about 2 percent of the known iron-formation in the district to a depth of 1,000 feet.

In 1937 the Geological Survey Division of the Michigan Department of Conservation initiated a program of detailed geologic mapping and magnetic surveying in an attempt to revitalize the mining industry in the district. Fieldwork was carried on for five field seasons on the south iron range of the district by Carl E. Dutton and Carl A. Lamey, and the results of their investigations were published in a series of progress reports (1939–42).

In 1943 a cooperative program was undertaken by the U.S. Geological Survey and the Michigan Geological Survey Division to remap and restudy the iron-bearing districts in Iron and Dickinson Counties, and in 1951 fieldwork leading eventually to this report was started on the north iron range by C. A. Lamey.

The purposes of the present work in the Menominee district were to consolidate the earlier mapping by Lamey and Dutton and new mapping, including that of areas studied for several Ph. D. theses, on modern topographic base maps; to affect map changes necessitated by new data; and to examine the possible economic potential of the iron-formations of the district in the light of recent advances in iron-ore beneficiation techniques.

LOCATION, ACCESSIBILITY, AND INDUSTRY

The region described is the Menominee iron-bearing district in southern Dickinson County, Mich., and parts of adjacent Wisconsin (see fig. 1). The Florence, Wis., mining district is to the west, and the Calumet and Felch Mountain districts are 3 and 6 miles, respectively, to the north.

The mapped area includes, from west to east, the Iron Mountain, the Norway, and the Vulcan 7½-minute quadrangles (pls. 1, 2, and 3, respectively) and an additional strip 1½ miles wide along the north border of these quadrangles. The total area is about 162 square miles, bounded approximately by meridians 87°45′ W. and 88°07′30″ W. and parallels 45°45′ N. and 45°55′ N.

Most of the Menominee district is easily accessible, because it is diagonally transected from east to west by U.S. Highway 2, which links the towns of Waucedah, Loretto, Vulcan, Norway, Quinnesec, and Iron Mountain. Further, a network of state and county roads connects with this main highway. Only the northeast quarter of the area is nearly uninhabited and accessible with difficulty.

Two railways serve this district. The main line of the Chicago and North Western Railway Co. parallels U.S. Highway 2 across the district. The line makes a loop at Iron Mountain and crosses again, passing along the north sides of Lake Antoine and Fumee Lake. The Chicago, Milwaukee, St. Paul and Pacific Railroad passes north-south through the west part of the district.

Each of the towns along the main highway grew up near producing iron mines, and in turn each has weathered the economic crises brought about by the abandonment of those mines. The communities are now sustained by a variety of small industries, by merchandising, and by a growing resort trade. Iron Mountain is the county seat of Dickinson County, and the twin cities of Iron Mountain-Kingsford, with a population of approximately 14,700, comprise the fourth largest community in the Upper Peninsula of Michigan. The city of Norway has a population of about 3,250 (1956), and Quinnesec, Vulcan, Loretto, and Waucedah are small settlements of only a few hundred families. Rural families, living mainly in the north and south parts of the district, tend approximately 20,000 acres of tillable land from which hay, oats, potatoes, and dairy products are produced.

TOPOGRAPHY

The topography of the Menominee district is typical of the glaciated upland that lies between Lake Superior and Lake Michigan. Low swampy tracts alternate with flat to gently rolling sand plains, containing hummocky and pitted drift areas, and rocky uplands. The average elevation above sea level is about 1,100 feet; the maximum, at Millie Hill (pl. 1), is about 1,570 feet; and the minimum, in the southwest corner of the Vulcan quadrangle (pl. 3), is about 840 feet.

The most conspicuous landscape features are two southwest-trending ranges of wooded hills which dominate the central part of the district, barren to partly wooded cliffs 100–200 feet high to the north of these ranges, and a high bluff along the Menominee River near Niagara to the south of the central ranges. The central ranges comprise the north and south iron ranges, composed of erosion-resistant Precambrian iron-formation and dolomite and generally capped by outliers of Cambrian and Ordovician sandstone or dolomite. Belts of flat to gently rolling low country underlain by easily eroded schist and slate, and almost completely covered by glacial deposits, flank the iron ranges. Being well drained and sparsely wooded, these belts furnish the townsites and agricultural tracts. This low glacial topography ends abruptly in the northeast and north, near Pine Creek (pls. 2, 3) against the northern cliffs, which are composed of quartzite and beyond which the area is rough and characterized by glacially rounded rock hills and ridges rising from heavily wooded swampland. The area that lies south of the Menominee River and the high bluff near Niagara, composed of metagabbro, is

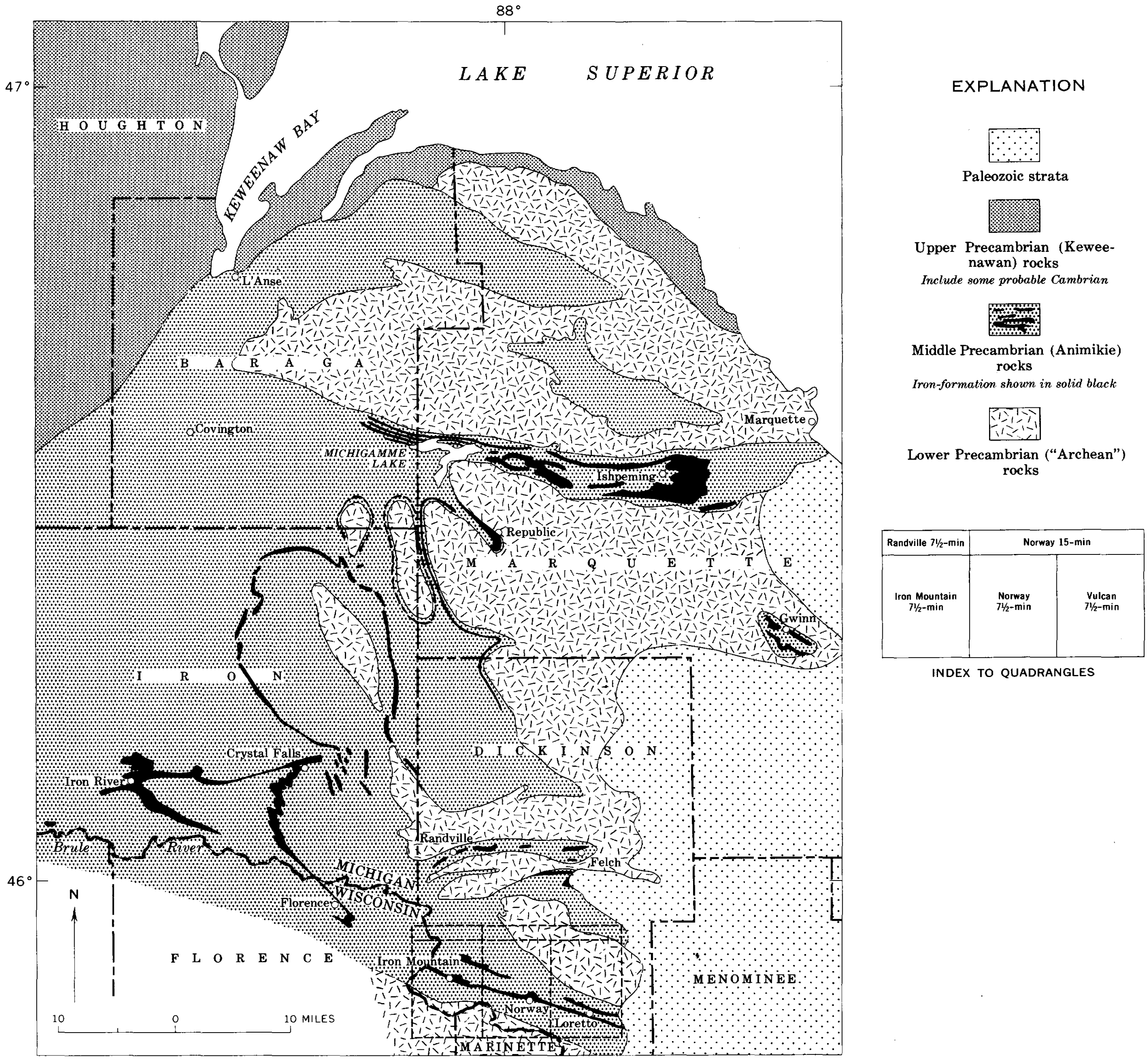


Figure 1.—Generalized geologic map of part of northern Michigan showing the location of the Menominee district with respect to the principal geologic features.

characterized by topography similar to that north of the quartzite cliffs.

W. S. Bayley (1904) formulated a fundamental concept bearing on the origin of the topography of the Menominee district. He postulated from the distribution of outliers of Cambrian strata that the main features of the present topography resulted from Precambrian erosion and that the features now shown are, for the most part, ones reexcavated in detail by post-Cambrian erosion. He stated (p. 128):

The major features of the topography of the district not only antedate the Glacial epoch, but they were determined before Upper Cambrian time. Not only are the hills of the region capped by the horizontal Upper Cambrian sandstone, but remnants of this formation have been found in the cross gorges at Iron Mountain, Quinnesac, and Norway, and in the longitudinal valleys at several places.

Although very resistant to weathering agencies, the sandstone has, nevertheless, been almost completely removed from the valleys, while it has escaped erosion mainly on the tops of the ridges. The topography which has resulted from the removal of this covering is therefore very similar to that which characterized the surface at the beginning of Upper Cambrian time. The present valleys correspond to the pre-Cambrian valleys, and the present hills to elevations that existed on the pre-Cambrian surface. The floors of the valleys may have been lowered somewhat since the sandstone disappeared, but since the bases of sandstone remnants are in several widely separated localities either at the present valley surfaces or only a few score feet above them, the differences in elevation between the pre-Cambrian valley floors and those of the present valleys can not be great.

The resurrected Precambrian topography has been modified greatly by glacial scour, by deposition of glacial deposits, and by postglacial erosion of such deposits. Large tracts, particularly to the north and northwest, are covered by glacial drift of variable thickness. In the northwest part of the Norway quadrangle (pl. 2) the drift forms a thin mantle 15–20 feet thick on gently undulating bedrock, but a short distance to the east, in the valley of Pine Creek, the drift is more than 250 feet thick. These large variations in the thickness of the glacial deposits have considerably modified and subdued the preglacial topography. Typical knob-and-kettle morainic terrain occurs in patches locally about the district. Large areas of relatively flat outwash plain composed of bedded sand and gravel as much as 200 feet thick occur at Iron Mountain, in the Lake Antoine-Fumee Lake area, south of Quinnesec, and southwest of Loretto. These gravel deposits have been terraced and dissected by postglacial erosion. Hummocky outwash, pitted outwash plain that contains a few small lakes and numerous ponds, covers an extensive area which extends into Wisconsin south and west of Iron Mountain, and two smaller areas, one south of Norway, and another south of Loretto and Waucedah.

DRAINAGE

The area is drained by the Menominee River and its tributary rivers, the Sturgeon, the Pine, and the Little Popple, and by Pine Creek, which is tributary to the Sturgeon. Each of these flows alternately on glacial deposits and on bedrock, and they therefore do not adhere strictly to preglacial stream courses. These streams are apparently antecedent, for each transects the nearly vertical rocks of the district at some place along its course.

Many references to various falls along the Menominee River at places where the waters traversed well-exposed bedrock are contained in the early geologic literature. These falls, the Sturgeon, the Little Quinnesec, the Big Quinnesec, and Twin Falls have been utilized as sites for hydroelectric plants. The falls of the Sturgeon River northeast of Loretto has also been adapted to the generating of electricity.

PREVIOUS WORK AND SUMMARY OF THE LITERATURE

Geological investigations on the Menominee Range in southern Dickinson County began in 1848 with the discovery of iron-formation by early explorers, J. W. Foster and S. W. Hill. In the years that followed, dozens of explorers and geologists examined the rocks of the district and wrote descriptions of their findings. Prominent among these were N. P. Hulst, Herman Credner, Charles Whittlesey, T. B. Brooks, C. E. Wright, Carl Rominger, A. D. Irving, G. H. Williams, C. R. Van Hise, C. K. Leith, and W. S. Bayley.

The prime reference for the geology of the district is W. S. Bayley, (1904). In it the major aspects of the geology of the district are established, and a summary of the literature published prior to 1904 is given.

Since the publication of W. S. Bayley's study, many reports have been published on aspects of the Menominee geology. These reports are listed chronologically, and some of them are discussed briefly in the following pages.

1905 Report of the special committee on the Lake Superior region: Jour. Geology, v. 13, no. 2, p. 89–104.

The committee proposes a threefold division of the Huronian rocks of the Marquette district, to be called Lower, Middle, and Upper Huronian. The Lower Huronian of previous reports is divided into Lower and Middle by an unconformity between the Wewe slate and the Ajibik quartzite.

1907 Lane, A. C., and Seaman, A. E., The pre-Ordovician, pt. 1 of Notes on the geological section of Michigan: Jour. Geology, v. 15, p. 680–695.

Writing about the Mid-Huronian, which included the Negaunee formation, the authors state (p. 685), "This we are inclined to believe is the main iron-bearing formation, not only of the Marquette range, but of the Menominee range as well."

1911 Conlin, Thomas, and O'Brien, P., Historical sketch of the Menominee range: Lake Superior Mining Inst. Proc., v. 16, p. 7.

1911 Hotchkiss, W. O., The general structure of the Florence iron district [Wisconsin; abs.]: Science, n.s. v. 33, no. 847, p. 464.

Writing about the Quinnesec schists of the Florence, Wis., and the Menominee, Mich., districts, Hotchkiss states, "Work of the past season shows that these schists overlie the iron-bearing series."

1911 Van Hise, C. R., and Leith, C. K., The geology of the Lake Superior region: U.S. Geol. Survey Mon. 52, 641 p.

Van Hise and Leith propose a threefold division of the Huronian rocks of the Menominee district, as shown in the comparison with W. S. Bayley's sequence (table 1). They state that the north area of Quinnesec schist, presumed to be of Archean age by Bayley (1904), is interbedded with the Michigamme slate and is therefore of Upper Huronian age. They also state that the south area of Quinnesec schist is probably of Upper Huronian rather than of Archean age because of observations made by Hotchkiss (1911) in Wisconsin, and, therefore, that the granite cutting the south Quinnesec schists is probably of Keweenawan age. Because the Hanbury slate was proved to be continuous with the Michigamme slate, they propose to use the latter name. Regarding several masses of iron-bearing rocks that were doubtfully referred to the Middle Huronian Negaunee by Bayley (1904), they state, "Subsequent work has demonstrated these to be Uupper Huronian."

To the Middle Huronian, not listed in Bayley's sequence, these authors allot quartzite, cherty quartz rocks, and conglomerate which were said to rest on the Randville dolomite but were not separated from it in mapping.

1911 Michigan Geological and Biological Survey, Mineral resources of Michigan with statistical tables of production and value of mineral products: Michigan Geol. and Biol. Survey Pubs, Geol. Ser., 1911, 1912, 1913, 1914, 1915, 1917, 1919, 1922, 1923, 1924.

Iron ore production figures and other statistical data.

1912 United States Geological Survey, annual volumes, Mineral resources of the United States, 1911–23: U.S. Geol. Survey.

1913 Leith, C. K., "Algonkian" vs. "pre-Cambrian": Econ. Geology, v. 8, no. 5, p. 507–508.

Discussion of the use of the terms "Algonkian" and "pre-Cambrian."

1914 Leith, C. K., Pre-Cambrian correlation from a Lake Superior standpoint: Internat. Geol. Cong., 12th, Toronto, Canada, 1913, Compte-rendu, p. 409–419.

1915 Allen, R. C., A revision of the correlation of the Huronian group of Michigan and the Lake Superior region, *in* Allen, R. C., and Barrett, L. P., Contributions to the geology of northern Michigan and Wisconsin: Michigan Geol. and Biol. Survey Pub. 18, Geol. Ser. 15, p. 21–32; also *in* Leith, C. K., and Allen, R. C., Discussion of correlation: Jour. Geology, v. 23, no. 8, p. 705–724.

1915 Allen, R. C.

Allen proposes a threefold division of the Huronian rocks in various areas of Michigan, including the Menominee district in southern Dickinson County. He correlates the Vulcan formation with the Negaunee formation of the Marquette district. Allen's proposed Huronian sections for the Menominee and Marquette districts (correlation table facing p. 30) are shown in table 2.

TABLE 1.—*Stratigraphic sequences for the Menominee district*

Van Hise and Leith (1911)		Bayley (1904)	
Cambro-Ordovician	Hermansville Limestone		
Cambrian system	Lake Superior Sandstone		Lake Superior Sandstone
Unconformity			Unconformity
Algonkian system:			
Keweenawan series	Granite(?)		
Huronian series: Upper Huronian (Animikie Group)	Quinnesec schist; Michigamme ("Hanbury") slate; Vulcan formation	Upper Huronian	Hanbury slate; Vulcan formation
Unconformity			Unconformity
Middle Huronian	Quartzite	Lower Huronian	Negaunee formation
Unconformity			
Lower Huronian	Randville dolomite; Sturgeon quartzite		Randville dolomite; Sturgeon quartzite
Unconformity			Unconformity
Archean system:			
Laurentian series	Granite and gneisses	Archean	Greenschist, granite, and gneiss
Keewatin series	Green schists		

TABLE 2.—*Stratigraphic sequences in the Menominee and Marquette districts*

[From Allen, 1915a]

		Marquette	Menominee
Upper Huronian		Greenschist intrusives and extrusives; Michigamme slate partly replaced by Clarksburg volcanics; Bijiki schist (iron bearing); Goodrich quartzite	
Middle Huronian (Animikie)		Negaunee iron formation; Siamo slate; Ajibic quartzite	Granite; Quinnesec schist; Hanbury slate; Vulcan iron formation
Lower Huronian		We-we slate; Kona dolomite; Mesnard quartzite	Quartzite; Randville dolomite; Sturgeon quartzite
Archean	Laurentian	Granite syenite; Peridotite; Palmer gneiss	Granite and gneiss
	Keewatin	Kitchi and Mona schist	Green schist

1915 Leith, C. K., Arguments for retaining the present correlation, *in* Leith, C. K., and Allen, R. C., Discussion of correlation: Jour. Geology, v. 23, no. 8, p. 724–729.

Leith stresses the inadvisability of correlating the middle Huronian Negaunee at Amasa, Mich., with the Vulcan at Crystal Falls, Mich., because of insufficient supporting evidence, and because lithologically and stratigraphically the Vulcan at Crystal Falls corrêlated better with iron-formation in the Upper Huronian of the Marquette district.

1916 Allen, R. C., Smith, R. A., and Barrett, L. P., Geologic map of Michigan: Michigan Geol. Survey Pub. 23, Geol. Ser., scale 1:750,000.

Map shows numerous changes based on fieldwork done by the personnel of the Michigan Geological Survey.

1919 Allen, R. C., Correlation of formation of Huronian group in Michigan: Am. Inst. Mining Metall. Engineers Mining and Metallurgy Bull. 153, p. 2579–2594; Trans., v. 63, p. 188–212, 1920.

Allen presents important new evidence of an unconformity between the Hanbury slate and the Vulcan iron-formation in the Menominee district of southern Dickinson County, Mich., and discusses the regional implications. He states (p. 201), "In section 13, T. 39, R. 29, and in sections 17 and 18, T. 39, R. 28 on the 'South Range,' east of the Sturgeon River, a large number of diamond drill holes penetrate the lower horizons of the Hanbury slate, the entire Vulcan group, and end in the Randville dolomite * * *. Half of them exhibit no evidence of unconformity between the Hanbury slate and the Vulcan group, a few show coarsely detrital beds at or very

TABLE 3.—*Stratigraphic sequence in Menominee district*

[From Allen, 1919]

Era	Sub-era	Period	Epoch	Stage	Description of formations
Proterozoic	Algonkian	Keweenawan			
		Epi-Huronian revolution, mountain building		Emergent interval	Granite
		Huronian	Upper Huronian	Quinnesec	Eruptive contact. Basic volcanic extrusives of great thickness, sills and dikes.
				Hanbury	Great slate series with beds of conglomerate, quartzite, graywacke, ferruginous chert and impure limestone. Thickness not known.
			Emergent interval		
			Middle Huronian	Loretto	Slate, 400 + ft. (122+ m).
				Curry	Iron-bearing member, 100 to 200 ft. (30 to 61 m).
				Brier	Ferruginous, siliceous, banded slate, 300 to 400 ft. (91 to 122 m).
				Traders	Conglomerate, quartzite and iron-bearing formation, 150 ft. (46 m).
			Emergent interval		
			Lower Huronian	Randville	Dolomite, cherty dolomite and quartzose and talcose facies, 1000 to 1500 ft (305 to 457 m).
				Sturgeon	Conglomerate, arkose, graywacke and quartzite. 1200 ft (366 m).
	Great Archeozoic interval				
Archeozoic	Laurentian				Granites and gneisses cut by dikes of granite and diabase.
	Keewatin				Eruptive contact. Green schists.

near the base of the Hanbury, and three show a well-developed Hanbury basal conglomerate lying on a slate which is conformable with the Curry member of the Vulcan group." The slate between the Curry member and the basal Hanbury conglomerate he calls Loretto slate because of its good development at the Loretto mine (p. 202). Allen's revised chronology of the Precambrian of the Menominee range is presented in table 3.

1920 Mining and Metallurgy, Correlation of formations of Huronian group in Michigan: Mining and Metallurgy, no. 157, sec. 12, p. 1–6.

Discussion of R. C. Allen's paper by W. O. Hotchkiss, E. F. Burchard, L. P. Barrett, Carl Zapffe, E. E. Siebenthal, E. C. Harder, and R. C. Allen.

1925 Royce, Stephen, Certain advances in geological information relative to the Lake Superior iron deposits: Lake Superior Mining Inst., 24th Ann. Mtg., Michigan and Ontario 1925, Proc., v. 24, p. 149–181.

Royce suggests that the Hanbury slate is an Upper Huronian formation equivalent in age to the Michigamme slate and that the associated iron-formations at Florence, Wis., and at Iron River and Crystal Falls, Mich., are therefore Upper Huronian. He states that these iron-formations may be equivalent to iron-formations in the Hanbury slate of the old Menominee district of southern Dickinson County.

1927 United States Bureau of Mines, 1927–34, annual volumes, Mineral resources of the United States, 1924–31: U.S. Dept. Interior, Bur. Mines.

1928 Bean, E. F., Geological map of Wisconsin (revised): Wisconsin Geol. and Nat. History Survey, scale 1:1,000,000.

1929 Leverett, Frank, Moraines and shore lines of the Lake Superior Basin: U.S. Geol. Survey Prof. Paper 154–A, p. 1–72.

Leverett outlines the glacial history of the Lake Superior region. On the general map, plate 1, the Menominee district is shown covered with drift and outwash gravels of glacial substage 4 of the Wisconsin stage.

1931 Lamey, C. A., Granite intrusions in the Huronian formations of northern Michigan: Jour. Geology, v. 39, no. 3, p. 288–295.

The author describes granite dikes that have intruded the Huronian formations. He states that the dikes resemble the porphyritic granite of the Southern complex (Marquette district), and that they may be related to that granite massif. "If this relationship is valid, some of the granite of the Southern complex is post-Huronian, is responsible for some of the metamorphism of the Huronian formations, and is probably related to the tectonic forces that caused folding in the area."

1933 Lamey, C. A., The intrusive relations of the Republic granite: Jour. Geology, v. 41, no. 5, p. 487–500.

A further description of the granite intrusives in Huronian formations is given. These post-Huronian intrusives are thought to be related to the porphyritic granite of the Southern complex. The latter Lamey named "Republic granite." The area of the Republic granite is extended southward to include the north granite and gneiss area of the Menominee district.

1933 United States Bureau of Mines, Minerals Yearbook 1932–33: U.S. Dept. Commerce, Bur. Mines, 819 p.

See note under U.S. Bur. Mines, 1934–48.

1934 Leith, C. K., The pre-Cambrian: Geol. Soc. America Proc. 1933, p. 151–180.

Leith summarizes the status of pre-Cambrian research and suggests many avenues for further study. He redefines the terms "Algonkian" and "Archean," the Algonkian as "that part of the pre-Cambrian sedimentary column in a particular region to which ordinary stratigraphic methods can be applied" and the Archean as "the underlying, more or less indivisible, basement complex, containing igneous or sedimentary rocks, or both, in which ordinary stratigraphic methods do not apply."

1934 United States Bureau of Mines, Annual volumes, Minerals Yearbooks, 1934–48: U.S. Bur. Mines.

Iron ore production figures and other statistical data.

1935 Leith, C. K., Lund, R. J., and Leith, Andrew, Pre-Cambrian rocks of the Lake Superior region; a review of newly discovered geologic features with a revised geologic map: U.S. Geol. Survey Prof. Paper 184, 34 p.

A new geologic map of the Lake Superior region is issued. Areas are discussed where the new map shows changes due to new geologic work or change of concept since the publication of Monograph 52 (Van Hise and Leith) in 1911. With two exceptions that are difficult to comprehend, the authors accept the stratigraphic column for the Menominee range as proposed by Allen (1915) and as modified by Allen (1919). (See table 4.)

TABLE 4.—*Comparison of stratigraphic sequences proposed by Allen (1919) and Leith and others (1935)*

	Allen (1919)	Leith and others (1935)
Upper Huronian	Quinnesec schist Hanbury slate	Michigamme slate (including iron-formation) South belt of Quinnesec greenstone [1]
Middle Huronian	Vulcan iron-formation	Vulcan iron-formation Greenstone (Lake Antoine region) [1]
Lower Huronian	Randville dolomite Sturgeon quartzite	Randville dolomite Sturgeon quartzite

[1] Doubtful stratigraphic position.

The exceptions involve the stratigraphic positions of the greenstone formations. The greenstone (Lake Antoine region), Bayley's north Quinnesec greenstone of Archean age (1904), is proved to be interbedded with the Upper Huronian Michigamme (Hanbury) slate, according to Van Hise and Leith (1911, p. 345). No new evidence is presented to support placing this greenstone formation between the Randville and Vulcan formations, where, as far as the present writers know, no greenstone is found.

The south belt of Quinnesec greenstone, although its stratigraphic position is uncertain, is apparently above the Michigamme (Hanbury) slate in the Menominee district, as indicated in the stratigraphic column by Allen (1919), and not below the Michigamme as indicated in the column by Leith, Lund, and Leith (1935).

1936 Dickey, R. M., The granitic sequence in the southern complex of Upper Michigan: Jour. Geology, v. 44, no. 3, p. 317–340.

Dickey contends that the bulk of the "Republic granite" as defined by Lamey (1932) is of Archean age. According to Dickey, Killarney granite of inconsiderable quantity cuts both the Archean granites and the Huronian formations. He suggests that the name "Republic granite" be restricted to the Killarney age granite.

1938 Dickey, R. M., The Ford River granite of the southern complex of Michigan: Jour. Geology, v. 46, no. 3, pt. 1, p. 321–335.

Dickey finds that the porphyritic granite of the southern complex (Republic granite of Lamey) contains quartzite inclusions believed to represent remnants of Lower Huronian formations and that the granite is overlain unconformably by Middle Huronian formations. He concludes that the granite is of post-Lower Huronian but of pre-Middle Huronian age. The granite (Lamey's Republic granite) is renamed "Ford River granite."

1938 Dickey, R. M., Present trends in studies of Michigan Huronian: Michigan Acad. Sci. Papers, 1937, v. 23, p. 419–426.

General discussion of problems related to Precambrian research in Michigan.

1938 Lake Superior Iron Ore Association, Lake Superior Iron ores: Cleveland, Ohio, Lake Superior Iron Ore Assoc., 364 p.

Mine directory and iron ore production statistics for Lake Superior mines, including mines of the Menominee range. A brief sketch of the Menominee geology is given by Stephen Royce.

1939 Dutton, C. E., and Lamey, C. A., Geology of the Menominee Range, Dickinson County [Michigan]: Michigan Dept. Conserv., Geol. Survey Div. Prog. Rept. 5, 10 p.

Dutton and Lamey report on the geology of that part of the south iron range between Norway and Quinnesec, Mich. An interpretive, planimetric, geologic map of the area is presented. The discussion emphasizes the importance of faulting as the determining factor in the structure of the area, and the probability that some of the rocks heretofore mapped as Hanbury slate are probably "footwall"— that is, rocks that underly the Vulcan iron-formation.

1940 Tyler, S. A., Marsden, R. W., Grout, F. F., and Thiel, G. A., Studies of the Lake Superior pre-Cambrian by accessory-mineral methods: Geol. Soc. America Bull., v. 51, no. 10, pt. 2, p. 1429–1537.

The results of investigations on accessory zircon in Precambrian granites of the Lake Superior region are reported.

In the pre-Huronian granite areas north of and contiguous with the Menominee district, the authors determine that there are two pre-Huronian granites—an older one characterized by hyacinth zircon, and a younger one characterized by malacon zircon. Granite dikes that cut the Huronian rocks also contain malacon zircon; therefore, it is not possible to distinguish some of the pre-Huronian granites from Huronian granites on the basis of zircon contained.

1941 Lamey, C. A., and Dutton, C. E., Geology of the Menominee range vicinity of Iron Mountain [Michigan]: Michigan Dept. Conserv., Geol. Survey Div. Prog. Rept. 6, 14 p.

A preliminary geologic map is presented for the area extending from Quinnesec west along the south iron range to the Menominee River.

Stratigraphy, structure, and exploration possibilities are discussed.

1941 Pettijohn, F. J., and Hildebrand, F. A., 1941, Archean-Huronian unconformity of the Menominee iron range, Michigan [abs.]: Geol. Soc. American Bull., v. 52, no. 12, pt. 2, p. 1927.

"The pre-Lower Huronian age of the granite and granite gneiss complex northeast of the Menominee iron range is reaffirmed by discovery of a hitherto undescribed exposure of basal conglomerate near Fern Creek in Dickinson County, Michigan. * * *. The conglomerate lies between the granite gneiss and the Sturgeon quartzite."

1942 Dutton, C. E., Economic geology of a part of the Menominee range. Dickinson County [Michigan]: Michigan Dept. Conserv., Geol. Survey Div. Prog. Rept. 9, 27 p.

Several areas believed by the writer to warrant exploration for iron ore are described. Reference is made to beneficiation tests run on some specimens of iron-formation from the Menominee range.

1942 Lamey, C. A., and Dutton, C. E., Geology of Menominee Range, Norway to Waucedah [Michigan]: Michigan Dept. Conserv., Geol. Div. Survey Prog. 8, 20 p

A preliminary geologic map is presented of the area between Norway and Waucedah, Mich. Details of structure and stratigraphy are discussed. A major fault is said to separate the north and south iron ranges.

1943 Pettijohn, F. J., Basal Huronian conglomerates of the Menominee and Calumet districts, Michigan: Jour. Geology, v. 51, no. 6, p. 387–397.

The author describes an unconformable relationship between the Fern Creek formation and the pre-Huronian gneiss complex in Dickinson County, Mich.

1947 Higgins, J. W., Structural petrology of the Pine Creek area, Dickinson County, Michigan: Jour. Geology, v. 55, no. 6, p. 476–489.

Higgins describes the results of petrofabric investigation done on the belt of Sturgeon quartzite and adjacent gneiss that extends from sec. 20, T. 41 N., R. 29 W. (in central Dickinson County), south and southeast for a distance of 15 miles.

1948 Trow, J. W., The sturgeon quartzite of the Menominee district, Michigan: Chicago Univ. Ph. D. thesis, 60 p.

Includes outcrop map of quartzite belt and also detailed maps of selected areas.

1949 Richardson, E. J., Jr., Some Lower Huronian stromatolites of northern Michigan: Chicago Nat. History Mus., Fieldiana—Geology, v. 10, no. 8, p. 47–62.

A description and a classification are given of algal structures in the Lower Huronian dolomites of Michigan.

1950 Dutton, C. E., Progress of geologic work in Iron and Dickinson Counties, Michigan: U.S. Geol. Survey Circ. 84, 7 p.

Dutton gives a brief review of current work of the U.S. Geological Survey in the region.

1952 Prinz, W. C., The geology of a portion of the Quinnesec igneous complex, south of Iron Mountain, Michigan: Columbus, Ohio State Univ. M.S. thesis, 90 p., 19 pls.

1952 Shapiro, Norman, The geology of a part of the [Precambrian] Quinnesec greenstone complex, Dickinson County, Michigan: Columbus, Ohio State Univ. M.S. thesis, 78 p., 17 pls.

1953 Froelich, A. J., The geology of a part of the Wisconsin granite-Quinnesec greenstone complex, Florence County, Wisconsin: Columbus, Ohio State Univ. M.S. thesis, 120 p., 22 pls.

1953 Wier, K. L., Balsley, J. R., Jr., and Pratt, W. P. Aeromagnetic survey in part of Dickinson County, Michigan, with preliminary geologic interpretation: U. S. Geol. Survey Geophys. Inv. Map GP–115.

Dickinson County was surveyed with an airborne magnetometer, and traverses were flown at 500 feet above the ground along north-south lines one-third of a mile apart. The magnetic profiles and a map showing the crests of magnetic anomalies accompany the report.

1955 James, H. L., Zones of regional metamorphism in the Precambrian of northern Michigan: Geol. Soc. America Bull., v. 66, no. 12, pt. 1, p. 1455–1487.

Zones of regional metamorphism are delineated for northern Michigan, including the Menominee district. The isograd map shows the west half of the Menominee district of Dickinson County in the biotite zone of regional metamorphism, and the east half in the chlorite zone. Most of the granite area to the north of the district is shown in the garnet zone, as in the granite area to the south of the district.

1955 Thompson, G. L., A study of part of the Pre-Cambrian granite-greenstone complex in southeastern Florence County and northwestern Marinette County, Wisconsin: Columbus, Ohio State Univ. M.S. thesis, 86 p., 8 pls.

1957 Fulweiler, R. E., The geology of a part of the igneous and metamorphic complex of southeastern Florence County, Wisconsin: Columbus, Ohio State Univ. M.S. thesis, 68 p., 14 pls.

1958 James, H. L., Stratigraphy of pre-Keweenawan rocks in parts of northern Michigan: U.S. Geol. Survey Prof. Paper 314–C, p. 27–44.

A revised stratigraphic column is presented for the pre-Keweenawan rocks of Iron and Dickinson Counties, Mich. The Precambrian rocks are placed in three categories: lower Precambrian, middle Precambrian, and upper Precambrian. A sequence of lower Precambrian rocks—Archean of older reports—is established on the basis of detailed mapping in central Dickinson County. Middle Precambrian rocks are said to consist of the Animikie series—"Huronian" of older reports—and post-Animikie but pre-Keweenawan igneous rocks. The term "Huronian," used in the district for nearly 60 years, is not used because of the uncertainty of correlation with the type Huronian of Canada. The revised stratigraphic column follows (table 5).

1959 Prinz, W. C., Geology of the southern part of the Menominee district, Michigan and Wisconsin: U. S. Geol. Survey open-file report, 221 p.

TABLE 5.—*Lithologic sequence of Precambrian rocks in Iron and Dickinson Counties, Michigan*

[From James, 1958]

[New stratigraphic names are indicated by asterisks]

Precambrian	Upper Precambrian	Keweenawan series	Diabase dikes and sills (probable age about 1,100 million years)			
		—Intrusive contact—				
	Middle Precambrian	Granitic intrusive rocks (probable age at least 1,400 million years)				
		—Intrusive contact—				
		Metadiabase and metagabbro				
		—Intrusive contact—				
		Animikie series	Paint River group*	Fortune Lakes slate*		
				Stambaugh formation*		
				Hiawatha graywacke*		
				Riverton iron-formation*		
				Dunn Creek slate*, with Wauseca pyritic member*		
			Baraga group*	Badwater greenstone*		
				Michigamme slate		
				Fence River formation	Amasa formation	
				Hemlock formation, with Mansfield iron-bearing slate member and Bird iron-bearing slate member*		
				Goodrich quartzite		
			—Unconformity—			
			Menominee group*	Vulcan iron-formation	Loretto slate member	
					Curry iron-bearing member	
					Brier slate member	
					Traders iron-bearing member	
				Felch formation		
			—Unconformity—			
			Chocolay group*	Randville dolomite	Saunders formation	
				Sturgeon quartzite		
				Fern Creek formation		
		—Unconformity—				
	Lower Precambrian	Gneissic granite and other crystalline rocks			Quinnesec formation (position uncertain)	Margeson Creek gneiss (position uncertain)
		—Intrusive or replacement contact??—				
		Dickinson group*	Six-Mile Lake amphibolite*	Hardwood gneiss* (position uncertain)		
			Solberg schist*, with Skunk Creek member*			
			East Branch arkose*			
		—Unconformity—				
		Granite gneiss				
		Quartzite and schist (small bodies included in granite gneiss)				

1961 James, H. L., Clark, L. D., Lamey, C. A., and Pettijohn, F. J., in collaboration with Freedman, Jacob, Trow, J. W., and Wier, K. L., Geology of central Dickinson County, Michigan: U.S. Geol. Survey Prof. Paper 310, 176 p.

This comprehensive report treats all aspects of the Precambrian formations of central Dickinson County. The general geologic map (pl. 2) covers an area of about 273 square miles which includes the Felch Mountain and Calumet troughs. The south boundary of this map corresponds with the north boundary of the several geologic maps of the southern Dickinson County area included in the present report.

FIELDWORK AND REPORT RESPONSIBILITY

The geologic fieldwork leading to this report was done by several geologists mainly in the 6-year period 1951–56, but partly over almost 20 years. The geologic and unpublished magnetic maps of the south iron range that were made by C. A. Lamey and C. E. Dutton in the period 1937–42 were fully consulted during the new mapping. Carl A. Lamey did geologic and magnetic surveying on the north iron range during the seasons 1951–53, assisted for individual seasons by Ronald A. Janc, Robert F. Novotny, and William C. Prinz. Prinz mapped the area of the southern complex in the period 1954–56. Samuel B. Treves mapped the area of the northern complex in the field seasons 1955–57. The outcrop area of the Sturgeon Quartzite, originally mapped by James W. Trow for a Ph. D. dissertation project, was plotted on the topographic base and field checked and adjusted by Prinz. The magnetometer surveys are by Prinz and Jacob E. Gair. Dip-needle surveys are by C. E. Dutton and C. A. Lamey (1937–42) and by Lamey (1951–53). Reexamination and mapping on the south iron range was commenced by R. W. Bayley and C. E. Dutton in the fall of 1955 and continued by Bayley in the summer and fall of 1956.

The map compilations and the writing of the report was chiefly the responsibility of R. W. Bayley with considerable assistance by C. A. Lamey and C. E. Dutton. The section of the report dealing with the geology of the north complex was done by S. B. Treves.

ACKNOWLEDGMENTS

This study has been a cooperative project between the U.S. Geological Survey and the Michigan Department of Conservation; the cooperation and active interest of the personnel of the Michigan Department of Conservation are gratefully acknowledged. We also acknowledge the assistance of the many iron mining companies who permitted use of and expedited acquisition of data concerning the abandoned mines and diamond-drill explorations of the district. The officers and employees of Picklands, Mather and Co. and the Oliver Iron Mining Division, United States Steel Corp., and Victor E. Kral of the Ford Motor Co. were particularly helpful.

GENERAL GEOLOGY

REGIONAL GEOLOGIC SETTING

The Menominee iron-bearing district as defined by W. S. Bayley (1904) is the southernmost of a series of similar districts of Michigan and adjacent Wisconsin (fig. 1) and is included within the Iron Mountain, Norway, and Vulcan quadrangles (pls. 1–3).

The district includes the southernmost of a series of nearly parallel, westward striking structural troughs that occur near the east margin of the Canadian Shield in northern Michigan. In general the troughs are infolded or downfolded areas of metasedimentary and metavolcanic rocks of middle Precambrian age which are separated by large islandlike masses of granitic rock and metasedimentary and metavolcanic rock of lower Precambrian age. Considerable geologic similarity exists between these major trough areas, and partly equivalent stratigraphic successions in each indicates preorogenic continuity of some of the formations. Each of the major troughs has a history of iron ore production from a major medial Animikie iron-formation. The very productive Marquette trough, 42 miles to the north, and the Republic trough, a few miles to the south, were described by Van Hise and Bayley and Smyth (1897) and by Van Hise and Leith (1911). The Felch trough, 12 miles north, was described by Clements and Smyth (1899), and the Calumet through, 9 miles to the north, was described by Van Hise and Leith (1911). A recent study of the geology of central Dickinson County by James and others (1961) includes descriptions of the Felch and the Calumet troughs.

Each of the trough areas mentioned above opens out to the west into a broad area of younger middle Precambrian rocks. The large triangular synclinorium embracing the Iron River-Crystal Falls districts, Michigan, and the Florence district, Wisconsin (fig. 1), is a northwest extension of the general Menominee trough structure. The geology of the Iron River-Crystal Falls district has been intensively investigated by the U.S. Geological Survey, in cooperation with the Geological Survey Division of the Michigan Department of Conservation, since 1943, and a geological map of the district has been published (James and others, 1959). Other published reports which deal with certain aspects of the geology of the district or with the geology of areas within the district are listed chronologically as follows: Dutton, Park, and Balsley (1945); Pettijohn and Clark (1946); Pettijohn (1947); James, Clark, and Smith (1947); Pettijohn (1948); James and Wier (1948); Balsley, James, and Wier (1949); Dutton (1949); James (1951); James and Dutton (1951); Pettijohn (1951, 1952). A geologic map and a brief description of the rocks of the Florence, Wis., district are given by Van Hise and Leith, (1911), and details of an unpublished investigation in the area by W. O. Hotchkiss (1910), of the Wisconsin Geological and Natural History Survey. A resurvey of the Florence area was begun by C. E. Dutton in 1955 under the joint auspices of the U.S. Geological Survey and the State of Wisconsin.

GENERAL STRATIGRAPHY

Precambrian rocks form the bedrock in most of the district. They are unconformably overlain by outliers of Cambrian and Ordovician sandstone and dolomite in some areas, and all rocks are concealed in part by unconsolidated glacial deposits of Pleistocene age in most of the area. Only the Precambrian rocks were investigated in this report. The unconsolidated glacial deposits were omitted from the maps in order to show the Precambrian bedrock geology; the lower Paleozoic rocks were also omitted, except for a transparent overprint to show the areal extent of those rocks.

The major divisions of the Precambrian rocks used in this report are those recommended by the Committee on Geological Names of the U.S. Geological Survey. The divisions are informal and are not intended to have time significance. They are:

Upper Precambrian = Keweenawan and other provincial series.

Middle Precambrian = Huronian, Animikie, and other provincial series.

Lower Precambrian = provincial series.

The lower Precambrian rocks as used herein are equivalent to the Archean rocks of previous reports, and the middle Precambrian rocks are equivalent to the Huronian rocks of previous reports. Diabase dikes are the only upper Precambrian (Keweenawan) rocks in the district.

In a recent discussion of the pre-Keweenawan stratigraphy of Iron and Dickinson Counties, Mich., James (1958) applied local group names to the rock sequences known previously as Lower, Middle, and Upper Huronian. In the present report these group names are used: Chocolay Group ("lower Huronian"), Menominee Group ("middle Huronian"), and Baraga and Paint River Groups ("Upper Huronian"). To the combined four groups James applied the previously established name "Animikie Series."

The sequence of Precambrian formations in the district is shown on table 6 below.

LOWER PRECAMBRIAN ROCKS

Rocks of early Precambrian age are exposed on both the northeast and south flanks of the Menominee trough proper. The northeast area (pls. 2, 3) is underlain chiefly by granitic gneiss—the Carney Lake Gneiss—which lies unconformably below the Animikie Series of the trough. The area flanking the trough on the south (pls. 1, 2) is underlain mainly by metavolcanic rocks of the Quinnesec Formation, and these are separated from the younger Animikie strata by a major east-west fault.

SOUTHERN AREA

GENERAL CHARACTERISTICS

The north part of the southern area is underlain by altered volcanic and sedimentary rocks which were designated "Quinnesec schist" by W. S. Bayley (1904). More recently these rocks have been designated "Quinnesec formation" (James, 1958, p. 32). The Quinnesec Formation is contained in a narrow northwest-trending belt which lies between the late Animikie Michigamme Slate to the north and the "Wisconsin batholith" to the south. The contact between the Quinnesec and Michigamme Formations is not exposed. The

TABLE 6.—*Sequence of Precambrian formations in the Menominee district, Michigan*

			Unconformity		
Precambrian	Upper Precambrian	Keweenawan Series	Fresh diabasic dikes		
			Unconformity		
	Middle Precambrian	Instrusive rocks	Pegmatite, aplite, and quartz dikes and veins Hoskin Lake Granite Marinette Quartz Diorite Metagabbro dikes and sills		
		Animikie Series	Paint River Group, not represented		
			Baraga Group	Badwater Greenstone Michigamme Slate	
				Unconformity	
			Menominee Group	Vulcan Iron-Formation	Loretto Slate Member Curry Iron-bearing Member Brier Slate Member Traders Iron-Bearing Member
				Felch Formation	
				Unconformity	
			Chocolay Group	Randville Dolomite Sturgeon Quartzite Fern Creek Formation	
				Unconformity	
	Lower Precambrian			*North area—Michigan* Sparse dikes and veins of syenite, granodiorite, aplite, and pegmatite Carney Lake Gneiss	*South area—Michigan-Wisconsin* Quinnesec Formation

line between the two formations on the maps separates sparse outcrops of Michigamme-type slate on the north and greenstone thought to belong to the Quinnesec on the south. The actual contact between the formations may be a thousand feet or more from the place indicated on the maps.

The formation is cut by the plutonic Marinette Quartz Diorite and the Hoskin Lake Granite (Prinz, 1965), by dikes and sills of metagabbro, by dikes of fresh diabase and of granite, and by abundant quartz and quartz-tourmaline veins. The dikes of granite may be related to the Hoskin Lake Granite. All the intrusive rocks appear to us to be of late or post-Animikie age.

QUINNESEC FORMATION

PREVIOUS INVESTIGATIONS

Most of the early references to the Quinnesec Formation and other rocks of the southern area were summarized by W. S. Bayley (1904). Descriptions of the Quinnesec rocks made by George H. Williams (1890) are of outstanding quality. Williams recognized the volcanic origin of the rocks, their metamorphosed condition, and their relation to the younger intrusives. More recent investigations of the rocks of the area were made by W. C. Prinz, Norman C. Shapiro, A. J. Froelich, G. L. Thompson, R. E. Fulweiler, and E. J. Lyons. Prinz and Shapiro, in 1951, mapped two small areas of Quinnesec and associated rocks in Michigan, whereas Froelich, Thompson, and Fulweiler, in 1952, mapped three small areas in Wisconsin. The results of these five studies are contained in theses for Master of Science degrees on file at the Ohio State University. They contain maps and petrographic descriptions, but, because of the limited sizes of the areas studied, no general conclusions were reached. Lyons did more extensive work in the vicinity of Niagara, Wis. He postulated (in Emmons and others, 1953, p. 107–110) that the intrusion of the Wisconsin granitic batholith into the Quinnesec basalts produced the porphyritic Hoskin Lake Granite as the end product of the granitization of the basalt. Stages of alteration between basalt and granite were believed to be represented by "plagioclase hornblendite," which corresponds to part of the late Animikie intrusive metagabbro and part of the metabasalt of this report, and "biotite-hornblende-quartz diorite," which corresponds to the intrusive Marinette Quartz Diorite of this report.

Prinz (1959) did the geologic mapping in the area underlain by the Quinnesec Formation and the younger intrusives, shown on plates 1 and 2 of the present report, and most of the descriptive material that follows has been summarized from his report.

GENERAL DISTRIBUTION

The area underlain by the Quinnesec Formation is shown on plates 1 and 2, and outcrop areas are indicated. The rocks are well exposed south and southeast of Aurora, Wis.; in the south part of sec. 7, T. 39 N., R. 30 W. (Mich.), in sec. 7, T. 38 N., R. 20 E. (Wis.), and in sec. 12, T. 38 N., R. 19 E. (Wis.); south, southwest, and west of Niagara, Wis.; at Pier Gorge (sec. 24, T. 39 N., R. 30 W., Mich.); and in secs. 19 and 20, T. 38 N., R. 21 E. (Wis.).

DESCRIPTION AND SPECIFIC DISTRIBUTION

The Quinnesec Formation consists mainly of dark-colored mafic and light-colored felsic volcanic flow and pyroclastic rocks which have been altered to greenstone, chloritic schist, hornblende schist, various mica schists, amphibolite, and, locally, hornblende hornfels. Metasedimentary rocks, now schist and metamorphosed iron-formation, form minor intercalations in the metavolcanic rocks. Original structures and textures are well enough preserved in many rocks to permit accurate classification as to origin, but they have been destroyed in others by deformation and recrystallization.

The formation consists of two distinct parts, a north belt comprising the rocks north of the northernmost meander of the Menominee River composed of felsic metavolcanic rocks and sparse metasedimentary rocks interbedded with greenstone and related schist, and a south belt composed largely of greenstone and related schist. The belt containing felsic rocks is limited to the north part of the Quinnesec Formation. It extends for 11 miles beyond the west border of the mapped area, but the eastern limit of the belt is unknown. The southern belt of greenstone broadens both to the east and west of the mapped area, and underlies a very extensive area in the north part of Wisconsin.

All the Quinnesec rocks are metamorphosed to some degree. In general the metamorphic grade, as indicated by the mineralogy of the rocks, increases southward toward the Hoskin Lake Granite, which has intruded the Quinnesec strata. Inclusions of Quinnesec rock in the granite are most altered; thus, it seems apparent that the granite was the source of the heat that caused the increase in the grade of metamorphism. This problem is discussed more fully on page 77.

Prinz (1959) also regarded the metamorphism as post-Animikie, but much later than the intrusion of the Hoskin Lake Granite which he thought to be of pre-Animikie age. Age determinations not available at the time of Prinz's mapping indicate that the granite is probably also post-Animikie (see p. 75).

The rocks of the Quinnesec will be described under three main headings: (1) basaltic metavolcanic rocks, (2) felsic metavolcanic rocks, and (3) metasedimentary rocks.

BASALTIC METAVOLCANIC ROCKS

The basaltic metavolcanic rocks comprise most of the Quinnesec strata which lie south of the Menominee River, as well as numerous intercalations in the belt of felsic rocks to the north. These mafic rocks, even though they comprise an assemblage of closely related types, have acquired differences in mineralogy and fabric because of differences in metamorphism and deformation; hence, they will be described under two broad subdivisions: (1) greenstone and related schist, which represent low-grade or greenschist-facies metamorphism, and (2) hornblende schist and hornfels, which represent middle-grade or amphibolite-facies metamorphism. The areas underlain by these two main groups are separated on the maps by the line between the greenschist facies and the oligoclase-amphibolite facies.

Greenstone and related schist (green schist)

The term "greenstone" is here used in the ordinary sense and applied to somewhat altered nonschistose mafic flow and pyroclastic rocks, the color being caused by an abundance of chlorite, amphibole, and epidote. The schists associated with the somewhat massive greenstone include foliated greenstone (green schist) and chlorite schist, the distinction between the two being made on the presence of abundant chlorite, epidote, and actinolite in the green schist, and chiefly chlorite in the chlorite schist.

The unweathered greenstone is green or greenish-gray, but weathered surfaces are brown. It is generally massive or slightly foliated, and aphanitic; locally it is porphyritic or fragmental. Relict igneous textures are discernible in most of the massive greenstone, although the primary minerals have been replaced. The textures are microdiabasic for the most part, but some are porphyritic. Primary volcanic structures, such as pillows and amygdules, shown on a few exposures indicate that the rocks are predominantly ancient lava flows, some of which probably are of marine origin. Individual flows are thick; bedding planes are commonly obscured by shearing, and attitudes are difficult to obtain. The most reliable determinations indicate that the flows are vertical or that they dip steeply south. The fragmental rocks, which are tuffs and volcanic breccias, make up only a small part of the formation; they are best exposed on the north bank of the Menominee River at Niagara, Wis.

The greenstone has been saussuritized or chloritized.

TABLE 7.—*Modes of typical greenstones*

[From Prinz, 1959, p. 37]

Mineral	1	2	3	4	5	6	7	8
Albite	25	15	10	40	45	50	45	35
Actinolite	45	40	35	30	15	10	0	30
Epidote minerals	25	40	20	15	Tr.	0	0	0
Chlorite	Tr.	Tr.	25	10	25	35	40	5
Calcite	Tr.	0	0	Tr.	Tr.	Tr.	5	20
Other	5	5	10	5	15	5	10	10

1. SW¼NE¼ sec. 21, T. 38 N., R. 21 E. (Wis.).
2. SE¼SE¼ sec. 14, T. 39 N., R. 30 W. (Mich.).
3. NE¼SE¼ sec. 15, R. 38 N., R. 20 E. (Wis.).
4. SW¼NW¼ sec. 24, T. 39 N, R. 30 W. (Mich.).
5. NW¼NW¼ sec. 20, T. 38 N., R. 21 E. (Wis.).
6. SW¼NW¼ sec. 24, T. 39 N., R. 30 W. (Mich.).
7. NW¼SE¼ sec. 14, T. 38 N., R. 20 E. (Wis.).
8. SW¼NW¼ sec. 24, T. 39 N., R. 30 W. (Mich.).

The common assemblage of alteration minerals includes actinolitic amphibole, chlorite, albite, quartz, epidote, and accessory ore minerals. Although none of the primary minerals persist, the textural relations indicate that actinolite has replaced a pyroxene and that albite has replaced a more calcic plagioclase. The mineral composition, relict textures, and general appearance of the rocks indicate they are metabasalts.

Modes of typical greenstone specimens as determined by Prinz are given in table 7.

Well-foliated chlorite schist occurs in most of the areas of greenstone. Quite generally it has formed in zones where the greenstone has been strongly sheared, especially on the margins of massive flows. An outcrop in the SW¼ of sec. 24, T. 39 N., R. 30 W. (Mich.), shows massive actinolitic greenstone which grades into well-foliated chlorite schist in which chlorite is the main mafic mineral. Williams (1890) observed the above relations at many places in the district.

The chloritic schist is well foliated, very fine grained, and about the same color as the greenstone. Chlorite, commonly in well-alined small plates, is the main constituent; minor and accessory constituents are muscovite, biotite, quartz, calcite, albite, leucoxene, and pyrite. Calcite is present in all of the chlorite schist, and is abundant in some, especially in that along the Menominee River east of Niagara.

Hornblende schist and hornfels

Altered basaltic volcanic rocks, similar to the greenstone in origin but more intensely metamorphosed, make up a belt 2,000–5,000 feet wide that flanks the Hoskin Lake Granite (pls. 1 and 2). They comprise hornblende schist, hornfelsic amphibolite, and minor hornblende-pyroxene hornfels. Hornblende schist showing vertical or steeply south-dipping foliation is most abundant. Hornfelsic amphibolite inclusions are common in parts of the granite, and the largest of these, in sec. 18, T. 38 N., R. 20 E. (pl. 1), shows distinct pillow structures. We have divided this group of rocks further, from mineralogical considerations,

TABLE 8.—*Modes of oligoclase-amphibolites*
[From Prinz, 1959, p. 43]

Mineral	1	2	3	4	5	6	7
Hornblende	50	35	40	40	45	30	30
Oligoclase-quartz	10	30	35	40	30	40	45
Epidote	40	25	25	20	15	12.5	Tr.
Biotite	0	10	0	Tr.	10	5	0
Chlorite	0	0	0	Tr.	Tr.	12.5	15
Calcite	Tr.	0	Tr.	Tr.	0	Tr.	5
Sphene-leucoxene	0	Tr.	Tr.	0	Tr.	Tr.	5
Anorthite in plagioclase	------	27	26	------	26	16	26

1. NE¼NW¼ sec. 16, T. 38 N., R. 20 E. (Wis.).
2. SE¼SW¼ sec. 7, T. 39 N., R. 30 W. (Mich.).
3. NW¼SW¼ sec. 16, T. 38 N., R. 20 E. (Wis.).
4. NW¼NE¼ sec. 8, T. 39 N., R. 30 W. (Mich.).
5. NW¼NE¼ sec. 8, T. 39 N., R. 30 W. (Mich.).
6. SW¼SE¼ sec. 7, T. 38 N., R. 20 E. (Wis.).
7. SW¼NW¼ sec. 3, T. 38 N., R. 19 E. (Wis.).

into two metamorphic facies[1]: (1) oligoclase amphibolite, characterized by the assemblage hornblende-oligoclase-quartz-epidote, with or without chlorite or biotite, and (2) andesine amphibolite, characterized by the assemblages hornblende-andesine-quartz or hornblende-andesine-diopside-quartz. Representative rocks of these facies fall into parallel belts on the maps and indicate a rise in metamorphic grade southward toward the Hoskin Lake Granite. The diopside-bearing rocks, which are represented by inclusions in the granite only, appear to be similar to rocks classed as hornblende hornfels by Fyfe, Turner, and Verhoogen (1958, p. 205–210, especially p. 209, basic assemblages).

The oligoclase-amphibolite is massive to well foliated, greenish gray to gray to almost black, and generally fine grained. Blue-green hornblende is the dominant mineral of these rocks. The hornblende is commonly alined, but in some specimens relict igneous textures persist. Equant crystalloblastic oligoclase and quartz make up 10–45 percent of the rock. Epidote is present in all specimens and is abundant in some. Accessory minerals are biotite, chlorite, calcite, sphene, and leucoxene. The modes of seven specimens of typical oligoclase-amphibolite, as determined by Prinz, are shown in Table 8.

The andesine-amphibolites make up the southernmost part of the Quinnesec Formation, the part closest to the Hoskin Lake Granite. They are similar to the oligoclase-amphibolites in general appearance, but are somewhat coarser grained and darker. Most of the rocks are hornblende schists which, close to the granite front, are cut by abundant granitic dikes, but some, mainly inclusions in the granite, are hornblende or hornblende-pyroxene hornfels.

Blue-green to brownish-green hornblende makes up 35–55 percent of these rocks, and andesine and quartz 40–60 percent. Minor alteration and accessory minerals include epidote, chlorite, sericite, magnetite, pyrite, sphene, biotite, zircon, hematite, calcite, apatite, tourmaline, and garnet. The mode of a typical specimen of andesine amphibolite, as determined by Prinz, is given in table 9.

The pyroxene-bearing hornfelses are restricted to a few inclusions of Quinnesec volcanic rock in the Hoskin Lake Granite in the NW¼ SW¼ sec. 16, T. 38 N., R. 20 E. (Wis.). One specimen examined consists mainly of green hornblende, andesine, colorless diopsidic pyroxene, and traces of epidote, sericite, sphene, pyrite, magnetite, and calcite. A second specimen shows andesine porphyroblasts 2.5 mm long set in a fine-grained mosaic of pale-green diopside, andesine, quartz, and green hornblende.

TABLE 9.—*Mode of andesine amphibolite*
[From Prinz, 1959, p. 51; specimen from SE¼NE¼ sec. 11, T. 38 N., R 19 E. (Wis.)]

Mineral	*Percent*	*Mineral*	*Percent*
Hornblende	51. 7	Biotite	0. 4
Quartz-andesine	43. 2	Apatite	. 2
Chlorite	2. 9	Tourmaline	Tr.
Magnetite	1. 6		

FELSIC METAVOLCANIC ROCKS

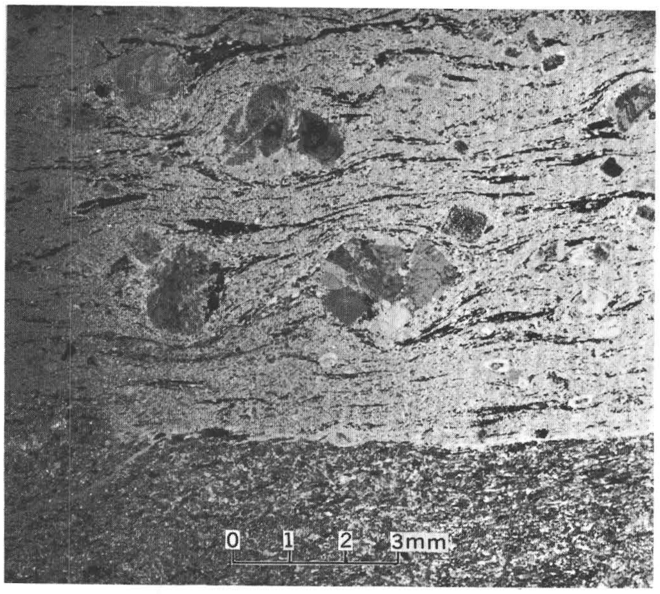
The felsic metavolcanic rocks of the Quinnesec Formation are confined wholly to the north part of the outcrop belt. Most of the exposures are along the banks of the Menominee River southeast of Niagara, northwest of the Big Quinnesec Dam, and northeast and west of the railroad bridge in sec. 7, T. 39 N., R. 30 W. Other exposures are in a roadcut north of Aurora, on a small hill southeast of Aurora, and at many places where these rocks occur as inclusions in younger metagabbro bodies. The felsic rocks occur in beds of varying thickness, interbedded and infolded with layers of greenstone and sparse beds of metasedimentary rock.

The fresh rock is generally gray, but weathered surfaces are variously colored; most of it is fine grained, more or less porphyritic, and exceptionally well foliated. The rocks in the south part of the belt are mylonitic, the foliation caused by granulation and smearing of quartz and feldspar grains, whereas the rocks to the north and east are schistose, the foliation caused by the formation of abundant metamorphic muscovite and chlorite. In general, the more southerly rocks are somewhat more metamorphosed than are those to the north and east.

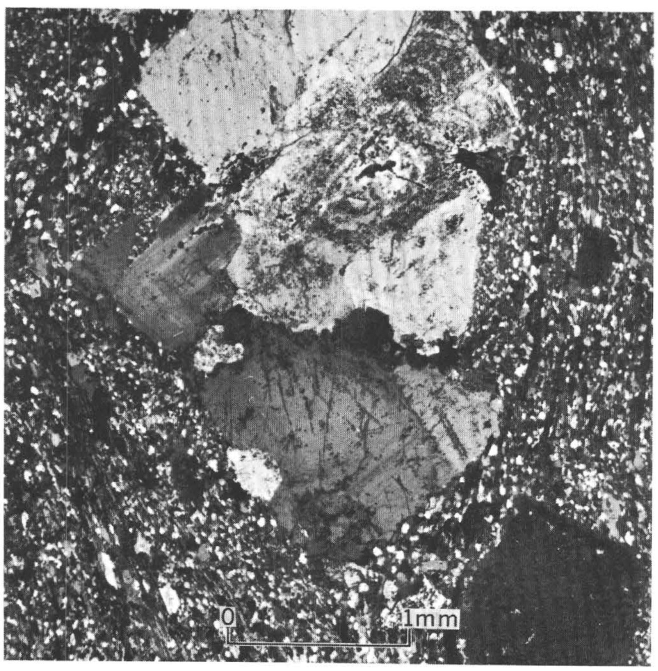
The felsic rocks are exposed along the banks of the river west of the railroad bridge in sec. 7, T. 39 N., R. 30 W., and at Big Quinnesec Falls were described in considerable detail by Williams (1890, p. 110–123). Williams regarded them as intrusive layers into the greenstone and classified them, with respect to their

[1] These two metamorphic facies are local facies which could be considered subfacies of the almandine-amphibolite facies of Fyfe, Turner, and Verhoogen (1958). They have been substituted for Prinz' (1959) epidote-amphibolite and amphibolite facies which, as applied, seemed less precise to us.

structures, as granite porphyry, augen gneiss, gneiss, felsite, and schistose porphyry. Prinz (1959) interpreted the schistose porphyry at Big Quinnesec Falls to be metarhyolite lava and classed the remaining rocks as gneissic porphyritic granite.



A



B

FIGURE 2.—*A*, Photomicograph of mylonitized "granite" showing augen of feldspar and quartz in a gneissic groundmass of feldspar, quartz, and biotite (dark streaks). Crossed nicols; × 7. *B*, Enlarged part of *A* showing glomeroporphyritic habit of plagioclase in augen. Crossed nicols; × 23.

The gneissic and porphyritic felsic rocks are composed chiefly of quartz, plagioclase, and biotite. Sericite, muscovite, and epidote are present, in most of the rocks as alteration products. Potassic feldspar is present in some as small phenocrysts, as rims of plagioclase crystals, and as small smeared-out clots in the groundmass. This feldspar is not easily determined in thin sections, but is clearly shown by X-ray diffraction patterns and on rock slices stained with sodium cobaltic nitrite. The accessory minerals are zircon, apatite, allanite, tourmaline, leucoxene, pyrite, and magnetite.

Quartz and plagioclase form phenocrysts and most of the fine-grained groundmass. The phenocrysts are distorted and granulated, and form augen around which the groundmass bends (fig. 2). In most of these rocks what appear to be single phenocrysts of plagioclase in hand specimen resolve into multicrystal clots when observed under the microscope (fig. 2*B*). The plagioclase crystals commonly show oscillatory zoning and albite twinning. Corroded anhedral phenocrysts of quartz as much as 3 mm in diameter occur in some specimens of the gneissic rock. Other specimens contain streaks, ribbons, and augen of recrystallized quartz.

Biotite, commonly partly altered to chlorite, forms segregated thin discontinuous wavy layers which serve to emphasize the gneissic or schistose character of the rocks.

A chemical analysis of a specimen of the above-described gneissic porphyry from sec. 7, T. 39 N., R. 30 W., is given in table 10. The analysis resembles most closely Nockolds' average analysis for rhyodacite-obsidian, which is given for comparison. Nockolds' average analysis (1954, p. 1014–1015) of muscovite-biotite granodiorite and muscovite-biotite tonalite also match the analysis of the gneissic porphyry fairly well.

In the felsic rocks that occur in the north part of the belt, particularly at Big Quinnesec Dam and down the river east of Niagara, the plagioclase phenocrysts are albite, the groundmass feldspar has been largely converted to muscovite, and the biotite to chlorite. In some of the rocks all the feldspar is gone, and the rocks are quartz-muscovite-chlorite schist, some of which contains glassy eyes of corroded and embayed quartz (fig. 3). Secondary calcite is a ubiquitous constituent of the schists and locally is abundant.

Williams (1890, p. 119) referred to this general group of rocks as schistose porphyry, and according to him (p. 121): "The microscope discloses in all thin sections of these rocks the typical structure of a quartz porphyry modified, however, by the action of a great pressure." The analyses and norms (table 10) show the virtual identity of Williams' schistose porphyry with

TABLE 10.—*Chemical analyses and norms of gneissic and schistose porphyry compared with calc-alkali rhyolite and rhyolite-obsidian*

Constituent	1	2	3	4
Chemical analyses				
SiO_2	67.77	66.27	66.69	73.66
TiO_2	------	.66	------	.22
Al_2O_3	16.61	15.39	16.69	13.45
Fe_2O_3	2.06	2.14	2.06	1.25
FeO	1.96	2.23	.93	.75
MgO	1.26	1.57	1.15	.32
MnO	------	.07	------	.03
CaO	1.87	3.68	1.40	1.13
Na_2O	4.35	4.13	2.46	2.99
K_2O	2.35	3.01	5.23	5.35
H_2O-	------	------	1.70	------
H_2O+	1.69	.68	------	.78
P_2O_5	------	.17	------	.07
CO_2	.19	------	1.42	------
Total	100.11	100.00	99.73	100.00
Norms				
qz	27.5	20.8	30.8	33.2
or	13.9	17.8	30.6	31.7
ab	36.7	35.1	21.0	25.1
an	8.1	14.5	------	5.0
c	4.0	------	7.0	.9
$CaSiO_3$	------	1.3	------	------
$MgSiO_3$	3.2	3.9	3.8	.8
$FeSiO_3$	1.8	1.3	------	------
mt	3.0	3.0	3.0	1.9
il	------	1.4	------	.5
ap	------	.3	------	.2
cc	.5	------	[1] 2.5	------

[1] Some excess CO_2 not included.

1. Gneissic porphyry from SW¼SW¼ sec. 7, T. 39 N., R. 30 W., Michigan (from Williams, 1890, p. 119): norm from Prinz (1959).
2. Average of 115 chemical analyses and norm of rhyodacite-obsidian (from Nockolds, 1954, p. 1014).
3. "Schistose porphyry" from Big Quinnesec Falls (after Williams, 1890, p. 121, slightly rearranged): norm from Prinz (1959).
4. Average chemical analysis and norm of calc-alkali rhyolite and rhyolite-obsidian (from Nockolds, 1954, p. 1012).

the average calc-alkali rhyolite and rhyolite obsidian of Nockolds (1954, p. 1012).

Widely different opinions have been expressed about the origin of this group of felsic rocks. Credner (1870, in Williams, 1890, p. 119) believed that these rocks represent altered sediments intercalated between beds of diabase, Williams (1890, p. 122) regarded all the felsic rocks as intrusive layers in the greenstone, and Prinz (1959) regarded some of the rocks as intrusive granite dikes, some of them as extrusive lava and tuff, and some of them of doubtful origin, either intrusive or extrusive, but igneous.

In an attempt to reconcile these differences of interpretation, Bayley and Dutton examined most of the exposures of these rocks in June 1960, and concluded that, overall, they most resemble felsic metavolcanic rocks, but that certain of the coarser grained varieties may be intrusives rather than extrusives. The conclusion is based on the following considerations: (1) the rocks themselves resemble very closely metarhyolites from other parts of the region; (2) the greenstone and felsic beds nearly everywhere are parallel; (3) the greenstone and felsic beds are tightly folded together into pitching folds; and (4) the area of felsic rocks in the map area falls on the strike of a belt of felsic volcanic rocks that can be traced, by scattered outcrops, for nearly 20 miles along the strike of the Quinnesec Formation.

In the mapped area, near Aurora, the belt of felsic rocks lies close to the Hoskin Lake Granite, and both Williams and Prinz assumed quite understandably that the granite and the granitic-appearing felsic rocks were related, even though it was not possible to show any direct connection between the two. Prinz (written commun., 1960) agreed that the recognition of most of the felsic strata as metavolcanic rocks rather than intrusive dikes serves to explain the wide textural and compositional differences observed in these rocks, which they probably would not show if they were all offshoots of the Hoskin Lake Granite or Marinette Quartz Diorite. He still regards some of the coarser felsitic rock as intrusive granite, however, and we agree on this point. He suggests further that the granite or granodiorite dikes could be related to large granitic elements of the general Wisconsin batholith which lie just south of the mapped area and which may be considerably older than the Marinette Quartz Diorite and Hoskin Lake Granite. To Prinz we add that they could also be dikes genetically related to the extrusive rocks and more or less restricted to the pile of volcanic rocks, as is often true in younger, better exposed volcanic series. It is generally agreed then that some of the coarser felsic rocks are intrusive dikes, but that their relation with the plutonic granitic rocks of the area has not been established.

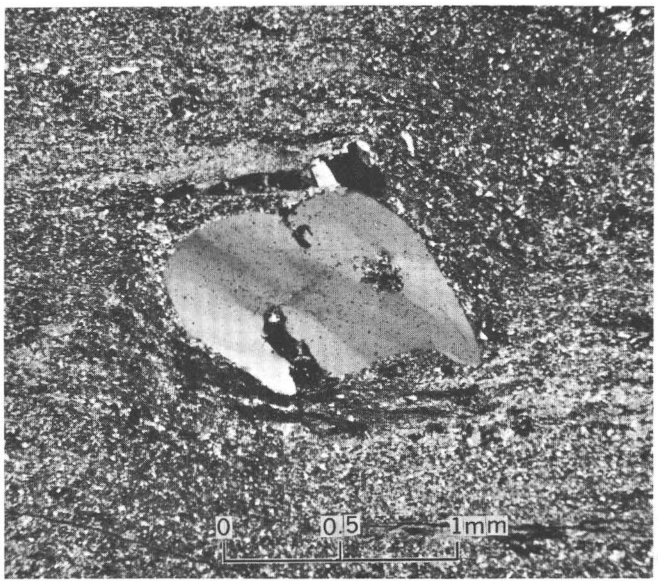


FIGURE 3.—Photomicrograph of quartz-muscovite schist showing embayed quartz eye. Crossed nicols; × 30.

METASEDIMENTARY ROCKS

The metasedimentary intercalations in the metavolcanic rocks are biotite schist, biotite-chlorite schist, quartz-magnetite-hornblende iron-formation and schist, impure quartzite, and arkose.

Two small outcrops of thin-bedded quartz-magnetite-hornblende iron-formation were noted, one in the NE¼SW¼ of sec. 12, T. 38 N., R. 19 E. (Wis.), and the other in NW¼SW¼ of sec. 7, T. 38 N., R. 20 E. (Wis.). The two outcrops lie on the strike of the enclosing andesine amphibolite, but no continuity between the outcrops could be established. The iron-formation is laminated and somewhat schistose. The layers are a few millimeters to a few centimeters in thickness. Layers rich in green or bluish-green hornblende and magnetite alternate with layers of granular quartz, probably originally chert. The hornblende occurs as anhedral grains a much as 0.5 mm long. It is highly pleochroic, X = yellow, Y = green, and Z = blue-green; $Z \wedge c = 17°-19°$; and $N_y = +1.667$. Magnetite forms grains as much as 0.5 mm wide which are dispersed between the hornblende grains, and smaller grains which are dispersed in the quartz layers. The iron-formation in most respects is similar to the Skunk Creek Member of the Solberg Schist of the Dickinson Group (lower Precambrian) in central Dickinson County (James, 1958, p. 32).

Biotite schist and biotite-chlorite schist thought to represent metamorphosed argillaceous sediments are interbedded with amphibolite in the SE¼SW¼ and the NE¼SW¼ sec. 7, T. 39 N., R. 30 W. (Mich.), and the NW¼ sec. 12, T. 39 N., R. 31 W. (Mich.); the SW¼NW¼ sec. 3 and the NE¼SE¼ sec. 12, T. 38 N., R. 19 E. (Wis.). The schists consist mainly of alined biotite and chlorite flakes in a granoblastic matrix of quartz and minor plagioclase. The plagioclase is albite (An_3) in a low-grade metamorphosed rock from sec. 3, T. 38 N., R. 19 E., and oligoclase (An_{29}) from a higher grade metamorphosed rock from sec. 12, T. 38 N., R. 19 E. The chlorite penetrates and mantles the biotite flakes—a feature that suggests that the chlorite is retrograde. Minerals present in the schist in accessory amounts include hornblende, calcite, tourmaline, magnetite, pyrite, hematite, zircon, and epidote. Much of the schist shows layering or lamination which appears to be original bedding. This layering and the mineral composition suggest that the original rocks were thin-bedded argillites. Some similar schists described by Froelich (1953, p. 64–65), by Thompson (1955, p. 60–61), and by Fulweiler (1957, p. 27–28) contain garnet.

Thin beds of impure quartzite and possibly arkose are associated with the biotite schist in the SE¼ and the NE¼SW¼ sec. 7, T. 39 N., R. 30 W.(Mich.). The quartzite is gray to brown, is massive, and consists chiefly of clear quartz grains (0.05–0.1 mm), but contains about 10 percent of brown biotite and accessory amounts of hornblende, magnetite, pyrite, chalcopyrite, hematite, zircon, epidote, sericite, muscovite, and feldspar.

The arkose forms thin beds which, according to Prinz (1959, p. 72), are difficult to distinguish from granite dikes and metarhyolite because of poor exposures and because deformation and metamorphism have produced similar-appearing gneissic rock from all. In hand specimen, the medium-grained arkose shows grains of clear quartz, red feldspar, biotite, and pyrite. In thin section the clastic character of the rock is clearly shown. Well-rounded to angular grains of the above minerals are embedded in a finer grained matrix of extremely variable grain size and composition. The minerals of the matrix are quartz, albite microcline, and biotite. Accessory minerals include apatite, calcite, pyrite, and leucoxene. The red feldspar observed in hand specimens is albite, which is present as angular grains as much as 3.4 mm across.

THICKNESS

It is not possible to arrive at a figure for the total thickness of the Quinnesec Formation in the area mapped, because of the intrusive contact on the south and because the north contact is hidden. No field evidence indicating that the rocks have been repeated by folding or faulting was observed, although minor folds and dislocations are present. If the assumption is made that the section is unbroken and continuous, direct map measurements show the thickness to be in excess of 10,000 feet.

ORIGIN

The Quinnesec Formation originated during an early Precambrian epoch of widespread volcanism during which many cubic miles of basaltic lava were poured out on the surface of the earth. The location of the center or centers of volcanism is unknown. The lava was laid down layer upon layer until a pile at least two miles thick had accumulated. Explosive volcanic activity at intervals provided some pyroclastic material, and during at least one short interval rhyolitic and rhyodacitic tuffs and flows were added to the pile. Pillow structures, which are common features in many of the mafic flows, indicate that some of the flows probably were extruded in a submarine environment. The metasedimentary schists of the Quinnesec probably represent incursions of terrigenous clastic materials into the volcanic area during lulls in the volcanic activity.

AGE AND CORRELATION

The age of the Quinnesec Formation cannot be exactly determined from the field evidence. Among the geologists who have worked in the area over the past 90 years, two widely divergent opinions have prevailed: (1) the Quinnesec strata are much older than the adjacent Michigamme strata and are probably pre-Animikie in age; (2) the Quinnesec strata are younger than the adjacent Michigamme strata and succeed the latter upward to the south.

Rominger (1881, in Bayley, 1904, p. 76) and W. S. Bayley (1904, p. 131) favored an early Precambrian (Archean) age for the Quinnesec rocks principally because of the striking similarity between them and other lower Precambrian (pre-Animikie) volcanic formations in the region. The Menominee district was envisioned by W. S. Bayley as a structural trough containing Huronian (Animikie) strata which were flanked on both north and south sides by pre-Animikie rocks. He explained the absence of Chocolay and Menominee Group rocks between the Michigamme Slate and the Quinnesec Formation by an unconformable overlap of the Michigamme on the Quinnesec. According to Van Hise and Leith (1911, p. 344), the possibility of a fault on this contact was also entertained.

Concepts regarding a late Animikie age were advanced by Brooks, Hotchkiss, Van Hise and Leith, and Allen. Brooks (1880, in Bayley, 1904, p. 64) and Hotchkiss (1910, in Van Hise and Leith, 1911, p. 345) shared the opinion that the Quinnesec strata succeeded the Michigamme strata. According to Hotchkiss, in Florence County, Wis., the Upper Huronian slate (Michigamme) is succeeded to the south by the volcanic formation (Quinnesec), and the two formations show an interbedded transition zone. On the basis of Hotchkiss' observations, Van Hise and Leith (1911) assigned the Quinnesec Formation to a post-Michigamme age, but with the reservation that "it is yet possible that the Quinnesec schist in the Menominee district may be really pre-Huronian [pre-Animikie of this report] for continuous exposures do not connect the two areas, and green schists of this type are known in at least three horizons in the pre-Cambrian of Michigan." (1911, p. 345). The late Animikie age of the Quinnesec was accepted by Allen (1920, p. 202–203), and he correlated the granitic intrusives that cut the Quinnesec strata with the Presque Isle and other unnamed granites of northwestern Wisconsin which show an apparent syntectonic relationship with respect to the post-Animikie orogeny.

Recently (1959–60) the evidence for Hotchkiss' conclusion that the Michigamme and Quinnesec Formations are interbedded was reexamined by Dutton during his resurvey of Florence County, Wis., and he has concluded that the two formations do not grade into one another but are separated by a major fault. Two lines of evidence place the fault at the Quinnesec-Michigamme contact: (1) he found that the ellipsoidal greenstones of the Quinnesec that lie closest to the projected contact between the Quinnesec and Michigamme Formations face north, whereas the graded graywacke beds of the Michigamme that lie closest to the contact face south: and (2) he found that the greenschist and oligoclase-amphibolite isofacies of regional metamorphism in the Quinnesec rocks end abruptly at the inferred faultline and that greenschist-facies rocks are juxtaposed to andesine-amphibolite facies rocks along the fault for several miles (pl. 1).

The above findings by Dutton completely refute any previous assignments of the Quinnesec made on the basis of its apparent superposition on the Michigamme Slate, but they do not clear all doubts as to the true relative age of the Quinnesec.

We believe that the evidence accumulated to date, though inconclusive, will best support W. S. Bayley's view that the Quinnesec Formation is of early Precambrian age. Our view is based on the following considerations:

1. Volcanic and sedimentary formations similar in general to the Quinnesec form an integral part of every pre-Animikie complex in the immediate region. Irving and Van Hise (1892) noted an old complex unconformably beneath Animikie strata of the Gogebic iron range of northwestern Wisconsin and Michigan. Weidman (1907, p. 383–384) showed that nearly all of the north-central Wisconsin in underlain by an ancient igneous and sedimentary complex, upon which strata that he interpreted as of Middle Huronian (Menominee Group) rest unconformably. The complex previously noted by Irving and Van Hise apparently is the west margin of the complex described by Weidman. Allen (1915a) and Hotchkiss (in Hotchkiss and others, 1915) determined that the region between the north-central area of Weidman and the Gogebic iron range was underlain by a series of northeast-trending alternating synclines of Animikie strata and anticlines of basement complex. Thus it was shown that a vast pre-Animikie complex is present in northern Wisconsin. Allen (1915a) also demonstrated that some granite younger than the Animikie Series is present in the north part of

this complex. Younger granite intrusions, such as these and the Hoskin Lake Granite that cuts the Quinnesec Formation of the Menominee district, apparently comprise only a small part of the exposed granite of this complex.

2. G. W. Bowen and C. F. Corey, working under the guidance of Van Hise and Leith (1911, p. 34), determined in the Menominee district that the Michigamme Slate is overlain by a volcanic formation (the Badwater Greenstone of present usage, and this relation was verified at several places during the recent general resurvey. The apparent Michigamme-Quinnesec succession in Florence County, Wis., obviously suggests that the Quinnesec Formation and the Badwater Greenstone are the same formation (Allen, 1915a, p. 30), but, during the recent resurvey of Florence County, Dutton was able to show for the first time that the Badwater Greenstone is a relatively thin (1,000 ft) formation sandwiched between the Michigamme Slate and the uppermost Animikie rocks, the Paint River Group, and that it does not compare well with the Quinnesec either in bulk or in content.
3. Relations in Florence County, Wis., indicate that a major fault separates the Michigamme Slate on the north from the volcanic formation (Quinnesec) on the south; further, the structure of the Florence district as it is now interpreted seems to require a major fault between the Michigamme Slate and the Quinnesec Formation. A short distance west of the Florence district, the youngest Animikie strata, the Paint River Group, rest on definite Badwater Greenstone, which in turn rests on the Saunders Formation. The Saunders Formation is probably partly equivalent in age to the Randville Dolomite and Sturgeon Quartzite of the Menominee district (Allen, 1910, p. 30); therefore, the Saunders-Badwater contact must be a fault contact or an unconformable contact, because the whole Menominee Group and the Michigamme Slate are absent. South of the Saunders Formation, apparently underlying it, is another greenstone series which could be the western extension of the Quinnesec Formation (Allen, 1910, p. 30). Thus it is possible to establish here that there are two greenstone belts, one north of the Saunders, called Badwater, and one south of the Saunders, called Quinnesec, and these belts are apparently younger and older than the Saunders, respectively. Although much of the area between the Florence district and the area south of Iron River is unmapped, the southernmost belts of greenstone in each of the mapped areas strike toward one another and they are very probably the same belt. If we accept the concept that the Saunders Formation is roughly the equivalent of the Randville Dolomite and Sturgeon Quartzite, as suggested by James (1958, p. 30, table 1), then the underlying greenstone series is probably of pre-Animikie age because the only other pre-Randville volcanic rocks known in the region are also pre-Animikie. If that greenstone series is of pre-Animikie age, then the Quinnesec of the Menominee district, which we believe to be directly connected to it, is also of pre-Animikie age.
4. Gabbro dikes and sill-like bodies cut the pre-Animikie rocks and all the Animikie formations, and the intrusions of this swarm is generally regarded as a late Animikie phenomenon. Several large sill-like plutons of this swarm cut the Quinnesec Formation. These plutons contain abundant stoped blocks of Quinnesec strata, mainly greenstone schist and mylonitic metarhyolite gneiss, which were clearly schist and gneiss before the gabbro was intruded. Because pregabbro deformation of the type indicated by these schist and gneiss inclusions has not been demonstrated outside the pre-Animikie complexes, we regard these deformed inclusions as further evidence for a pre-Animikie age of the Quinnesec Formation.

The mineral ages determined on biotite by rubidium: strontium and by potassium:argon ratios yielded respectively, for Quinnesec greenstone in Wisconsin, 1,350 and 1,340 million years (Davis and others, 1960, table 10, p. 153), but these ages we consider to represent a time of metamorphism rather than the time of origin of the greenstone.

The Quinnesec Formation, as a lower Precambrian formation, correlates best with the Dickinson Group of central Dickinson County (James, 1958, p. 31–32) and the Dickinson Group undivided of eastern Iron County (Bayley, 1959b, p. 11). The Dickinson Group is older than major bodies of pre-Animikie granite, such as the Carney Lake Gneiss. It is thick and extensive, and contains all the elements found in the Quinnesec, plus some others, such as arkosic conglomerate.

NORTHERN AREA

CARNEY LAKE GNEISS AND ASSOCIATED ROCKS

BY SAMUEL B. TREVES

GENERAL RELATIONS

The Carney Lake Complex of Treves (1960), herein adopted and renamed the Carney Lake Gneiss, underlies about 50 square miles of generally uninhabited and poorly accessible country in the northeast part of the mapped area (pls. 2, 3). The east half of T. 40 N., R. 29 W. is the type area.

The fieldwork in the gneiss area was done during the summers of 1955–57. Areas of probable outcrops were delimited on aerial photographs, then visited, and the outcrops located by pace and compass traverses from known points and by direct reference to topographic features or survey points. Traverses spaced 250–500 feet apart were made over most of the rest of the area. An attempt was made to visit each 40-acre plot, but it was impossible to reach some areas because of widespread impassable swamps.

The petrographic descriptions of the rocks are based on the examination of approximately 300 thin sections and more than 1,000 hand specimens. Feldspar determinations are based on measurements of the maximum extinction angle of albite twinning, refractive index, optic sign, and differential staining. Volume percentages of minerals were determined with a point counter, and grain sizes were measured with an ocular scale. Five chemical analyses of rocks from the complex were made in the laboratories of the University of Minnesota. These analyses were made possible by a grant from the Geological Society of America, which is gratefully acknowledged.

GNEISS AND GRANITIC ROCKS

Granitic gneiss constitutes about 85 percent of the Carney Lake Gneiss; of the remainder, about 5 percent is granodiorite and syenite dikes, and about 10 percent is inclusions of older rock. The gneiss is not uniform in composition or appearance, but varies from a gray plagioclase-biotite gneiss to red microcline-biotite gneiss. For the purpose of discussion these types will be designated gray gneiss, composite gneiss, and red gneiss, respectively. The several types of gneiss will be discussed first.

Gray gneiss

The gray gneiss probably constitutes about 25 percent of the complex collectively called the Carney Lake Gneiss. It is most abundant in the northern half of the complex where it contains many amphibolite inclusions (fig. 4). Some of the best and most easily accessible exposures are about a mile beyond the north limit of the area shown on plate 3, along the Merimam Truck Trail in sec. 25, T. 41 N., R. 29 W., and sec. 30, T. 41 N., R. 28 W., where the road passes along the northeast edge of the complex.

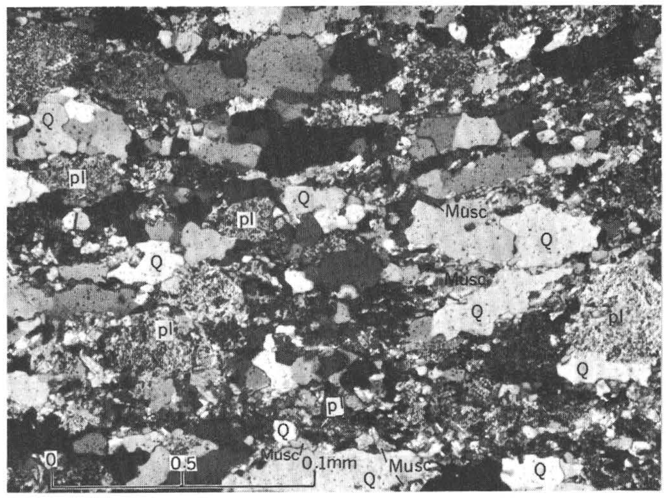


FIGURE 4.—Photomicrograph of fine-grained gray gneiss showing linear arrangement of quartz stringers, and extensive alteration of plagioclase. Crossroad nicols; × 45. Q, quartz; pl, plagioclase; musc, muscovite.

The grain size of the gray gneiss ranges from fine to coarse. Fresh specimens are gray, whereas weathered specimens are darker gray and dull. In thin sections the gray gneiss shows abundant plagioclase, quartz, and biotite (fig. 4). The foliation is well shown by alined biotite, and also by the plagioclase and quartz, which are arranged in subparallel elongated grains and in lenticles. Cataclastic structures are common.

The plagioclase is generally oligoclase, which is commonly the most abundant mineral present. The oligoclase is subhedral and most of it is unaltered, but locally it is extensively replaced by sericite or muscovite, and some large crystals have been replaced by granular aggregates of quartz and albitic oligoclase.

The quartz, which is next abundant, is anhedral and ranges from very small interstitial crystals to large elongate, lobate grains. It occurs as discontinuous lenses and layers which bifurcate and enclose lenses of plagioclase and biotite. The quartz of the layers is generally a mosaic of anhedral crystals which are strained and extensively fractured.

Biotite, the only other abundant mineral, is subhedral and ranges from reddish brown to brownish green. It occurs in the interstices between plagioclase crystals and appears to form an interconnected network which outlines the plagioclase. Some biotite is altered to chlorite.

Epidote-clinozoisite, ilmenite, magnetite, pyrite, apatite, sphene, allanite, and zircon are common acces-

TABLE 11.—*Modal analyses of gray gneiss from the Carney Lake Gneiss*

Mineral	Specimen										
	101157	33556	14655	1155	7155	1655	5755	27056	7857	4855	(1)
Plagioclase	60.3	44.5	44.3	34.9	58.8	37.9	36.3	39.3	41.5	22.9	43.2
Quartz	26.2	28.4	30.6	38.1	21.7	26.0	35.2	19.6	20.6	35.2	27.4
Biotite	12.5	13.2	13.1	3.4	6.2	10.7	8.6	10.9	18.4	17.5	11.5
Epidote-clinozoisite	Tr.	8.4	.6	4.1	2.7	11.2	4.2	9.6	4.6	20.5	5.1
Muscovite and sericite	Tr.	1.2	4.9	19.7	6.2	13.7	15.7	11.5	5.4	1.8	9.5
Opaque minerals [2]	.5	2.1	Tr.	Tr.	1.3	Tr.	Tr.	4.1	6.6	.7	1.0
Apatite	Tr.	.5	Tr.	Tr.	Tr.	Tr.	Tr.	2.1	2.9	.5	Tr.
Sphene	----	----	----	----	----	Tr.	----	----	----	----	Tr.
Allanite	Tr.	Tr.	----	Tr.	Tr.	Tr.	----	----	----	----	Tr.
Zircon	----	----	Tr.	Tr.	Tr.	Tr.	----	----	Tr.	Tr.	Tr.
Chlorite	----	2.7	6.5	----	----	----	----	2.9	----	.9	1.3
Carbonate	----	Tr.	Tr.	----	----	Tr.	----	Tr.	----	----	Tr.
Microcline	----	----	Tr.	Tr.	2.4	----	----	----	----	----	Tr.
Albite	----	----	----	----	.7	----	----	----	----	----	Tr.

[1] Average of 20 specimens.
[2] Magnetite, ilmenite, and pyrite.

Location of specimen:

101157. NE¼ sec. 25, T. 41 N., R. 29 W.
33556. NE¼ sec. 13, T. 40 N., R. 29 W.
14655. NW¼ sec. 35, T. 40 N., R. 29 W.
1155. SW¼ sec. 36, T. 40 N., R. 29 W.
7155. SW¼ sec. 1, T. 39 N., R. 29 W.
1655. NE¼ sec. 1, T. 39 N., R. 29 W.
5755. SE¼ sec. 36, T. 40 N., R. 29 W.
27056. SE¼ sec. 6, T. 40 N., R. 28 W.
7857. SE¼ sec. 26, T. 41 N., R. 29 W.
4855. SW¼ sec. 1, T. 40 N., R. 29 W.

sory minerals. Muscovite, microcline, and albite also occur in the gray gneiss, but are rare. The muscovite occurs as large skeletal plates in plagioclase and appears to represent reorganization of sericite into large muscovite crystals. The microcline occurs along the borders of quartz pods and layers and appears to replace plagioclase. Albite occurs as rims on plagioclase and as clear anhedral crystals at the borders of quartz layers. The occurrence of muscovite, microcline, albite, and most of the quartz suggests that these were the last minerals to form.

Modal analyses of 10 specimens of gray gneiss, which exhibit some of its variability and which were selected from widely separated areas, and the average of 20 modal analyses, are presented in table 11. The table does not include the modes of hornblende-biotite gneiss or biotite-rich gneiss, which are varieties of gray gneiss that occur adjacent to disrupted inclusions and which pass transitionally into the gray gneiss. These rocks are virtually the same as the gray gneiss except that they contain hornblende and more biotite. Two modal analyses of these transitional rocks are presented in table 12.

TABLE 12.—*Modes of hornblende-biotite gneiss and biotite-rich gneiss from the Carney Lake Gneiss*

Mineral	*Hornblende-biotite gneiss*	*Biotite-rich gneiss*
Biotite	24.1	31.5
Hornblende	15.6	Tr.
Plagioclase	36.5	36.4
Quartz	8.3	15.2
Epidote group	7.2	9.3
Apatite	Tr.	Tr.
Opaque minerals [1]	3.1	4.2
Allanite	Tr.	Tr.
Muscovite	Tr.	3.4
Chlorite	5.2	Tr.
Zircon	Tr.	Tr.

[1] Magnetite, ilmenite, and pyrite.

The modal analyses shown in tables 11 and 12, when compared with the modal analyses of the biotite schist and amphibolite inclusions (tables 13 and 11), appear to exhibit a complete gradation from amphibolite to biotite schist to hornblende and biotite-rich gneisses to the gray gneiss.

A chemical analysis of the gray gneiss and, for comparison, the chemical anlyses of the composite gneisses from central Dickinson County, and the average of six analyses of granite and granitic gneiss from the southern border of the Felch trough are presented in table 13. The table shows that the analyses of the gray gneiss and the gneiss from the north are similar, but that the gray gneiss contains more FeO, MgO, and CaO, and less K_2O.

TABLE 13.—*Chemical analyses of the gray gneiss and composite gneiss, and gneisses and granite from central Dickinson County, Mich.*

Constituent	1	2	3	4	5
SiO_2	72.95	73.26	70.92	73.39	73.13
Al_2O_3	14.02	13.89	16.95	13.68	14.19
Fe_2O_3	.34	.37	.82	.65	.82
FeO	2.12	1.64	1.78	1.18	.72
MgO	.79	.63	.07	.20	.45
CaO	3.00	1.35	1.47	1.20	.79
Na_2O	3.78	2.73	3.90	4.72	3.12
K_2O	1.34	4.68	3.16	4.12	5.64
H_2O+	.40	.51	.04	.07	.15
H_2O-	.03	.08	.08	.27	.46
TiO_2	.30	.23	.21	.16	.19
P_2O_5	.07	.05	.18	.16	.02
MnO	.02	.05	.03	.02	Trace
CO_2	.22	.03	.06	.34	----
Total	99.38	99.50	99.67	100.16	99.68

1. Gray gneiss, specimen 101157, from NE¼ sec. 25, T. 41 N., R. 29 W., Carney Lake Gneiss.
2. Composite gneiss, specimen 100857, from SE¼ sec. 26, T. 40 N., R. 28 W.
3. Granite gneiss from SE¼SW¼ sec. 33, T. 42 N., R. 28 W., central Dickinson County (after James and others, 1961, p. 27).
4. Granite gneiss from the NE¼SW¼ sec. 22, T. 41 N., R. 30 W., central Dickinson County (after James and others, 1961, p. 27)
5. Average of six analyses of granite and granitic gneiss from near the south border of the Felch trough (after Clements and Smyth, 1899b, p. 389, 391).

Composite gneiss and red gneiss

The composite gneiss constitutes at least 70 percent of the complex. It is present almost everywhere, but is more abundant in the southern half of the area, where it contains minor patches of red gneiss and many inclusions of biotite schist. The composite gneiss is well exposed in outcrops along the roads of the southern half of the complex and in the pastures in sec. 1, T. 39 N., R. 29 W., and sec. 36. T. 40 N., R. 29 W. (pl. 3)

The grain size of the composite gneiss ranges from medium to coarse. The gneiss is streaky and consists of red and gray elements, the red parts composed of pink microcline and quartz, the gray parts chiefly of plagioclase and biotite, which are the same minerals that constitute the gray gneiss. At some places the red part forms patches and streaks within the gray, at others the gray is enveloped by the red, and at still others the two elements form alternating layers. The red part commonly occurs as veins or layers of coarse pegmatite which cut across the foliation or bifurcate and join other layers. Here and there veins and layers of the red material swell and form large pods of pegmatite which fade transitionally into the gray rock. Pegmatite pods may also pinch and swell along the strike of the foliation of the gneiss. Locally they stop abruptly against the gray rock, only to appear again further along the strike.

Thin sections of the composite gneiss show small replicas of features shown by the outcrops (fig. 5). The thin sections show two elements, an altered part that appears older, and an unaltered part that appears younger.

The older part consists chiefly of plagioclase and biotite, except where it is much altered. The plagioclase, generally oligoclase, is subhedral, is usually flecked with sericite, and locally is extensively replaced by epidote, clinozoisite, and sericite. Many plagioclase crystals are almost completely replaced by masses of anhedral quartz, albitic oligoclase, clinozoisite, and skeletal plates of muscovite that commonly replace plagioclase where sericitization is extensive. The biotite is greenish brown to reddish brown, and occurs as wisps and shreds that outline the plagioclase crystals. Accessory minerals are magnetite, ilmenite, pyrite, apatite, sphene, zircon, and allanite.

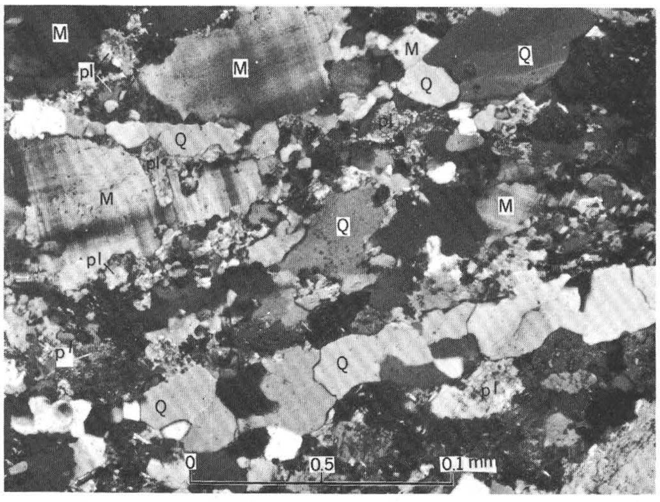


FIGURE 5.—Photomicrograph of composite gneiss showing elongated quartz layers, Q; microcline, M; and extensively altered plagioclase, pl. Crossed nicols; × 45.

The younger part consists of quartz, microcline, and albite. Most of the quartz and all of the microcline and albite appear to be closely related in time of formation. These minerals occur in layers and veins that enclose islands of plagioclase and biotite. The quartz of these layers occurs as large, sublobate crystals that are commonly strained or fractured. Small anhedral crystals of quartz also occur as individuals or in clusters of a few crystals in the interstices between the plagioclase and biotite of the older part. The microcline characteristically occurs at the borders of the layers of quartz as twinned anhedral crystals that replace the adjacent plagioclase, but also occurs as anhedral patches and large subhedral crystals in the quartz layers. In some rocks the microcline appears as patches in the parts of the thin section that are rich in plagioclase and biotite, where it replaces plagioclase and contains optically continuous remnants of the plagioclase and inclusions of quartz and chlorite, all of which are clearly unreplaced parts of the older part of the rock. The albite of the quartz layers is not abundant. It is present only at the borders of the layers, where it occurs as discrete anhedral crystals that appear to be earlier than either the quartz or microcline.

The paragenetic sequence of minerals in the composite gneiss, although not entirely clear in all thin sections, seems to be (1) oligoclase, biotite, small interstitial quartz, and probably the accessory minerals, (2) albite, (3) sericite and muscovite, and (4) microcline and quartz of the layers. The extensive sericitization of the plagioclase and the formation of the muscovite apparently are an early phenomenon that occurred prior to the formation of the microcline.

The percentages of the chief minerals of the composite gneiss are extremely variable, and this variation is probably governed by the amount of late quartz, microcline, and albite that was introduced. The modes of 10 specimens, which show the variability of the gneiss, are presented in table 14, along with the average of the modal analyses of 20 specimens of the gneiss.

TABLE 14.—*Modal analyses of composite gneiss from the Carney Lake Gneiss*

Mineral	Specimen										
	11857	100857	4857	2757	2657	11657	25056	15256	33356	9655	(1)
Plagioclase	34.7	35.2	36.6	39.5	40.1	42.7	37.4	26.5	31.2	36.1	35.8
Quartz	28.5	27.1	34.5	29.1	23.2	23.1	29.2	21.6	35.5	31.8	21.8
Microcline	15.7	28.4	18.1	14.8	11.3	11.3	16.0	25.6	23.9	20.8	17.6
Biotite	8.7	6.7	3.2	6.9	14.8	12.8	6.9	2.1	2.7	6.1	7.1
Muscovite-sericite	10.2	2.0	6.4	7.3	4.3	5.1	6.7	10.8	3.3	3.4	5.7
Albite	---	---	Tr.	---	4.2	2.5	---	4.3	---	Tr.	1.1
Epidote-clinozoisite	Tr.	Tr.	1.2	Tr.	Tr.	Tr.	2.1	6.5	2.5	Tr.	Tr.
Chlorite	1.7	Tr.	---	1.3	.7	1.7	Tr.	---	Tr.	.7	Tr.
Opaque minerals [2]	.5	.6	Tr.	.7	1.4	1.8	1.9	.6	.6	1.1	.9
Apatite	Tr.	Tr.	Tr.	---	Tr.	Tr.	.8	Tr.	.4	Tr.	Tr.
Zircon	Tr.	Tr.	---	Tr.	---	---	Tr.	Tr.	Tr.	Tr.	Tr.
Allanite	Tr.	Tr.	Tr.	---	---	---	Tr.	Tr.	Tr.	Tr.	Tr.
Sphene	---	---	---	---	---	---	---	Tr.	---	Tr.	Tr.
Carbonate	---	---	Tr.	---	---	---	---	---	---	---	Tr.

[1] Average of 20 specimens.
[2] Magnetite, ilmenite, and pyrite.

Location of specimen:
11857. SE¼ sec. 6, T. 40 N., R. 28 W.
100857. SE¼ sec. 26, T. 40 N., R. 28 W.
4857. NE¼ sec. 25, T. 41 N., R. 28 W.
2757. SE¼ sec. 6, T. 40 N., R. 28 W.
2657. SE¼ sec. 6, T. 40 N., R. 28 W.
11657. NE¼ sec. 2, T. 40 N., R. 30 W.
25056. SE¼ sec. 13, T. 40 N., R. 29 W.
15256. NW¼ sec. 6, T. 39 N., R. 28 W.
33356. NW¼ sec. 13, T. 40 N., R. 29 W.
9655. SE¼ sec. 1, T. 39 N., R. 29 W.

A chemical analysis of the composite gneiss is presented in table 13, and also, for comparison, the analysis of the gray gneiss and some other gneisses, and the average of six specimens of granite and gneiss from central Dickinson County, Mich. The analysis of the composite gneiss differs from that of the gray gneiss chiefly in having a smaller amount of FeO, CaO, and Na_2O, and a larger amount of K_2O.

No easily accessible outcrops of the red gneiss are present in the complex, but the character of such gneiss can be observed most readily in the SE¼NE¼ sec. 13, T. 40 N., R. 29 W., where an outcrop of composite gneiss shows an irregularly shaped body of red gneiss having a diameter of about 4 feet that fades

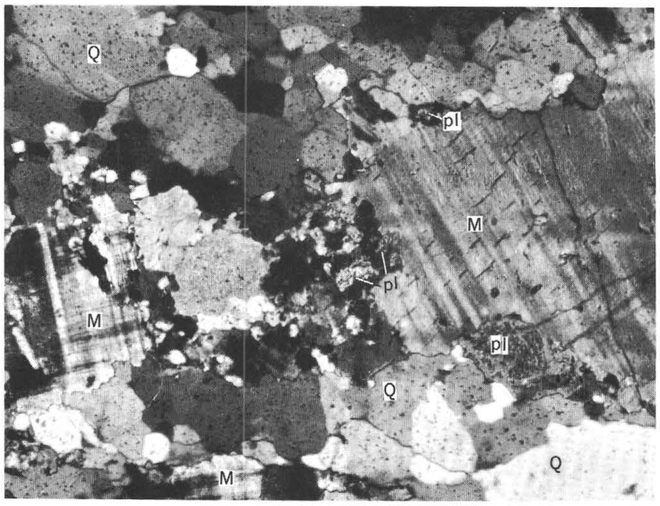


FIGURE 6.—Photomicrograph of red gneiss showing abundant microline, quartz, and altered plagioclase. Note the unreplaced remnants of plagioclase at the margins of the large microcline crystal. Crossed nicols; × 45. M, microcline; Q, quartz; pl, plagioclase.

transitionally into the composite gneiss. The same relation is shown in the NW½SE¼ sec. 1, T. 39 N., R. 29 W., and the NE¼NW¼ sec. 7, T. 40 N., R. 28 W. Red gneiss containing abundant biotite occurs in the NW¼SW¼ sec. 25, T. 40 N., R. 29 W.

The red gneiss is medium grained and weakly foliated. Fresh specimens are pink or red, whereas

TABLE 15.—*Modal analyses of red gneiss from Carney Lake Gneiss*

Mineral	Specimen		
	30655	31756	33756
Quartz	38.5	33.6	41.5
Microcline	24.2	26.2	28.1
Plagioclase	16.5	20.3	7.3
Biotite-chlorite	8.7	4.2	3.6
Muscovite	7.1	7.9	11.7
Albite	3.6	5.3	4.3
Epidote-clinozoisite	.8	1.8	2.7
Opaque minerals [1]	.6	.7	.8
Apatite	Tr.	Tr.	Tr.
Zircon	Tr.	---	---
Allanite	Tr.	---	---
Sphene	Tr.	---	---
Carbonate	Tr.	---	---

[1] Magnetite and pyrite.
30655. Spotted red gneiss from the SE¼ sec. 23, T. 40 N., R. 29 W.
31756. Red gneiss from the NW¼ sec. 7, T. 40 N., R. 29 W.
33756. Red gneiss from the NW¼ sec. 18, T. 40 N., R. 29 W.

weathered specimens are brownish pink. Thin sections exhibit abundant quartz and lesser amounts of microcline, plagioclase, and muscovite (fig. 6). Biotite, which is abundant in the gray and composite gneisses, is scarce in the red gneiss. Like the composite gneiss, the red gneiss consists of two components of different age, an older part composed of extensively altered plagioclase, and a younger part that consists of quartz, microcline, muscovite, and minor amounts of albite, but the red gneiss generally contains more quartz and microcline and less biotite and plagioclase than the other gneisses. Three modal analyses of red gneiss are presented below in table 15; one specimen the spotted red gneiss, is a biotite-rich variety. The variation in the mineralogy is related to the introduction of the younger minerals and the replacement and alteration of the older ones.

GRANODIORITE

Granodiorite occurs as rare dikes that are most abundant in the southern half of the complex and are more likely to be closely associated with the composite gneiss than the other types. Easily accessible outcrops are present in the SE¼SE¼ sec. 23, T. 40 N., R. 29 W., and the NW¼NE¼ sec. 26, T. 40 N., R. 29 W., along the fire-control road. An outcrop of a magnetite-rich variety is present below the spillway of the Sturgeon River Dam in the SE¼NW¼ sec. 8, T. 39 N., R. 28 W.

The granodiorite is massive, equigranular, pink, medium to fine grained, and brownish-pink weathering. Hand specimens show abundant pink feldspar, quartz, and biotite. Thin sections show hypidiomorphic-granular texture, abundant plagioclase, lesser amounts of potassium feldspar, quartz, and biotite, and a few accessory minerals. The plagioclase is oligoclase which occurs as almost euhedral crystals that commonly are zoned. The potassium feldspar is usually subhedral orthoclase, but plaid twinning on the borders of some crystals indicates microcline. The orthoclase may be unaltered or sericitized to varying degrees, and inclusions of quartz are common. Quartz also occurs as anhedral crystal clusters and individual grains in the interstices between the feldspar crystals. The biotite is brown or greenish brown and occurs as poorly terminated interleaved plates and individual laths. Epidote, apatite, magnetite, pyrite, microcline, muscovite, allanite, and zircon are present in accessory amounts. Secondary chlorite, carbonate, and hematite derived from magnetite and pyrite are present in some thin sections. Dikes of the granodiorite from widely separated areas have virtually the same composition. The mode of a specimen from the SE¼SE¼ sec. 23, T. 40 N., R. 29 W., is given in table 16.

TABLE 16.—*Mode of granodiorite*

[Specimen from the SE¼SE¼ sec. 29, T. 40 N., R. 29 W., Michigan]

Mineral	*Percent*	*Mineral*	*Percent*
Plagioclase	34.7	Apatite	1.9
Orthoclase	21.9	Opaque minerals [1]	2.5
Quartz	16.2	Chlorite	2.1
Biotite	12.0	Allanite	Tr.
Muscovite	5.9	Zircon	Tr.

[1] Pyrite and magnetite.

SYENITE

Red syenite occurs only in one part of the complex. It constitutes a series of low, flat outcrops in the E½ sec. 34, T. 40 N., R. 28 W., about one-quarter mile east of the edge of the area shown on plate 3. The syenite, which is unlike any other rock of the complex, is faintly foliated and ranges from medium to coarse grained. Pink feldspar and muscovite are the only constituents that can be identified in hand specimens. In thin section the syenite is hypidiomorphic-granular, and locally exhibits cataclastic structure. The rock contains abundant plagioclase and lesser amounts of potassium feldspar, perthite, and muscovite. The plagioclase is albite, characteristically unaltered and slightly pink in plane polarized light; it occurs as blocky subhedral crystals, not all of which are twinned. The potassium feldspar is orthoclase, which also occurs as blocky subhedral crystals. The perthite, which consists of orthoclase and albite(?), exhibits much the same characteristics as the other feldspars. Muscovite occurs as sheaves and fan-shaped aggregates; terminations of the crystals are often ragged and penetrate the feldspars. Quartz is rare; it occurs as individual anhedral crystals in the interstices between feldspar crystals.

A modal analysis of the syenite is given in table 17, and a chemical analysis is presented in table 18.

TABLE 17.—*Mode of syenite*

Mineral	*Percent*	*Mineral*	*Percent*
Albite	58.7	Muscovite	8.5
Orthoclase	19.3	Quartz	5.8
Perthite	7.1	Accessories [1]	.7

[1] Magnetite, pyrite, apatite, and zircon.

TABLE 18.—*Chemical analysis of syenite*

[Specimen 100557, from the NE¼ sec. 33, T. 40 N., R. 28 W., Michigan]

Constituent	*Percent*	*Constituent*	*Percent*
SiO_2	67.51	H_2O-	0.01
Al_2O_3	18.76	TiO_2	.08
Fe_2O_3	.13	P_2O_5	.03
FeO	.19	MnO	.01
MgO	.82	BaO	----
CaO	.26	CO_2	.11
Na_2O	9.66		
K_2O	1.95	Total	99.74
H_2O+	0.22		

INCLUSIONS IN THE GNEISS

Inclusions in the Carney Lake Gneiss constitute less than 10 percent of the complex, but bear importantly on the character and origin of the gneiss. They consist of amphibolite, biotite schist, and metasedimentary rocks. The inclusions are clearly older than the gneiss and may represent (1) engulfed parts of the pre-gneiss Dickinson Group which occurs to the north of the complex and consists, in part, of a metamorphosed series of basic tuffs, graywacke-type deposits, and basaltic flows (James and others, 1961), or (2) engulfed parts of the Quinnesec Formation which occurs to the south of the complex and consists of metavolcanic rocks, greenstone, amphibolite, and schist.

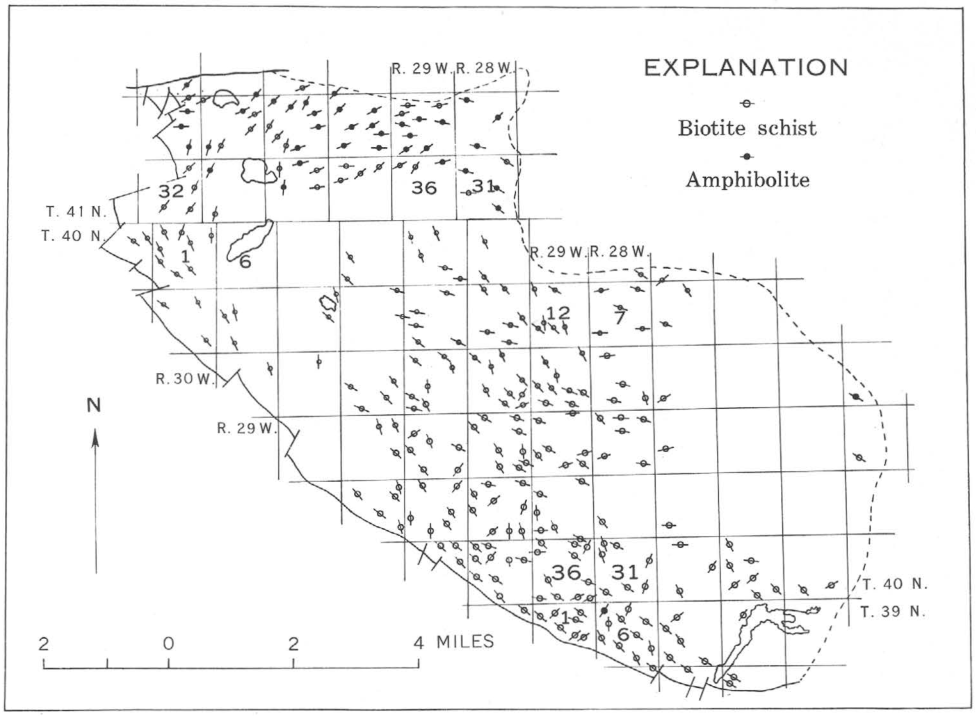


FIGURE 7.—Distribution and orientation of inclusions in the Carney Lake Gneiss.

Amphibolite

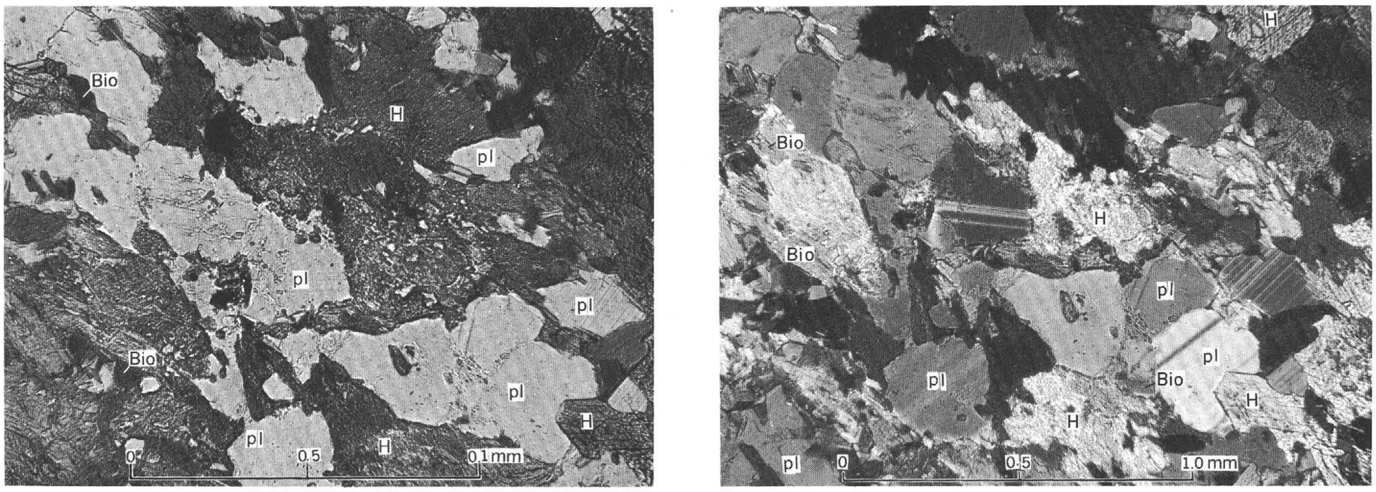
Amphibolite inclusions are abundant in the gneiss north of the correction line between T. 40 N. and T. 41 N., but occur only rarely south of this line (fig. 7). Easily accessible outcrops showing abundant inclusions of amphibolite are present in sec. 25, T. 41 N., R. 29 W., and sec. 30, T. 41 N., R. 28 W., where the Carney Truck Trail passes close to the northern boundary of the complex. The inclusions range from small lenses, only inches long, to large ellipsoidal bodies having an average diameter of 30 feet. The elongation is usually parallel to the strike of the foliation of the adjacent gneiss.

The amphibolite is commonly massive and medium grained. Fresh surfaces exhibit abundant, easily identified hornblende, plagioclase, and pyrite. In thin section the minerals show a subparallel arrangement, and hornblende or plagioclase form an interconnected network which resembles either the ophitic or diabasic texture of basalt (fig. 8). The hornblende is usually subhedral and is strongly pleochroic (X = dark green, Y = yellow-green to pale green, Z = blue-green). The angle between *c* and Z ranges from 9° to 25°. Large crystals are commonly poikilitic and include blebs of quartz and lesser amounts of feldspar, magnetite, ilmenite, pyrite, and apatite. Opaque needles of rutile and crystals of magnetite(?) are often symmetrically arranged along cleavage planes and at right angles to the prismatic cleavage. Basal sections show lines of opaque inclusions arranged at right angles to each other. Plagioclase is the only other abundant mineral in the amphibolite. It is generally calcic oligoclase and only rarely albite. The plagioclase commonly shows some sericitic alteration, and locally is completely replaced by clinozoisite and epidote.

The amphibolite also contains biotite, quartz, apatite, sphene, allanite, zircon, pyrite, ilmenite, and magnetite. Narrow veins and fractures cut all the abundant minerals of the amphibolite. The veins contain quartz, albite(?), epidote, and carbonate.

The modes of 10 specimens of amphibolite and the average of 20 modal analyses are presented in table 19. The variation in mineralogy probably results from the differences in the original composition and subsequent alteration and replacement. For example, in some amphibolite, biotite replaced hornblende, and both of these altered to chlorite; plagioclase altered to sericite, epidote, and clinozoisite. Much of the abundant quartz of specimens 13056 and 25656 (table 19) was probably introduced during the formation and emplacement of the gneiss.

A chemical analysis of an amphibolite inclusion from the complex is presented in table 20 together with the analysis of an amphibolite from the Felch complex and the average analysis of the Hawaiian and Indian basalts.



A *B*

FIGURE 8.—Photomicrographs of amphibolite from inclusion in the Carney Lake Gneiss. *A*, Plane-polarized light; × 60. *B*, Crossed nicols; × 60. pl, plagioclase; H, hornblende; Bio, biotite. Chemical analysis of material from same inclusion shown on table 20.

Biotite schist

Biotite schist inclusions are abundant in the gneiss south of the correction line between T. 40 N. and 41 N. (fig. 4). Easily accessible outcrops which show inclusions of schist are located near the southern boundary of the complex. These inclusions occur in much the same manner as the amphibolite, but they also occur as thin sheets or dikelike bodies that send branches into the gneiss and also are intruded by the gneiss.

The biotite schist ranges from a very fine grained greenish-black dense rock to a medium-grained black or green rock. Fresh surfaces of the finer grained varieties exhibit widely spaced plates of black biotite and specks of bright-yellow pyrite set in a green-black matrix of finer biotite and chlorite; coarser varieties show abundant alined foils and plates of shiny-black biotite which swirl around aggregates or individual crystals of altered feldspar or lenses of gray quartz. Thin sections show abundant biotite and lesser amounts of plagioclase. The biotite occurs as intergrown subparallel laths which form a matrix in which the other minerals are set, and it contains inclusions of quartz, magnetite, epidote, apatite, rutile, sphene, and zircon. Both zircon and apatite exhibit haloes. The plagioclase is most commonly oligoclase, but sodic andesine and albite are present in some specimens. Generally the plagioclase shows incipient to nearly complete sericitic or saussuritic alteration. Veins and fracture fillings of quartz, quartz-epidote, quartz-carbonate, and carbonate are abundant and cut all the major minerals of the biotite schist.

TABLE 19.—*Modes of amphibolites from Carney Lake Gneiss*

Mineral	Specimen										
	7357	3257	257	1156	13056	33056	25656	157	10357	7657	(1)
Plagioclase	30.2	27.8	3.9	33.5	13.3	32.8	20.2	11.2	21.3	24.5	17.2
Hornblende	59.7	41.8	60.9	45.6	64.7	49.0	57.8	42.1	46.7	58.5	53.8
Biotite	6.4	3.7			8.3	Tr.	Tr.	6.5	12.3	4.9	3.8
Apatite	.6	Tr.	1.5	3.7	3.1	3.5	1.9	2.8	.6	Tr.	1.9
Opaque minerals [2]	.7	Tr.	2.1	2.5	7.2	6.0	6.5	5.6	8.2	1.9	4.3
Quartz	1.3	Tr.	2.3	3.0	11.3	5.2	10.9		8.4	9.7	4.8
Epidote-clinozoisite	.9	18.5	12.4	19.2		Tr.		23.3	2.5	Tr.	10.1
Sphene	.4	Tr.	3.8	1.5	2.1	.9	Tr.	8.4		.5	2.4
Carbonate		3.6	.8								.2
Sericite		4.6	Tr.								.3
Chlorite			12.3			2.6	2.7			Tr.	1.1
Allanite				Tr.							Tr.
Zircon					Tr.		Tr.				Tr.

[1] Average of 20 specimens.
[2] Magnetite, ilmenite, and pyrite.

Location of specimen:

7357. NE¼ sec. 26, T. 41 N., R. 29 W.
3257. SE¼ sec. 30, T. 41 N., R. 28 W.
257. SE¼ sec. 11, T. 40 N., R. 29 W.
1156. SW¼ sec. 26, T. 40 N., R. 29 W.
13056. NW¼ sec. 15, T. 40 N., R. 29 W.
33056. NW¼ sec. 13, T. 40 N., R. 29 W.
25656. NW¼ sec. 19, T. 40 N., R. 28 W.
157. SE¼ sec. 11, T. 40 N., R. 29 W.
10357. NE¼ sec. 28, T. 41 N., R. 29 W.
7657. SE¼ sec. 26, T. 41 N., R. 29 W.

TABLE 20.—*Chemical analyses of amphibolites from the Carney Lake Gneiss and averages of basalt analyses*

Constituent	1	2	3	4
SiO_2	50.36	49.26	49.58	50.61
Al_2O_3	13.26	12.91	13.19	13.58
Fe_2O_3	6.30	1.95	2.04	3.19
FeO	9.34	9.10	9.49	9.92
MgO	5.55	10.67	8.30	5.46
CaO	7.85	11.26	10.69	9.45
Na_2O	2.11	1.37	2.25	2.60
K_2O	1.14	.57	.55	.72
H_2O^-	.16	.06	------	} 2.13
H_2O^+	1.55	1.44	------	
TiO_2	1.77	.49	3.17	1.91
P_2O_5	.20	.05	.26	.39
CO_2	------	.08	------	------
MnO	Tr.	.26	.12	.16
Total	99.59	99.47	99.64	100.12

1. Amphibolite, SE¼ sec. 23, T. 42 N., R. 28 W. (after Clements and Smyth, 1899b, p. 397).
2. Amphibolite, NE¼ sec. 26, T. 41 N., R. 29 W. (this report).
3. Average of 11 olivine basalt analyses, Hawaii (after Daly, 1944, p. 1365).
4. Average of 11 tholeiitic basalt analyses (after Washington, 1922, p. 774).

The modes of 10 specimens of biotite schist and the average of 15 modal analyses are presented in table 21. Hornblende-bearing varieties were deliberately omitted because they probably represent transitional rocks. They are similar to the biotite-rich amphibolites but contain more biotite than hornblende.

A chemical analysis of a biotite schist inclusion is given in table 22.

Metasedimentary rocks

Inclusions of metasedimentary rocks are not abundant, nor do they appear to be systematically distributed throughout the gneiss. They are, however, clearly xenoliths and therefore are probably representative of some of the rocks which once covered this area.

Quartzite, sericitic quartzite, and quartz-biotite schist inclusions have been mapped. The inclusions of quartzite and sericitic quartzite are rare. An inclusion of gray quartzite about 1 foot in diameter occurs in the NE¼SE¼ sec. 1, T. 40 N., R. 30 W. The quartz is strained and recrystallized. Inclusions of sericitic greenish-white quartzite are small and rare. One occurs in the SW¼ sec. 35, T. 40 N., R. 29 W. The quartz in this inclusion is strained, somewhat recrystallized, and set in a matrix of fine-grained sericite.

Inclusions of quartz-biotite schist are present in most sections. They are generally small, but near the southeast corner of the NE¼NE¼ sec. 34, T. 40 N., R. 29 W., an inclusion about 10 feet long and 4 feet wide occurs in the gneiss. The schist is light gray and fine grained. Thin sections show marked schistosity and abundant quartz, plagioclase, and biotite. Muscovite, epidote, magnetite, microcline, zircon, and carbonate are rare constituents. The muscovite and microcline appear to replace other minerals and possibly represent the addition of material during the emplacement of the gneiss.

A modal analysis of a specimen from the SE¼ sec. 23, T. 40 N., R. 29 W., is shown in table 23.

Origin of the inclusions

The original material from which the inclusions of the metasedimentary rocks were formed is clear. These rocks are obviously remnants of sandstone, impure sandstone, and shale or graywacke, or of quartzite,

TABLE 22.—*Chemical analysis of biotite schist inclusion from Carney Lake Gneiss*

[Specimen from SE¼ sec. 1, T. 40 N., R. 29 W.]

Constituent	*Percent*	*Constituent*	*Percent*
SiO_2	49.75	K_2O	4.61
Al_2O_3	13.20	H_2O+	1.71
Fe_2O_3	7.08	H_2O-	.03
FeO	9.26	TiO_2	2.54
MgO	5.15	P_2O_5	.33
CaO	3.71	MnO	.17
NaO_2	1.76	CO_2	.08
		Total	99.38

TABLE 21.—*Modes of biotite schist inclusions in Carney Lake Gneiss*

Mineral	Specimen										
	28756	27156	18655	8255	6255	2255	25955	26055	17055	100657	(1)
Biotite	43.8	47.2	73.7	61.5	52.1	58.2	27.1	43.1	51.2	41.4	49.2
Plagioclase	7.4	22.1	Tr.	4.3	11.7	6.3	41.8	40.8	3.2	25.4	16.3
Quartz	9.1	6.3	3.0	7.0	11.5	5.1	5.0	5.0	8.5	7.1	10.2
Epidote-clinozoisite	12.4	17.4	------	11.2	17.2	19.2	7.5	2.5	17.2	10.5	9.7
Opaque minerals [2]	6.1	Tr.	11.5	1.3	2.3	1.2	16.3	3.4	Tr.	8.3	3.5
Carbonate	5.7	3.4	8.2	6.3	------	3.3	Tr.	3.5	14.5	Tr.	5.5
Muscovite	7.3	------	3.6	Tr.	------	------	Tr.	Tr.	5.4	Tr.	1.5
Chlorite	8.2	Tr.	------	9.4	Tr.	------	7.3	2.7	Tr.	7.3	3.0
Sphene	------	2.5	------	Tr.	2.6	3.2	Tr.	Tr.	------	Tr.	.7
Apatite	------	1.1	------	Tr.	2.6	1.7	Tr.	Tr.	------	Tr.	.4
Allanite	------	Tr.	------	------	Tr.	Tr.	------	------	------	------	Tr.
Zircon	------	------	------	------	------	Tr.	------	------	Tr.	------	Tr.
Actinolite	------	------	------	------	------	1.8	------	------	------	------	Tr.
Rutile	------	------	------	------	------	------	------	Tr.	------	------	Tr.
Leucoxene	------	------	------	------	------	------	------	Tr.	------	------	Tr.

[1] Average of 15 specimens.
[2] Magnetite, ilmenite, and pyrite.

Location of specimen:

28756. SE¼ sec. 27, T. 40 N., R. 29 W.
27156. SE¼ sec. 6, T. 40 N., R. 28 W.
18655. NW¼ sec. 28, T. 40 N., R. 29 W.
8255. SW¼ sec. 35, T. 40 N., R. 29 W.
6255. SE¼ sec. 36, T. 40 N., R. 29 W.
2255. SW¼ sec. 36, T. 40 N., R. 29 W.
25955. NW¼ sec. 8, T. 39 N., R. 28 W.
26055. NW¼ sec. 8, T. 39 N., R. 28 W.
17055. SW¼ sec. 22, T. 40 N., R. 29 W.
100657. SE¼ sec. 1, T. 40 N., R. 29 W.

TABLE 23.—*Mode of quartz-biotite schist inclusion in Carney Lake Gneiss*

[Specimen from SE¼ sec. 23, T. 40 N., R. 29 W.]

Mineral	*Percent*	*Mineral*	*Percent*
Quartz	35.3	Epidote	4.9
Biotite	30.2	Microcline	3.4
Plagioclase (An_{12})	16.9	Carbonate	Tr.
Muscovite	7.2	Zircon	Tr.
Magnetite	2.1		

sericitic quartzite, and quartz-biotite schist which once covered this area.

The original nature of the inclusions of biotite schist and amphibolite is not readily apparent. It is best indicated by the field relations, petrographic descriptions, and chemical analyses.

Inclusions of amphibolite and biotite exhibit both sharp and gradational borders with the gneiss. In the NW¼SW¼ sec. 25, T. 41 N., R. 29 W., the amphibolite grades into hornblende gneiss within a distance of 3 feet. In the NE cor. SE¼NW¼ sec. 35, T. 41 N., R. 29 W., an outcrop of biotite schist is cut by pegmatite. Approximately in the center of the outcrop, small pods of amphibolite less than 4 inches in diameter in the biotite schist grade into the schist over a distance of about one-half inch. Although hornblende is present in several of the specimens of biotite schist and biotite is a common accessory in several of the specimens of amphibolite, it is only in the above-mentioned outcrop that it can be clearly demonstrated that the amphibolite is an unreplaced remnant and that the biotite schist was formerly amphibolite.

The chemical analyses and textures of both the amphibolite and biotite schist (tables 20 and 22) indicate that the original rock from which they were derived might have been basalt which reacted differentially and in steps with invading granitic material which eventually gave rise to the Carney Lake Gneiss. The distribution of inclusions in the gneiss suggests that these xenoliths have been metamorphosed at different levels within the gneiss.

Associated middle and upper Precambrian intrusive rocks

The lower Precambrian Carney Lake Gneiss is cut by metadiabase dikes, pegmatites, and unaltered diabase dikes. The metadiabase dikes are clearly of late middle Precambrian age, as they cut both Animikie strata and the gneiss of the complex. They are fully described on page 65.

The pegmatites that are here tentatively assigned a middle Precambrian age are unlike the pegmatites associated with the composite gneiss both in appearance and mineralogy. They occur not only as dikes cutting the gneiss, but also as veins cutting the metadiabase dikes. Further, they exhibit sharp contacts against the gneiss. The pegmatite dikes are composed of large crystals of pink potassium feldspar, white platy plagioclase, thick tablets of biotite, and minor amounts of muscovite. The veins consist of abundant quartz and lesser amounts of extensively altered white feldspar, and muscovite. An outcrop which shows one of the pegmatite dikes is in the SW¼NE¼ sec. 1, T. 39 N., R. 29 W., and an easily accessible outcrop which shows a pegmatitic vein cutting a metadiabase dike is in the SE¼SE¼ sec. 23, T. 40 N., R. 29 W.

Unaltered diabase dikes occur in a few outcrops along the backwaters of the Sturgeon River. The dikes are narrow and short and consist of very fine grained dense black basalt and fine-grained diabase. They are composed of labradorite, augite, and olivine and are similar to other diabase dikes that occur along this area. They are clearly later than the post-Animikie (late middle Precambrian) period of metamorphism and are therefore classified as upper Precambrian (Keweenawan) rocks in this part of northern Michigan.

Smaller structural features of the complex

The gneisses of the complex are strongly foliated and are broken by joints that appear to be related to the foliation; the attitudes of the joints are shown on the geologic maps (pls. 2, 3). Lineation is not well shown by the rocks of the complex, but where present it tends to be almost horizontal in the plane of the foliation. Faults were observed only at the borders of the complex, where they cut middle Precambrian rocks.

Statistical analysis shows that the dominant direction of strike of the foliation is to the northwest and the dip is to the northeast at intermediate angles. This northwest trend is very clearly shown by the alinement of inclusions in the southern half of the complex (fig. 7). In the extreme northern exposures of the complex the alinement of inclusions is almost east-west, and in the center of the complex it is somewhat chaotic but tends to be dominantly to the northwest. The foliation in general and the alinement of the inclusions is subparallel to the contact of the complex with the Animikie strata.

Joints are conspicuous features of the gneisses of the complex. Analysis indicates that the major trend is northwest, parallel to the major trend of the foliation. These longitudinal joints are common, and characteristically they dip more steeply than does the foliation. Cross and diagonal joints are also present but not common, and flat-lying joints are rare.

AGE OF THE CARNEY LAKE GNEISS

The stratigraphic relations and dates determined from radioactive constituents of the gneiss (Davis and others, 1960) indicate that the Carney Lake

Gneiss, granodiorite dikes, and syenite are all probably of early Precambrian age. Along the southern and western borders, the rocks of this complex are unconformably overlain by the lower Animikie (middle Precambrian) Sturgeon Quartzite; to the north they are in fault contact with or are unconformably overlain by the upper Animikie Michigamme Slate; in sec. 6, T. 40 N., R. 28 W., the gneiss is overlain by quartzite which is probably correlative with the Sturgeon Quartzite. The following absolute ages have been determined from minerals in a sample of the Carney Lake Gneiss (Davis and others, 1960, p. 153–154):

Mineral	*Age (million years)*
Biotite	1,380 (Rb:Sr)
Biotite	1,360 (K:Ar)
Microcline	2,300 (Rb:Sr)
Zircon	1,735 ($U^{238}:Pb^{206}$)
Zircon	2,160 ($U^{235}:Pb^{207}$)
Zircon	2,590 ($Pb^{207}:Pb^{206}$)
Zircon	1,240 ($Th^{232}:Pb^{208}$)

From these results, and others for rocks of the adjacent areas of Michigan and Wisconsin, after making proper allowance for the probable lowering of apparent age as a result of loss of daughter products by diffusion, and other factors, Davis and others concluded (1960, p. 154) that the Carney Lake Gneiss is about 2,700 million years old.

AGE AND ORIGIN OF THE GNEISSES AND GRANITIC ROCKS

The concept of the age and origin of the gneisses and granitic rocks, as here presented, is based on an interpretation of field relations and on petrographic and chemical data.

The field relations show that (1) the gray gneiss grades into composite gneiss, into inclusions of amphibolite, and, more rarely, into inclusions of biotite schist, (2) this gneiss exhibits sharp contacts against the inclusions and appears as dikes in them, (3) the composite gneiss grades into red gneiss and into biotite schist, and (4) both the composite gneiss and the red gneiss contain inclusions of biotite schist and occur as dikes and stringers in some of the inclusions. Further, the gneisses are cut by red granodiorite dikes; one of these dikes, in turn, is cut by a late middle Precambrian metadiabase dike, and another metadiabase dike contains an inclusion of granodiorite. The relations of the syenite, which is known only in the southeast corner of the complex, and the grandiorite dikes are not clear, but the lack of foliation of the granodiorite dikes and the slight foliation of the syenite may indicate that the syenite was emplaced before the granodiorite dikes.

Mineralogical changes that accompany the transition of amphibolite into gray gneiss, of gray into composite gneiss, and finally of composite gneiss into red gneiss are well known by petrographic study of hand specimens and thin sections. The change of amphibolite into gray gneiss is marked by (1) the conversion of some hornblende into biotite and the appearance of abundant oligoclase, (2) an increase in amounts of biotite and oligoclase, accompanied by some quartz, and (3) the appearance of quartz and small amounts of albite and microcline that form discontinuous lenses and layers that bifurcate and enclose lenses of plagioclase and biotite and mark the final stage in the conversion to gray gneiss. The change from gray gneiss to composite gneiss is marked by a considerable increase in the quartz-rich parts of the rock, the presence of more microcline, the alteration of plagioclase to sericite, muscovite, and clinozoisite, and the replacement of plagioclase by quartz, albitic oligoclase, and microcline. These changes give rise to the formation of rock composed of two parts, an older one composed chiefly of plagioclase (some of it extensively altered) and biotite, and a younger part composed chiefly of quartz, microcline, and albite. The rock is then a typical composite gneiss. Similarly, the change from composite gneiss into red gneiss is marked by increased amounts of quartz and microcline, accompanied by moderate amounts of albite and muscovite. All these changes are shown not only by thin sections and hand specimens, but also by outcrops, some of which show mixing of components on a grand scale.

A study of the chemical analyses of the amphibolite, gray gneiss, and composite gneiss (tables 13 and 20) shows the chemical changes that must have accompanied these transitions into various types of gneiss. The most significant chemical aspects of the transition are presented in table 24. This table shows that, in terms of bulk chemistry, the conversion of amphibolite into gray gneiss was accompanied by a large increase in SiO_2, considerable increase in Na_2O and K_2O, and large decrease in ferric and ferrous iron oxides, MgO and CaO. Mineralogically these changes are reflected in the change from calcic oligoclase and hornblende in the amphibolite to more sodic

TABLE 24.—*Chemical changes involved in transition of amphibolite into gneisses, in volume percent*

[From tables 12, 16, and 19]

Constituent	Amphibolite	Gray gneiss	Composite gneiss
SiO_2	49.26	72.95	73.26
Fe_2O_3	1.95	.34	.37
FeO	9.10	2.12	1.64
MgO	10.67	.79	.63
CaO	11.26	3.00	1.35
Na_2O	1.37	3.78	2.73
K_2O	.57	1.34	4.68

oligoclase and minor albite, biotite, orthoclase and minor microcline, and quartz in gray gneiss. The chemical changes that accompanied the conversion of the gray gneiss into the composite gneiss consist chiefly of a slight increase in SiO_2, a marked increase in K_2O, a slight decrease in combined ferric and ferrous iron oxide and in MgO and a somewhat greater decrease in CaO and Na_2O. These changes again are in harmony with the mineralogical changes, the large increase in K_2O reflects the increase in sericite and muscovite, and the moderate decreases in CaO and Na_2O reflect the destruction of more plagioclase. Although no chemical analysis of the red gneiss is available, the modal analysis of that rock (table 15) indicates that its formation from the composite gneiss would be accompanied by an increase in K_2O and probably also in SiO_2. The chemical analyses, therefore, indicate the same changes as those observed in thin sections, hand specimens, and outcrops.

The gneisses of the complex are obviously mixed rocks. They were probably formed during the invasion of a chiefly basaltic terrain by granitic magma. The gray gneiss represents a phase that was formed by magmatic assimilation of or reaction with an older mixed series of basaltic and sedimentary rocks which are now represented by inclusions in the gneiss. Further invasion of this early phase by the same magma, or perhaps a later pegmatitic magma or differentiate, formed the composite and red gneisses. Still later syenitic and granodioritic differentiates intruded the earlier formed gneisses. That such reactions are possible when a granitic magma invades a mixed series of sedimentary and volcanic rocks has been demonstrated by the work of Bowen (1922; 1928, p. 175–223) and by Nockolds (1933, 1934, and 1935). The origin of the granitic magma, whether by anatexis or syntexis or by differentiation of basalt, is not indicated by the field evidence.

MIDDLE PRECAMBRIAN SEDIMENTARY ROCKS

ANIMIKIE SERIES

The middle Precambrian is represented in northern Michigan by a thick sequence of mostly sedimentary but partly volcanic rocks which constitutes the Animikie Series. The series has been divided into four groups, which are, from oldest to youngest, the Chocolay, Menominee, Baraga, and Paint River (James, 1958). The Chocolay, the Menominee, and the combined Baraga and Paint River Groups correspond respectively to the Lower, Middle, and Upper Huronian of previous reports (Leith and others, 1935). In the Menominee district the middle Precambrian rocks occupy the central troughlike area, which trends northwestward between the two areas of lower Precambrian rocks. This central trough includes about 60 square miles, within which outcrops are few and widely scattered. The Paint River Group has not been recognized in this district, but rocks of equal age may have been included in the Baraga Group.

The subdivisions of the groups, as recognized in the Menominee district, and their correlatives in central Dickinson County and the Marquette district, are shown on table 25.

CHOCOLAY GROUP

The Chocolay Group is equivalent to the Lower Huronian of past literature (Leith and others, 1935). It is made up, in ascending order, of the Fern Creek Formation, the Sturgeon Quartzite, and the Randville Dolomite. The group rests with profound unconformity on the Carney Lake Gneiss and is separated from the overlying Menominee Group by an unconformity of uncertain magnitude. The name of the group refers to Mount Chocolay near Marquette, Mich., where the Mesnard Quartzite and Kona Dolomite are extensively exposed (James, 1958). The Sturgeon and Randville Formations of the Menominee district are

TABLE 25.—*Subdivisions of the Animikie Series*

Group	Menominee district	Central Dickinson County	Marquette district
Baraga ("Upper Huronian," in part)	Badwater Greenstone Michigamme Slate	Badwater Greenstone Michigamme Slate Hemlock Formation	Michigamme Slate, including Clarksburg Volcanics Member Goodrich Quartzite
	Unconformity	Unconformity	Unconformity
Menominee ("Middle Huronian")	Vulcan Iron-Formation Felch Formation	Vulcan Iron-Formation Felch Formation	Negaunee Iron-Formation Siamo Slate Ajibik Quartzite
	Unconformity	Unconformity	Unconformity
Chocolay ("Lower Huronian")	Randville Dolomite Sturgeon Quartzite Fern Creek Formation	Randville Dolomite Sturgeon Quartzite Fern Creek Formation	Kona Dolomite Mesnard Quartzite
	Unconformity	Unconformity	Unconformity

virtually identical to the two formations at Mount Chocolay and are presumed to be directly correlative.

FERN CREEK FORMATION

General features

The contact between the middle Precambrian rocks and the underlying Carney Lake Gneiss is exposed in but three places. These are (1) at the Sturgeon River Dam in sec. 8, T. 39 N., R. 28 W., (2) on Black Creek near the S¼ cor. sec. 6, T. 39 N., R. 28 W., and (3) on Fern Creek in sec. 34, T. 40 N., R. 29 W. At each of these exposures the basal strata are arkosic or graywacke conglomerate, arkose, and argillite. The conglomerates rest upon the truncated folia of the Carney Lake Gneiss and contain randomly oriented pebbles and boulders of similar gneiss; thus they show a clear-cut unconformable relation of the first magnitude.

These basal Animikie rocks constitute a lithologic assemblage easily distinguishable from the overlying vitreous quartzite of the Sturgeon Quartzite. At the Sturgeon River site, exposures on the west side of the river exhibit a transition, by intermixing and interbedding, from the arkosic material which makes up the bulk of the basal rocks into the clean well-sorted quartz sands that characterize the Sturgeon Quartzite.

Brooks stated in 1880 (see Bayley, 1904, p. 66) that the conglomerates and arkosic rocks exposed along the Sturgeon River represent a basal phase of the Sturgeon Quartzite, and most geologists who have subsequently visited these exposures have agreed with this conclusion. Pettijohn (1943) cast some doubt on this early interpretation by stating that some of the graywacke conglomerate was probably tillite and that a closely associated laminated argillite containing scattered (rafted) cobbles was probably also of glacial origin. Pettijohn concluded that the evidence for glacial origin was compelling enough to warrant setting off these basal rocks as a separate unit, which he designated Fern Creek Formation. More recently Trow (1948), after studying and mapping the basal material and the quartzite, applied the name Fern Creek to all the conglomeratic and feldspathic strata lying below the typical Sturgeon Quartzite.

The map distribution of the Fern Creek strata (pls. 2, 3) was taken from Trow's outcrop maps that were adjusted to fit the topographic quadrangle maps. This compilation required considerable field observation, but no critical resurvey of the area was attempted. The distribution shows that the Fern Creek rocks are chiefly confined to areas that apparently were valleys on the pre-Animikie erosion surface. The Fern Creek Formation may also underlie the covered low ground between the Sturgeon Quartzite ridge and the gneiss front, but these areas could equally well be underlain by the soft sericitic basal part of the Sturgeon Quartzite, which is exposed near the gneiss in this vicinity.

Description

Conglomerates make up the bulk of the formation at most places. Some contain matrices of arkose, and some of graywacke. Massive and schistone arkose, massive and laminated argillite, sericitic schist, and quartzite are present also.

The maximum thickness of the formation occurs at the Sturgeon River site, where Trow (1948) measured 260 feet of composite section distributed on both sides of the river. That section in abbreviated form follows. Note that the transition to Sturgeon Quartzite takes place in the upper 80 feet of the formation, that few, if any, pebbles occur in the upper half, and that most of the conglomerate is in the lower 90 feet.

	Feet
Sturgeon Quartzite: gritty, green, sericitic quartzite.	
Fern Creek Formation:	
1. Quartzite, reddish-gray, feldspathic	10
2. Quartzite, mottled green and reddish-gray, slaty	13
3. Quartzite, buff to reddish-gray, arkosic	4
4. Quartzite, greenish-gray and dark-green, laminated, slaty	53
5. Arkose, red	32
6. Arkose and silvery-gray schist in alternating beds; arkose layers poorly graded	17
7. Arkose, pink; contains gneiss and quartz pebbles	8½
8. Same as unit 6	5½
9. Slate, laminated; graded bedding	1½
10. Graywacke, red and gray, laminated; graded bedding	30
11. Graywacke conglomerate; 40–60 percent cobbles and pebbles of gneiss and vein quartz	15
12. Graywacke, gray, chloritic, laminated	9
13. Graywacke, gray and red, feldspathic, poorly laminated; contains scattered boulders as much as 9 in. in diameter	17
14. Graywacke conglomerate, 80 percent boulders and cobbles of gneiss and vein quartz	9
15. Graywacke, red and gray, feldspathic; contains scattered cobbles of gneiss	8
16. Graywacke and arkosic conglomerate, 50 percent cobbles and boulders of gneiss and vein quartz, angular blocks near base	28
Total	260½
Unconformity.	
Carney Lake Gneiss.	

The preceding section represents the thickest known sequence of Fern Creek strata. Elsewhere the close proximity of the Sturgeon Quartzite to the gneiss indicates that if the Fern Creek is present at all, its average thickness cannot be greater than about 100 feet. At Browning Creek, sec. 20, T. 41 N., R. 30 W., central Dickinson County, the Fern Creek is not present, and the sericitic basal part of the Sturgeon Quartzite is separated from the subjacent gneiss by a mere wisp of quartz pebble conglomerate (Pettijohn, 1943).

The best available petrographic descriptions of the Fern Creek rocks are those of W. S. Bayley (1904, p. 179–182) quoted below.

Conglomerates

The conglomerates occur only in a few places very near the granite-schist complex * * * In all these places the rock is composed of gneiss, granite, and quartz pebbles and boulders in the groundmass that is sometimes an arkose, sometimes a graywacke, and occasionally a quartzite. The fragments vary in size from a few inches to a foot and a half in diameter. Some of them are well rounded, while others are sharp edged. The matrix in which they are embedded is sometimes a white quartzite, but usually it is dark colored, being either dark red or dark gray, according to the predominance in it of red feldspar grains or of chloritic or micaceous particles derived from the decomposition of this mineral. The darker-colored matrices are also schistose, the schistosity being generally parallel to their contacts with the granite-gneiss complex.

The character of the conglomerates varies mainly with respect to the groundmass. This may have the composition of a quartzite, an arkose, or a graywacke, and it may either be massive or schistose. The conglomerates with a quartzite matrix are usually massive in structure, while those with an arkose or graywacke matrix are always more or less schistose. The included pebbles are in all instances fragments of gneiss, granite, basic schists, and vein quartz, identical with the corresponding rocks in the Archean complex.

The quartzite conglomerates are simply quartzites, like the normal rock of the formation described below, containing pebbles and boulders of granite, vein quartz, and crystalline schists, together with irregular grains of the individual components of the first-named rock. . . .

The larger pebbles in these conglomerates possess their original character. The smaller ones, however, show plainly the effects of mashing. The quartz pebbles are crushed into fragments, which are sometimes separated from one another by portions of the groundmass, but which are as frequently aggregated into groups of variously oriented grains, differing in outline from the original pebbles in being much elongated, often to the extent of becoming almost vein-like. The particles of the aggregates are crossed by strain shadows, and the whole appearance of the quartz suggests pressure phenomena.

The smaller granite and gneiss pebbles are likewise shattered. Their individual components occur as sharpedged fragments scattered through the matrix at intervals. The plagioclase and the microcline are still fresh, but the orthoclase has been changed to a micaceous aggregate of secondary products, in which particles of muscovite, kaolin, chlorite, and quartz may be detected.

The matrix surrounding the pebbles is composed of isolated grains of quartz, microcline, and plagioclase in an extremely fine-grained groundmass that has nearly the composition of an altered feldspar. In the darker-colored phases of the rocks there is in addition to the usual decomposition products of feldspar, quite a little chlorite, and other greenish alteration products of biotite. All these components are arranged in a rudely parallel direction, which is approximately parallel to the direction of the elongation of the crushed quartz fragments derived from the quartz pebbles.

The coarser grains of the matrix evidently represent sand grains in a sediment that consisted largely of the material from which the finer portion of the groundmass was produced. This must have been a mud or silt which was made up of the finer detritus of the Archean rocks, and which was therefore comparatively rich in ferromagnesian components.

Arkoses and Graywackes

The conglomerates pass rapidly into nonconglomerate beds, whose composition is like that of the conglomerate groundmass. In some places the nonconglomerate rock is a red massive or schistose arkose; in other places it is a dark-gray schistose graywacke; and in still other places, it is a massive or schistose light-gray quartzite, according as the matrix of the conglomerate associated with it is coarse or fine, feldspathic or quartzitic, massive or schistose. These rocks are interbedded with the conglomerates in comparatively thin layers. They are more common in the upper portion of the conglomerate beds than at lower horizons and constitute gradation phases between the conglomerates and the typical quartzites * * *.

There are very few of the arkoses that are not schistose. Their feldspathic constituent furnished abundant material for the production of micaceous products, and these, under the influence of shearing stresses, readily took on the parallel arrangement which gave rise to the schistose structure. In their present condition they are feldspathic sericite-schists. The few massive arkoses met with appear as thin beds of fairly coarse-grained pink rocks lying between thicker beds of quartzite. The more purely quartzite phases contained but little feldspar; consequently there was less solution of their constituents and therefore a smaller production of mica. These rocks are therefore only rarely schistose, except along their shearing zones bounding joint cracks along which readjustment took place during folding.

In thin section the arkoses are seen to be composed of quartz grains and fragments of orthoclase, various plagioclases, microline, and microperthite, in a matrix of finer grains of the same substances colored red by iron oxides. Magnetite crystals, clumps of rutile, an occasional crystal of zircon, and a few grains of epidote are the only other constituents noticeable, except certain micaceous decomposition products of the feldspars. These are especially abundant in the finest-grained portions of the matrix, which in many places is composed almost exclusively of kaolin, sericite, and quartz. The larger feldspar grains are often quite fresh, the plagioclases and the microcline being almost devoid of alteration products of any kind. The orthoclase, on the contrary, is usually much altered, even in the largest grains. None of the grains are as completely waterworn as those in the typical quartzite. They are usually much more angular than the latter, like grains of sand that have not traveled far from the coast along which they were formed.

The graywackes are darker than the arkoses. They are always schistose and fine grained. The schistosity can be seen to arise from the parallel arrangement of tiny shreds and plates of chlorite and some light-colored mica, and in all cases where it has been carefully examined it is parallel to the bedding. The graywackes are always finer grained than the arkoses. A few large quartz grains are scattered through them, but the main portion of the rocks consists of small grains of quartz and decomposed feldspars, spicules of muscovite, chlorite, and green biotite, crystals and grains of magnetite, little nests of calcite, and occasionally a small prism of tourmaline. Much of the quartz of the groundmass seems to be secondary, as it is in the form of interlocking areas traversed by the micaceous spicules.

The schistose structure is produced in part by the parallel arrangement of the chloritic and micaceous minerals and in part by original sedimentation, for there is often noticeable in the sections alternating layers of finer- and coarser-grained components and layers containing more or less of the quartz and feldspathic constituents.

A few specimens collected from the Fern Creek and Black Creek sites during the present study show in general the features described by W. S Bayley, but secondary greenish-brown biotite is probably a more abundant constituent than he implied. The plagioclase, particularly in the graywacke-type rocks, is albite, although originally it may have been more calcic. The albite appears unaltered, whereas the potassic feldspars including microcline are partly or completely replaced by sericite.

Structure

The Fern Creek beds are nearly vertical, and some may be slightly overturned, but in general they dip steeply away from the Carney Lake Gneiss. Many of the beds near the contact with the gneiss are extremely schistose and in places they are plicated, but some are more massive. Some of the fine-grained rocks show an oblique slaty cleavage. The unconformable contact of the conglomerate with the gneiss is generally tight and unsheared. The attitude of the Fern Creek rocks apparently results chiefly from an upward and outward thrust of the gneiss, and the nature of the contact with the underlying gneiss indicates that the primary foliation of the gneiss has been tilted (Pettijohn, 1943, p. 391). Apparently the margins of the gneissic complex were rotated outward, and much of the differential movement between the structurally competent rocks—the gneiss and the vitreous Sturgeon quartzite—was taken up by shearing within the Fern Creek rocks and the sericitic rocks near the base of the Sturgeon Quartzite.

The Fern Creek beds are disrupted by a longitudinal fault in the NW¼ sec. 8, T. 39 N., R. 28 W., along which there has been thrusting from the north. More faults of this type are thought to be present, but only the one above is certainly known. The formation is offset by many diagonal, normal or tear faults, along some of which altered diabase dikes have been injected.

Numerous small quartz veins and pegmatite dikes cut the Fern Creek rocks, but some pegmatite dikes that cut the Carney Lake Gneiss are truncated at the Fern Creek-Carney Lake contact.

Conditions of deposition

The Fern Creek strata are composed chiefly of clastic materials derived from the Carney Lake Gneiss and contiguous rocks. In gross aspect these strata appear to represent a conglomeratic basal phase of the marine Sturgeon Quartzite. Their isolation in restricted areas between the gneiss and the Sturgeon apparently results from relief on the pre-Animikie erosion surface.

The considerable variation in lithologic types within the formation indicates that the conditions of deposition changed from time to time. The poorly sorted conglomerates in the lower part of the formation suggest fluviatile deposition, which was at times torrential; the graywacke and slate suggest deposition in quiet waters that were lagoonal or estuarine; and the even-bedded quartzite and arkose in the upper part of the formation, which appear to grade upward into the Sturgeon Quartzite, suggest deposition in a nearshore marine environment.

STURGEON QUARTZITE

General features

At most places in the Menominee district, and in the Felch and Marquette districts, a thick formation composed of light-colored vitreous quartzite rests unconformably on the lower Precambrian ("Archean") rocks and is overlain by a dolomite formation. The stratigraphic equivalence of the quartzite and dolomite couplet in these several areas has been a tenet of the regional geology for more than 50 years. The quartzite formation is called Mesnard Quartzite, and the dolomite formation is called Kona Dolomite in the Marquette district, and Sturgeon Quartzite and Randville Dolomite, respectively, in the Felch and Menominee districts. The type exposures of Sturgeon Quartzite are along the Sturgeon River in the Felch district (Clements and Smyth, 1899b, p. 398).

In the mapped area the Sturgeon Quartzite is exposed in a nearly continuous belt along the northwest margin of the Carney Lake Gneiss (pls. 2, 3). The quartzite belt extends beyond the mapped area to the northwest into central Dickinson County where it has been described by James and others (1961, p. 31–33). To the southeast the quartzite is overlapped by Paleozoic rocks. Along the north margin of the Carney Lake Gneiss (pl. 3) an erosional remnant of Sturgeon Quartzite lies between the gneiss and a broad area of no outcrop which we infer is underlain by Michigamme Slate. An isolated area of quartzite and quartz schist in sec. 24, T. 40 N., R. 31 W. (pl. 1), seems to represent an anticlinal node of Sturgeon Quartzite, but correlation with the Sturgeon is uncertain.

The faulting and secondary folding of the Animikie rocks in the interior of the Menominee trough have exposed the Sturgeon Quartzite at only one place, as noted above. The Randville Dolomite, however, is extensively exposed, and it is inferred that the Sturgeon lies immediately below it.

Description

The Sturgeon Quartzite consists chiefly of flaggy to thick-bedded vitreous quartzite. The quartzite is light colored, mostly white, but gray, pale-green, and pale-pink varieties are not uncommon. Ripple marks and crossbedding are abundant throughout the formation, and they indicate that the top is away from the Carney Lake Gneiss. The basal 100–200 feet of the formation is composed of pale-green sericitic quartzite and sericitic schist, which grade downward into the arkosic Fern Creek Formation at the Sturgeon River damsite. Because these sericitic rocks are less resistant to erosion than is the quartzite, they are not widely exposed, but underlie low covered ground between the quartzite ridge and the Fern Creek rocks or the Carney Lake Gneiss. They are exposed on the west bank of the Sturgeon River at the Sturgeon River Dam; between the east and west branches of Black Creek in sec. 6, T. 39 N., R. 28 W., and between the gneiss and quartzite ridges in the NW¼SE¼ sec. 28, T. 40 N., R. 29 W., and the NE¼SE¼ sec. 2, T. 40 N., R. 30 W.

Typically the Sturgeon Quartzite is composed almost entirely of quartz, a small percentage of sericite, and accessory zircon and opaque dust. The clastic character of the quartz, has been entirely obliterated by recrystallization in the very pure quartz rocks, but the rounded shapes of the grains are still discernible in rocks containing a few percent of intergranular sericite. The average grain size of the quartz ranges from 0.08 to 0.6 mm; for nine thin sections representing about 1,000 feet of exposed beds along the east side of the Sturgeon River gorge the maximum grain size is about 1.0 mm, and the estimated silica content of these specimens is 96 percent.

The quartzite generally shows no obvious internal deformation. A perceptible alinement of elongate quartz grains does show in some specimens, but no more than would be expected as a primary sedimentary feature. The basal sericitic rocks, however, are schistose, and some thin beds of quartzite interlayered with the schist do show a marked elongation of quartz grains parallel to the strike of the formation; Higgins (1947) determined that the "c" axes of the grains generally share the along-the-strike attitude.

Quartz veins and altered diabase dikes cut the Sturgeon Quartzite at many places. North of the Norway quadrangle, along the quartzite belt, the diabase is present in transverse faults along which the quartzite is displaced; some of the dikes show marginal shearing, but massive igneous-textured dikes are most common.

Some of the basal beds of the Sturgeon contain abundant magnetite and black tourmaline, and some pyrite. The magnetite forms euhedral crystals as much as 1.0 mm. wide; the tourmaline forms small acicular crystals which are preferentially alined along the bedding in some specimens but randomly oriented in others. Some thin quartzite beds have been partly replaced by massive-appearing tourmaline along joints and fractures.

Thickness

Because the Sturgeon strata are nearly vertical, the width of the outcrop belt, which ranges from 1,000 to 2,000 feet, is the approximate thickness of the formation.

Relation to other formations

The boundary between the Sturgeon Quartzite and the Fern Creek Formation is a gradational one. The contact between the quartzite and the overlying Randville Dolomite is not known to be exposed any place in the district. These two formations are assumed to be conformable because of their close association in several districts and their apparent affiliation with a single marine transgression.

Origin

Deep erosion of the pre-Animikie gneissic basement produced much arkosic material throughout Fern Creek time. In places outside the district, where relief probably was greater, arkosic material accumulated also throughout Sturgeon time and well into Randville time (R. W. Bayley, 1959b). It appears, therefore, that the well-sorted quartz sands of the Sturgeon Quartzite must have been derived from a deeply weathered terrain lying beyond the limits of the district—indeed, perhaps even beyond the limits of the northern peninsula of Michigan. A tabulation of crossbedding data by Trow (1948) shows that the prevailing direction of transport of the Sturgeon sands was from the northwest. More recently Pettijohn (1957) has shown that many of the Precambrian orthoquartzites of the Lake Superior region, some of which may be equivalent to the Sturgeon, were derived from a source area lying to the west or northwest.

The Sturgeon Quartzite, therefore, is considered the product of an early Animikie marine transgression on the Canadian Shield, similar in most respects to the much younger Late Cambrian transgression. The record is fragmentary, but the available data indicate a transgression toward the northwest.

RANDVILLE DOLOMITE

Next in conformable sequence above the Sturgeon Quartzite is a thick formation composed predominantly of dolomite. This formation is similar in its lithology and geologic position to the Randville Dolomite of the Felch district and to the Kona Dolomite of the Marquette district. W. S. Bayley (1904, p. 200) regarded the dolomite formation of the Menominee district as a direct correlation of the Randville Dolomite of the Felch district, and so designated it. The formation takes its name from exposures near Randville, Mich. (Clements and Smyth 1899b, p. 406).

Distribution

Within the mapped area the Randville Dolomite occurs in the three southeast-trending belts: a north belt which lies adjacent to and strikes parallel with the Sturgeon Quartzite; a central belt along the strike of the north iron range; and a south belt along the south iron range. The locations of all known dolomite exposures in these belts are shown on the accompanying maps (pls. 1–3). At many places the outcrop data were supplemented with information from mine workings and diamond-drill-hole records, but in some places the data are inadequate to locate contacts accurately.

Contacts between the Randville Dolomite and adjacent formations along the north belt are obscured, but some reasonable inferences regarding the position of the contact are possible.

Although nowhere exposed, the contact between the dolomite and the Sturgeon Quartzite, as it appears at most places on plates 2 and 3, has been fixed to coincide with the foot of the topographic slope made by the resistant Sturgeon Quartzite.

The contact between the Randville Dolomite and the Michigamme Slate, along the south side of this belt, is based chiefly on the records of a fortuitous scattering of diamond-drill holes, most of which are in the north part of the area shown on plate 2. The only known exposures on the belt are near the common line between secs. 11 and 14, T. 40 N., R. 30 W., where marble and quartzite protrude above the glacial deposits along the banks of a small stream, and to the northwest in secs. 2, 3, and 10. Elsewhere along the north belt the dolomite is covered by glacial deposits as much as 285 feet thick.

The north-south limits of the central belt of dolomite, which extends southeast along the north iron range for nearly 7 miles (pls. 1 and 2), are fairly well supported in the east half by outcrop, drill-hole, and magnetic data; supporting evidence is available for only the south contact of the west half of the belt because most of that area is covered by a Cambrian outlier.

The south belt of dolomite extends southeast along the south iron range from the Menominee River on the west to the east limit of the present mapping and beyond. Cambrian outliers cover a large part of the belt, but dolomite exposures are plentiful enough in a few places to establish closely the north and south limits. It is inferred that the Randville-Michigamme fault contact lies immediately north of the northernmost dolomite outcrops of the belt because the dolomite has proved to be generally more resistant to erosion than the Michigamme Slate, but the contact could be moved 1,000–1,500 feet to the north before crossing outcrops of Michigamme Slate.

Description

Massive clastic dolomite makes up a large part of the Randville Dolomite and is closely associated with thick- and thin-bedded sandy dolomite, dolomitic and quartzose slate and phyllite, and pebbly dolomite conglomerate. Thick beds of nearly pure crystalline dolomite are present in some areas and probably make up an important part of the formation. A most distinctive rock type in the formation shows algal structures (stromatolites). These are domical, 1–3 inches high, 3–12 inches in diameter, and composed of nested laminae of pure dolomite. The algal structures occur nearly every place in the district where the dolomite is exposed. They form reefs as much as 50 feet thick and of great but undetermined linear extent. They are also present in the Randville Dolomite of central Dickinson County (James and others, 1961) and in the Kona Dolomite of the Marquette district. As pointed out by James, stromatolite structures are also reported in nearly all dolomite of late Precambrian age—in the western United States and Canada, Australia, South Africa, and Fennoscandia—and most geologists now accept the view that they represent fossil algal colonies.

In the mapped area the algal dolomite is usually associated with thin-bedded sandy and conglomeratic dolomite of shallow-water deposition. This general association may be best observed in the outcrop area southeast of Lake Antoine (pl. 1), where algal dolomite, ripple-marked sandy dolomite, and thin dolomite beds showing mud cracks occur together.

A typical exposure of Randville Dolomite—for example, the roadside exposure 1,000 feet south of the abandoned Loretto mine (pl. 3)—is made up of three distinct rock types: massive thick-bedded dolomite, sandy thin-bedded dolomite, and algal dolomite. The sandy and algal phases are confined to the south part of the outcrop area under discussion; the thin sandy beds occur chiefly on the north edge of the outcrop but are also interbedded and with algal dolomite at other places in the outcrop. Some of the surfaces are blackish, are lichen covered, and show a thin, somewhat limonitic weathering rind, but other surfaces are clean and show the effects of glacial scouring. The dolomite is generally light colored, gray, buff, or white, in places pink or purple and white mottled. The sandy and argillaceous rocks are commonly gray or brown.

The massive dolomite shows beds 1–3 feet thick separated by stylolite seams; some of it has been intricately veined by quartz along joints and minute fractures. In outcrops the massive dolomite appears very fine grained and porcellaneous; in the thin section it shows a fine-grained crystalline mosaic of dolomite broken only by scattering of silt-sized quartz grains and a few quartz veinlets.

The sandy beds are generally less than 6 inches thick, and some of them show oscillation ripple marks which indicate that the tops of the beds face south, in agreement with the convexity of the adjacent algal structures. The quartz sand which pervades the thin dolomite beds also forms lenses and interdigitations in the dolomite, as well as discrete thin beds of dolomitic quartzite. The latter are commonly interlayered with thin laminae of dolomite, some of which have been broken and incorporated as platy fragments in the sandy layers. The sand in the dolomite is composed chiefly of well-rounded quartz grains, but grains of potassic feldspar, plagioclase, and a few shreds of muscovite and chlorite are also present, and are locally abundant.

The algal dolomite consists of nearly pure laminated dolomite having an average maximum index of refraction of 1.680. The laminae are only a few millimeters thick and appear in thin section as layers of crystalline dolomite of differing grain size. Specimens of Randville Dolomite examined from other parts of the district also show indices of pure dolomite. A chemical analysis of the Randville Dolomite made by E. Brewster and reported by W. S. Bayley (1904, p. 215) follows.

Chemical analysis and normative mineral composition (weight percent) of Randville Dolomite from the Chapin mine

Chemical analyses		*Norms*	
CaO	30.97	$CaCO_3$	55.20
MgO	20.48	$MgCO_3$	42.84
Fe	.80	$FeCO_3$	1.66
P_2O_5	.05	Apatite	.11
Al_2O_3	.20	Al_2O_3	.20
Residue	.73	Mostly silica	.73

The analyzed rock is thus almost normal dolomite of the composition $CaCO_3 \cdot MgCO_3$.

In addition to the three rock types that occur south of the abandoned Loretto mine, others that are present throughout the district are sericitic phyllite, sandy clastic dolomite, pebbly dolomite conglomerate, and silicified dolomite breccia.

The phyllite forms only a minor part of the exposed formation. It is associated with the thin-bedded sandy phase of the formation and may occur interbedded with dolomite, siltstone, or quartzite. The phyllite is gray, green, or black, and shows a sericitic sheen on cleavage surfaces and bedding planes. It is composed chiefly of quartz silt grains (0.01 by 0.05 mm) which are somewhat elongated in a schistose matrix of sericite. Brown euhedral tourmaline (0.07 by 0.03 mm) and an undetermined prismatic mineral, probably rutile (0.01 by 0.02 mm), are present as randomly oriented crystals in some of the layers.

Clastic dolomite makes up the most abundant part of the exposed formation and includes sandy dolomite and pebbly dolomite conglomerate. The clastic dolomite is composed almost completely of rounded detrital fragments of dolomite and of quartz grains (fig. 9). These clastic rocks have been recrystallized, as a rule, and some have undergone strong internal deformation which has obliterated the original clastic character of the carbonate fragments; at some places the original texture of the rocks is still evident. Typical clastic dolomite is exposed in a roadcut 900 feet north of the S¼ cor. sec. 35, T. 40 N., R. 30 W., (pl. 2), and an extremely deformed example, now a tectonite, showing perfectly alined spindle-shaped quartz grains in a re-

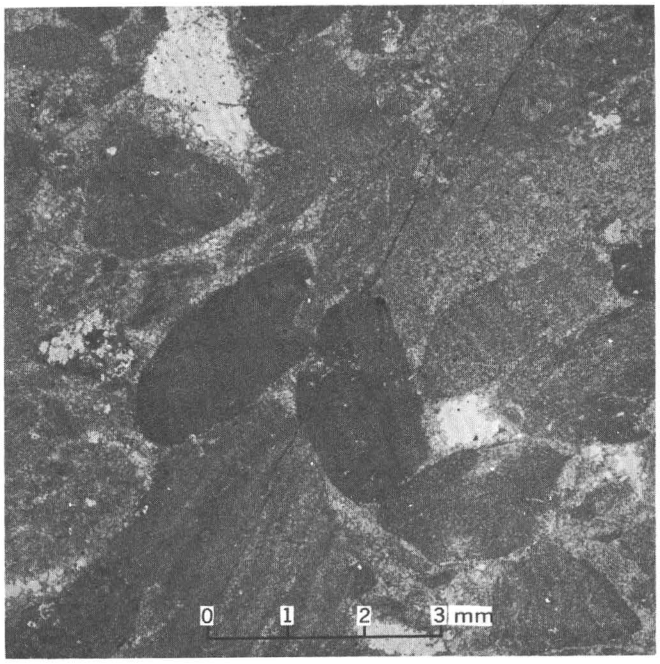


FIGURE 9.—Photomicrograph of clastic dolomite. Plane light; × 10.

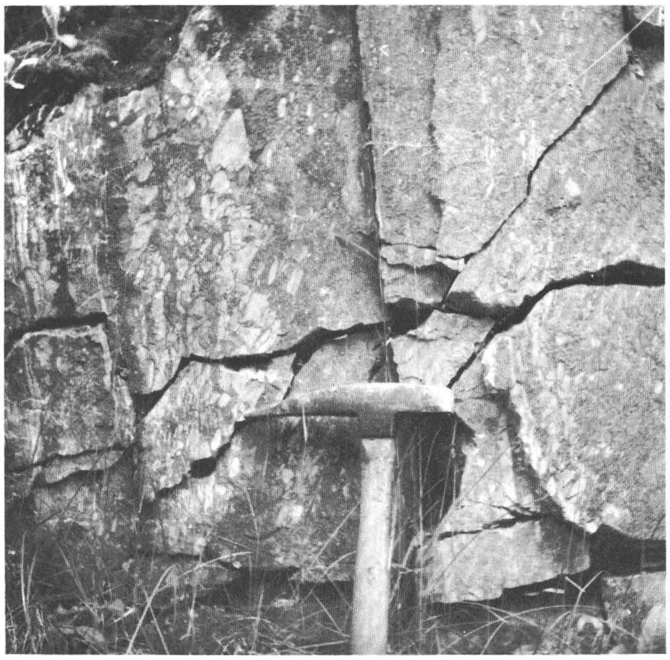


FIGURE 10.—Intraformational conglomerate in the Randville Dolomite, NE¼ sec. 34, T 40 N., R. 30 W.

crystallized dolomite matrix, is exposed near the N¼ cor. sec. 9, T. 39 N., R. 29 W. (pl. 2). Dolomite conglomerate showing rounded fragments of fine-grained dolomite as much as 3 inches long in a sandy dolomite matrix makes up a considerable part of the exposed formation near Iron Hill in sec. 32, T. 40 N., R. 29 W. (pl. 2). The conglomerate is extremely deformed internally, and the pebbles are alined. Probably the best exposure of the intraformational conglomerate beds is in an abandoned railroad cut in the NE¼ sec. 34, T. 40 N., R. 30 W. (pl. 2) where they are numerous and have been quarried normal to the strike (fig. 10).

The dolomite has been brecciated in many places in the area, particularly where the Randville strata are faulted or involved in tight folds, and most of the breccia has been silicified. The breccia is composed almost completely of angular dolomite matrix. Where silicified, the whole is replaced by cherty-appearing quartz which is commonly red. Both silicified and nonsilicified breccia occur in the NE¼ sec. 9, T. 39 N., R. 29 W. Some of the breccia in sec. 9 forms lenticular bodies which lie in the plane of the bedding, but most of it is related to two major crush zones which trend parallel to the axial planes of the folds. Other breccia exposures are north of the highway in the NE¼ sec. 2, T. 39 N., R. 30 W., and north of the Norway mine in the NE¼SE¼ sec. 5, T. 39 N., R. 29 W.

In the Iron Hill area (pl. 2) the silicification of the dolomite is extensive and not confined to the brecciated beds. The silicified rocks are almost wholly quartz, but contain here and there a few tiny hexagonal plates of a micaceous mineral whose optical properties best fit dickite.

Relation to other formations

The contact between the Randville Dolomite and the Sturgeon Quartzite has not been seen at any place in the district, but we infer that the two formations are conformable on the basis of their nearly parallel regional distribution. However, there is evidence to suggest that the contact of the Randville with the overlying Felch Formation is an unconformable one; further discussion of this subject is deferred until the Felch Formation has been described.

Thickness

The only place in the district where the thickness of the formation may be estimated is north of the Loretto mine (pl. 3). If it be assumed that the dolomite underlies the covered interval between the Sturgeon Quartzite and the Vulcan Iron-Formation and that the attitude of the beds is nearly vertical, the thickness of the dolomite is about 2,000 feet. For the purpose of comparison with this estimate, thickness values from other areas follow:

Felch district, 400–600 feet, (James and others, 1961).
South part Sagola basin, 2,000 feet, (James and others, 1961).
East Iron County (Kiernan quadrangle), 1,800 feet (Gair and Wier, 1956).
East Iron County (Lake Mary quadrangle), 1,950 feet (Bayley, 1959b).
Marquette district, 200–700 feet (Van Hise and Leith, 1911).

Conditions of deposition

In its broadest aspect the Randville Dolomite represents the carbonate facies of a stable shelf suite of rocks which includes the orthoquartzite of the Sturgeon Quartzite. The Randville Dolomite, like the Sturgeon Quartzite, apparently was deposited in shallow water, in a sea that was at times advancing and at times retreating over a broad continental shelf. Sedimentary structures in the dolomite, such as oscillation ripple marks and mud cracks, suggest that much of the deposition took place in very shallow water, which at times withdrew from the area. The clastic dolomite beds in the formation indicate that, at times, well-indurated Randville Dolomite was exposed to erosion along local shorelines. A survey of the environments of modern stromalites by Rezak (1957, p. 141–147) suggests that unless modern algae live quite differently from the Precambrian algae, the environment postulated for Randville time is almost ideally suited for the growth of flourishing algal colonies.

MENOMINEE GROUP

The Menominee Group includes the Felch Formation and the Vulcan Iron-Formation. The name was proposed for the Menominee district, Dickinson County, Mich. (James, 1958, p. 35). The Felch schist and the Vulcan iron-formation of the Menominee district were included in the Middle Huronian by Leith and others (1935, table facing p. 10).

FELCH FORMATION

The type locality of the Felch Formation is in the Felch Mountain district of central Dickinson County, Mich., where the designated strata lie between the Randville Dolomite and the Vulcan Iron-Formation (James, 1958, p. 35). At the type locality it was designated Felch schist (Van Hise and Leith, 1911, p. 303).

In the Menominee district the term "Felch Formation" is here applied to a distinctive sericitic slate and quartzite sequence which similarly lies above the Randville Dolomite and below the Vulcan Iron-Formation. W. S. Bayley (1904, p. 295) regarded these strata as forming a part of the Traders Member of the Vulcan Formation and referred to them as either "Traders quartzite" or "Traders slate," depending on the predominant lithology. More recently Dutton and Lamey (1939) used the informal term "footwall slate" to designate the same strata. The Felch Formation as defined above is probably correlative with the Ajibik Quartzite and Siamo Slate of the Marquette district, the Palms Quartzite of the Gogebic district, and the Pokegama Quartzite of the Mesabi district (Leith and others, 1935, table facing p. 10).

Distribution

As far as known, Felch strata form the stratigraphic footwall to the iron-bearing Vulcan Iron-Formation wherever the latter is present. At a few places where the Vulcan Iron-Formation is absent, the Michigamme slate rests unconformably on the Felch rocks.

Outcrops are very scarce, but the rocks have been found in numeorus test pits and mines, and the distribution and general character of the formation are fairly well known. Places where the rocks may be examined are:

On the north side of the Federal mine pit, NW¼ sec. 25, T. 40 N., R. 31 W. (slate and quartzite).

On the north side of the Bradley pit, SE¼ sec. 25, T. 40 N., R. 31 W. (slate).

On the north side of the Quinnesec mine pit, SE¼ sec. 34, T. 40 N., R. 30 W. (slate and quartzite).

At several places in the Munro open pit, SW¼ sec. 6, T. 39 N., R. 29 W. (slate).

On the north side, near the west end of the Norway mine pit, SE¼ sec. 5, T. 39 N., R. 29 W. (slate).

At the Norway city dump, NW¼ sec. 9, T. 39 N., R. 29 W. (slate and quartzite).

At the east end of the Globe Iron Co. open pit, NE¼ sec. 20, T. 40 N., R. 30 W. (slate and quartzite).

North of the railroad grade, north of the Indiana mine, SE¼ sec. 22, T. 40 N., R. 30 W. (slate, quartzite, and quartzite conglomerate).

In the SE¼ sec. 25, T. 40 N., R. 30 W., and the SW¼ sec. 30, T. 40 N., R. 29 W. (silicified breccia).

Near Iron Hill in the SE¼ sec. 32, T. 40 N., R. 29 W. (conglomerate and silicified breccia).

An isolated area containing quartzite is in the north-central part of sec. 14, T. 40 N., R. 30 W. These rocks are assigned to the Felch Formation chiefly on lithologic similarity, but have not been separated from the Randville Dolomite in the mapping.

Description

The lithology of the Felch Formation is remarkably uniform throughout the length of the south iron range but variable along the north range. On the south range the formation is about 100 feet thick and consists of thin-bedded sericitic slate and phyllite, and intercalated thin-bedded quartzite. The quartzite layers appear to be prevalent in the upper part of the formation, and a thin (4 in. to 3 ft) key bed of dark ferruginous quartzite, the so-called "Traders quartzite," is commonly present near the top of the formation. Observations made along the north iron range at localities listed above, particularly in the vicinity of the Indiana and Forest mines and near Iron Hill, indicate that the formation changes character and thickens to about 500 feet toward the north. Thick-bedded vitreous quartzite, quartzite conglomerate, dolomite conglomerate, and silicified quartzite breccia are the best exposed and therefore the best known rock types of this northern area. The fine-grained material will be described first, then the quartzitic and conglomeratic material.

The fine-grained clastic rocks which make up the major part of the formation on the south iron range include slate, phyllite, siltstone, and schist. All these rocks show minor differences imposed during deposition and modifications imposed by later deformation and low-rank metamorphism, but they bear a close outward resemblance to one another and show a common mineralogy. They are predominantly thin-bedded rocks, the layers commonly less than 1 cm thick, and most show bedding-plane fissility. Cleavage surfaces of the phyllite are lustrous and spangled with tiny plates of white mica. On fresh surfaces the rocks are gray to greenish gray, but where weathered they may be pale green, red, or light buff, or almost white and mottled with red; color banding is not conspicuous. The chief mineral components of all fine-grained types are pale-green sericite and quartz; minor components are feldspar, chlorite, biotite, hematite, and magnetite. Most of the rock layers are composed of about equal parts of the two chief components, but layers composed predominantly of one or the other are common. Medium to coarse well-rounded grains of quartz and potassic feldspar, commonly visible to the unaided eye, are scattered throughout many specimens of the slaty rock and form wafer-thin discontinuous quartzite stringers between the slaty layers. These latter characteristics of the slate are useful but not infallible criteria for identifying Felch strata in the field. The prevailing texture of the rocks is microschistose. In some specimens the quartz grains as well as the sericitic groundmass are elongated in the plane of schistosity, which at most places parallels the bedding.

The formation contains a variety of quartzose rocks. Thin layers of light-colored vitreous quartzite occur interbedded with the slate at many places. In addition to these, massive heavy-bedded vitreous quartzite is present in a few localities. Vitreous quartzite conglomerate was found in test pits north of the Indiana mine, and brecciated and silicified quartz rock, probably derived from an original quartzite, occurs near the Forest mine and east of the Forest mine near Iron Hill.

The quartzite is composed chiefly of well-rounded quartz grains as much as 3 mm in diameter, commonly set in a groundmass of finer grained angular quartz, sericite, chlorite, and magnetite. Well-rounded grains of potassic feldspar, mostly microcline, and small granite fragments are sparingly present in most specimens. The above materials, except magnetite, are similar to those that comprise the sandy part of the Randville Dolomite; it is assumed that they had a common source.

Massive silicified breccia, composed of angular quartz blocks in a quartz matrix, is exposed at the NW cor. sec. 31, T. 40 N., R. 29 W., and in two small knobs lying between that corner and the Forest mine (pl. 2). The rock is white and commonly mottled with red; it is fine grained, vitreous, and devoid of any sedimentary structures except possibly bedding planes in a few places. Reticulating veins of quartz abound throughout the rock, and open fractures are lined with tiny quartz crystals. An outcrop of this brecciated rock near the SW cor. sec. 30, T. 40 N., R. 29 W., and one in sec. 25, T. 40 N., R. 30 W., contain a few fragments of sericitic slate, but elsewhere the breccia consists of a single rock type. Partly silicified rock on the south side of the large outcrop in the SE¼ SE¼ sec. 25, T. 40 N., R. 30 W., shows the granular texture of quartzite. Microscopically the silicified breccia presents a fine even-grained quartz mosaic showing local spots and veins of coarser grained crystalline quartz. All evidence of detrital origin has been obliterated.

Directly southeast of the Forest mine the silicified breccia lies between the Vulcan Iron-Formation and the Randville Dolomite, in the position of the Felch Formation. Farther southeast of the Forest mine, beyond the east limit of the Vulcan Iron-Formation, in the NW¼ sec. 31, T. 40 N., R. 29 W., a north-directed diamond-drill hole which started in the Michigamme Slate passed into the quartz breccia at a depth of about 1,500 feet. This suggests that the brecciated unit is probably stratigraphical and that it continues east between the Michigamme and Randville Formations where the Vulcan is absent.

Near Iron Hill, in sec. 32, T. 40 N., R. 29 W., similar quartz breccia and massive beds of red cherty-appearing quartzite rest unconformably on the Randville Dolomite. This relation was discovered and illustrated by Van Hise and Leith (1911, p. 335), and has been verified by the writers. The unconformable contact may be seen in the outcrop 400 feet west of the southeast corner of the small outlier of Cambrian sandstone which is on the west side of the road in the east-central part of sec. 32; it may be seen also along the southfacing dolomite escarpment, 200 feet southwest of the other exposure. The first-mentioned outcrop consists of low-dipping beds of dense cherty quartzite which rest on steeply dipping beds of dense dolomite. A spring issues from the unconformable contact. The second-mentioned outcrop shows the quartzite breccia plastered on the south side of and folded over the top of the steeply dipping and contorted sandy dolomite which forms the south-facing escarpment. Figure 11, reproduced from Van Hise and Leith (1911, p. 335, fig. 45), shows the geological relations of the formations at Iron Hill, relations thought to obtain by the present writers, even though some of the structural detail shown could not be verified in the field. Briefly, the structure is anticlinal and plunges east. The fold is defined by the dense quartzite and silicified breccias which dip steeply away from the dolomite on the north and south, which lie nearly horizontal on the dolomite near the center of the outcrop area, and which wrap around the dolomite on the east end of the outcrop area. The fact that the minor drag folds in the dolomite commonly plunge east probably reflects the present anticlinal structure, even though the dolomite shows nearly right-angular discordance beneath the mantling quartzite and breccia.

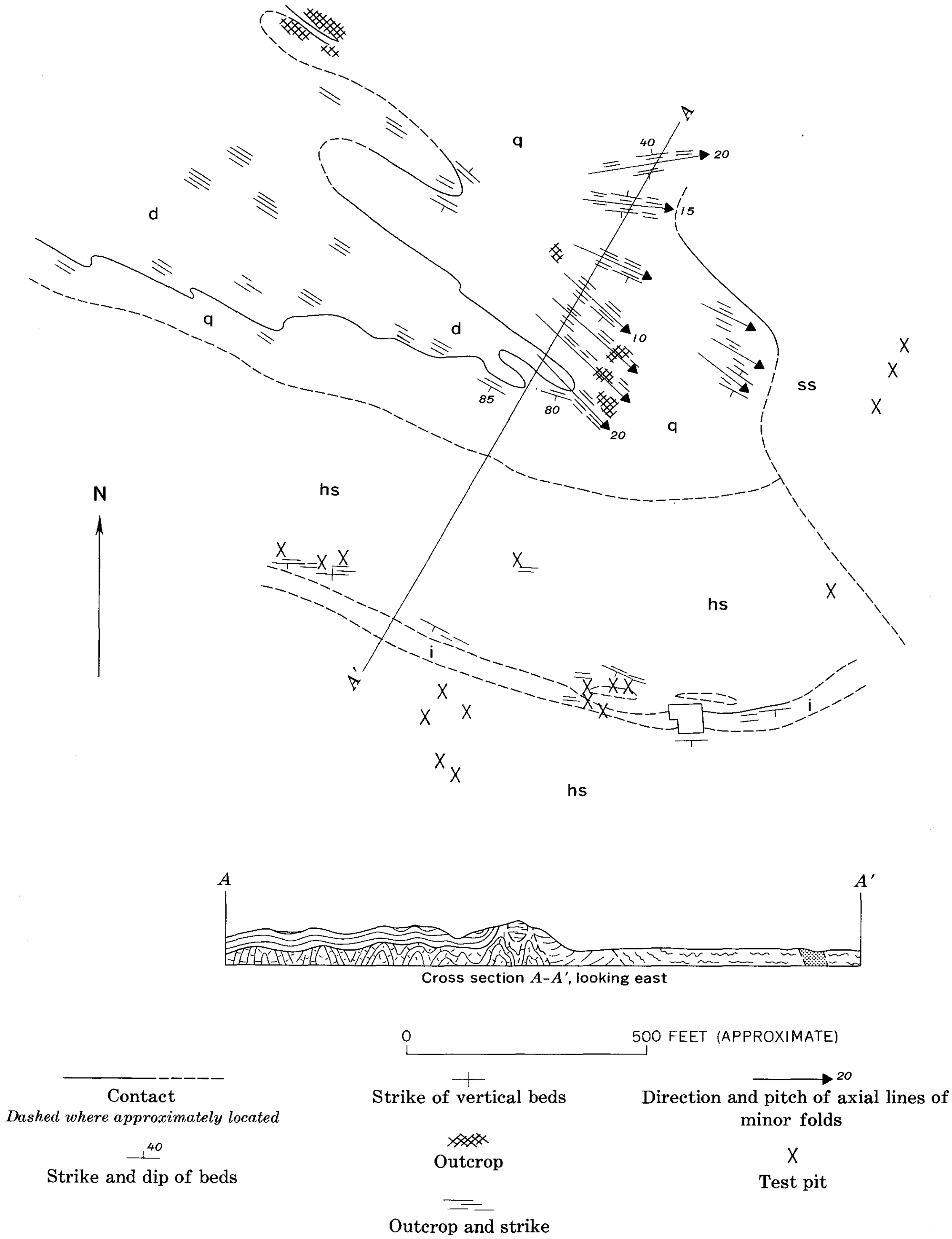


FIGURE 11.—Geologic map and section of the Iron Hill area, secs. 32 and 33, T. 40 N., R. 29 W., modified from Van Hise and Leith (1911, fig. 45). Lower Huronian dolomite, d; middle Huronian quartzite, q; upper Huronian Michigamme (Hanbury) slate, hs; containing interbedded iron-formation, i; Cambrian sandstone, ss.

A belt of dolomite conglomerate 50–100 feet wide is exposed along the south side of the long dolomite ridge near Iron Hill in sec. 32, T. 40 N., R. 29 W. It is composed chiefly of well-rounded to angular pebbles and boulders of dolomite, but also some of gray quartzite, in a groundmass of quartz sand and comminuted dolomite (fig. 12). The conglomerate is massive, and its attitude has not been determined, nor was it seen to be in contact with the rocks above or below. The close association of this conglomerate with Felch strata and the fact that it seems not to contain rocks younger than the Randville are the basis for our inclusion of it in the Felch Formation.

Relation to other formations

The relation of the Felch slate and quartzite to the Vulcan Iron-Formation appears to be one of gradation. The change from chiefly clastic deposition to chiefly chemical deposition seems to take place in the upper few feet of the Felch Formation and is marked by the appearance of iron minerals and cherts and the gradual dimunition of clastic grains. This gradation may be seen in the outcrops in the NW¼ sec. 9, T. 39 N., R. 29 W., (pl. 2).

The relation of the Felch Formation to the Randville Dolomite is not so clearly defined. The sequence Vulcan-Felch-Randville has been cut by diamond-drill holes at many places. The information from the drill holes, from mine workings, and from many test pits dug in the vicinity of the Vulcan Iron-Formation indicates that the Randville is succeeded almost everywhere by the distinctive strata of the Felch Formation, which almost everywhere conform structurally to it. Despite the apparent structural accord between the two formations at most places, several lines of evidence suggest that the Randville and Felch Formations are really unconformable. The evidence from the surrounding region is as follows:

FIGURE 12.—Dolomite conglomerate from exposure in sec. 32, T. 40 N., R. 29 W.

In the Marquette district, a post-Kona (Randville) period of uplift and erosion caused the removal of most of the Chocolay Group from the district before the succeeding Menominee Group was deposited (Van Hise and Leith, 1911, p. 261). In the Penokee-Gogebic district, the Bad River Dolomite, which has been correlated with the Randville Dolomite, rests unconformably beneath the Palms Quartzite (Leith and others 1935, table facing p. 10). In the Felch district a post-Randville disturbance is thought to be indicated by the abrupt change from dolomite deposition in Randville time to clastic deposition in Felch time and by gross differences in the thicknesses of the Randville and Felch Formations from place to place in the district, which probably reflect pre-Felch erosion of Randville strata (James and others, 1961).

The evidence found in the Menominee district to support a post-Randville disturbance is as follows:

1. Felch strata rest on Randville strata near Iron Hill with a marked discordance of bedding.
2. In Felch time, coarse gravel, sand, and mud were deposited where chiefly chemical sediments (dolomite) were being deposited previously. Some of the coarse gravel is composed of boulders of dolomite and quartzite.
3. The Felch strata become thicker and more conglomeratic toward the north.
4. A precise uniformity of Felch and Vulcan facies along the length of the Menominee district indicates that the strike of sedimentation was eastward, not northward as in the previous Sturgeon-Randville epoch.

Collectively the preceding items, including the regional considerations, indicate a post-Randville disturbance of wide scope. The evidence does not indicate orogeny, but broad and gentle warping of the crust. Some areas were uplifted and Chocolay Group rocks were eroded from them, but the thinness of the Felch Formation and the return to chemical sedimentation in Vulcan time indicate that these uplifted areas were not mountainous, nor long enduring.

The most important result of the disturbance was that the Animikie basin was shaped to nearly its present outline; what was open seaway in Chocolay time was, at the advent of Felch time, a narrow area of the sea, probably having restricted circulation and thus, eminently suited to concentrate vast quantities of iron to be deposited in Vulcan time and later.

Conditions of deposition

We regard the Felch rocks as having been laid down near shore during a slow and fluctuating marine transgression. The deposition of clastics continued until the local source area became submerged. The thickening and coarsening of the Felch strata toward the north and the local unconformity near Iron Hill suggest to us that the Carney Lake Gneiss area was probably one of the post-Randville high areas. The fact that the strike of sedimentation in the Menominee trough was eastward during Felch and Menomineee times indicates the area to the south in Wisconsin was also a tectonic high.

The chemical precipitation of iron and silica started in late Felch time. With the submergence of the local sources of clastic materials, these precipitates continued to build up, undiluted by clastics, to form the Vulcan Iron-Formation.

VULCAN IRON-FORMATION

The Vulcan Iron-Formation is the major iron-bearing unit in the district. It was named by Van Hise in Clements and Smyth (1899a, p. 25–26) for exposures near the West Vulcan mine in the Menominee district. Van Hise correlated the Vulcan with the Negaunee Iron-Formation of the Marquette district, and although he and W. S. Bayley (1900, 1904) and he and Leith (1911) later assigned it to a younger age, his original correlation is now generally accepted—the Vulcan Iron-Formation is now correlated with the Negaunee Iron-Formation of the Marquette district, the Ironwood Iron-Formation of the Penokee-Gogebic district, and the Biwabic Iron-Formation of the Mesabi district (Leith and others 1935).

Divisions of the formation

Throughout most of the Menominee district the Vulcan Iron-Formation divides naturally into three litholog units which W. S. Bayley (1904, p. 280) designated members. From oldest to youngest they are the Traders Iron-Bearing Member, the Brier Slate Member, and the Curry Iron-Bearing Member. Allen (1920, p. 202) discovered a ferruginous slate unit in diamond-drill core from an exploration near Loretto (pl. 3), and later underground at the Loretto mine, which he called Loretto Slate. The slate was said to rest conformably on the Curry Member and beneath a conglomerate at the base of the overlying Michigamme Slate. This slate unit is now known to be widely though sparingly distributed within the district, and is here designated as the fourth and uppermost member of the Vulcan Iron-Formation, the Loretto Slate Member. The Traders Iron-Bearing Member was named from the old Traders mine in sec. 17, T. 40 N., R. 30 W. (Bayley, 1904) but subsequent work has shown that the iron-formation exposed there is the Curry Member (R. B. Hall, U.S. Geol. Survey, unpub. data). The Brier Slate Member takes its name from the Brier Hill area near Norway (Bayley, 1904, p. 291). The Curry Iron-Bearing Member was named because of its occurrence in the Curry mine at Norway (Bayley, 1904, p. 291).

General distribution

The known and inferred distribution of the Vulcan Iron-Formation is shown on plates 1, 2, and 3. Aeromagnetic and ground magnetic surveys have proved reliable tools for locating these very magnetic rocks even where they are deeply covered by younger deposits; in fact, they have been easily delineated by magnetic surveying to the east of the mapped area where the overlying Paleozoic rocks are in excess of 500 feet thick. The likelihood, therefore, that any sizeable area of Vulcan Iron-Formation was missed in mapping is very remote.

The distribution of the Vulcan Iron-Formation parallels closely the distribution of the underlying Felch and Randville Formations, except in a few places where the Vulcan strata have been dislocated by faulting or removed by post-Vulcan but pre-Michigamme erosion. Removal by pre-Michigamme erosion is indicated along the Randville-Michigamme contact of the north dolomite belt (pls. 1 and 2), from the north border of the area southeast to sec. 3, T. 39 N., R. 24 W., and in the area between the Forest mine and Iron Hill along the south edge of the central dolomite belt (pl. 2). Post-Michigamme faulting best explains the absence of the Sturgeon, Randville, Felch, and Vulcan Formations along the Quinnesec-Michigamme contact in the south part of the district. Pre-Michigamme erosion of only the upper members of the Vulcan is indicated on the south iron range by the sporadic distribution of the Loretto Slate Member, and by the absence of the Loretto and Curry Members and part of the Brier Member in the area between the Pewabic and Vivian mines, secs. 31–34, T. 40 N., R. 30 W. (pls. 1, 2). The Vulcan has been removed from the bedrock surface by faulting in secs. 1 and 2, T. 39 N., R. 30 W. (pl. 2), and in sec. 18, T. 39 N., R. 28 W. (pl.

3). At each of these places magnetic surveying and diamond drilling have shown the iron-formation to be present at depth.

The magnetic survey shows unquestionably that the Vulcan Iron-Formation is not present at the bedrock surface on the north side of the Randville Dolomite of either the north or south iron ranges (see plates). It is absent north of the south range because the south range is a fault block—an upfaulted repetition of the downdip part of the north range; it is absent north of the north iron range because it was removed by pre-Michigamme erosion.

W. S. Bayley (1904), using an interpretation of the iron-bearing members that differs from the one here presented, concluded that the Curry Member was the most widely distributed of the two iron-bearing members and that where only one of them was present it was the Curry Member. This conclusion was a natural outgrowth of a previously made assumption that the Vulcan was deposited on an irregular erosion surface, which limited the distribution of the lower members. It has since been shown by Dutton and Lamey (1939–42) and by the present mapping that the Traders Member is the more widespread of the two iron-bearing members, probably chiefly because the pre-Michigamme erosion removed the members in normal sequence.

TRADERS AND CURRY IRON-BEARING MEMBERS

Description

Dutton (1942, p. 4) suggested that the Curry and Traders Members of the Vulcan Iron-Formation could be differentiated locally by color and degree of granularity, but the present consensus is that the rocks are too similar to be separated without other geological criteria. The following description of iron-formation applies equally well to the Curry and to the Traders Members.

The rocks of the Traders and Curry Members are iron-formation, which has been defined by James (1954, p. 239) as "a chemical sediment, typically thin-bedded or laminated, containing 15 percent or more of iron of sedimentary origin, commonly but not necessarily containing layers of chert."

The iron-formation of the Vulcan is thin bedded and commonly laminated, but it does not display uniformity in the thickness of the beds. Individual beds generally range from 1 mm to 30 cm (about 12 in.) in thickness. As a rule, beds of granular jasper alternate with beds composed chiefly of oxides of iron, principally hematite Fe_2O_3 (69.94 percent iron), and a lesser amount of magnetite Fe_3O_4 (72.4 percent iron). Almost all of the iron-rich layers contain a small amount of crystalline quartz, and at some places dolomitic carbonate and chlorite as well.

The iron-formation usually is dark. Viewed from a distance it commonly appears dark gray or reddish brown, but at close range it appears as a medley of deep red or maroon, metallic gray, and black. If much oxidized, hues of orange and red are dominant. Most jasper beds are maroon (liver colored) or red. They are generally thicker than adjacent iron-rich beds and most are uniformly straight bedded, but irregular beds and lenticular beds are common.

The jasper beds are composed chiefly of red jasper granules, specular hematite, magnetite, and metachert (a fine-grained mosaic of crystalline quartz). The granular character of most jasper beds can be seen by the unaided eye, but a wetted surface and a hand lens are helpful. In their primary state the jasper granules are a mixture of amorphous silica and red iron oxide (fig. 13). In their characteristic crystallized state, the iron oxide is specular hematite, magnetite, or both, and the silica is crystalline quartz (fig. 14). Most jasper beds contain, in addition to jasper granules, ooliths which are made up of concentric layers of red amorphous hematite and silica about a nucleus of quartz or jasper. Recrystallized ooliths form the same products as the granules. The granules, in shape, size, and appearance, resemble the greenalite granules that are so characteristic of the Biwabic Iron-Formation of the Mesabi iron range. They may represent the analogue of the greenalite granules, formed under oxidizing conditions.

The crystallization of the granules does not appear to be a result of age or of diagenesis, but a result of the metamorphism accompanying elevated temperature. The crystallization was studied by James (1955, p. 1473–1480) over a broad range of metamorphic temperatures and conditions. Iron-formation composed dominantly of unaltered jasper granules and iron-formation composed dominantly of altered jasper granules are both present in the Menominee district. The metamorphic grade in the area underlain by the iron-formation, as determined from the mineralogy of slate and greenstone, ranges from low chlorite to low biotite grade. Throughout this area all degrees of crystallization of granules are found, but without exception crystallization has been most active in the areas of higher grade metamorphism. Once formed, the essential minerals of the crystallization of the granules, quartz and hematite or magnetite, persist as stable minerals to the highest grades of regional metamorphism (James, 1955, p. 1475; Yoder, 1957, p. 233). At temperatures higher than the chlorite grade, only enlargement of the mineral grains takes place (Van Hise and Bayley, 1897, p. 272–273; James, 1955, p. 1473).

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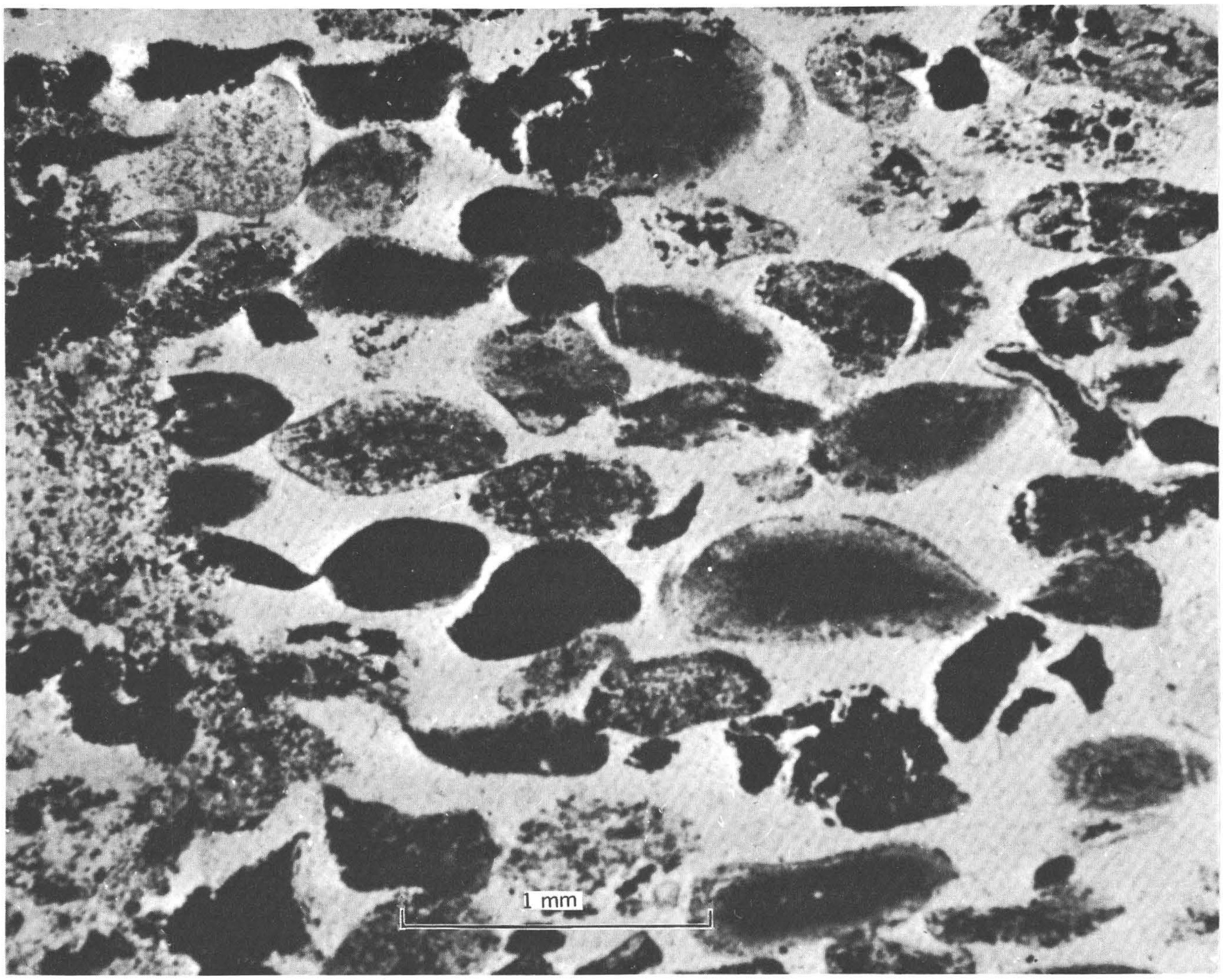

FIGURE 13.—Photomicrograph, showing jasper granules in a metachert matrix. The very dense granules are bright-red noncrystalline jasper. The gray (salt-and-pepper) granules show an early stage of crystallization—segregation of iron oxide as hematite or magnetite from silica. Ordinary light.

Undeformed granules are roughly ellipitical, but some are nearly round. Commonly they do not exceed 1 mm in length or diameter. In apparently undeformed rocks the primary elongate shape of the granules is reflected in a preferred elongation, as shown in figure 13. In slightly deformed rocks, the elongation is marked, the granules appear drawn out, pointed on the ends; in rocks that have been subjected to extreme internal deformation, the granular structure has been obliterated, and the rocks appear schistose (fig. 15). The ultimate product is quartz-specularite schist.

The average size of crystalline quartz grains in 20 specimens of the jasper layers is 0.10 mm. The sizes range from about 0.02 to 0.34 mm. In the same specimens the average size of the ore grains is 0.04 mm, a little less than one-half the size of the quartz grains, and the sizes range from 0.01 to about 0.23 mm.

The tenor of the iron in the jasper layers varies considerably, from very little in those that are mainly cherts, to very much in those that are rich in hematite or magnetite. Some granules are composed of only slightly ferruginous jasper, whereas others are composed chiefly of amorphous or finely crystalline iron oxide. Granules showing wide variations in the relative quantities of these constituents occur side by side in the jasper beds.

Everywhere the iron-formation contains some thin-bedded, so-called slaty iron-formation—that is, iron-formation without conspicuous jasper beds. Whereas the thick-bedded iron-formation is generally coherent and massive in outcrop, the thin-bedded iron-formation is generally fissile and separates along bedding planes into plates 1 mm to several centimeters thick. Most of it is laminated. The laminae consist of alter-

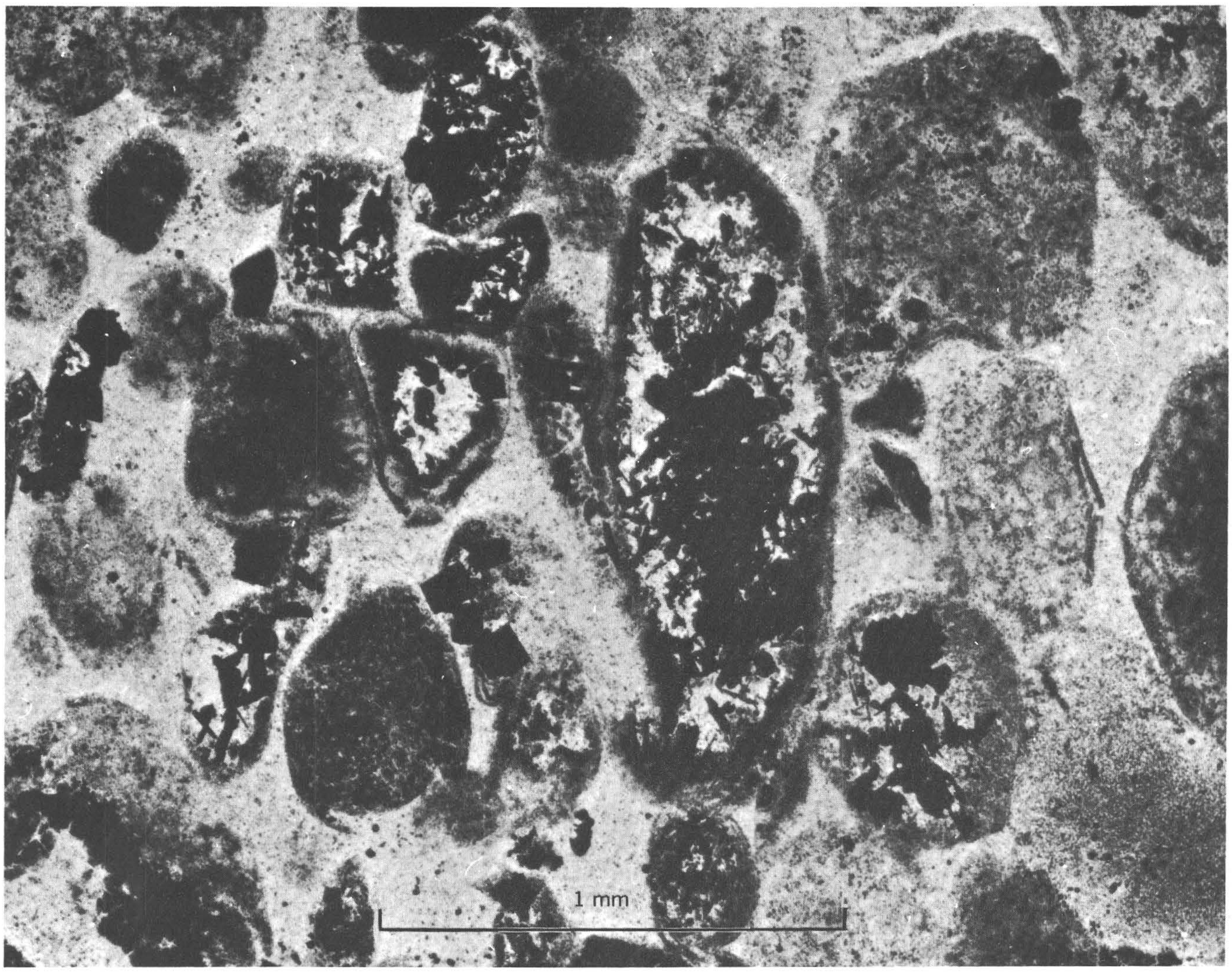

FIGURE 14.—Photomicrograph showing several types of crystallization of jasper granules. Quartz, white; specular hematite and magnetite, gray or black. Note the wide variation in the iron to silica ratios from one granule to another. The black spicules are specular hematite, and the square to rectangular sections are magnetite. Most granules pictured contain both iron minerals. The mottled granule (top center) represents an early stage of crystallization and segregation. The shapes of the iron oxide segregations suggest incipient specular hematite. Plane-polarized light.

nating beds of relatively iron-free nongranular chert and beds of iron ore minerals. Adjacent laminae may or may not be gradational. Some specimens of thin-bedded iron-formation are composed entirely of magnetite and nongranular chert; some contain carbonate (probably dolomite) and a small amount of pale-green chlorite. Beds adjacent to granular jasper beds usually contain a few jasper granules and some specular hematite. The thin-bedded iron-formation, with its various modifications, can be classified as magnetite-banded iron-formation of the oxide facies of iron-formation as defined by James (1954, p. 261). Accordingly, the magnetite of these beds may be considered a primary or diagenetic mineral (James, 1954, p. 257). The average size of chert grains in 11 specimens of the thin-bedded iron-formation is 0.04 mm, and the range is 0.03 to 0.11 mm. The average size of the ore grains is 0.05 mm, and the range is 0.03 to 0.11 mm.

A few specimens of thin-bedded iron-formation examined contain clastic quartz and plagioclase in addition to chert, magnetite, and chlorite. These clastic rocks resemble the Brier Slate Member, and undoubtedly represent minor incursions of clastics into an area where chemical sedimentation was in progress.

Curry Member iron-formation containing thin layers of dolomitic carbonate occurs sparingly in the Millie pit at Iron Mountain (NW¼ sec. 31, T. 40 N., R. 30 W.). The dolomite is closely associated with chert, chlorite, and magnetite, and appears to be a primary mineral. Quantitatively the carbonate is not important.

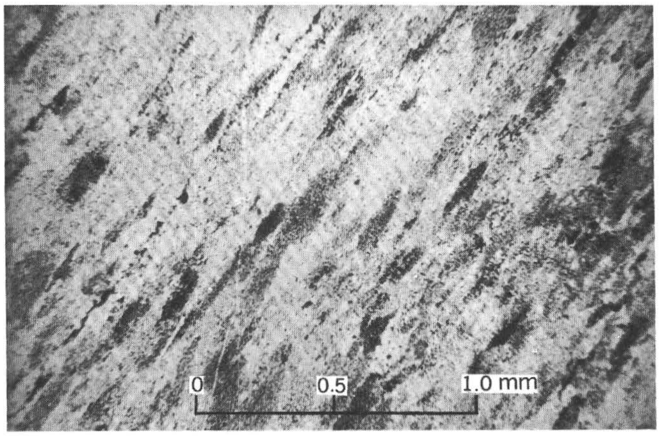

FIGURE 15.—Photomicrograph of slightly schistose granular jasper layer from the Traders Member of the Vulcan Iron-Formation. Plane-polarized light; × 36.

Its significance as a mineral in iron-formation has been discussed by James (1954, p. 261).

Chemical composition

Most of the iron-formation in the district has been to some degree altered by oxidation and enriched by removal of silica or by transportation and redeposition of iron. Numerous chemical analyses of highly enriched rocks (the iron ores) are available in the literature, but no analysis of unaltered iron-formation is available. Table 26 shows an analysis of Curry Member iron-formation representing 13,417 tons of siliceous ore mined from the Millie pit at Iron Mountain by the North Range Mining Co. in 1936. Although analyzed rock probably approaches the composition of unaltered iron-formation, examination of the Millie pit by the writers indicates that the analyzed sample probably contained some high-grade ore. The iron content shown in the analysis is therefore probably too high and the silica content too low for typical unaltered iron-formation.

TABLE 26.—*Chemical analysis, in percent, of iron-formation from the Millie pit, Iron Mountain, Mich.*

A		*B*	
Fe	36.26	SiO_2	40.1
P	.03	Al_2O_3	.8
SiO_2	40.15	Fe_2O_3	50.1
Mn	.13	FeO	1.6
Alumina	.77	MgO	2.0
Lime	1.39	CaO	1.4
Magnesia	1.97	Na_2O	------
S	.009	K_2O	------
Ignition loss	2.60	P_2O_5	.07
		MnO	.2
		S	.00
		CO_2	2.6
		Total	98.87

A. Analysis of shipments from Millie mine, Iron Mountain, Mich. for 1936 (Lake Superior Iron Ore Association, 1938). Samples dried at 212°F.
B. Analysis *A* recast by James (1954, p. 260) in terms of standard oxides. The iron is divided arbitrarily into 35 percent ferric and 1.26 percent ferrous. All loss on ignition assumed to be CO_2.

The average iron content of the Traders Member is 34 percent, as compiled from the logs of 15,620 feet of diamond drilling in that member. The average iron content of the Curry Member is 31 percent for 5,959 feet of drilling. Both averages include a considerable amount of enriched iron-formation showing iron values above 40 percent. The original iron content of the iron-formation therefore was undoubtedly lower than the above averages, probably 25–30 percent. Inasmuch as the iron most commonly forms oxides, on the average the Vulcan iron-bearing members contain 44–50 percent iron oxides, 46–50 percent quartz, and negligible amounts of minor constituents.

Weathering

At many places the iron-formation has been changed to great depth by oxidizing surface and ground waters. Degrees of alteration vary, but commonly the processes have been the same from place to place. Most rocks containing magnetite show evidence of the oxidizing nature of the solutions, in that all or part of the magnetite has been altered to hematite, and is now martite. Some of the rocks have been enriched in iron, and the chert and clastic components have been replaced by hematite. During early stages of alteration, feldspar is converted into kaolin and perhaps other clay minerals, and these and other micaceous minerals are stained and partly impregnated by hematite, whereas chert and other granular minerals are stained around the margins. As the process continues, the grains are replaced from the margins inward until the rock is composed mainly of hematite. At a few places a knife-sharp line may be seen between the unaltered and the replaced rock, but generally the alteration front is not observed. In structurally favorable areas, iron enrichment and silica leaching by groundwater circulation has led to the production of high-grade iron ore which at many places contains more than twice as much iron as the original iron-formation.

The ore bodies generally plunge along the troughs of synclines or occur along faults or in brecciated zones at the ends of folds. These places are especially favorable for the circulation of ground water. Formation of ore bodies was particularly favored where a relatively impervious rock lay beneath the iron-formation.

Iron ores

Only minor pockets of iron ore were accessible for inspection during the present survey. The full description by W. S. Bayley (1904, p. 373–391) of the iron ores of the district are greatly condensed below. The ores mined in the past may be divided into two broad classes; (1) siliceous ores and (2) direct-shipping ores. Each class may be divided further according to the chemical and physical characteristics of the ore. The siliceous ore retains the general aspects of unaltered iron-formation. In some places it represents originally rich parts of the iron-formation, and at others slightly enriched iron-formation. The direct-shipping ore represents greatly enriched iron-formation. The ore may be massive or show the original bedding of the iron-formation in minute detail. These ores are commonly soft blue hematite, and in some places they may be schistose and specular. Brecciated ores of both classes occur. They are composed of iron ore and jasper fragments in a matrix which may be drusy hematite or calcite or dolomite. The material of the ore fragments in the hematitic breccias is like that of the hematite bands interstratified with the jasper bands. These ore fragments in the calcareous breccias are fragments of banded rich ores. In the banded rich ores the replacement of silica by hematite had been practically completed before the ore was brecciated, while in the calcareous breccias deposition of ore continued after brecciation.

A representative analysis, in percent, of the direct-shipping iron ores of the district is given below.

Loretto mine, 1920 shipments

[Analysis by Lake Superior Iron Ore Association, 1938]

Constituent	*Dried at 212° F*	*Natural*
Iron	61.10	56.09
Phosphorus	.018	.017
Silica	5.03	4.62
Manganese	.08	.07
Alumina	2.39	2.19
Lime	1.05	.96
Magnesia	2.31	2.12
Sulfur	.007	.006
Loss	1.83	1.68
Moisture	----------	8.20

Thickness

Because the rocks of the district stand nearly vertically at most places, it is possible to arrive at approximate figures for the thicknesses of the formations and members by averaging a series of direct map measurements. A series of 15 such measurements across the Traders Member averaged 135 feet, but ranged from 80 to 200 feet. Measurements were made at places thought to be uncomplicated by folding or faulting; it is presumed, therefore, that the range of thickness represents chiefly primary differences in the thicknesses of the member. Eleven measurements across the Curry Member averaged 130 feet, and ranged from 60 to 200 feet. If there are no structural complications, the wide variations in the thickness of the Curry Member may be due either to differences in the original thickness or to removal of part of the member by pre-Michigamme erosion.

BRIER SLATE MEMBER

The Brier Slate Member is best exposed in the Brier Hill area near Norway, but it has been recognized everywhere in the district where the Vulcan Iron-Formation is present.

Description

The Brier strata are chiefly siliceous iron-rich slate, much of which could be classified as iron-formation as defined by James (1954, p. 239–240), inasmuch as the rock is well laminated and contains an average of more than 15 percent iron of sedimentary origin. Fine grain, thin laminae, and dark-brown color are characteristic features. The Brier strata may not easily be confused with any other rocks in the district except the Loretto strata, which they closely resemble.

The unstained (unoxidized) slate is gray or greenish brown, but almost everywhere it has been iron-stained to some extent and is brown or reddish brown. Most of the slate, though well laminated, is not fissile, and makes blocky rhombic talus, but some of it is fissile and parts on bedding planes. Most commonly only the closely spaced laminae are visible to the naked eye, and when examined with a hand lens the rocks generally show no distinct mineral grains, but they do show the sparkle of microscopic magnetite crystals. The remarkable similarity of the rocks of the Brier Slate Member throughout the district indicates nearly constant conditions of deposition.

Although uniformity of characteristics is rather general, some variations occur. At Vulcan, and near the east end of the district, and presumably in the area between, the member contains near its middle a zone of silty slate which is exceptionally coarse for the Brier Member and which shows glassy quartz grains under the hand lens. At Quinnesec, Norway, and Vulcan, the member also includes thin beds of cherty dolomite that contain magnetite. Near the contact between the Brier and Traders Members, thin cherty layers occur in the Brier Member, and some layers contain a little pyrite, but the Brier strata close to the Curry Member are neither cherty nor pyritic.

Many of the specimens examined are microschistose and show varying degrees of alinement of mineral grains, but other specimens show no pressure effects. The schistose rocks have generally not recrystallized, but a few specimens show granoblastic textures. Segregation of light and dark minerals, mainly quartz-feldspar and magnetite, into layers (fig. 16) is the chief cause of the lamination shown by the slate.

The principal minerals present are, in order of abundance, quartz, feldspar, magnetite (or martite), sericite, chlorite, dolomite, chert, and pyrite; uncommon detrital minerals are muscovite, zircon, tourma-

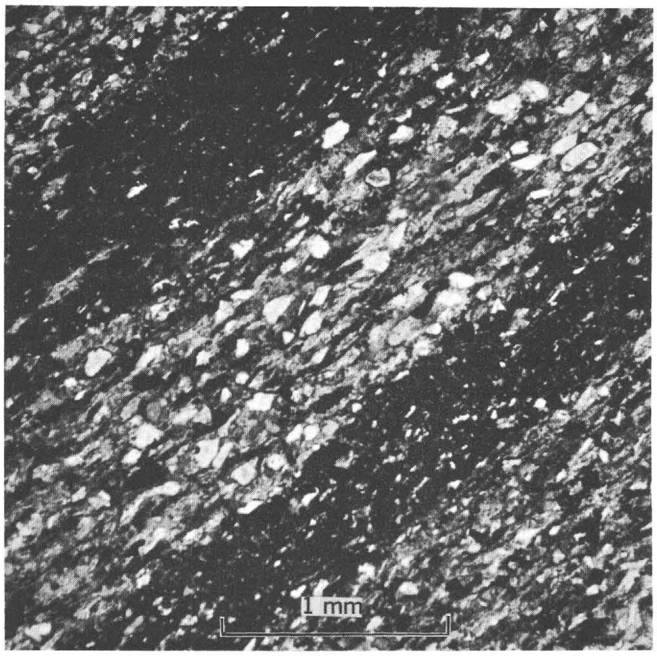

FIGURE 16.—Photomicrograph of Brier Slate Member of the Vulcan Iron-Formation showing layering. Dark layers are chiefly magnetite and quartz, and light layers are quartz and altered feldspar in a sparse chert matrix. Plane-polarized light; × 30.

line, and apatite; the chief minerals formed by metamorphism are biotite, chlorite, sericite, and magnetite; and minerals formed by weathering are hematite, limonite, and kaolinite and perhaps other clay minerals.

The quartz and feldspar grains are slightly rounded to very angular. Some quartz grains are ribbon shaped, as if derived from a gneiss or a schist, and all show undulatory extinction, probably caused by stress in the source rocks. Plagioclase, chiefly albite, and potassic feldspar, both microcline and orthoclase, are present in about equal amounts. The feldspar forms clear angular grains in some specimens, but it shows a well-advanced stage of alteration to sericite, or clay minerals and iron oxide (fig. 17), in most specimens. The size of the clastic quartz and feldspar grains ranges from 0.11 to 0.01 mm (22 specimens) and averages 0.05 mm.

Magnetite is abundant but generally restricted to certain layers in the rock; it is commonly associated with chlorite, biotite, or dolomite. In many of the specimens examined the grains show red translucent edges that indicate replacement by hematite. The size of magnetite crystals ranges from 0.10 to 0.01 mm and averages 0.03 mm.

Micaceous minerals form an important part of all the rocks examined. The chief micaceous mineral present is chlorite, green biotite, or greenish-brown biotite,

TABLE 27.—*Chemical analysis of slate from the Brier Member, from sec. 8, T. 39 N., R. 29 W., Michigan*

[From W. S. Bayley, 1904, p. 330; analyst: E. T. Allen]

Constituent	*Percent*	*Constituent*	*Percent*
SiO_2	50.15	K_2O	4.38
TiO_2	.52	H_2O at 105°	.81
Al_2O_3	6.55	H_2O above 105°	1.43
Fe_2O_3	33.80	P_2O_5	.08
FeO	.94	S	Tr.
MgO	.94	Cr_2O_3	Tr.
CaO	.16		
Na_2O	.31	Total	100.07

the differences in type possibly resulting from differences in degree of metamorphism. Sericite is present in all the rocks, but it is an important constituent only in the more feldspathic rocks in which the feldspar has been altered, probably by metamorphism. Weathering and circulation of ground water form iron-stained clay minerals instead of sericite.

An analysis of typical slate from the Brier Member, from sec. 9, T. 39 N., R. 29 W., is given in table 27.

In the analysis shown in table 27, the total iron indicated, approximately 24 percent, is a little high for slate from the Brier Member, but not excessively so. The iron content of the Brier Member as compiled from drill logs representing 4,505 feet of diamond drilling averages 18 percent, and ranges from 7 to 34 percent. The combining ratio possible between Fe^{+2} and Fe^{+3} (if correct) permits the presence of only a very small amount of magnetite in the analyzed sam-

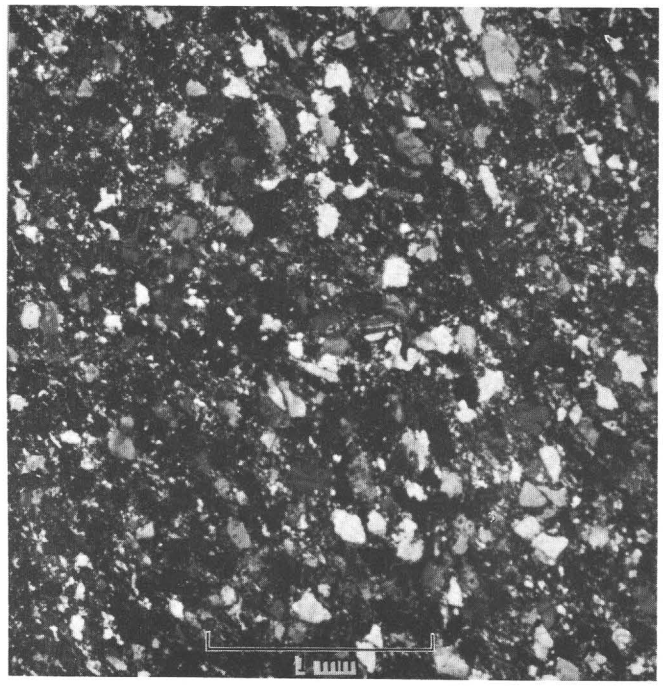

FIGURE 17.—Photomicrograph of Brier Slate Member of the Vulcan Iron-Formation showing plainly the detrital character of the quartz and feldspar grains. The twinned plagioclase grains are albite. Crossed nicols; × 30.

ple. The nearly perfect octahedra in many specimens of Brier Member indicate the presence of magnetite. Oxidation of the analyzed rock, and formation of hematite (martite), therefore appears likely. In other respects the analysis is about what would be expected from the mineralogy. The high percentage of K_2O reflects the abundant clastic feldspar and the potassium-bearing micas in the rock.

The carbonate-bearing slate is distinguished in the field by conspicuous light-colored carbonate layers as much as 1 cm thick, which alternate with dark layers of magnetite and the usual clastic minerals. In thin section the carbonate layers show a crystalline mosaic of carbonate admixed with variable quantities of metachert. The neighboring layers, which are composed predominantly of magnetite and clastic quartz, also contain scattered areas of carbonate and some metachert. The maximum index of refraction of the carbonate is approximately 1.680, the high index of dolomite. An analysis of the carbonate, separated by solution in nitric acid from 1 g of slate taken from a test pit north of the Curry mine shaft at Norway, was reported by W. S. Bayley (1904, p. 328) as follows: MgO, 0.0579 g; CaO, 0.0810 g; and CO_2 in about the proportion necessary to saturate the two bases. The molecular proportions of the two bases are as 1:1, and the carbonate is therefore a typical dolomite, containing 0.12159 g $MgCO_3$ and 0.14474 g $CaCO_3$, or a total of 0.2663 g dolomite in 1 g of rock.

Thickness

The thickness of the Brier Member differs from place to place in the district. On the west end of the north iron range it is unusually thin, about 100 feet, which appears to be the true thickness there, whereas the average thickness on the south iron range is about 300 feet. In areas where the Brier Member belt greatly exceeds 300 feet in breadth, repetition by folding and faulting is suspected.

LORETTO SLATE MEMBER

Distribution

At a few places in the district the Loretto Slate Member conformably overlies the Curry Iron-Bearing Member. Although named from the Loretto area, it does not crop out there. During the present examination of the Loretto mine property, ferruginous slate, probably of the Loretto Member, was found on the dumps of the test pits south of the Curry Iron-Bearing Member (pl. 3), but it was not possible to delineate the member on the diamond-drill-hole records examined; the distribution of the member in the Loretto area therefore is still poorly known.

The Loretto Member is best exposed on the south iron range in the Vulcan mine area, and poorly exposed near Millie Hill (pl. 2). The exposures are in gullies, test pits, and subsidence cracks in the south slope of Brier Hill in secs. 10 and 11, T. 39 N., R. 29 W. The gradational zone between the Curry Iron-Bearing Member and the Loretto Slate Member may be seen in a caved adit entrance in the SW¼ sec. 11, and near the south end of a large subsidence crack in the SE¼ of sec. 10. The total outcrop area is small, and the top of the unit is not exposed. To expedite the geologic mapping, the Loretto Member was included in the underlying Curry Member, but letter symbols on the map indicate the areas of occurrence (pls. 1, 3). A grass-roots outcrop of ferruginous schist on the south slope of Millie Hill in the NE¼ sec. 31, T. 40 N., R. 30 W. (pl. 1), constitutes the only other known outcrop of the Loretto Slate Member in the south iron range.

On the north iron range, ferruginous schist exposed along the south side of the Clifford and Traders pits in secs. 17 and 20, T. 40 N., R. 30 W., overlies and is gradational with the Curry Member, and has been assigned to the Loretto Slate Member. (Robert B. Hall, 1948, unpub. data).

Description

The lithology of the Loretto Slate Member is similar to that of the Brier Slate Member. Microscopic examination of seven specimens of Loretto Slate Member shows that all are fine grained and all contain magnetite or martite; six contain clastic grains of quartz in a schistose matrix composed chiefly of sericite and chlorite; one contains laminated cryptocrystalline quartz (metachert) and blebs of chlorite. A specimen from the south side of Millie Hill (pl. 1) contains dark-brown stilpnomelane in addition to the usual constituents. The iron content of the member reported in drill-hole logs is 16–30 percent.

In the Vulcan mine workings, 42 feet of slate in fault contact with the upper part of the Curry Member and in fault contact with the Michigamme Slate was tentatively assigned to the Loretto Member by L. M. Scofield (geologist of the Penn Iron Mining Co., 1932; unpub. data) who described the unit in the following terms:

> Firm, blocky, thinly laminated (⅛"); fine to medium-fine grained ferruginous slate, generally oxidized to chocolate brown but at places partly oxidized by laminae, leaving a banded (½") rock of green and brown. Oxidized portions show a reddish-tan streak. Bedding laminations are distinct and grain size varies by laminae. Near the top the slate sometimes shows a glassy sheen from many fine quartz grains. Drill hole data in West Vulcan suggests that higher horizons (here eroded or faulted out) are much more quartzose and low (10%) in iron content.

TABLE 28.—*Facies of sedimentary iron-formations and their metamorphic equivalents*

[From James, 1955]

Sedimentary			Metamorphic		
Facies		Composition [1]	Low grade [2]	Intermediate [3]	High grade [4]
Sulfide		Ferrous sulfide Organic carbon "Clay"	Pyrite Carbon (amorphous) Quartz Sericite	Pyrite Graphite Quartz Micas Garnet (rare)	Pyrite Graphite Pyrrhotite(?) Micas-garnet
Carbonate		Carbonate Chert Greenalite	Carbonate Quartz Stilpnomelane Minnesotaite	Grunerite Quartz Magnetite Carbonate	Grunerite Quartz Magnetite Pyroxene
Silicate	Nonclastic	Greenalite Chert Carbonate Magnetite	Minnesotaite Stilpnomelane Quartz Carbonate Magnetite	Grunerite Quartz Magnetite	Grunerite Quartz Magnetite Pyroxene
Silicate	Partly clastic	Iron-rich clay Chert Carbonate Magnetite(?)	Chlorite Stilpnomelane Quartz Carbonate Magnetite Biotite	Grunerite Quartz Magnetite Epidote Garnet Carbonate Mica	Grunerite Quartz Magnetite Garnet Hornblende Pyroxene
Oxide	Magnetite-banded	Magnetite Carbonate Greenalite Chert	Magnetite Stilpnomelane Minnesotaite Carbonate Quartz	Magnetite Grunerite Quartz Garnet	Magnetite Grunerite Quartz Pyroxene Garnet
Oxide	Hematite-banded	Ferric oxide Chert Magnetite Calcite	Hematite Quartz Magnetite Calcite	Specular hematite Quartz Magnetite Calcite	Specular hematite Quartz Magnetite Calcite
			Less than 0.10 mm	0.10 to 0.20 mm	More than 0.20 mm
				Quartz grain size [5]	

[1] Inferred.
[2] Chlorite and biotite zones.
[3] Garnet and staurolite zones.
[4] Sillimanite zone, essentially. Inferred for some rocks.
[5] Diameter of typical grains in relatively pure layers of chert.

Relations to other formations

At the few places where the Loretto strata are well exposed, they rest conformably on the Curry Iron-Bearing Member and are gradational with it. The upper contact of the Loretto Slate Member is nowhere exposed, but it is assumed that the overlying Michigamme Slate is unconformable. At the few places where the Michigamme Slate may be seen in contact with the Vulcan strata the Loretto Slate Member is absent. The present distribution of the Loretto Member suggests that it remains only in areas structurally protected from post-Vulcan, pre-Michigamme erosion. The only known conglomerate between the Michigamme Slate and the Loretto Member was reported by Allen (1920, p. 201–202) from a diamond-drilling exploration in sec. 13, T. 39 N., R. 28 W., east of the Sturgeon River (see p. 58). Two of the southernmost drill holes in sec. 13, which were drilled steeply toward the north, penetrated the conglomerate, one at bedrock surface and the other after cutting about 100 feet of Michigamme-type slate. Beneath the conglomerate, which contained pebbles of Menominee and Chocolay Group rocks, the holes cut through 25–65 feet of banded ferruginous slate. According to Allen, the discovery of this conglomerate proved beyond any question that the periods of deposition of the Vulcan and the Michigamme (Hanbury) were separated by an interval of emergence and erosion. More recent drilling has penetrated conglomerate beds at the base of the Michigamme Slate in other parts of the area, but the most compelling evidence for post-Vulcan, pre-Michigamme unconformity is the present distribution of the Vulcan and Michigamme Formations.

CONDITIONS OF ORIGIN

There is a very extensive literature dealing with the origin of sedimentary iron-formations, and although a controversy continues about the sources of the iron and silica, there is general agreement about the chemical sedimentary origin of these iron-rich rocks. The iron-formation members of the Vulcan are chemical sedimentary rocks whose present characteristics reflect the chemical and physical conditions in the basin in which they were deposited and later diagenetic and metamorphic changes. The predominance of the magnetite and hematite indicates furthermore that the iron formations are of the oxide type, as distinguished from the several other recognized types: sulfide, carbonate, and silicate, indicated in table 28 (James, 1955). The availability of iron and silica being assumed, each of the above facies require that special chemical conditions have prevailed in the sedimentary environment. For instance, it has been experimentally determined that strong reducing conditions must prevail on the ocean bottom for sulfide iron-formation to form, whereas strong oxidizing conditions must prevail to form the oxide types. The various chemical parameters of the sea-water environment, in terms of Eh and pH, have been examined by Krumbein and Garrels (1952) and by Huber (1958).

According to James and others (1961, p. 45), the iron-formations are of marine origin, and "They probably were deposited in relatively shallow basins adjacent to land areas of low relief that were being intensively weathered under tropical or subtropical conditions. Under these conditions the iron and silica content of streams is many times that in streams of the temperate climates, and clastic load is very light. The basins of deposition probably were restricted—that is, separated from the open ocean by barriers, not necessarily emergent, which would inhibit circulation and permit development of abnormal oxidation potentials." The concept of the barred basin is required to explain the deposition of the sulfide and ferrous carbonate iron-formations, which require a reducing environment, inasmuch as the waters of almost all of the open ocean are oxidizing from surface to bottom. The well-aerated margins of such a restricted basin would seem to be the site for the deposition of oxide iron-formation of the type found on the Menominee range. The existence of an elongate basin in the district during Vulcan deposition was deduced from the apparent strike of deposition as indicated by the Felch and Vulcan strata. Similar basins apparently existed at the sites of the other iron districts in the region, but these were likely restricted from one another by open ocean or slightly submerged land barriers; thus, though the iron deposition in each basin was probably contemporaneous, no unit-by-unit correlation between them is possible.

The Brier and Loretto Slate Members represent interludes in Vulcan sedimentation during which the deposition of clastic material was dominant over the chemical deposition of iron and silica which characterizes the adjoining iron-formation members. Iron deposition continued throughout Brier time and into Loretto time, but was much diluted by detrital materials. The location of the source of these materials is conjectural. The angularity of the grains and the abundant feldspar indicate that the source area was probably close to the district and was composed chiefly of metamorphic and igneous rocks (probably granite) which were undergoing rapid mechanical erosion.

The fact that the Brier and Loretto Members are not recognizable units of the Vulcan Iron-Formation in the adjacent iron districts to the north (Felch Mountain and Marquette) probably means that the clastic material of the Brier Member came from one or more local source areas, possibly from the areas of the Carney Lake Gneiss and Wisconsin complexes. This local condition could have resulted from slight uplift of these areas and regression of the sea. This uplift and regression would expose to erosion lower Precambrian rocks that lost their cover of Chocolay rocks during the previous post-Randville uplift and erosion.

DISTRIBUTION AND RELATIONS ALONG THE NORTH IRON RANGE

At the west end of the north iron range (pl. 1) all the Vulcan members are present and show their characteristic lithologies. The Traders Iron-Bearing Member is exposed in the open pit of the Globe Iron Co., sec. 20, T. 40 N., R. 30 W.; the Curry Iron-Bearing Member is exposed between the iron-bearing members on the north wall of the Clifford and Traders pits; and the Loretto Slate Member is exposed above the Curry along the south sides of the last-mentioned pits. A few hundred feet west of the Traders pit, typical Michigamme strata are revealed in test pits. Traders Member iron-formation occurs in test pits east of the Traders pit and also east of the Globe pit; quartzose Felch slate is found in pits still farther east.

The general area is marked by a strong magnetic high (pl. 4),[2] the crest of which is over the Curry Member in the Clifford pit. The magnetic high splits into two crests at the northwest end of that pit. The major crest turns west and follows the plunge direction of an anticlinal structure westward to the vicinity of Bass Lake, sec. 13, T. 40 N., R. 31 W., whereas the minor (lower intensity) crest continues northeast into the Traders pit, but the intensity diminishes rapidly.

An east-west magnetometer traverse made at the north end of the Traders pit indicates that the magnetic anomaly, and therefore the iron-formation, does not continue north of the inferred fault that marks the north boundary of the range.

The fact that in the 1½ miles between the Clifford pit and Bass Lake the main anomaly decreases in magnitude by 60° on the dip needle and becomes broadly lobate probably indicates that the iron-formation has plunged to great depth. If a constant dip of the iron-formation is assumed, this depth is nearly 8,000 feet. Whatever the depth, the Vulcan almost certainly underlies the Michigamme strata between the iron ranges, near Bass Lake, and probably as far east as the Fumee Lake area (pl. 2; pl. 1, section *A–A′*).

The magnetic crest that marks the position of the Vulcan Iron-Formation is markedly linear and extremely strong and has been traced eastward to the north shore of Lake Fumee. The crest lies south of the abandoned Indiana mine (pl. 2) where iron-formation from the Traders Member was mined, and diamond drilling indicates that the Brier and Curry Members are in normal succession south of the mine. The crest of the anomaly therefore lies over the Curry Member. East of the Indiana mine, on the north shore of Fumee Lake, the magnitude of the magnetic anomaly over the Curry Member falls off sharply toward the east, although the width of the anomaly remains nearly constant to the east end of the lake. As here interpreted, the Curry Member has been removed by pre-Michigamme erosion in the area east of the center of the lake, although it may be present at depth as far east as the east end of the lake.

The position of the Traders Member is much more difficult to determine by means of magnetic anomalies in this part of the area than is that of the Curry Member. North of the Indiana mine there is a narrow anticline of Felch and Randville strata, and north of that, a shallow syncline of Traders Iron-Bearing Member. The northwest part of the syncline is obscured by Cambrian sandstone and is inferred. The Traders Member at this place is not magnetic enough to create a minor crest above the general background of the larger anomaly that crests over the Curry Member, it was therefore not possible to outline the syncline by magnetic surveying.

At the Forest mine, sec. 25, T. 40 N., R. 30 W. (pl. 2), the abandoned mine shaft is about 100 feet south of an outcrop of Randville Dolomite. Iron-formation and Brier-type slate are the only rocks identified on the mine dump; the iron-formation mined therefore was very likely the Traders Member. Its location is marked by a feeble anomaly of 30–260 gammas (by vertical magnetometer), which is very nearly submerged in the greater anomaly caused by the same member to the south. In the SE¼ sec. 25, east of the Forest mine, the succession of formations from north to south is Randville Dolomite, Felch Formation, iron-formation (presumably the Traders Member), and Michigamme Slate. The narrow linear character of the magnetic anomaly, which is similar to that observed elsewhere over the Traders Member, indicates that the Traders Member is present under the Pleistocene gravels as far west as the W¼ cor. sec. 25. Its location has been proved by diamond drilling over part of this area. The duplication of the member in the area of the Forest mine may be due to faulting, as indicated on the map (pl. 2). A short distance east of the SE cor. sec. 25, T. 40 N., R. 30 W., the aforementioned anomaly on the Traders Member ends, and diamond drilling in the NW¼ sec. 31, T. 40 N., R. 29 W., show that the Michigamme Slate rests directly on the Felch Formation; the total Vulcan Iron-Formation is absent. This relation is interpreted to indicate pre-Michigamme erosion of the Vulcan strata. A similar structural relation could result from faulting, but the absence of any magnetic anomaly rules against faulting as the only cause.

At the east end of the north iron range (pl. 3) a normal succession of Animikie rocks is present in the area of the abandoned Loretto and Appleton mines, but outcrops are very scarce. The geology of this area has been compiled from new surface mapping, existing mine maps, and interpreted diamond-drill-hole logs; mapping credit therefore is due chiefly to the geologists of the mining companies. The structure of the mined area is primarily homoclinal, complicated chiefly by right-lateral faults, but minor folding exists also.

The iron-formation at the Loretto mine is marked by a moderate magnetic anomaly that strikes roughly east-west through the mine property (pl. 4). To the

[2] Plate 4 compiled from original magnetic data sheets, which were plotted at 1 inch=400 feet and which show station locations and values. The data sheets were placed on open file for public inspection in March 1960 at U.S. Geol. Survey Offices, Washington 25, D. C., and Madison 6, Wis.; Michigan Depart. Conservation Geol. Survey Div., Lansing, Mich.; Michigan College of Mining and Technology Dept. Geology, Houghton, Mich. The results of an aeromagnetic survey of the Menominee district are contained in Wier and others (1952).

west the anomaly decreases sharply in magnitude, but, though broadened and weakened, it may be traced by dip needle westward to the NW¼ sec. 12, T. 39 N., R. 29 W., where Vulcan Iron-Formation, flanked on both sides by Michigamme Slate, was found in a north-inclined diamond-drill hole. The drill-hole log by Stephen Royce, of Pickands, Mather and Co., indicates that fault breccia separates the Vulcan from the Michigamme in the north. Vulcan Iron-Formation shown in secs. 2 and 3, T. 39 N., R. 29 W., is inferred from a magnetic anomaly on the projected trend of the anomaly in sec. 12, T. 39 N., R. 29 W., but not clearly a continuation of it (pl. 4). Near the west end of the anomaly, in the central part of sec. 3, T. 39 N., R. 29 W., a north-inclined diamond-drill hole cut a section of conglomerate, which may represent the basal Michigamme Slate, and penetrated the Randville Dolomite at depth. A fault breccia, reported to contain iron-formation fragments, lay between the conglomerate and the dolomite, but no iron-formation was found.

East of the Loretto mine, in the Appleton mine area, the Vulcan Iron-Formation steps down to the south through a series of right-lateral faults and then trends a little south of east to the east edge of the mapped area. The repetition of the iron-formation in the Appleton area is suggested by the double-crested magnetic anomaly. To the east of this area the anomaly is linear, broken only by one minor offset. The area underlying the crest of the anomaly in sec. 16, T. 39 N., R. 28 W. has been drilled, and iron-formation was found "at ledge" beneath Cambrian sandstone, but the information obtained does not warrant dividing the Vulcan Iron-Formation in this eastern belt.

DISTRIBUTION AND RELATIONS ALONG THE SOUTH IRON RANGE

At the Norway city dump at Brier Hill (NW¼ sec. 9, T. 39 N., R. 29 W., pl. 2), in a series of low outcrops, trenches, and test pits, one can trace the classic Vulcan section as originally set down by W. S. Bayley (1904). On the north is a bold ridge of south-dipping dolomite; south of the dolomite is 100 feet of swamp, and, on the north bank of the swamp, a low ledge of south-dipping Traders Iron-Bearing Member which shows a selvage of Felch slate and quartzite on the north side. A traverse southeast from the Traders Member ledge will cross the other members of the formation except the Loretto Slate Member in normal sequence (see pl. 2). The Brier Slate Member is particularly well exposed, and it is from this area that it takes its name.

East of the Brier Hill area to the vicinity of the Sturgeon River (pl. 3), the Vulcan strata lie close to the surface along the south slope of the Brier Hill, where they have been mined and extensively explored by test pits. In part of this area the Traders Member and the lower part of the Brier Member are covered by Cambrian sandstone. In the Vulcan mine area, SW¼ sec. 11, T. 39 N., R. 29 W., the Loretto Slate Member is present.

Farther to the east, beyond the Sturgeon River, the Vulcan strata are deeply covered by glacial deposits for the next 3 miles, but extensive diamond drilling permits a fairly good concept of the structure and distribution of members. Drilling in the SE¼ sec. 17 indicates that structural complications persist that far east, but their exact nature could not be determined, and therefore the Vulcan members remain undifferentiated. From the location of the above-mentioned drilling east to the first test pits in Vulcan rocks, near the east border of the mapped area, the location of the Vulcan is plotted from magnetic data (pl. 4).

To the west of the Brier Hill area, for more than one-half mile, the Vulcan rocks are not exposed but they are again exposed in places westward to the Munro mine (pl. 2). The geology of this covered half-mile segment has been compiled from maps of the upper level of the abandoned Aragon mine and the records of surface drill holes. This segment ends against a northeast-trending fault zone, west of which exposures of the Vulcan rocks are adequate to trace the various members through the complicated structural pattern in the Norway and Cyclops mines area. The magnetic survey in the area east of the above-mentioned fault zone at Norway gives a general picture of the distribution of the Vulcan Iron-Formation (pl. 4). As in most of the district, the Curry Member underlies the crest of the anomaly, but at several places the Traders Member produces a small shoulder anomaly of a few degrees. The broad elliptical area of high magnetics in the north part of Norway, NE¼ sec. 8, reflects the location at depth of the folds of iron-formation which plunge toward the west in the Aragon mine area. In the vicinity of shaft 5, Aragon mine (2,500 feet W., 1,250 feet S. of the NE cor. sec. 8), the crest of the anticlinal fold in the Curry Member lies beneath approximately 500 feet of Michigamme strata.

Magnetic surveying in the triangular fault block north of Norway gives no clue to the bedrock geology because the iron-formation is oxidized and nonmagnetic, but near the apex of the block the Curry Member is marked by a distinct anomaly that can be traced westward to the vicinity of the abandoned Munro mine (pl. 2).

Magnetic anomalies at the Munro mine present a difficult problem. East of the mine the crest of the magnetic anomaly is definitely centered over the Curry

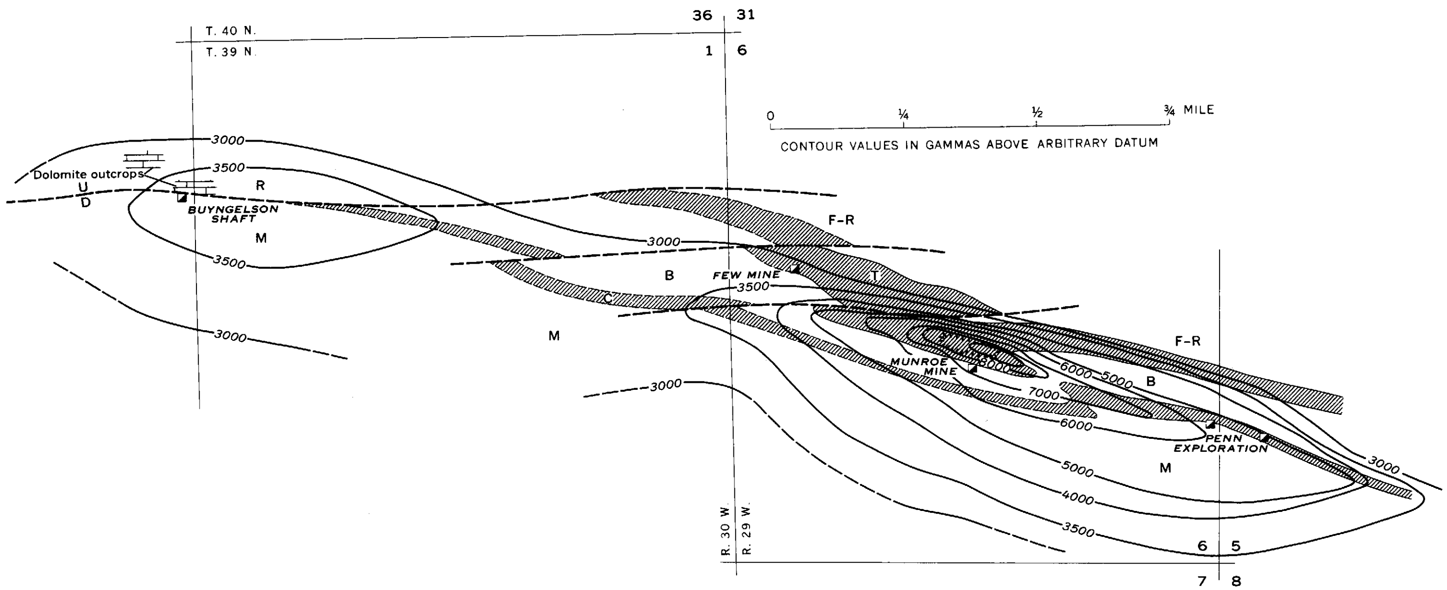

Figure 18.—Geologic and magnetic map of the Munroe Mine area. F-R, Felch and Randville Formations: T. Traders Iron-Bearing Member; B, Brier Slate Member; C, Curry Iron-Bearing Member; M. Michigamme Slate. Magnetometer survey by W. C. Prinz.

Member, and what appears to be the same crest is centered over the iron-formation at the Munro mine (fig. 18), but the iron-formation at the mine is clearly underlain by the Felch Formation and overlain by the Brier Slate Member, and therefore is the Traders Member. Apparently the Traders Member, which ordinarily is much less magnetic than the Curry, has been greatly thickened by folding and faulting at the mine, and the resultant increase in magnetic intensity has caused a shift of the magnetic crest to the north. The Curry Member very likely lies south of the mine in its normal stratigraphic position, but is not detectable magnetically above the background of the larger anomaly to the north. At the west end of the Munro open pit the magnetic crest and the Traders Member diverge; the Traders Member trends a little north of west through the Few mine, whereas the magnetic crest trends nearly due west and diminishes rapidly in magnitude. It is presumed that the magnetic crest has again shifted to the Curry Member, as indicated by the inferred structure on the geologic map (pl. 2 and fig. 18), but alternative interpretations could be made.

The Vulcan Iron-Formation apparently is absent at the surface in the Bryngelson shaft area (fig. 18). The Bryngelson shaft, in the NE¼ sec. 2, T. 39 N., R. 30 W., was sunk in sheared graphitic slate of the Michigamme Slate which is in fault contact with the Randville Dolomite. A diamond-drill hole 1,800 feet east of the Bryngelson shaft was reported to have cut 30 feet of Vulcan Iron-Formation lying between the Randville Dolomite and the graphitic slate of the Michigamme. Magnetometer surveying in this section indicates that the iron-formation is present at depth along the fault as far west as the Bryngelson shaft. Theoretically, both iron-formation members of the Vulcan should be found at depths south of the fault anywhere in sec. 2. W. S. Bayley (1904, p. 372) thought that the dolomite at the Bryngelson shaft area formed a topographic high during Vulcan deposition and that the Vulcan strata were not deposited. His interpretation still remains a possibility, but, in the writers' opinion, an unlikely one on the basis of the fault patterns displayed at the Munro mine, east of the Bryngelson shaft, and also beneath the surface at the Fumee Creek exploration in the NW¼ sec. 2, T. 39 N., R. 30 W., west of the above shaft. Section *B–B′*, (pl. 2) shows the fold and fault relations in the explored area.

West from the Fumee Creek area the Vulcan strata form a simple homoclinal belt to the vicinity of Quinnesec. The main magnetic crest along this belt is, as usual, above the Curry Member, but at Quinnesec there is a distinct shoulder anomaly over the Traders Member as well.

The anomaly marking the location of the Curry Member ends approximately 3,000 feet west of the NE cor. sec. 3, T. 39 N., R. 30 W., and as far as can be determined the Curry Member is absent from there west to the vicinity of the Pewabic mine, NE¼ sec. 31, T. 40 N., R. 30 W. (pl. 1). At the abandoned Keel Ridge mine in the SE¼ sec. 32, T. 40 N., R. 30 W., the Michigamme Slate rests on a subnormal thickness of Brier Slate Member, and south of the Vivian mine, SW¼ sec. 34, T. 40 N., R. 30 W., a similar relation is inferred. The dip-needle survey along the belt shows a well-defined linear anomaly that rises to a crest over the Traders Member (pl. 4), but no anomaly indicating the presence of the Curry Member is recognized.

The absence of the Curry Member along this belt may be due to faulting, nondeposition, or pre-Michigamme erosion. The Vulcan rocks, wherever they may be seen along the belt, are overturned and dip north (pl. 4). Under these conditions, a vertical or north-dipping fault through the Brier Member could place the Curry Member beneath the Traders Member in such a way that both would be represented by a single magnetic crest over the Traders Member. The asymmetry of the profile of this magnetic anomaly with its low-angle side dipping into the area of dolomite indicates that this type of fault is a real possibility. The writers, although they do not wish to appear too emphatic, prefer to consider the absence of the Curry Member to be a result of pre-Michigamme erosion. This concept is based on the following observations:

1. The pre-Michigamme erosion appears to have regional importance. Within the district it best explains the absence of the total Vulcan Iron-Formation in the area of Iron Hill (pl. 2) and along the north belt of the dolomite (pls. 2 and 3).
2. Where the contact of the Michigamme Slate and Brier Member has been seen at the Keel Ridge mine, it does not appear to be a fault contact. Neither, however, could it be interpreted as an uncomformable contact.
3. At Quinnesec, 2,100 feet west and 400 feet south of the NE cor. sec. 3, T. 39 N., R. 30 W., there is an exposed contact between the Curry Member and the Michigamme Slate that does appear unconformable. The Curry Member crops out on the north side of an old railroad grade, and the south surface of the outcrop (top of formation) is quite irregular. At one place on this irregular surface, an argillaceous conglomerate fills holes

in the underlying iron-formation, and graphitic pyritic slate overlies the argillaceous conglomerate. Moreover, the upper few feet of this iron-formation are partly pyritized, and the pyrite has replaced the original oxides of iron. The foregoing relations are interpreted to mean that the Curry Iron-Bearing Member which was deposited in an oxidizing marine environment, was consolidated, raised above sea level, eroded slightly, and subsequently was again submerged, this time into the very reducing environment of the pyritiferous graphitic slate. The iron oxides of the iron-formation, being unstable in the new environment, became pyritized.

Between the Quinnesec and Vivian mines (pl. 2) the Traders Member is present in three belts of overturned rocks which dip north. The north belt is the north limb of a west-plunging syncline, the middle belt is anticlinal and converges to the west with the north belt, and the south belt is anticlinal and plunges beneath the Brier Slate Member to the west (pl. 2, section *A–A′*). Refinement of previous geologic mapping done in this area by Lamey and Dutton (1941) indicates that the several faults inferred by them probably do not exist. Two of the proposed faults, one at either end of the Vivian mine, would cross mine workings which show no offset in the beds. The increased thickness of the Traders Member in the Quinnesec area causes a broad magnetic high having an asymmetrical profile which reflects the overturned north dip of the iron-bearing beds. To the west the profile narrows, but the fact that the asymmetry is the same indicates that the overturn persists at least as far west as the Keel Ridge mine (pl. 1).

Logs of widely spaced diamond-drill holes along the iron-formation belt between the Vivian mine and the Keel Ridge mine, particularly from holes in sec. 33, indicate that the iron-formation is unusually thin but continuous, and nearly everywhere underlain by slate and quartzite of the Felch Formation. Descriptions of the rocks south of (above) the iron-formation indicate they represent Brier Slate Member; the iron-formation therefore is probably the Traders Member and not the Curry Member as indicated on an earlier map (Lamey and Dutton, 1941), and the inferred fault shown on the above-mentioned map between the Randville Dolomite and the Curry Member (now interpreted as Traders) probably does not exist. To reiterate a point previously made, the Curry Member is probably absent from this belt owing to post-Vulcan but pre-Michigamme erosion. The Michigamme strata overlie the Brier strata along this belt, but the exact location of the contact between them can be determined only at the Keel Ridge mine.

At the Keel Ridge mine the exposed section is homoclinal, is overturned, and dips north. The succession from north to south is Randville Dolomite, Felch Formation, Traders Iron-Bearing Member of the Vulcan (mapped as Curry Member by W. S. Bayley, 1904), Brier Slate Member, and Michigamme slate. This succession is well exposed on the north side of U.S. Highway 2 in the SE¼ sec. 32 (pl. 1). The contact between the Brier Member and the Michigamme Slate is shown on excavated slopes on the north side of what is now an auto yard.

West from Keel Ridge the Traders Member strikes approximately N. 60° W. into the area of the Pewabic mine, but south of the Pewabic mine it bends sharply around to the east in a well-defined west-plunging anticline-syncline couple. Only the extreme west area of the anticline is inferred. If the attitude of the iron-formation is projected from the Keel Ridge area, the south limb of the anticline is overturned and dips about 80° N. The north limb of the anticline is known from the Pewabic mine maps to be in normal position and to dip about 50° N., and both limbs are underlain stratigraphically by the slate and quartzite of the Felch Formation, which in turn is underlain by dolomite. The Pewabic mine is in the synclinal part of the couple. Mine maps indicate that the structure is somewhat complicated by secondary folding and shearing in the trough. The south limb (north limb of adjacent anticline) dips about 50° N.; the north limb is nearly vertical. The fold plunges about 35° W. and iron-ore bodies mined in the trough of the fold showed a similar direction of plunge. Vertical drilling on the lowest level of the mine showed that Randville Dolomite lay beneath, as would be expected from the structure.

From the Pewabic mine the Traders Member trends westward into the area of the Walpole and Chapin mines. At the Walpole mine it forms another anticline-syncline couplet similar to that at the Pewabic mine, but it is joined by the Curry Member, which forms parallel structures. The exact place where the Curry enters the section is not known and, as far as the writers know, there has been no exploration in the critical area. The Curry Member is not present within the Pewabic syncline in the mine area. This member must terminate in a synclinal hook approximately as shown on the map, but its shape and exact location are unknown.

In the area of the Walpole, Millie, and Chapin mines, the several members of the Vulcan make a parallel series of very systematic anticlines and synclines which are overturned, for the most part, and which plunge 20°–30° W. The geologic map of this area represents a compilation of new surface mapping and judicious use of mine maps; caved areas have been reconstructed to represent the conditions prior to mining. The fold pattern as indicated at the surface persists to the lowest level of the Chapin mine, at a depth of approximately 1,800 feet. Only two small exposures of Traders Member are known in this area, one at each end of the Chapin caved area along the south side, both in contact with Brier Slate Member. Traders Member iron-formation is also present on the dumps of shafts and test pits in the Walpole mine area. The Brier Slate Member is exposed at three places along the south side of the Chapin caved ground and in a pit near the Millie mine shaft. Test pits in the Brier strata are confined chiefly to the area immediately east of the Millie mine. Shaft 3 of the Walpole mine (1,300 feet S., 500 feet W. of NE cor. sec. 31) was sunk in Brier Slate Member. and a test pit, 150 feet south of that shaft, is in the same rock.

The Curry Member underlies all the west part of Millie Hill and is well exposed at the west end of the hill in a small opencut locally referred to as Millie pit. The Curry Member is also exposed in patches along the south side of that hill and in pits near the Millie mine shafts. Numerous test pits have ended in Curry Member iron-formation in the trough of the syncline east of the Millie mine and along the southeasternmost extent of the Curry Member. Near the east end of the south slope of Millie Hill (NW¼NE¼ sec. 31) a small outcrop of magnetic ferruginous slate above the Curry Member is tentatively placed in the Loretto Slate Member of the Vulcan Iron-Formation.

The magnetic anomaly which marks the Traders Member at the Keel Ridge mine dies out toward the northwest in the area of the Pewabic anticline (pl. 4). The Pewabic syncline shows no distinct magnetic crests, nor does that part of the Walpole syncline underlain by the Traders Member. The high magnetic anomaly is caused by the Curry Member in the Millie Hill area, as at other places where the Curry Member is present. The anomaly is high and broad, and at the east end of the hill it is bilobate—a feature that reflects the fold structure. The anomaly over the Curry Member ends south of the east end of the Bradley open pit, and the Curry Member appears to be very thin and discontinuous west of there.

As the magnetics over the Curry Member wane, an anomaly of lesser magnitude emerges on the shoulder over the Traders Member near the west end of the Chapin caved ground and continues westward to the end of the range at the Menominee River. Along this belt the Traders Member is well exposed at the Bradley pit and in the pit of the Federal mine (west Chapin mine). In both of these open pits the Traders Member, although folded internally, stands in nearly vertical position between the Felch Formation on the north and the Brier Slate Member on the south. Test pits west of the Federal mine indicate that the Traders and Brier Members continue near the surface at the south edge of the Cambrian sandstone as far west as the north-south road through sec. 26. On the east side of the road, at the place where the belt crosses, surface rubble about a grass-roots outcrop shows the approximate location of the Traders-Brier contact. Logs of diamond-drill holes put down along the belt west of the Bradley pit show that the Traders and Brier Members are present at least to the NW¼ sec. 26. The first two holes west of the Federal mine cut a thin iron-formation above the Brier, which may be the Curry Member. Drilling in the NE¼ sec. 26 shows that the Traders Member is repeated, probably by faulting. West of the Menominee River, in sec. 22, T. 39 N., R. 19 E. (Wisconsin), the fact that diamond-drill holes spaced in a line perpendicular to the strike of the Vulcan belt all ended in Michigamme strata indicates that the Vulcan does not continue at the bedrock surface very far beyond the river. The same relation is indicated by the magnetic survey. The high anomaly over the Vulcan rocks ends at the river. The broad gentle slope to the west shown by the magnetic survey possibly indicates a west-plunging anticlinal hook in the Vulcan rocks, as depicted on the map (pl. 1).

BARAGA GROUP

The Baraga Group, named for Baraga County, Mich. (James, 1958, p. 30), comprises most of the strata referred to the Upper Huronian of previous reports. In the Marquette district the group includes, in ascending order, the Goodrich Quartzite and the Michigamme Slate; the lower part of the Michigamme includes the Bijiki Iron-Formation Member, the Clarksburg Volcanics Member, and the Greenwood Iron-Formation Member. In the Menominee district the group includes the Michigamme Slate and the overlying Badwater Greenstone.

MICHIGAMME SLATE

The Michigamme Slate underlies an enormous area in the northern Peninsula of Michigan and forms a valuable stratigraphic connecting link between the widely separated iron-bearing districts. The type area is the vicinity of Lake Michigamme in Baraga County

(Van Hise and Leith, 1911, p. 267). In the Menominee district the Michigamme Slate was previously designated "Hanbury Slate" by W. S. Bayley (in Van Hise and Leith, 1911, p. 340).

Distribution

The Michigamme Slate is by far the most widespread of the sedimentary formations in the district, but so poorly exposed that its depicted areal limits are more often based on the distribution of neighboring formations and structure than on any physical evidence of its presence.

Three main belts are underlain by Michigamme strata; one south of the south iron range, one between the iron ranges, and one north of the north iron range; the first two extend the length of the district. The north belt divides into two belts where it passes beneath the younger infolded Badwater Greenstone (pls. 1, 2). The belt south of the Badwater Greenstone and the other two belts of Michigamme Slate coalesce west of the iron ranges into a single belt which continues into the Florence, Wis. district.

Description

The Michigamme Slate consists chiefly of slate, especially quartzose, micaceous, and graphitic types, but it also includes large quantities of graywacke, subgraywacke, quartzite, conglomerate, dolomite, dolomitic slate, dolomitic quartzite, and some iron-formation. The conglomerates are described first, then the fine- and medium-grained nondolomitic clastic rocks, then dolomite and dolomitic rocks, and lastly the iron-bearing rocks.

Conglomerate.—Thin conglomeratic argillite is present at the base of the Michigamme Slate in two of the three places where the base of the formation is exposed at the surface, at the Cundy mine at Quinnesec, and in the West Vulcan mine pit at Norway (pl. 2). A similar conglomeratic bed is exposed above the Curry Member of the Vulcan Iron-Formation, but not in contact with it, southwest of the Cyclops mine at Norway. No conglomerate is present at the third place, near the Keel Ridge mine east of Iron Mountain (pl. 1), where the basal beds are sericitic slate that rests with apparent conformity on the Brier Slate Member of the Vulcan Iron-Formation.

The conglomeratic bed at the Cundy mine and in the West Vulcan mine pit is probably only a few feet thick. It is fine grained, dense, gray green, and mottled with hematite-red streaks that parallel a poorly formed schistosity. The pebbles are sparse and small (½–2 in. in diameter) and are composed of argillite or, rarely, of quartz or chert. None of the pebbles seen is definitely distinctive of any older formation. The matrix of the rock is so fine grained that it is not clearly resolved even under the microscope. It appears to be composed of randomly oriented flakes of pale-green chlorite and angular grains of quartz in a reddish clay base. Throughout this fine-grained matrix are some larger particles, among which are many euhedral crystals of magnetite, some cubic crystals of pyrite spaced about one-half inch apart, and a few silt-size grains. A few of these small grains are similar to the matrix; a few consist of angular particles of albite and material that resembles reddish altered potassic feldspar. These feldspathic grains closely resemble some parts of the Brier Slate Member. Under the microscope most grains resolve into spheres of reddish clay and quartz surrounded by granules of magnetite. The spheres may represent altered grains of some ferruginous rock, such as slate of the Brier Member or iron-formation.

A conglomerate bed 50–70 feet thick, penetrated in diamond drilling east of the Sturgeon River at Loretto, was assigned to the basal Michigamme by Allen (1919). According to Allen, the conglomerate was cut in three drill holes, but we have been able to find the records for only two of these. They are the southernmost holes in sec. 13, T. 39 N., R. 29 W. (pl. 3). The holes were apparently inclined toward the north. One cut the conglomerate beneath 83 feet of Pleistocene gravel, and the other (westernmost hole) cut the conglomerate after passing through 161 feet of Pleistocene gravel and 110 feet of Michigamme-type slate. In both holes the conglomerate was underlain by banded gray slate (Loretto Slate Member of the Vulcan Iron-Formation) and the other members of the Vulcan Iron-Formation in their proper stratigraphic order. Allen (1920, p. 201) described the conglomerate as being "composed of fragments of white dolomite, chert, and quartz derived from the Lower Huronian dolomite (Randville), intermixed with red, black, and brown jasper, ferruginous slate and quartzite, in which are represented all of the members of the underlying Vulcan Group of the Middle Huronian."

No similar conglomerate is known from anywhere else in the district, even though the base of the Michigamme is exposed at three places along the south iron range and has been drilled through at more than 40 other places. If basal Michigamme, the conglomerate of Allen is unique in composition with respect to known basal Michigamme strata elsewhere in the district and must be restricted to a very small area.

In the northwest part of the district, in secs. 13, 14, 15, and 24, T. 40 N., R. 30 W. (pl. 1), six diamond-drill holes of the Pine Creek exploration (Pickands, Mather and Co.), as recorded by Stephen Royce, passed through typical Michigamme Slate and graywacke into a 20- to 30-foot bed of basal conglomerate, and thence into Randville Dolomite. The logs of each of these holes show that the conglomerate contains dolomite and quartzite pebbles in an argillaceous or graywacke matrix similar to the Michigamme strata immediately above it.

An isolated group of outcrops of quartz pebble conglomerate in sec. 3, T. 39 N., R. 29 W. (pl. 3), has been included in the Michigamme Slate by the writers, but the assignment is tentative. A diamond-drill hole directed north near the west end of the group of outcrops passed through the conglomerate and into Randville Dolomite below. The conglomerate is composed entirely of quartz and quartzite pebbles in a quartz matrix and is unlike any other conglomerate seen in the area. Its composition suggests that the material was derived from the Sturgeon Quartzite. The decision to assign it to the Michigamme Slate rests chiefly on its location with respect to the dolomite, and to the Vulcan Iron-Formation east of it, the latter inferred from the magnetic survey. An alternative interpretation, favored by Lamey, is that it belongs to the Felch Formation. Not enough information is now available to assign the conglomerate firmly to either formation.

Nondolomitic clastic rocks.—Most of the formation consists of nondolomitic clastic rocks of fine or medium grain. Slate of various kinds is by far the most abundant type, but locally graywacke, subgraywacke, and quartzite are also abundant. (The term "slate" is used in the provincial sense to denote most thin-bedded, laminated or fissile, fine-grained rocks, whether or not they show slaty cleavage; slightly schistose rocks, such as phyllite, are also included.)

The chief varieties of nondolomitic slate are quartzose (the most usual type), sericitic, and graphitic. The slate is mostly dark colored, gray or black, but some is light colored, white or pale red, or white and red mottled, the "calico slate" of W. S. Bayley (1904, p. 463). The quartzose slate is composed chiefly of quartz, sericite, chlorite, and biotite. The principal accessory minerals are hematite, magnetite, apatite, pyrite, graphite, and tourmaline. The sericitic slate contains, of course, a large amount of sericite along with some other micaceous minerals, whereas the graphitic slate contains considerably more than the usual amount of graphite. The sericitic and graphitic slates are commonly associated with subgraywacke and quartzite, and rarely with ferruginous chert or iron-formation.

The quartzite and subgraywacke form beds as much as a few feet thick, are fine to medium grained, and are generally dark colored. They are composed predominantly of quartz and varying amounts of feldspar, sericite, chlorite, biotite, and magnetite. A carbonaceous quartzite was uncovered in a test pit of the Traders mine. Some of the subgraywacke is phyllitic, and several specimens examined under the microscope are microschistose and show a marked alinement of the quartz grains between mica folia.

All the slate is strongly deformed. In almost any outcrop the beds and folia strike nearly parallel to the iron ranges and dip northeast or southwest at steep angles, and they commonly show plications and minor folds which may plunge northwest or southeast.

The distribution of the various types of slate is somewhat sporadic. The sericitic slate is sparingly exposed over a broad area and apparently is not characteristic of any particular part of the formation. Typical exposures may be seen (1) along the banks of the Menominee River south of the railroad bridge north-northwest of Iron Mountain and at the Keel Ridge mine (pl. 1), (2) in the group of outcrops near Iron Hill (pl. 2), and (3) along the banks of the Sturgeon River west of the Loretto mine (pl. 3). The graphitic slate appears to be most abundant near the base of the formation, within 500 feet of the Vulcan Iron-Formation. Outcrops of these black rocks are few, but they have been exposed by numerous test pits and penetrated in many drill holes. In the north part of the district, typical graphitic slate may be seen, from west to east: where the Michigamme Slate is in contact with the Badwater Greenstone along the Menominee River, in test pits west of the Traders mine (pl. 1), in the groups of outcrops near Iron Hill (pl. 2), and on the south bank of the Sturgeon River west of the Loretto mine (pl. 3). Along the south iron range graphitic slate is exposed at the Bryngelson shaft (NE¼ sec. 2, T. 39 N., R. 30 W.), in test pits south of the Vivian mine at Quinnesec, in the Cundy mine shaft at Quinnesec, on the south side of the West Vulcan mine pit (pl. 2), and at a few other places.

In the northwest part of the district, where the bedrock is almost completely covered by Pleistocene deposits, extensive diamond drilling has revealed that the Michigamme strata there are chiefly gray to black banded slate and interbedded graywacke, and graphitic slate. Banded calcareous slate, dark-colored dolomite, and brown medium-grained quartzite occur locally. These strata grade downward into a basal conglomerate.

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Dolomitic rocks.—Dolomitic quartzite, dolomitic shale, and dolomite occur chiefly in the broad belt of outcrops south of Hanbury Lake (pls. 2 and 3).

Dolomitic quartzite occurs south of Hanbury Lake only, where it is associated with dolomitic slate and dolomite and with intrusive metagabbro. The quartzite beds appear to be confined to the eastern three-fourths of the group of outcrops south of the lake, probably because the quartzite beds lens out to the west or are doubled back in a fold. Numerous minor folds in the slate show small areas where the beds strike north across the overall northwest foliation, and folding is thus indicated as the more likely cause of the limited distribution of the quartzite.

The dolomitic quartzite is dark gray or, if encrusted with limonite, brown. The beds are 1–10 feet thick and commonly show quartz-filled cross fractures that do not enter the adjacent slate beds. A distinctive characteristic of the quartzite is the presence of chips of black slate as much as 6 inches long in most beds. The rock is made up of about equal parts of well-rounded and well-sorted quartz grains and dolomite, and trace amounts of carbonaceous dust, magnetite, and pyrite. The quartz grains all show undulatory extinction when viewed under the microscope, a feature probably inherited from the source rocks inasmuch as the quartzite does not appear to be deformed internally.

The only exposed dolomite in the formation is confined to the belt of outcrops trending northwest from south of Hanbury Lake (pl. 2). The northwesternmost rocks on the belt are dolomitic slates which crop out in secs. 4 and 5, T. 39 N., R. 30 W. (pl. 1). South of Hanbury Lake the dolomite is light colored, banded, and somewhat slaty; it occupies the north part of the group of outcrops. The best exposed rock is at the lakeshore. The beds are folded, and in the northernmost outcrop the general strike is nearly normal to the trend of the outcrop belt; the dips are low to the southeast, because these beds are on the crest or in the trough of a minor fold. The lesser plications and folds on the beds plunge at low angles, less than 30° SE. West of Hanbury Lake, in the south part of secs. 7 and 8, T. 39 N., R. 29 W., are outcrops of siliceous and dolomitic gray slate and rather thick-bedded siliceous gray dolomite. A rind of limonite that coats the exposed surfaces of the dolomite indicates that the carbonate is probably ferruginous, an observation previously made by W. S. Bayley, who reported the chemical analysis shown in table 29. On the assumption that all the iron, magnesia, and lime form carbonates, W. S. Bayley gives the composition of the carbonate as about 9 percent $FeCO_3$, 41 percent $MgCO_3$, and 50 percent $CaCO_3$.

TABLE 29.—*Analysis, in percent, of ferruginous dolomite in Michigamme Slate*

[From W. S. Bayley, 1904, p. 480; analyst; George Steiger]

Constituent	*Entire rock*	*Soluble in HCl (dil)*
SiO_2	36.71	2.28
Al_2O_3	5.34	1.27
Fe_2O_3	.35	3.61
FeO	3.37	
MgO	10.78	8.83
CaO	15.11	15.20
Na_2O	.12	
K_2O	2.40	
H_2O at 105°	.55	
H_2O above 105°	1.61	
TiO_2	.27	
CO_2	23.22	
P_2O_5	.05	
MnO	.23	
Total	100.11	

Microscopic examination of a specimen of the thick-bedded dolomite shows that 80 percent is crystalline dolomite and the remainder mostly tiny widely spaced quartz grains, a few euhedra of pyrite, and enough graphitic dust to color the rock gray.

Lamey and Dutton (1939, p. 6) pointed out that the dolomitic strata south of Hanbury Lake resemble parts of the Randville Dolomite and that the associated slates and quartzites resemble Felch Formation rocks; they suggested that these older formations may have been faulted into their present position along a fault running through Hanbury Lake. No direct evidence has been found to indicate definitely that the strata south of Hanbury Lake are not in fact Randville and Felch strata. Their assignment to the Michigamme is chiefly due to R. W. Bayley's view that the ferruginous and graphitic gray dolomite and graphitic quartzite are not characteristic of the Randville. He concedes that the dolomitic rocks are unusual in the Michigamme, but regards them as representing a different facies from the slate-graywacke facies that characterizes the formation at most other places.

Ferruginous cherts and iron-formation.—Iron-bearing rocks of the Michigamee Slate occur at four localities. From northwest to southeast these are: (1) 2,100 feet north, 1,100 feet east of the SW cor. sec. 15, T. 40 N., R. 30 W. (near Badwater Greenstone, pl. 1); (2) 1,150 feet north, 300 feet west of the SE cor. sec. 32, T. 40 N., R. 29 W. (near Iron Hill, pl. 2); (3) west of the road near the center of sec. 19, T. 39 N., R. 28 W. (Turner's exploration); (4) 1,100 feet south, 500 feet east, and 2,200 feet south, 3,000 feet east of the NW cor. sec. 26, T. 39 N., R. 29 W. (just south of the border of the mapped area, pl. 3). The rocks will be discussed in the order stated.

In the first locality there is one shallow test pit in ferruginous rocks. Material collected from the dump is

FIGURE 19.—Photomicrograph showing thuringite(?) granules in quartz-carbonate nodules from near Iron Hill. Quartz, white; carbonate, gray; muscovite, gray rectangles; thuringite(?), black. Plane-polarized light; × 36.

gray chloritic quartzite, quartz-sericite slate, and quartz-martite iron-formation. Most of the rocks are stained red by hematite. The iron-formation is laminated, having layers 1 cm or less thick. It is composed of about equal parts of clastic quartz and euhedral martite, and a minor quantity of red iron-stained chlorite. Some of the iron-formation has a cherty appearance, and a few specimens show thin dark-gray chert layers which display a simple quartz mosaic when viewed under the microscope.

The iron-formation at this locality causes a distinct magnetic anomaly which, though discontinuous, seems to be traceable along much of the Michigamme-Badwater contact in Dickinson County and southeastern Iron County, and several explorations along this belt of anomalous magnetism have found ferruginous rocks of various kinds. These strata appear not to be anomalously magnetic to the west in Florence County, Wis., but diamond drilling and test pitting near the Michigamme-Badwater contact has revealed a thin iron-formation at one locality; there is, therefore, a zone of ferruginous rocks near the top of the Michigamme Slate over a very wide area.

At the second locality, near Iron Hill (pl. 2), the iron-bearing rocks are exposed in test pits in a narrow east-trending ferruginous zone. The zone has been explored for about 300 feet, most of it west of the north-south road. The rocks revealed in the pits are (1) red to gray vuggy ferruginous cherts showing rare seams and zones of rather pure hematite, (2) seams and zones of white and pale-gray chert-carbonate rock, which are remnants of the original rock that have not been iron-stained or replaced by hematite, and (3) rare zones of banded chert-magnetite rock. The ferruginous rocks are underlain by sericitic and graphitic slate, and overlain by quartzite, subgraywacke, and slate. The original material from which the highly ferruginous rock formed appears to have been a cherty, somewhat ferruginous and siliceous dolomite, containing an intergranular pale-green cryptocrystalline silicate mineral that resembles chlorite. The present lean iron deposits appear to have formed by oxidation, replacement, and enrichment by iron as a result of weathering and circulation of meteoric water. The deposit is not more than 50 feet thick and is probably about 10 feet thick. No iron ore has been produced from it.

Some unusual nodules are contained in sheared graphitic slate which lies close to the ferruginous rocks at the Iron Hill locality. They are generally biscuit shaped, a few inches in diameter, gray or white, and fine to medium grained, and are composed of carbonate, quartz, and a green chlorite mineral, possibly thuringite. The thuringite(?) forms granules in a quartz and carbonate groundmass, and most granules also include some quartz and carbonate and, rarely, plates of muscovite (fig. 19). The quartz, both within and without the granules, shows a fine-grained mosaic fabric and probably is recrystallized chert. The nodules resemble Mesabi greenalite iron-formation, and presumably they had a similar origin as chemical precipitates (Van Hise and Leith, 1911, p. 521–522).

The ferruginous rocks at the third locality are exposed in a few test pits and trenches (Turner's exploration, sec. 19, T. 39 N., R. 28 W.) in an open field west of the north-south road. The rock exposed in the pit is dense reddish-brown vuggy chert cut by numerous white quartz veins. Most of the rock is composed of quartz and hematite only, but one specimen contains minute grains of carbonate scattered throughout the quartz grains. The hematite in all the rocks appears to be secondary, replacing some other mineral, possibly the carbonate or an iron silicate. Schistose chloritic quartzite is the only other rock exposed by the test pits. In the quartzite the chlorite is very much altered and iron-stained.

In sec. 26, T. 39 N., R. 29 W., south of Vulcan, chert-siderite iron-formation crops out along the road at two places (Prinz, 1959, pl. 2). This rock is typical banded carbonate iron-formation, essentially the same as that which is particularly characteristic of the Iron River, Crystal Falls, and Florence iron districts. It is gray, but commonly coated with limonite or hematite, and here and there stained black by manganese oxide on fracture surfaces. It is well bedded and shows color banding. Layers of siderite or siliceous siderite, and layers of chert, as much as 1 inch thick, make up most of the rock. Fine-grained magnetite and a few grains of pyrite are present in some of the rock. The thickness of the formation cannot be determined from the outcrop because the flanking rocks are not exposed.

A magnetic anomaly caused by the magnetic part of the iron-formation can apparently be traced to the northwest, on the projected strike of the beds, for several miles, but with reasonable assurance for only about 1½ miles. We did not attempt to trace the formation to the southeast, but probably it extends in that direction also.

Relation to other formations

Several lines of evidence lead to the conclusion that the Michigamme Slate rests unconformably on the older formations. The more important of these are (1) the present limited distribution of the Menominee Group, (2) the apparent discordancies between post-Vulcan and post-Michigamme folds, (3) the regional aspect of the post-Vulcan but pre-Michigamme diastrophism, and (4) the presence of Michigamme basal conglomerates.

As for the limited distribution of the Menominee Group, along parts of the south iron range the two upper members of the Vulcan Iron-Formation are absent, and the Michigamme Slate rests on a subnormal thickness of Brier Slate Member of the Vulcan, whereas on the north iron range the Michigamme Slate rests on all the older units down to the Randville Dolomite.

Along the north belt of the dolomite, bordering the Carney Lake Gneiss, and in the northwest part of the mapped area, the Michigamme Slate rests on the Randville Dolomite; in the extreme northeast part of the area (pl. 3) the Michigamme rests on the Sturgeon Quartzite and the Carney Lake Gneiss. In the southern parts of the area covered by plates 1 and 2, it rests against the Quinnesec Formation, but the contact probably is a fault. The above-cited relations indicate a diverse pre-Michigamme terrane on which all the formations, down to the very oldest, were exposed. Whether all this terrane was once covered by the Menominee Group cannot be determined, but because the group is present in the Felch, Republic, Marquette, and Penokee-Gogebic iron-bearing districts of the region, which are widely separated, we assume that it was originally more widespread than at present and that great parts of it were removed by pre-Michigamme erosion.

Discordances between post-Michigamme folds and older folds also indicate unconformable relations between the Michigamme and the underlying strata. The homoclinal iron ranges are almost certainly related to a major pre-Michigamme anticlinal structure centering in the area of the Carney Lake Gneiss. The north iron range is part of the south limb of that structure, and the south iron range is but an upfaulted repetition of that limb. North of the north iron range the Michigamme strata rest unconformably on the Randville Dolomite. The dolomite apparently formed a broad west-pitching anticlinal nose in post-Vulcan time from which the overlying Vulcan strata were stripped, and the north iron range probably is the scarp or south limit of erosion from this anticline. The west, down-pitch limit is not known. At present, the Badwater Greenstone and the Michigamme Slate form a major post-Animikie syncline superposed on this pre-Michigamme anticlinal nose, and the older formations of the north iron range apparently form an adjacent faulted anticlinal ridge.

The regional aspect of the pre-Michigamme but post-Vulcan diastrophism is important here only insofar as it indicates the pattern to which the local tectonic events may be compared. If we accept the stratigraphic equivalence of the Vulcan of the Menominee district and the major iron-formations of the Marquette, Republic, and Penokee-Gogebic districts (Leith and others, 1935), then the apparently synchronous post-iron-formation diastrophism, so well illustrated in these widely separated districts, was regional in scope, and probably disturbed the rocks of the Menominee district as well.

The basal conglomerates of the Michigamme comprise inconclusive but permissive and supporting evidence for post-Vulcan but pre-Michigamme elevation and erosion. The conglomerate in sec. 3, T. 39 N., R. 29 W. (pl. 3), described on page 59, appears to be either basal Michigamme or basal Felch, and the age is too uncertain to be useful. The conglomeratic argillite at Quinnesec, Norway, and West Vulcan (p. 58) contains no fragments definitely known to have been derived from an older formation, and hence also is of little use as an indicator of a major unconformity. The conglomerate at Loretto (p. 58) described by Allen (1920) from drill holes in sec. 13, T.

39 N., R. 29 W., apparently contains fragments derived from most of the formations below the Michigamme Slate. This conglomerate apparently is unique and restricted to a very small area, but if it marks the base of the Michigamme Slate, as Allen states, it is an important indicator of a period of emergence and erosion before the deposition of the Michigamme strata. The conglomerate disclosed by diamond drilling in the northwest part of the district (p. 58) lies at the base of the Michigamme Slate, rests on Randville Dolomite, and contains dolomite and quartzite pebbles in a matrix similar to the Michigamme Slate. This conglomerate indicates that the strata down to the Randville Dolomite, as well as part of the dolomite, were removed by erosion before deposition of the Michigamme Slate.

BADWATER GREENSTONE

The Badwater Greenstone is a thick pile of basaltic volcanic rocks which overlies the Michigamme Slate. It occupies the area shown in the north part of plate 1 near Badwater Lake, hence the name (James, 1958, p. 37).

These volcanic rocks were correlated with the Quinnesec Formation and assigned to the Archean by W. S. Bayley (1904), but according to Van Hise and Leith (1911, p. 345), "Later work by G. W. Corey and C. F. Brown (unpub. field notes, 1905) has shown that they are really intrusive and extrusive in the Upper Huronian or in part contemporaneous."

Detailed mapping in southeastern Iron County and adjacent areas of Wisconsin south of the Brule River by C. E. Dutton, H. L. James, and K. L. Wier has demonstrated that the Badwater is stratigraphically above the Michigamme Slate and below the Paint River Group. The Badwater now includes all the Mastodon-Spread Eagle-Antoine belt of greenstone, which Leith and others (1935, p. 4) designated as doubtfully Middle Huronian, and the greenstone belt that underlies the Paint River Group and surrounds the general synclinal structure of the Iron River-Crystal Falls iron district.

Description

Texturally and mineralogically the Badwater Greenstone resembles Quinnesec greenstone, but contains no amphibolites, amphibole schists, or hornfelses, nor any rhyolitic material. The Badwater rocks were described in part by Williams (1890, p. 123–133), chiefly along the Menominee River, and in part by W. S. Bayley (1904, p. 159–167) as the Quinnesec schists of the western area, and most of his descriptions also are of rocks along the Menominee River. A few thin sections were studied during the present investigation, especially of rocks exposed some distance from the Menominee River, to supplement these earlier descriptions.

The formation comprises thick basaltic lava flows mainly and some chloritic slate. The rocks are greenish gray to black and form rock ridges and minor small outcrops that protrude above the general cover of glacial deposits. In some exposures the greenstone is schistose. The easternmost exposures (pls. 1 and 2) are ellipsoidal basalt; the attitudes of the ellipsoids in this area, as determined by C. A. Lamey, indicate the superposition of the Badwater on the Michigamme, and a major synclinal fold.

The massive greenstone examined in the sections is fine to medium grained and shows relict diabasic or ophitic textures, which are indicated by partially preserved crystals of plagioclase or by the outlines of completely altered ones. The chief constituents of most of these rocks are plagioclase, actinolitic amphibole, and minerals of the epidote group (epidote, clinozoisite, and zoisite). Remnant grains of sodic andesine and some sodic labradorite were recognized in about half of the specimens examined, and the former presence of the grains is indicatd by the shapes of areas filled with alteration minerals in the others. The alteration minerals of plagioclase are albite, epidote group minerals, and sericite. In addition, all specimens examined contain sphene, and some contain chlorite, calcite, ilmenite, and pyrite. Most of the amphibole is nearly colorless and only faintly pleochroic; the colors commonly exhibited are pale green, pale greenish gray or grayish green, very pale gray, and very pale blue. The maximum extinction angle is 13°–19° in most specimens. Most of the specimens are cut by small veins composed of quartz, chlorite, epidote minerals, or calcite, either alone or in various combinations of two or three of these minerals.

The schistose greenstone is composed chiefly of (1) chlorite and minerals of the epidote group, and thus is chlorite schist; or (2) of actinolitic amphibole and minerals of the epidote group, and some chlorite, sphene, and minor brown biotite, and can thus be classed as green schist. In only one of the three specimens of schistose greenstone were the outlines of feldspar recognized, although this specimen was completely altered to epidote minerals, chiefly clinozoisite.

The two specimens of slaty greenstone examined are very fine grained, and the original texture is not apparent. They show a definite parallel alinement of the constituents and an arrangement of material in small lenses, some of which join and form layers; also, many small grains, about 0.02–0.04 mm across, are present, at least some of which are detrital. Neither amphibole nor plagioclase is present, and the amount of epidote minerals is small. The fine grain of the rocks precludes positive identification of minerals, but these specimens seem to be composed almost wholly of angular grains and lenses of quartz, lenses of a carbonate (probably calcite), wisps of chlorite, many small grains of an opaque material, and a small amount of biotite. In addition to the carbonate in the lenses, both rocks are cut by minute veins of carbonate.

Thickness

The thickness of the greenstone is estimated to be 5,000–8,000 feet for the district. The greenstone is several miles thick to the northwest, but due west (in Wisconsin) the south limb of the greenstone syncline thins to less than 1,000 feet within a few miles. The variations in thickness probably reflect differences in the original dimensions of the lava pile.

Relation to other units

Over a very extensive area the Badwater Greenstone rests on the Michigamme Slate and beneath slates of the Paint River Group. Both upper and lower contacts are conformable, or apparently so. The uncommonly thin Michigamme Slate beneath the Badwater syncline, as shown on sections *A–A′* (pl. 1) and *A–A′* (pl. 2), suggests that erosion of some of the Michigamme strata may have preceded the Badwater volcanism, but we think that the thin Michigamme section more likely was caused by (1) deposition of a less than normal thickness on a pre-Michigamme topographic high, (2) greater than normal compaction of the slate by the overlying volcanic rocks, and, possibly, (3) removal of beds by submarine slumping in Michigamme time.

CONDITIONS OF DEPOSITION OF THE BARAGA GROUP

The Michigamme and Badwater Formations at most places are of the eugeosynclinal facies; that is, these formations were deposited in deep water in a structurally unstable basin. Thus, they contrast sharply with the underlying Animikie formations, which consist chiefly of well-sorted terrigenous clastics and chemical precipitates—products of shallow-water deposition in a very stable basin.

In the iron districts, which lie mostly on the margins of the Animikie basin, the sudden geologic change from stable shelf to eugeosynclinal sedimentation indicates that, after the post-Vulcan disturbance and moderate subareal erosion in some quarters, the basin foundered generally, and marginal areas that were shallow shelf in previous epochs became areas of deep water. Contemporaneous uplift of marginal lands is indicated by the more than 5,000 feet of Michigamme strata deposited in Iron and Dickinson Counties before the Badwater volcanics were extruded into the sea floor. One of these marginal land areas probably lay to the south of the Menominee district, in Wisconsin. Two lines of evidence lead to this conclusion: (1) The Michigamme Slate pinches out toward the south end and is not present south of Iron River, Mich., and (2) from northwest to southeast, in the Menominee district, the Michigamme strata became more shelflike in character, as if a land mass were being approached. This shoreward facies of the Michigamme is a unique remnant; doubtless similar rocks skirted the basin at one time, but were removed by post-Animikie erosion. In summary, we believe the Baraga Group was deposited in a subsiding basin having considerable bottom relief. Minor depressions and troughs within the basin were the sites of graphitic and pyritic slate deposition. Slate deposition was continuous over the floor of the basin but penetrated at close intervals by turbidity flows of quartzose clastics from the basin slopes. On a shelving area near the south margin of the basin, in the south part of the Menominee district, mainly slate was deposited but also abundant subgraywacke, quartzite, dolomitic slate, and ferruginous dolomite, which are shelf equivalents of the slate-graywacke suite in the basin bottom. Near the end of Michigamme sedimentation the basin had stabilized to the extent that the influx of clastics nearly stopped, and some chemical iron-formation was deposited; but this interlude of quiescence was soon shattered by the extensive outpouring of submarine lavas of the Badwater. After Badwater time, sedimentation continued in at least the south part of the basin until the Paint River Group was deposited.

PAINT RIVER GROUP

Rocks of the younger Paint River Group, which overlies the Badwater Greenstone in the neighboring Florence, Wis., district, are not definitely known to be present in the Menominee district, but they could be present in the southernmost part of the south belt of Michigamme Slate as mapped by us. Outcrops are so scarce along that part of the belt that the character of the rocks is much in question. The fact that a chert-siderite iron-formation is exposed along the belt (p. 62) enhances the possibility that the Paint River rocks are present, because the main iron-formation of that group is of the same type; however, iron-formation of that type could as well lie within the Michigamme Slate. If Paint River strata are present along the south belt of Michigamme Slate, then the Badwater Greenstone, which lies between the Michigamme Slate and the Paint River Group, could be present as well. A single outcrop of greenstone, which we have assigned tentatively to the Michigamme, but which could possibly represent the Badwater Greenstone, occurs in sec. 3, T. 39 N., R. 31 W. (pl. 1).

The simple homoclinal structure of the south iron range suggests a normal sequence of formations southward, toward the major fault against the Quinnesec Formation. There appears to be room enough between that range and the above-mentioned fault for the Michigamme Slate, for the Badwater Greenstone, and for at least part of the Paint River Group, but the present evidence is too scanty to warrant a more definite statement.

UPPER MIDDLE PRECAMBRIAN INTRUSIVE ROCKS

Four main types of intrusive rocks that appear to be of late Animikie or post-Animikie age have been recognized in the district. From oldest to youngest they are metagabbro dikes and sills, quartz diorite, granite, and pegmatite dikes. Of these the metagabbro dikes antedate the post-Animikie folding and metamorphism, the quartz diorite and granite are probably syntectonic, and the pegmatite dikes are postgranite. Also present are abundant quartz veins.

METAGABBRO DIKES AND SILLS

Metamorphosed gabbro intrusives are widely but sparingly distributed throughout the Menominee district and adjoining areas. The largest bodies have been indicated on the several plates, but many smaller bodies are not shown. The major occurrence is in the south part of the district where long sill-like masses, several thousand feet thick, have penetrated the Quinnesec strata (pls. 1, 2). Some large dikes cut the Carney Lake Gneiss and Sturgeon Quartzite (pls. 2, 3), and many small dikes have been noted in the Randville, Vulcan, Michigamme, and Badwater Formations.

DIKES

The dikes display considerable variation of texture and mineral composition, some the result of original formation, some the result of deformation, metamorphism, and other alteration. The rocks are mostly greenish black but weather brown, are massive and medium to coarse grained, and commonly exhibit diabasic or ophitic texture; some have been strongly deformed and are fine-grained schists.

Two groups of dikes are recognized on the basis of whether their present chief constituents are (1) actinolite and albite or (2) hornblende and oligoclase. The dikes of the first group are confined to the Animikie strata and the south edge of the Carney Lake Gneiss, whereas the dikes of the second group are mainly in the north part of the Carney Lake Gneiss. The boundary between these two groups is shown on the maps by the greenschist oligoclase-amphibolite facies line.

The massive dikes of the first group are nearly the same everywhere. They are mainly diabase or ophite of medium to coarse grain, composed chiefly of pale-green actinolite and albite or saussurite. The term "saussurite" as used here indicates an aggregate of fine-grained epidote minerals and small amounts of other minerals.

The actinolite has replaced an original pyroxene, probably augite; the albite (pseudomorphic, showing relict twinning) or the saussurite has replaced an original calcic plagioclase, probably labradorite. These main constituents are present in nearly equal amounts and constitute more than 90 percent of the rocks. Minor constituents are titaniferous magnetite, augite, chlorite, and drab-green or brown biotite. The titaniferous magnetite has altered in several ways. It may be represented by skeletal crystals of ilmenite and magnetite, by leucoxene, by sphene and magnetite, or by blebs and rods of magnetite mantled by leucoxene or sphene. The magnetite-sphene couplet and brown biotite are present in the dikes of the second group in the north part of the area.

Sheared dikes of both groups are now schists. Those of the greenschist facies occur chiefly in the Michigamme Slate, where they have been subjected to the stresses of close folding. Extremely deformed dikes are dark-green chlorite schist, which usually contains relict albite, magnetite, leucoxene, apatite, and biotite. A dike of this type cuts the slate and quartzite of the Michigamme in the group of outcrops south of Hanbury Lake (pl. 2). Sheared dikes of the oligoclase-amphibolite facies are hornblende-oligoclase schists that have the same general appearance in outcrops as the dikes of the greenschist group, but are easily differentiated in thin sections. The main differences are the presence of the oligoclase (An_{10-30}) instead of albite, common blue-green hornblende instead of pale-green actinolite, and brown biotite instead of chlorite. Epidote is present in some specimens but absent in others.

METAGABBRO SILLS

Large sill-like bodies of metagabbro intrude the Quinnesec strata in the south part of the district. They were mapped and described in detail by Prinz (1959, p. 128–173), and the following material is mainly a summation of his findings.

The parts of the sills exposed in the mapped area range in length from about 1,200 feet to 6.5 miles and in width from about 200 to 3,400 feet. Most of the larger ones have been designated by informal names, but others remain unnamed. The dimensions of those for which Prinz suggested informal names are:

Sill	*Length (miles)*	*Width (feet)*
Sturgeon Falls	1.25	2,600
Niagara	6.5	3,200
Horserace	2.0	2,000
Western	2.5	3,400

An unnamed sill north of the Western sill (pl. 1) extends to the west a distance of 3 miles beyond the edge of the mapped area. Among other unnamed dikes or sills, the largest is south of Niagara in sec. 15, T. 38 N., R. 20 E., and is about 4,000 feet long and 800 feet wide. Two others lie north of the Niagara sill. The northernmost, in the Michigamme Slate, is represented by a single small outcrop of magnetic metagabbro in the SW¼ sec. 12, T. 39 N., R. 30 W., but an elongate positive magnetic anomaly which centers on the outcrop suggests that the unexposed part is at least 2,000 feet long. The other, the southernmost, in the NE¼ sec. 14, T. 39 N., R. 30 W., is about 1,200 feet long and 200 feet wide.

The metagabbro is very well exposed and can be found almost anywhere within the designated limits of the sills (pls. 1, 2). It is resistant to erosion and forms relatively rugged topography, hills as high as 100 feet being common. At the Horserace Rapids the Menominee River has excavated a steep-walled gorge 100 feet deep in the Horserace sill and, downstream from Niagara, cliffs of metagabbro 200 feet high face the river.

The rock types in the sills can be divided into seven groups: normal and magnetite-rich metagabbro, pegmatitic metagabbro, meta-anorthosite, granophyre, serpentinite, pyroxenite, and schist and amphibolite. Not all these groups are represented in every sill, and the normal and magnetite-rich metagabbros make up more than 90 percent of all sills. The normal and magnetite-rich metagabbros and the serpentinite have been separated in mapping, and some of the minor types are shown on the maps by letter symbols, as are inclusions in the sills of schistose Quinnesec strata.

The Sturgeon Falls sill is composed chiefly of normal metagabbro, but serpentinite and pyroxenite occur as a selvage on the north side, and small patches and lenses of meta-anorthosite are also present. According to Prinz (1959, p. 174), the metagabbro cuts the serpentinite.

The Niagara still is composed chiefly of normal and magnetite-rich metagabbro, distributed as indicated on plates 1 and 2. Some meta-anorthosite and porphyritic metagabbro are present in the normal metagabbro, and pegmatitic metagabbro is present as small segregations

FIGURE 20.—Layered porphyritic metagabbro from the Niagara sill. After Prinz (1959, pl. 16-B, p. 133).

and dikes in both the normal and magnetite-rich types.

The Horserace sill is chiefly normal metagabbro and granophyre, the latter rock confined to the north edge, whereas the Western sill is chiefly normal metagabbro but contains minor amounts of meta-anorthosite.

Along some internal shear zones and along the margins of each of these four major sills, the metagabbro has been deformed and variously reconstituted, and it is now chloritic or hornblendic schist.

The sills generally are massive; structures indicating crystal transport or sedimentation are rare, but two areas of layered rock are known. The porphyritic metagabbro in the Niagara sill at Niagara shows layering that is parallel to the edge of the sill and dips steeply northeast (fig. 20). Prinz ascribes this layering to the constriction of the sill at this point, which he says may have caused currents or the interference of currents in the magma to produce flow banding. R. W. Bayley interprets the bands to be gravity segregations. The other layered rocks are in the Western sill, in the SE¼ sec. 10, the layers of which strike northeast, nearly parallel to the margin of the sill, and dip 45° NW. These layered rocks are similar to rhythmical gravity-segregated ones observed in the well-differentiated Kiernan sill of eastern Iron County (Bayley, 1959a, p. 414).

Primary parallel structures, which indicate flowage of partly crystallized magma, occur at a few places, particularly around large inclusions of country rock.

NORMAL AND MAGNETITE-RICH METAGABBRO

Both the normal and the magnetite-rich metagabbro are medium to coarse grained and commonly are characterized by ophitic or diabasic textures that are preserved in varying degrees by numerous alteration minerals and, in some specimens, by relict plagioclase and, more rarely, by relict augite. The relict minerals of some specimens indicate that the unaltered normal gabbro was composed of bytownite-anorthite or labradorite, augite, minor amounts of titaniferous magnetite, and possibly some olivine. The alteration minerals formed include andesine, oligoclase, albite, hornblende, actionalite, chlorite, epidote, clinozoisite, some biotite, and quartz. The only relict mineral recognized in the magnetite-rich metagabbro is labradorite. These rocks may contain as much as 10 percent of titaniferous magnetite.

Six distinct mineral assemblages have been established for normal metagabbro, and two assemblages have been recognized in the magnetite-rich metagabbro. These assemblages are shown in the following table. The relict minerals are in italic, and minor minerals, not always present, are shown in parentheses:

Group	Mineral assemblage	Chief occurrence
Normal metagabbro		
1	Actinolite-*augite*-saussurite (chlorite-quartz).	Sturgeon Falls and Niagara sills.
2	Actinolite-saussurite-albite-(chlorite-quartz).	Do.
3	Chlorite-albite-calcite-sassurite (quartz).	Do.
4	Actinolite-saussurite-albite-(*labradorite*-quartz-chlorite).	Do.
5	Hornblende-oligoclase-*labradorite*-epidote (quartz-biotite).	Horserace sill.
6	Hornblende-*labradorite* or *bytownite* (-andesine-epidote-quartz).	Western sill.
Magnetite-rich metagabbro		
A	Hornblende-chlorite-epidote-albite and *labradorite*-magnetite.	Niagara sill.
B	Hornblende-epidote-albite and *labradorite*-magnetite-chlorite.	Do.

The minerals are listed in their approximate order of abundance. Rocks of assemblages 1–4 are confined mainly to the Sturgeon Falls and Niagara sills; they thus occur within the area which Prinz has designated greenschist facies on the basis of the metamorphic minerals in the Quinnesec Formation and the metagabbros. Assemblage 5 rocks in the Horserace sill lie within the oligoclase-amphibolite facies area, and assemblage 6 rocks are mainly in the Western sill, in the andesine-amphibolite facies area.

The modes of selected specimens of normal and magnetite-rich metagabbro are given in table 30. Specimens were selected to show the mineral composition of assemblages, 2, 4, 5, 6, A, and B. Assemblage 1 is similar to assemblage 2 except that it contains relict augite. Assemblage 3 rocks are very fine grained and their modes therefore uncertain. They contain approximately 40 percent chlorite, 35–45 percent combined albite and relict plagioclase, 5–25 percent calcite, and as much as 10 percent minerals of the epidote group.

The modes show fairly constant total quantities of mafic minerals whether actinolite or hornblende or chlorite, and wide variations in quantities of plagioclase and epidote minerals, which indicate chiefly the extent of alteration of the original plagioclase, the greater amounts of epidote occurring with the lesser amounts of plagioclase. The combined quantities of epidote and plagioclase, plus the amphibole and chlorite, range from 90 to nearly 100 percent of the rock; the computed average ratio of mafic minerals to plagioclase for the unaltered rocks is 52:42, which is close to the common ratio for diabasic gabbros.

Both plagioclase and amphibole are significant in the history of these rocks; therefore, a brief petrographic description of these minerals is summarized from the work of Prinz.

PLAGIOCLASE

Plagioclase or its alteration products constitute about one-half of the total bulk of the rocks. The grains are lath-shaped or stubby, nearly euhedral, and 1.0–2.0 mm long.

In assemblages 1 and 2, aggregates of saussurite generally fill the slots formerly occupied by the plagioclase, but some secondary clear albite, quartz, chlorite, calcite, and sericite occur in the plagioclase sites in some specimens. In rocks of assemblage 3 the sites of the plagioclase laths are commonly rimmed by clear albite and filled by chlorite or saussurite, or they may contain relict grains of labradorite. In rocks of assemblage 4, albite is commonly present as outer mantles on grains of relict labradorite or andesine that show considerable saussuritization, but in some specimens the original plagioclase has been completely saussuritized, and the saussurite aggregates contain euhedral grains of albite. In the porphyritic metagabbro of the Niagara sill, assemblage 4, saussurite has completely replaced the plagioclase.

In metagabbro of assemblage 5, the plagioclase in the sites of the original feldspar grains generally is either oligoclase or andesine-labradorite, but one speci-

TABLE 30.—*Modes of typical metagabbro and magnetic-rich metagabbro*

[From Prinz, 1959, p. 137, 163]

Specimen	1	2	3	4	5	6	7	8	9
Assemblage	2	4	4	A	B	5	5	6	6
Actinolite	48.5	51.4	36.4	------	------	------	------	------	------
Hornblende	------	------	------	37.5	48.7	66.9	46.2	43.6	58.3
Chlorite	------	2.4	6.0	22.3	3.6	------	------	------	------
Plagioclase [1]	6.5	18.6	25.6	15.1	17.8	23.0	27.8	55.8	38.4
Epidote group	43.8	20.0	28.8	15.6	20.8	4.6	23.4	.4	.5
Biotite	------	------	------	------	------	2.9	------	------	.2
Sericite	.1	------	------	------	------	------	1.8	.1	1.8
Quartz	.7	1.7	2.3	.4	------	------	------	------	.5
Magnetite	------	1.3	.4	6.1	9.1	.1	------	------	.3
Leucoxene	.3	4.6	------	2.5	------	2.4	------	------	------
Sphene	------	------	------	------	------	.1	------	------	------
Calcite	.1	0	.5	------	------	------	.5	.1	------
Pyrite	------	------	------	.3	------	------	.3	------	------

[1] 1 and 2, albite; 3, albite and *andesine;* 4 and 5, albite and *labradorite;* 6 and 7, oligoclase and *labradorite;* 8 and 9, andesine and *labradorite* or *bytownite* (italics indicate relict minerals).

men contains both oligoclase and andesine; granoblastic mosaics of oligoclase border some of the larger crystals. Saussurite is rare, and instead, epidote and sericite are the common alteration products.

Most of the rocks composed of minerals of assemblage 6 contain two varieties of plagioclase: clear andesine-sodic labradorite and relict calcic labradorite-bytownite-anorthite. Both varieties occur in the same grains, the more sodic one as mantles on or zones in a more calcic one; the grains therefore appear mottled. Sericite is present but not abundant, and epidote is almost entirely lacking.

AMPHIBOLE

Amphibole is the most common mineral in the metagabbro and is either actinolite or hornblende. It is greenish black to black in the fresh rock and, in diabasic types, it forms anhedral plates, about 4 mm long and nearly as wide, that are interstitial to plagioclase. In most rocks the amphibole is clearly pseudomorphic after augite and preserves the original augite forms, but in some rocks of assemblages 1–4 the original augite boundaries have been destroyed.

Prinz' microscopic study of the metagabbro indicates that the common amphibole of the rocks that contain mineral assemblages 1–4 is pale-green actinolite. The actinolite may be dense and more or less confined to the sites of previous augite, or fibrous and matted and interlaced with adjacent alteration minerals. In some rocks it is altered to chlorite, which occurs as irregular patches within actinolite grains or intergrown with fibrous actinolite.

Hornblende occurs in the magnetite-rich metagabbro, assemblages A and B, and also in rocks in assemblages 5 and 6. The amphibole of the magnetite-rich metagabbro is blue green with a brownish tinge and appears to be hornblende, possibly aluminous. If so, these rocks are the only albite-hornblende metagabbros recognized in the district. Deep-blue-green to brownish hornblende is characteristic of the rocks containing mineral assemblages 5 and 6. It is generally confined to the sites of previous augite, but in some specimens it fills cracks in the plagioclase and also occurs as small inclusions in the plagioclase. Some of the hornblende is poikiloblastic and contains grains of plagioclase and quartz. Only rarely is chlorite associated with hornblende in

TABLE 31.—*Chemical analyses and norms of metagabbro*

Specimen	1	2	3	[1] 4	5	6
Assemblage	1	2	4	6	A–B	
Chemical analyses						
SiO_2	51.46	47.96	48.35	47.6	43.80	48.36
TiO_2	------	------	------	.2	------	1.32
Al_2O_3	14.35	16.85	15.40	22.2	16.08	16.84
Fe_2O_3	3.90	4.33	4.04	1.1	9.47	2.55
FeO	5.28	4.17	4.63	4.4	10.50	7.92
MnO	------	------	------	.10	------	.18
MgO	9.54	9.15	11.61	7.1	6.54	8.06
CaO	9.08	13.25	10.38	12.5	7.81	11.07
Na_2O	2.92	1.25	1.87	1.8	1.96	2.26
K_2O	.24	.30	.35	.50	.34	.56
H_2O	3.30	2.89	3.60	2.0	3.99	.64
P_2O_5	------	------	------	.01	------	.24
CO_2	.20	.08	.08	.18	.08	------
Totals	100.27	100.23	100.31	99.7	100.57	100.00
Norms [Prinz, 1959, p. 150]						
qz	1.6	1.4	------	------	0.5	------
or	1.1	1.7	2.2	2.8	1.7	3.3
ab	24.6	10.5	15.7	15.2	16.8	18.9
an	25.3	39.7	32.5	51.1	34.2	34.2
$CaSiO_3$	7.8	10.7	7.8	3.9	1.6	8.0
$MgSiO_3$	23.8	22.9	24.1	11.0	16.4	14.0
$FeSiO_3$	6.6	4.1	4.2	4.2	11.5	7.4
Mg_2SiO_4	------	------	3.4	4.9	------	4.3
Fe_2SiO_4	------	------	.7	2.0	------	2.5
mt	5.6	6.3	5.8	1.6	13.7	3.7
il	------	------	------	.5	------	2.4
ap	------	------	------	------	------	.6
cc	.5	.2	.2	.5	.2	------

[1] Sample also contains 0.0050 percent Cu, 0.003 percent Co, 0.017 percent Ni, and 0.058 percent Cr.

Specimen:
1. "Gabbro" from Sturgeon Falls. Williams (1890, p. 76, anal. I).
2. "Gabbro-diorite" from Little Quinnesec Falls. Williams (1890, p. 89, anal. I).
3. "Gabbro" from Big Quinnesec Falls. Williams (1890, p. 104, anal. I).
4. Metagabbro from NW¼NW¼ sec. 15, T. 38 N., R. 19 E., Wisconsin (Prinz, 1959, p. 150). Analysts: Paul L. D. Elmore, Katrine E. White, Samuel D. Botts, and Harry J. Rose, U.S. Geological Survey.
5. "Dark massive greenstone" from Little Quinnesec Falls. Williams (1890, p. 91, anal. I).
6. Average gabbro. Nockolds (1954, table 7, anal. I).

these rocks, but in some specimens it occurs in late veinlets.

Hornblende and actinolite both are fairly common in the rocks of the Sturgeon Falls sill, and either one may form a mantle on relict pale-green augite.

CHEMICAL COMPOSITION OF METAGABBRO

The chemical makeup of the metagabbro is shown by table 31, which includes four analyses from Williams (1890), one new one from Prinz (1959), and, for comparison, an analysis of average gabbro from Nockolds (1954, table 7, analysis I).

The analyses of table 31 show that the Menominee metagabbros are approximately the average type, though they contain slightly more Mg0 than the average, are somewhat hydrated, and may be slightly oxidized. The higher average Mg0 content is probably original, whereas the hydration and oxidation are undoubtedly due to alteration.

The norms indicate an approximate 50:50 ratio of mafic minerals to plagioclase, and also suggest a composition range of An_{50-79} for the original plagioclase, which is close to the observed range in the rocks. The observed variations in the composition of the relict plagioclase probably are inherent and closely related to variations in the original $Ca0:Na_20$ ratios from one sill to another and in different parts of the same sill.

PEGMATITIC METAGABBRO

Very coarse grained ophitic and pegmatitic metagabbro forms zones, schlieren, and dikes as much as 6 inches thick in the normal and magnetite-rich metagabbro. The pegmatitic metagabbro probably was formed by the intrusion of a late-cooling liquid fraction of magma into fractures in crystalline gabbro.

The mineral composition of this rock is about the same as that of the host rocks, and those masses that are in the magnetite-rich metagabbro are also rich in magnetite. Bladed crystals of actinolite or hornblende as much as 2½ inches long, and somewhat smaller crystals of plagioclase, are the chief constituents. Accessory and minor alteration minerals are the same as those in the normal metagabbro. Myrmekitic intergrowths of quartz and plagioclase occur in one specimen. The mode of a specimen of pegmatitic metagabbro is shown in table 32.

TABLE 32.—*Mode of pegmatitic metagabbro from sec. 15, T. 39 N., R. 30 W., Michigan*

[After Prinz, 1959, p. 168]

Mineral	*Percent*	*Mineral*	*Percent*
Amphibole	47. 8	Magnetite	1. 3
Chlorite	4. 3	Sphene-leucoxene	1. 8
Plagioclase (An_{13})	29. 8	Apatite	. 2
Epidote group	13. 7	Calcite	. 4
Quartz	. 7		

META-ANORTHOSITE

Metamorphosed anorthosite is widespread in the metagabbro sills but is nowhere abundant. It forms small zones or lenses as much as 20 feet thick, of undetermined length, the ends being commonly covered. It is most abundant in the Niagara sill but occurs also in the Sturgeon Falls and Western sills.

The meta-anorthosite is a medium-grained white to greenish-gray rock that consists mainly of plagioclase or its alteration products, chiefly minerals of the epidote group, and contains minor quantities of the mafic minerals that occur in the normal metagabbro. In thin sections the rock displays an interlocking mesh of lath-shaped plagioclase crystals whose average length is about 3 mm. Where the rock occurs in the area of lowest grade metamorphism, the plagioclase is albite or sodic oligoclase; saussuritic alteration is common, and any mafic minerals present are nearly colorless chlorite and actinolite. Where it occurs in areas of intermediate metamorphic intensity as in the western sill, the meta-anorthosite is composed chiefly of relict calcic plagioclase (about An_{70}) and clear secondary andesine (about An_{34}), both of which form parts of the same mottled crystals as in the normal metagabbro. The mode of a typical specimen of this rock is shown in table 33.

The origin of the meta-anorthosite is uncertain, but it may have formed by differential crystal setting—that is, the heavier and possibly larger crystals of pyroxene gravitated downward in the magma, and left a plagioclase-rich residuum, as in rhythmical layers but on a larger scale.

TABLE 33.—*Mode of meta-anorthosite from the SW¼SE¼ sec. 10, T. 38 N., R. 19 E., Wisconsin*

[After Prinz, 1959, p. 168]

Mineral	*Percent*	*Mineral*	*Percent*
Hornblende	12. 0	Biotite	0. 6
Chlorite	1. 6	Sericite	Tr.
Plagioclase (An_{34} and An_{70})	85. 8	Calcite	Tr.

GRANOPHYRE

Quartz-rich granophyric-textured rocks occur along the north contact of the Horserace sill in the SE¼ sec. 12, T. 39 N., R. 31 W. The zone of granophyric rock is about 100 feet thick. It seems to be gradational with normal metagabbro on the south and gradational with Quinnesec amphibolite and metarhyolite on the north. There is no chilled zone of metagabbro between the sill and the country rock.

The granophyre is fine to medium grained, massive, and greenish gray. Glassy quartz and striated plagioclase are visible in all specimens, and dark-greenish-black hornblende in most. Thin sections show that quartz and feldspar are the predominant minerals, and these are intergrown in characteristic granophyric habit. These intergrowths form a mesh within which are abundant equidimensional or lathlike crystals of oligoclase or microcline, crystals and crystal clots of blue-green hornblende, and scattered plates of brown biotite.

The origin of the granophyre is uncertain. Prinz (1959, p. 181) regards it as a segregation from the gabbroic magma, and it may be. Certainly it is very much like the granophyric segregations from other sills. But located as it is between normal metagabbro and the Quinnesec strata, some of the latter being close to the granophyre in composition, the possibility remains that the granophyre may be of metasomatic origin as in the Hangnest sill of South Africa (Walker and Poldervaart, 1941, p. 445–446).

SERPENTINITE AND PYROXENITE

Serpentinite and pyroxenite are present in several places along the north side of the Sturgeon Falls sill, and, within the mapped area, are exposed in sec. 21, T. 38 N., R. 21 E. Generally the pyroxenite lies between the metagabbro and the serpentinite or forms narrow bands in the metagabbro. Near the pyroxenite, the metagabbro is unusually fine grained; a chilled contact is thus suggested. The contact between the pyroxenite and serpentinite is hidden. Serpentinite is also exposed about a mile southeast of the outcrops in sec. 21, beyond the edge of the mapped area; there, according to Prinz (1959, p. 174), the metagabbro cuts the serpentinite.

The pyroxenite is a medium-grained greenish-black rock in which greenish-black pyroxene crystals are visible in hand specimens. The anhedral pyroxene crystals are roughly equidimensional, 3–10 mm wide, and make up nearly all of the rock. The only other minerals noted are magnetite and serpentine, which are quantitatively unimportant, and which occur along pyroxene grain boundaries. In thin sections the pyroxene is nearly colorless and appears to be a diopsidic type having a large 2V, an extinction angle $Z \wedge c = 38°$, and $Ny = 1.678$.

The serpentinite is fine grained, greenish black, and massive or very slightly foliated. It is cut by veinlets of white carbonate (dolomite or magnesite) that is weakly effervescent in dilute HCl and veinlets of cross-fiber asbestos as much as one-fourth inch wide. The serpentine is chiefly antigorite, which comprises 75–80 percent of the rock. Other constituents are 15–20 percent of carbonate, about 5 percent magnetite and chromite, and trace amounts of chrysotile and colorless chlorite. The antigorite occurs in mats of small fibers that average about 0.2 mm long. In one specimen the antigorite is separated into roundish cells surrounded by magnetite which suggests that the original mineral was olivine. In another specimen the antigorite fibers show the same orientation over a considerable area, as in bastite; the serpentinized mineral may have been a pyroxene, but the alteration area is irregular in shape and not distinctive of any mineral.

TABLE 34.—*Chemical analysis and norm of serpentinite from the SW¼NW¼ sec. 26, T. 39 N., R. 29 W., Michigan*

[After Prinz, 1959, p. 178. Analysts: Paul L. D. Elmore, Katrine E. White, Samuel D. Botts, and Harry J. Rose, U.S. Geological Survey]

Constituent	Chemical analysis	Chemical analysis, recalculated [1]	Norm [2]	
SiO_2	40.8	47.7	$CaSiO_3$	3.5
TiO_2	.02	.02	$MgSiO_3$	40.2
Al_2O_3	(3)	----	$FeSiO_3$	1.6
Fe_2O_3	3.1	3.6	Mg_2SiO_4	47.1
FeO	3.3	3.8	Fe_2SiO_4	1.9
MnO	.08	.09	mt	5.3
MgO	36.9	43.0		
CaO	1.5	1.7		
Na_2O	.04	.05		
K_2O	.04	.05		
H_2O	11.5	0		
P_2O_5	.01	.01		
CO_2	2.3	0		
Cu	.0020	0		
Co	.020	0		
Ni	.23	0		
Cr	.42	0		
Total	100.26	100.0		

[1] Recalculated to 100 percent less H_2O, CO_2, Co, Cu, Ni, and Cr.
[2] Norm from recalculated analysis.
[3] Less than 0.5 percent.

A chemical analysis of the serpentinite from sec. 26, T. 39 N., R. 29 W., is given in table 34, and, for comparison, a recalculated analysis in which water, carbon dioxide, and some minor constituents are omitted, and the norm from this recalculated analysis.

The analysis, the norm, and the petrographic characteristics of the serpentine suggest strongly that the original rock was composed chiefly of magnesium olivine and magnesium pyroxene. Probably the latter was diopside, as in the associated pyroxenite. The analyzed rock is a little unusual in that it contains so little iron oxide and alumina compared to other ultramafic rocks. The analysis compares best with Nockolds' (1954, p. 1023) average analysis of dunite, but the average dunite contains nearly twice as much iron oxide as the serpentinite. The diopside in the pyroxenite suggests that the iron-poor characteristic of the rock is probably inherent and not due to iron loss during serpentinization.

The origin of the serpentinite and pyroxenite is in doubt. A similar serpentinite body was found at the base of the Kiernan metagabbro sill in eastern Iron County, Mich., by Bayley (1959a, p. 412), who interpreted it as being formed by gravity settling of early formed crystals from the gabbro magma. However, metagabbro of the Kiernan sill is not known to intrude the serpentinite as it does in the Sturgeon Falls sill. The intrusive relations in the latter indicate that the ultramafic rocks were emplaced before the metagabbro. The chilled borders on the metagabbro dikes indicate further that a considerable time may have elapsed between intrusive episodes. To explain these observations, Prinz speculated that the ultramafic rocks may represent a fraction, separated from the gabbroic magma at depth, which was pressed into Quinnesec strata and followed closely by the normal gabbro magma. We suggest that the metagabbro and ultramafic rocks may not be genetically related at all, but that the ultramafic rocks are of early Precambrian age as are similar rocks in the Marquette district (Van Hise and others, 1911, p. 255).

METAMORPHISM OF SILL ROCKS

The formation of alteration minerals in these rocks, and the partial or complete destruction of original textures, may have resulted partly from late-stage emanations at the time of emplacement of the sills (deuteric alteration), partly from hydrothermal solutions that permeated the rock after its emplacement, and partly from dynamothermal metamorphism. Which one of these causes was the most important in each sill is extremely difficult to state, because many of the secondary minerals present may be formed in several ways. The fact that the mineral compositions of the different groups do, nevertheless, fit in a general way into the rather well-defined metamorphic pattern of the area suggests that they are generally related to the post-Animikie orogeny and regional metamorphism. The sills that apparently had the lowest degree of metamorphism (greenschist facies) are characterized by assemblages which contain newly formed minerals that are various combinations of actinolite, albite, chlorite, saussurite, and calcite, and may contain relict labradorite or augite; those affected by an intermediate degree of metamorphism are characterized by an assemblage which contains newly formed oligoclase and hornblende and some relict bytownite-anorthite. It is Lamey's view that the combinations of hornblende with plagioclase are the ones most indicative of metamorphism, and that the other combinations could have resulted from alterations of other types; that is, gabbros altered by deuteric processes to chloritic and actinolitic rocks containing albite would pass into hornblendic rocks containing more calcic plagioclase at the appropriate degree of metamorphism.

Prinz' view, shared by R. W. Bayley, is that the metmorphism was not progressive; that is, it did not progress from low grade to high grade as suggested by Lamey. The fact that the highest grade rocks show the best preserved igneous textures and the greatest amounts of high-temperature relict minerals seems to negate the possibility that these rocks could have been deuteric greenstones before regional metamorphism. Rather, it seems, there has been a regression of original igneous minerals—the reconstitution has been most extensive in the low-grade rocks. The fact that relict augite is found only in the lowest grade rocks suggests that these rocks were metamorphosed at temperatures and other conditions low enough to fall within the stability range of that mineral, and that extensive deuteric alteration probably did not precede the metamorphism.

ATTITUDE OF THE METAGABBRO SILLS

Ideally, true sills of the dimensions of those in the Quinnesec could be expected to show petrographically the effects of differentiation by gravity settling and crystal fractionation. A well-differentiated metagabbro sill, the west Kiernan sill of Iron County (Bayley, 1959a), presumably of the same igneous cycle, illustrates the principal relations. From the base to the top (about 5,000 ft) five zones are recognized: (1) a basal ultramafic zone 600–1,200 feet thick; (2) a zone of diabasic and ophitic metagabbro 2,000–5,000 feet thick, which commonly shows rhythmic layering; (3) a zone of iron-rich transition rocks 200–1,000 feet thick, present locally; (4) a zone of granophyre 200–500 feet thick, present locally near the top of the sill; and (5) a zone of diabasic metagabbro 100–500 feet thick, probably representing the upper chilled margin of the sill, present locally.

The sills in the Quinnesec strata show some of the features of the Kiernan sill, but nowhere is there a complete sequence of zones. On the basis of the iron-rich zones, which favor the north side of the Niagara sill, and the granophyre zone on the north side of the Horserace sill, Prinz concluded that the tops of the sills probably face north, and that they were probably rotated from a horizontal position by the post-Animikie folding. If the iron-rich and granophyric zones of the above-mentioned sills are accepted as being analogous to similar zones in the Kiernan sill, they do impart a sense of tops north of these sills, but they do not necessarily imply that the original attitudes of all the sills were strictly horizontal. The Western sill, for example, contains rhythmical layers, presumably deposited in a horizontal position, which now dip 45° NW. The schist inclusions in that sill show nearly the same attitude as the schist bordering it on the north, 65°–75° S. If the rhythmical layers were rotated back to the horizontal position, the schist inclusions and the bordering schist would become nearly vertical; the so-called Western sill is therefore apparently not a sill but a dike or pluton. The discordant character of this body is indicated along its northeast margin where it truncates southwest-striking Quinnesec strata. If it differentiated while cooling, and the layering suggests that it did, then the acidic upper part has been eroded away.

The only clue to the amount of rotation that the Niagara and Horserace sills have undergone is the steeply dipping porphyritic metagabbro layers in the Niagara sill, which, if formed by crystal sedimentation rather than by flow, as contended by Prinz, indicate 80°–90° of rotation. Depending upon local differences in their original attitudes and the attitudes of the host schist and also upon their locations with respect to other tectonic elements, each of the sills or plutons probably was rotated or folded different amounts during the post-Animikie orogeny.

AGE OF THE METAGABBRO BODIES

Two swarms of diabasic gabbro dikes occur in the Menominee district and in the surrounding areas, one generally older than the post-Animikie folding and metamorphism, and one younger. The metagabbro dikes and sills such as those previously described comprise the older swarm, whereas only undeformed and unaltered diabasic dikes, thought to be of late Precambrian (Keweenawan) age, are included in the younger swarm.

The older swarm includes all metamorphosed dikes and sills. Most of these cut Animikie strata, including the youngest, the Paint River Group, some lie isolated in the pre-Animikie rocks, a few cut both the pre-Animikie and overlying Animikie rocks, and some of those that cut both the pre-Animikie and overlying Animikie rocks are in fault zones and are somewhat sheared. No dikes of this swarm are known to be truncated at the erosion surface between pre-Animikie and Animikie rocks. The intrusion of the swarm as a whole, therefore, is broadly limited in time between the beginning of Animikie sedimentation and the post-Animikie deformation. But because the dikes commonly occur in the younger Animikie formations and because those in fault zones suggest that they were intruded at the time of the post-Animikie folding, we believe that the swarm generally is of very late or post-Animikie age.

MARINETTE QUARTZ DIORITE

GENERAL RELATIONS

The Marinette Quartz Diorite is an intrusive igneous body that underlies approximately 3 square miles near the south border of the district in Marinette County, Wis. (pls. 1, 2), from which it takes its name (Prinz, 1959, 1965). It is exposed in isolated and scattered rock knobs and low hills. The unit extends to the south beyond the limits of the mapped area, but its overall size and distribution are imperfectly known. Lyons (Emmons and others, 1953, pl. 16) shows about 7½ square miles of quartz diorite to the south, but reconnaissance mapping by R. W. Bayley indicates that the extent of this particular unit is probably much less. Within the mapped area the quartz diorite forms two separate bodies, one thin elongate body lying between Quinnesec strata and the Hoskin Lake Granite, in secs. 15, 16, and 22, T. 38 N., R. 20 E., and a second larger elongate body lying south of the granite. Whether these were originally two bodies or one body divided by the intrusion of the granite could not be determined. The quartz diorite shows sharp intrusive contacts against the Quinnesec rocks and is somewhat finer grained and more feldspathic near the contacts. The Hoskin Lake Granite has intruded the quartz diorite and has produced a contact zone of mixed rocks as much as 1,000 feet thick. The contact zone contains many rock types gradational between the two end members, quartz diorite and granite, and broad areas composed predominantly of one or the other of the members. Quartz diorite inclusions in the granite are cut by granite dikes.

DESCRIPTION

(From Prinz, 1959, 1965)

The quartz diorite is generally massive, but locally it is foliated. It is dark gray to black and medium grained, most commonly hypidiomorphic granular but porphyroblastic at some places near the Hoskin Lake Granite. The average rock contains 40–50 percent plagioclase, 0–20 percent potassic feldspar, 10–30 percent quartz, and 20–30 percent biotite, chlorite, or hornblende. Accessory minerals are sphene, magnetite, zircon, apatite, pyrite, calcite, and blue tourmaline.

The average length of plagioclase laths is 2.0 mm, but exceptions are as much as 5.0 mm. Most of the plagioclase is oligoclase, which shows normal zoning having cores as calcic as An_{40}. Twins according to the albite and combined albite-carlsbad laws are most common. Most plagioclase crystals show alteration to biotite, sericite, or epidote-clinozoisite. Where epidote-clinozoisite alteration is prevalent, it is assumed that the original plagioclase was considerably more calcic than at present.

Most of the potassic feldspar is microcline, which is generally clear, unaltered, and finer grained than the oligoclase. It is associated with quartz and myrmekite, interstitial to the oligoclase crystals, and some of it partly replaces oligoclase.

The quartz is clear and ranges from 0.2 to 1.0 mm. Much of it occurs as sievelike blebs in clots of biotite plates, and many of the clots of biotite contain remnants of hornblende crystals showing identical sieve structure, but most hornblende remnants are dense.

The sievelike quartz in the clots of biotite probably represents excess silica from the alteration of hornblende to biotite.

Light-brown biotite is an abundant constituent. In some specimens it is associated with and rarely intergrown with bright-green chlorite ($N_{y=z} = 1.646$), probably prochlorite, which appears to be an alteration product of biotite.

Isolated crystals of hornblende and remnants of hornblende in clots of biotite are the common blue-green variety, $Z \wedge c = 15°–19°$. Most of the original hornblende has been replaced by biotite and byproduct quartz, epidote, sphene, and calcite.

Sphene is the most common accessory mineral. It forms euhedra, rounded blebs in hornblende and biotite, and rims around magnetite. Apatite and calcite are ubiquitous, but in minor amounts. The other accessory minerals are rare.

A specimen of the finer grained border phase of the quartz diorite from near the contact with the Quinnesec Formation in sec. 15, T. 38 N., R. 20 E., consists almost wholly of a mosaic of equant, sericitized oligoclase grains (0.2–0.8 mm) and some chlorite, epidote, magnetite, and sphene.

Fine-grained mafic inclusions probably representing pieces of mafic volcanic rocks from the host Quinnesec Formation are common in parts of the quartz-diorite. The inclusions show the same mineralogy as the quartz diorite, but contain more mafic minerals and less quartz.

ORIGIN

The Marinette dioritic body is composed of quartz diorite, but the mineralogical relations, particularly the replacement of hornblende by biotite, the sericitization and epidotization of plagioclase, and introduction of microcline, which in part replaces earlier plagioclase, indicate that the original composition might have been that of a hornblende diorite containing calcic andesine, or even that of a hornblende or augite gabbro. Intrusion of the Hoskin Lake Granite into such a rock could bring about metasomatic alteration that would produce the mineralogy and characteristics of the present Marinette Quartz Diorite.

Lyons (Emmons and others, 1953, p. 108) has suggested that the quartz diorite may have had the composition of basalt and that it represents a stage in the granitization of basalt. The writers do not accept this theory of the origin because field relations indicate clearly that the quartz diorite has intruded the metabasalts of the Quinnesec and that it in turn has been intimately intruded by the Hoskin Lake Granite. We believe the Marinette Quartz Diorite, therefore, is an intrusive igneous body.

HOSKIN LAKE GRANITE

GENERAL RELATIONS

The name Hoskin Lake Granite has been applied by Prinz (1959, 1965) to a large body of coarse-grained porphyritic granite which underlies an area of approximately 6 square miles in the south part of the district in Wisconsin, and which we believe to be of post-Animikie age. It was named for Hoskin Lake in sec. 23, T. 38 N., R. 19 E. Rocks of this unit are well exposed and, except for a few areas, outcrops occur almost everywhere within the indicated mapped limits (pls. 1, 2).

The granite forms an arcuate mass ½–3 miles across which lies south of the Quinnesec Formation. This mass is narrow near its eastern extremity, where it is flanked to the west and to the east by quartz diorite, and thick near its western extremity, where it lies between the Quinnesec Formation and the quartz diorite.

Reconnaissance mapping by R. W. Bayley south of the mapped area shows that the overall shape of the porphyritic granite body is that of a dike or sill 1–2 miles wide. Only a part of a north-convex loop of the dike is in the area mapped by Prinz. The east limb extends south and east of the mapped area for about 7 miles, and the west limb extends southwest for 4 miles, west for 4 miles, and northwest for about 6½ miles, into Florence County, Wis., where it has been mapped by C. E. Dutton along its southern edge. The porphyritic granite is separated from nonporphyritic granitic rocks of the main Wisconsin batholith by a belt of Quinnesec greenstone 1–5 miles wide.

The granite shows intrusive relations toward both the Marinette Quartz Diorite and the Quinnesec Formation. The contact between the granite and the Quinnesec is remarkably sharp, in contrast with the broad, mixed-rock contact with the quartz diorite. Very narrow migmatitic contact zones occur at a few places, but at other places the granite and Quinnesec are in contact along shear zones. Satellitic granite dikes cut sharply across the grain of the Quinnesec at many places close to the granite contact. The dikes are best exposed in the NE¼ sec. 16, T. 38 N., R. 20 E., and near Spike Horn Creek, sec. 25, T. 38 N., R. 21 E., south of Niagara. Numerous inclusions of Quinnesec metabasalt are scattered through the main body of the granite. The inclusions are generally sharp walled. The largest one is in sec. 18, T. 38 N., R. 20 E.

Most of the granite exposures are cut by minute ramifying veinlets of black tourmaline, some of which also contain pyrite and arsenopyrite.

Small dikes of fresh diabase cut the granite in sec. 12, T. 38 N., R. 19 E., and dikes of aplite and pegmatite cut the granite in many places. Dacite porphyry dikes cut the granite in secs. 16 and 21, T. 38 N., R. 20 E.

MAIN GRANITE MASS

Hoskin Lake Granite is generally light colored, usually pale gray, but it is red in some places. Most of it is massive and quite homogeneous. The main mass of the exposed rock is porphyritic granite, but locally, close to the contact with the Quinnesec Formation, granite grades into finer grained, nearly nonporphyritic quartz monzonite, which is a chilled border phase. Commonly the granite shows a poorly formed primary linear arrangement of feldspar phenocrysts,

FIGURE 21.—Hoskin Lake Granite. Magnet is 1½ inches wide. Edge of pegmatite dike upper left.

but at some places it shows a secondary augen structure, and locally a smeared mylonitic structure. The augen structure is usually poorly formed or not present at all in the central part of the main mass.

The granite is medium to coarse grained, hypidiomorphic granular, and porphyritic. Large rectangular phenocrysts of microcline and oligoclase as much as 7 cm long dominate the texture (fig. 21). These phenocrysts are set in a groundmass of oligoclase, quartz, microcline, and biotite. The microcline, oligoclase, and quartz are present in about equal amounts; biotite and chlorite make up 5–15 percent of the rock, and the accessory minerals zircon, monazite, allanite, and sphene, less than 3 percent.

The microcline phenocrysts appear unaltered, and some contain laths of sericitized plagioclase; many show irregular or granulated borders, and some show myrmekitic border areas which penetrate adjacent oligoclase crystals. Much of the microcline is perthitic and contains stringers of oligoclase. The microcline of the groundmass occurs in small subhedral crystals associated with quartz and myrmekite, all of which appear to have formed later than the oligoclase.

The biotite is deep brown. In the massive rock it is mainly dispersed in the groundmass, but in the foliated rock the crystals are alined and form mantles on augen of quartz and feldspar. The biotite contains haloed inclusions of zircon, monazite, allanite, and sphene. Some of it has been replaced by chlorite.

The quartz-monzonite phase, near the contact of the granite with the Quinnesec Formation, consists of sericitized anhedral oligoclase crystals in a medium-grained groundmass of quartz, microcline, oligoclase, and biotite. Some of the biotite has been replaced by chlorite. The accessory minerals are the same as those in the porphyritic granite.

The mylonitic granite shows the usual features of such a rock. It is well foliated, composed of granulated and recrystallized minerals that are present in the granite, and shows eyes of quartz and crushed plagioclase.

GRANITE DIKES

Porphyritic and nonporphyritic granite dikes related to the Hoskin Lake Granite cut both the Marinette Quartz Diorite and the Quinnesec Formation, but as far as has been determined they cut no formation younger than the Quinnesec within the area mapped. They vary in size from thin stringers resembling lit-par-lit injections to thick well-defined dikes and irregular bosses as much as 10 feet across. Pegmatite and aplite dikes, which perhaps represent the late-stage fluids from the granite, also cut the granite itself, the quartz diorite, the Quinnesec strata, and possibly the late-Animikie metagabbro.

A few dikes which cut hornblende schist near the main granite contact show evidence of having been deformed slightly since they were emplaced (NW¼ sec. 16, T. 38 N., R. 20 E.), but most dikes do not.

ORIGIN AND SUBSEQUENT CHANGES

The small size of the area underlain by the Hoskin Lake Granite and the limited amount of study given to this phase of the general problem of the Menominee district permit only a few generalizations about its origin.

At the present level of erosion the granite appears to have been injected as magma. It shows undeniable intrusive contacts with adjoining older rocks. This type of origin appears to be suggested also by sharp-walled inclusions of Quinnesec Formation in the main granite, which seem to be stoped blocks rather than unreplaced material that has survived a period of granitization. No field evidence noted by Prinz (1959) or by the writers was interpreted as indication that the Hoskin Lake Granite was produced by granitization of Quinnesec basalt, as suggested by Lyons (Emmons and others, 1953, p. 108).

AGE OF THE MARINETTE QUARTZ DIORITE AND HOSKIN LAKE GRANITE

The times of intrusion of the Marinette Quartz Diorite and the Hoskin Lake Granite can be determined approximately from the following field observations and isotopic dates:

1. Each of the intrusives cuts the Quinnesec Formation and is therefore post-Quinnesec in age.
2. The quartz diorite cuts the late- or post-Animikie metagabbro, and pegmatite and aplite dikes probably related to the granite cut the metagabbro; both intrusives are therefore likely to be younger than the metagabbro.
3. Primary flow structures and deformational structures in the Hoskin Lake Granite suggest that it may have been intruded during the post-Animikie orogeny, that is, it is syntectonic.
4. Ages of zircon, as determined by lead-alpha and isotopic lead methods, from the Marinette Quartz Diorite and the Hoskin Lake Granite suggest that these intrusives are 1,850 and 1,650 million years old respectively (T. W. Stern, oral commun., 1959; Davis and others, 1960, p. 154). These dates bracket the age of the post-Animikie folding and metamorphism which Goldich and others (1957, p. 550) suggest occurred in Michigan and Minnesota about 1,700 million years ago. The Hoskin Lake Granite is apparently the same age as a known syntectonic post-Animikie pegmatite in Eastern Iron County whose age, by rubidium-strontium on microcline, is 1,650 million years (Davis and others, 1960, p. 153). But both of the intrusives appear to be very much younger than the pre-Animikie Carney Lake Gneiss, which may be as much as 2,700 million years old (Davis and others, 1960, p. 154).

As we interpret the field evidence and the above dates, the Marinette Quartz Diorite and the Hoskin Lake Granite must be post-Animikie intrusives. Nearly every major rock unit in the district and the surrounding area in Michigan has been dated by one or more of the isotope methods. In almost all samples, discordant ages have been found which suggest more complicated histories for the minerals tested than the known geologic history would indicate. The discordancies are due primarily to the loss of radiogenic daughter products, often differentially between different minerals of the same sample; less than the true age is therefore indicated by most minerals. The numbers given represent maximum indicated ages, which we hope are somewhere in the neighborhood of the true ages.

Dates for other Animikie rocks from beyond the district suggest that the post-Animikie folding and metamorphism may have begun as much as 1,900 million years ago in this part of Michigan.

UPPER PRECAMBRIAN INTRUSIVE ROCKS

KEWEENAWAN SERIES

DIABASE DIKES

Unmetamorphosed and undeformed dikes of diabasic gabbro of probable Keweenawan age are found at several widely separate localities. They are found in the Carney Lake Gneiss near the Sturgeon River Dam (secs. 4 and 9, T. 39 N., R. 28 W.), as already noted by Treves (p. 127), and in the Hoskin Lake Granite in sec. 12, T. 38 N., R. 19 E. These dikes are narrow (few feet), dense, and black, and commonly show chilled margins against the country rocks. Diabasic texture is most common. The minerals are labradorite, augite, and olivine. The dikes are clearly younger than the post-Animikie regional metamorphism but are Precambrian and are therefore classified as Keweenawan.

METAMORPHISM

The rocks of the Menominee district, except the Carney Lake Gneiss and part of the Quinnesec Formation, are only slightly metamorphosed, although they may be highly deformed. In general, the metamorphism is regional in type and fits the metamorphic pattern for the surrounding area, as described by James (1955), but very locally, around the Hoskin Lake Granite, it is contact metamorphism; to some extent some rocks have been affected by retrograde metamorphism.

Two epochs of metamorphism have been rather clearly established for the general region, one pre-Animikie, the other post-Animikie. The effects of the pre-Animikie metamorphism in the Menominee district are shown clearly only by the Carney Lake Gneiss, although the Quinnesec Formation, if the postulated age is correct, could have been affected.

The rocks of the Menominee district divide naturally into three groups or facies, which, on the basis of mineralogy, correspond to low, moderate, and high-moderate regional metamorphism, and locally to contact metamorphism. By far the most widespread of these groups is the low grade or greenschist facies (Eskola, 1915, 1920; Turner and Verhoogen, 1951, 1960, p. 533–534) which includes most of the rocks of the district. Boundary lines (isofacies) on the geologic maps show the approximate extent of the area underlain by the rocks of this facies. The metasedimentary rocks in the greenschist facies area are characterized by the mineral assemblages muscovite (sericite)-chlorite-quartz or muscovite-biotite-quartz, and thus correspond to both chlorite and biotite zones of regional metamorphism. The greenstone and the mafic dikes in that area are characterized by the mineral assemblage albite-chlorite-actinolite-quartz and commonly by some relict minerals. The two higher facies are, in order of increasing metamorphism, the oligoclase-epidote-amphibolite and the andesine-amphibolite facies. These facies were determined wholly on the basis of the mineralogy of the mafic rocks. Both are characterized by plagioclase and hornblende, instead of albite and actinolite as in the greenschist facies rocks, and are separated from each other by the more calcic plagioclase in higher grade facies. The two higher facies are local facies which seem to fit the Menominee rocks; they could be considered as subfacies of the almandine-amphibolite facies of Fyfe, Turner, and Verhoogen (1958, p. 228–235). The oligoclase-amphibolite lies both north and south of the greenschist in belts which correspond almost exactly with the garnet zones of James (1955, pl. 1). The andesine amphibolite is confined to the south part of the area, close to the Hoskin Lake Granite. Very close to the granite, or within it as xenoliths, are some metavolcanic rocks that are hornblende or hornblende-diopside hornfelses. These seem to be contact metamorphic rocks, which likely correspond to the hornblende-hornfels facies of Fyfe, Turner, and Verhoogen (1958, p. 205–210); however, these rocks are included with the andesine amphibolites on the geologic maps.

METAMORPHISM OF THE SEDIMENTARY ROCKS

The degree of metamorphism of all the sedimentary rocks, except some minor intercalations in the Carney Lake Gneiss and the Quinnesec Formation, is very low. The minerals formed from originally fine-grained detrital material are chiefly chlorite and sericite, and, much less abundantly, biotite. The chemical precipitates were generally recrystallized to some extent; dolomite and other carbonates to finely crystalline aggregates, chert to finely crystalline quartz, and iron oxide to euhedral octahedra of magnetite or, less commonly, plates of specular hematite. The finer grained and more argillaceous parts of most formations were converted into phyllites or fine-grained schists, in which the foliation was caused partly by rearrangement of fine-grained material and partly by recrystallization of it into chlorite, sericite, and some muscovite and biotite. Exceptions to these general results are apparent at places in most formations, and at those places the rocks are virtually unmetamorphosed, although they are deformed.

METAMORPHISM OF THE IGNEOUS ROCKS

The igneous rocks of the Menominee district that could have been affected by pre-Animikie metamorphism are those from which the Carney Lake Gneiss was formed and those of the Quinnesec Formation. Those that could have been affected by the post-Animikie metamorphism are the two foregoing groups plus the Animikie sedimentary and volcanic rocks, and the late-Animikie or post-Animikie gabbroic dikes and sills. The Marinette Quartz Diorite and the Hoskin Lake Granite apparently were emplaced about the same time as the post-Animikie metamorphism, the quartz diorite somewhat earlier than the granite; thus, the granite caused some metamorphism of the quartz diorite as well as of the adjacent Quinnesec strata. The sparse granite, pegmatite, and aplite dikes that cut the Quinnesec, the Animikie rocks, and the late- or post-Animikie metagabbro were emplaced after the climax of the deformation and metamorphism, and the Keweenawan diabase dikes distinctly later than the metamorphism. To facilitate discussion, the rocks are divided into three groups: (1) the Carney Lake Gneiss; (2) the mafic rocks, which group includes the largely volcanic Quinnesec and Badwater Formations, and the gabbroic dikes and sills; and (3) the Marinette Quartz Diorite.

THE CARNEY LAKE GNEISS

S. B. Treves showed (p. 20) that the Carney Lake Gneiss consists of gray gneiss, composite gneiss, and red gneiss; that it contains inclusions of amphibolite and biotite schist; that the gray gneiss grades into composite gneiss, into inclusions of amphibolite, and, more rarely, into inclusions of biotite schist; and that the composite gneiss grades into biotite schist. He also showed that the original rock of the area now underlain by the Carney Lake Gneiss probably was basaltic lava and some intercalated sediments.

Treves' detailed study indicates clearly that in this area during the time of pre-Animikie metamorphism the original basaltic lavas and intercalated sediments were invaded by granitic magma. The basalt was changed to amphibolite, then progressively to biotite schist, gray gneiss, composite gneiss, and red gneiss; the minor sedimentary intercalations were changed chiefly into quartz biotite schist and then into gneiss.

The effect of post-Animikie metamorphism on the Carney Lake Gneiss is obscure. No definite petrographic changes can be ascribed to such metamorphism, and apparently the gneiss was little affected. In the gneiss area, therefore, the late-Animikie metagabbro dikes are used as indicators of the metamorphic grade.

THE MAFIC ROCKS

The mafic igneous rocks, except the Keweenawan diabase dikes, have all been altered, and similar mineral assemblages have been formed from all rocks, whether lava flows, dikes, or sills. These mineral assemblages differ, however, from place to place, some probably because of the intrusion of the Hoskin Lake Granite, others as a result of changes in the intensity of regional metamorphism. In the main the various mineral assemblages formed by alteration appear to fit into the general metamorphic pattern of the region, as outlined by James (1955, pl. 1).

THE QUINNESEC AND BADWATER FORMATIONS

The Quinnesec and Badwater Formations are composed chiefly of lava and tuffs of general basaltic composition, in which the original minerals probably were chiefly labradorite and augite, inasmuch as relics of these minerals have been identified in both formations. The Badwater and much of the Quinnesec consist of typical greenstone, green schist, and chlorite schist, but southern and western parts of the Quinnesec Formation are hornblende schist, amphibolite, and, rarely, hornblende-pyroxene hornfels.

The minerals commonly present in varying amounts in all the greenstone, whether of the Quinnesec or of the Badwater, are chlorite, actinolitic amphibole, and minerals of the epidote group; also, most of the Quinnesec greenstone contains albite, whereas in the Badwater greenstone albite is less common. The presence of quartz, usually in minor amounts, is not uncommon. The minerals usually present in the schist derived from greenstones of both formations are: in the greenschist, chlorite, epidote, and actinolite, and in the chlorite schist, chiefly chlorite.

The greenstone, green schist, and chlorite schist of the Quinnesec Formation give place to hornblende schist, amphibolite, and hornblende hornfels in the southernmost exposures. The usual minerals present in the hornblende schist and amphibolite of the oligoclase-amphibolite facies are oligoclase, epidote, which in places is abundant, and usually some quartz. Either biotite or chlorite may be present also, the chlorite apparently formed by alteration of hornblende. The common minerals in similar rocks but of the andesine-amphibolite facies are blue-green to brownish-green hornblende, andesine, and some quartz; some diopsidic pyroxene is present in rocks very close to the Hoskin Lake Granite. Xenoliths of the Quinnesec Formation in the granite are composed chiefly of green hornblende, diopsidic pyroxene, and andesine.

THE LATE-ANIMIKIE METAGABBRO DIKES AND SILLS

Some of the late-Animikie metagabbro dikes are massive, others are sheared. Either type is characterized by assemblages of minerals that differ from one part of the area to the other. The massive dikes that cut the Animikie strata and also those that lie within the southernmost part of the Carney Lake Gneiss (greenschist facies) are composed chiefly of actinolitic amphibole and albite, or chiefly of minerals of the epidote group, whereas the sheared dikes are chiefly chlorite schists; however, the massive dikes that lie farther north within the Carney Lake Gneiss (oligoclase-amphibolite facies) are composed chiefly of hornblende and oligoclase, and the sheared ones of the same area are hornblende-oligoclase schist. Most of the oligoclase-bearing metagabbro contains some biotite.

The large late-Animikie metagabbro sills of the district lie chiefly within the Quinnesec Formation, and, like that formation, they span several facies of regional metamorphism. The different mineral compositions of the various sills and along a single sill, as in the Horserace sill (pl. 1), reflect the regional metamorphic grade. The principal mineral assemblages present in the sills are summarized in the table on page 67. The eight groups of mineral assemblages shown in that table may be condensed to four, each characterized by a few minerals that constitute the major components of the sills, although in different combinations and proportions. These are: (1) actinolite-chlorite-albite-epidote, (2) hornblende-chlorite-albite-epidote, (3) hornblende-oligoclase-epidote, and (4) hornblende-andesine. Assemblages 1 and 2 correspond to greenschist facies, and assemblages 3 and 4 correspond to the oligoclase-amphibolite facies and andesine-amphibolite facies, respectively. The distribution of these assemblages is shown by isofacies on plates 1 and 2.

The meta-anorthosite associated with metagabbro in the sills shows the same general type of alteration. The newly formed plagioclase is albite or sodic oligoclase where the meta-anorthosite is associated with metagabbro composed chiefly of actinolite, chlorite, albite, and epidote; it is andesine where the meta-anorthosite is associated with metagabbro composed chiefly of hornblende, andesine, and epidote.

Serpentine and pyroxenite are present in places along the north side of the Sturgeon Falls sill, which is characterized by the mineral assemblage actinolite-chlorite-albite-epidote. The composition of the serpentinite, which is chiefly antigorite, is compatible with the pattern of alteration of the associated metagabbro, the difference being caused by a different composition of the original rock, which was probably chiefly olivine and pyroxene. The present composition of the pyroxenite seems to be anomalous, being made up chiefly of unaltered diopsidic pyroxene. A more detailed examination of the pyroxenite localities is needed.

THE MARINETTE QUARTZ DIORITE

The Marinette Quartz Diorite, which is thought to be of post-Animikie age and to have been intruded about the time of the general post-Animikie metamorphism, may have been little affected by regional metamorphism. A zone of mixed rocks as much as 1,000 feet thick that occurs near the junction of the quartz diorite and the granite shows, however, that the quartz diorite was intruded by the slightly younger Hoskin Lake Granite and was considerably affected by contact metamorphism and metasomatism. Many rocks that are gradational between granite and quartz diorite are present in this zone, but not enough of these mixed rocks were studied to permit a detailed description of the alteration of the quartz diorite.

The present study of the Marinette Quartz Diorite shows that hornblende has been partially replaced by biotite; that very locally biotite, in turn, has been partially replaced by chlorite; that sericite and epidote-clinozoisite have formed from plagioclase; and that microcline has been introduced interstitially, and locally it has partly replaced plagioclase. These features indicate that the present composition of the quartz diorite may be a result of contact metamorphism and metasomatism, and that the original rock, before intrusion of the granite, might have been a hornblende diorite or perhaps a hornblende- or augite-gabbro.

TIMES AND CAUSES OF METAMORPHISM

Only two times of orogeny accompanied by regional metamorphism have been identified on the basis of geologic relations and petrography. The older occurred long before Animikie time, and a Pb^{207}:Pb^{206} age for zircon from the Carney Lake Gneiss indicates that it was probably as much as 2,590 million years ago (Davis and others, 1960, p. 154).

The younger orogeny and regional metamorphism occurred at the end of Animikie time and before Keweenawan time. All the Animikie formations were affected by the folding and to varying degrees by the metamorphism. The late- or post-Animikie intrusive diabase dikes and gabbro sills were emplaced before metamorphism, and at least the last of them after some deformation had taken place. The Hoskin Lake Granite is somewhat deformed and is considered to be syntectonic. The exact age in years of this late- or post-Animikie event is difficult to fix. Numerous minerals from metamorphosed Animikie rocks and crosscutting igneous rocks have been dated by Davis and others (1960, p. 153–154), but these mineral dates are generally discordant and range from 1,100 to 2,010 million years. If maximum numbers are used, the post-Animikie metamorphism would seem to have occurred in the period 1,650 to 2,010 million years ago. In Minnesota and northwestern Wisconsin, where the post-Animikie metamorphism appears to be the last thermal event recorded by the rocks, Goldich and others (1961) have concluded that the post Animikie orogeny and metamorphism occurred about 1,700 million years ago. Ages younger than 1,650 million years which characterize most minerals from Dickinson County and eastern Iron County do not correlate with any known geologic event in the district and are not understood. Davis and others (1960, p. 154) have suggested that epochs of regional metamorphism 1,100 and 1,400 million years ago caused the common occurrence of these mineral dates. If regional metamorphism was the cause, it is a very subtle type that would not have been detected but for the mineral dating. Precise data on the thermal stability of the various minerals involved and the diffusion rates of the daughter products of the radioactive atoms are needed before even a qualitative estimate of the meaning of the mineral dates is possible.

The causes of the post-Animikie metamorphism in the Menominee district must be the same as those throughout the surrounding area, which have been discussed by James (1955). For the general region, metamorphism of low grade (chlorite and biotite zones) appears clearly to have been associated with intense pressures and somewhat elevated temperatures, and perhaps also widespread hydrous emanations that accompanied the post-Animikie orogeny. The higher grade nodes of metamorphism probably resulted from local surges of heat and perhaps emanations, the cause of which is not entirely clear, but appear most likely to have been associated with local bodies of subjacent magma. James (1955, p. 1485–1486) states:

> Evidently the heat for metamorphism was derived from subjacent bodies of magma, by means of which heat acquired at a greater depth was transferred by mass movement to higher levels in the crust. The central part of the metamorphic node presumably lies above the highest part of the subjacent body. On the basis of an estimated temperature of 600°C or more for the sillimanite zone and a thermal gradient that would have been a minimum of 1°C/50 feet, the magma was not more than a mile, and possibly was considerably less, below the present surface at the core region of a node.

James (1955, p. 1480–1483) discussed the relations between regional metamorphism and intrusive rocks and between metamorphism and structure in the northern Michigan area; he reached the conclusion (p. 1455) that,

* * * the metamorphic patterns show no clear relation to regional structure or to the exposed granitic intrusions of the same orogenic cycle. The granite is most abundant in the areas of strongest metamorphism, but field evidence shows that most of the masses were intruded after the metamorphism. Field evidence also shows that deformation and metamorphism are independent variables in the orogenic scheme for this particular region.

We agree with James on all points. The metamorphism and deformation were broadly synchronous; they acted independently, however, and with different timing in different parts of the affected area. For example, in central Dickinson County the metamorphism followed the deformation, in eastern Iron County the deformation continued after the metamorphism (Bayley, 1959b, p. 88–94), and in the Menominee district they were apparently contemporaneous.

Most of the post-Animikie intrusive igneous rocks occur in the areas of moderate- to high-grade metamorphism. The granite dikes, pegmatites, and aplites were intruded after most of the deformation and probably after most of the metamorphism; minor plutons, such as the Peavy Pond Complex of eastern Iron County (Bayley, 1959b) and the Marinette Quartz Diorite and Hoskin Lake Granite of the Menominee district, were intruded during the deformation and metamorphism—that is, were syntectonic. The exposed areas of these plutons is only a few square miles. They show contact metamorphic effects on the adjacent rocks, but the plutons are very small with respect to the area regionally metamorphosed; we believe that they are pinnacles or cupolas that enlarge downward and join much larger igneous masses that caused the regional metamorphism.

STRUCTURE

The Precambrian rocks of the Menominee district have been strongly deformed. The present major structural features are largely the result of a post-Animikie period of folding and faulting—the Penokean orogeny, as redefined by Goldich and others (1961)—but two earlier Animikie diastrophic epochs and a pre-Animikie orogeny are recorded in the rocks.

MAJOR STRUCTURAL FEATURES

Three major structural elements compose the fundamental framework of the district: (1) the lower Precambrian anticlinal block of Carney Lake Gneiss in the northeast, (2) the middle Precambrian Menominee trough in the central part of the district, and (3) the lower Precambrian anticlinal Wisconsin complex in the southwest (fig. 22). This basic three-part framework is reminiscent of the several other Precambrian iron districts to the north, all of which are east-trending troughlike areas of Animikie rocks infolded or infaulted between anticlinal masses of older Precambrian rock. However, each possesses internal peculiarities not shared by the other districts.

The Menominee trough is basically a south-facing homocline—half of a syncline. We presume that it was once a syncline, the south limb of which was removed by erosion from the upthrown (south) block of the major fault that separates the Animikie strata from the Quinnesec Formation in the south part of the district. The homocline is severely plicated internally, and is interrupted near the middle by a major longitudinal fault along which a downdip segment of the homocline was elevated to form the south iron range.

Only the flaring west end of the Menominee trough lies within the district. By far the greater part lies to the east, beneath Paleozoic rocks, where it has been traced by magnetic surveying and diamond drilling eastward for more than 40 miles (Allen, 1915b). To the west of the district, the faulted south margin of the trough continues its northwestward trend for many miles, whereas the north margin flares out northward following the northwest margin of the Carney Lake Gneiss. This expanded northwest quadrant of the trough is dominated by the east end of a west-plunging syncline whose general outline is indicated by the distribution of the Badwater Greenstone, and whose axis must lie between the south-facing gneiss front and the north iron range fault block. This syncline, though superficial, is a major structural feature which assumes synclinorial proportions westward where it encompasses the upper Animikie Paint River Group of the Florence iron district, Wisconsin, and the Crystal Falls and Iron River iron districts, Michigan.

The important structural features of three major structural elements will now be briefly discussed in this order: (1) the Carney Lake anticline, (2) the Quinnesec (Wisconsin complex) area, and (3) the Menominee trough.

CARNEY LAKE ANTICLINE

The broad Carney Lake anticline trends northwest and is composed chiefly of ancient granitic gneiss which contains inclusions and partly assimilated clots of yet older volcanic and sedimentary rocks. The approximate internal structure of the gneiss anticline is shown on figure 22 by form lines based on hundreds of observations of strike and dip of layering, foliation, and other structural elements which vary in a perplexing way throughout the gneiss area. The figure shows that the gneiss is intricately folded. Moreover, it shows that the folds are truncated by the overlying Animikie formations, and therefore could not possibly be related to the broad anticline in which they are exposed. The gneiss and its internal folds

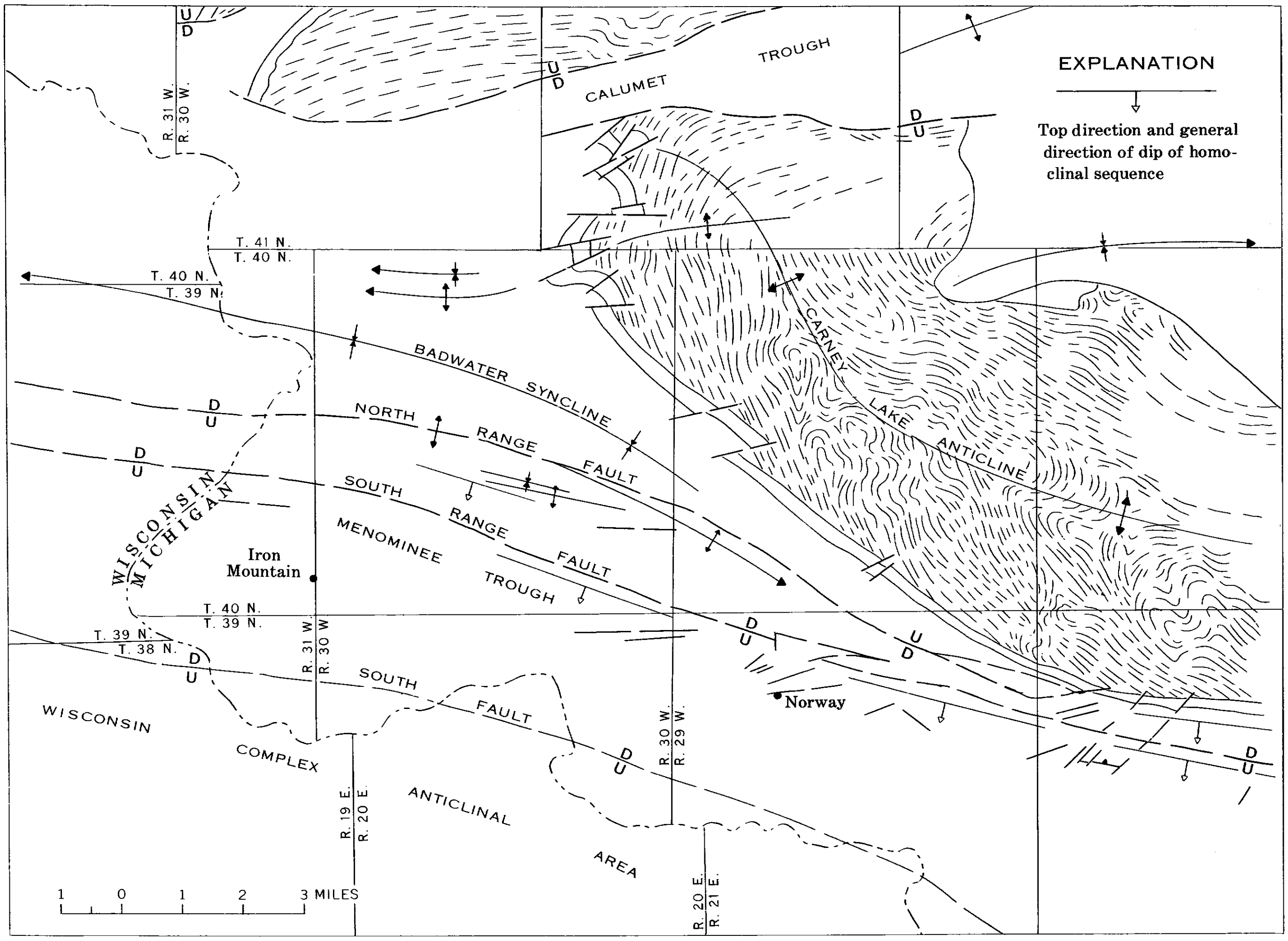

Figure 22.—Map of the Menominee district, extended north to include the Calumet trough, showing major structural elements, and, by form lines, the internal folding of the Carney Lake anticline.

formed long before Animikie time, and the nature of the rock indicates it formed at great depth. On the basis of this postulated depth and the age dating, we infer that a very extended period of uplift, erosion, and sedimentation intervened between the formation of the gneiss and the formation of the present gneiss anticline. According to local geochronology, this period lasted 700–800 million years, only a small part of which is needed to account for Animikie sedimentation, because equally thick series have been deposited within Tertiary time.

The axis of the gneiss anticline, though dominantly northwest trending, is somewhat arcuate, convex toward the west. Crossfolding on an east-west axis roughly parallel to the line between T. 40 N. and T. 41 N. is indicated by the synclinal reentrant of Michigamme Slate on the east flank, and the complementary bulge on the northwest flank.

The south limb of the gneiss anticline is vertical or slightly overturned and dips under the gneiss as if the gneiss had moved up and out over the mantling formations, whereas the single available exposure of Sturgeon Quartzite indicates that the north limb is steeply dipping but apparently not overturned. Pettijohn (1943, p. 390–391) observed that the conglomerate beds of the Fern Creek Formation are steeply dipping or overturned, yet adhere tightly to the underlying gneiss; and he concluded from this feature that the margins of the gneiss must have rolled out a similar amount. In addition to the outward rotation of the gneiss block, there is evidence that the gneiss moved upward and outward by shearing along high-angle marginal faults. James Trow (written commun., 1955) mapped a fragment of such a reverse fault in a schist zone between the base of the Sturgeon Quartzite and the underlying Fern Creek Formation in sec. 8, just west of the Sturgeon River Dam (fault not shown on pl. 3), and we infer that this general schist zone, which is poorly exposed, was the site of considerable fault movement. At one place, about 2 miles beyond the north edge of the mapped area along the Sturgeon Quartzite belt, the quartzite was repeated along such a marginal fault before the advent of the cross faults that displace the gneiss and quartzite (James and others, 1961, pl. 1).

A survey of the evidence concerning the history of the Carney Lake anticline suggests that it has been a buoyant tectonic element for a very long time. The general area of the anticline was above sea level and being eroded in pre-Animikie time, twice again in Animikie time, and at least twice more after Animikie time; in alternate periods it was submerged and was receiving sedimentary materials. Its first uplift in Animikie time is speculative, and the exact area of uplift is unknown, but that uplift occurred close to the district is indicated by the conglomerates that make up significant parts of the Randville and Felch Formations.

The second Animikie uplift of the gneiss is better recorded. It occurred after the deposition of the Vulcan Iron-Formation and before the deposition of the Michigamme Slate. The basic anticlinal shape of the uplift formed during this disturbance, and this uplift remained a positive tectonic element during the remainder of Animikie time. The pre-Michigamme anticline probably looked very much like the present one except that the limbs were not so steeply dipping. The crest trended east-west in approximately the area now underlain by the Badwater Greenstone, and from this crest area and about the margins of the anticline the pre-Michigamme Animikie formations were stripped away by erosion.

During Michigamme and later Animikie sedimentation the crest area of the gneiss anticline was relatively high, and thus it was covered late, and then only by a thin layer of late Animikie sediments. The Michigamme Slate, for example, which is commonly 4,000–5,000 feet thick in areas close by, is less than 500 feet thick on the crest of the anticline, and the Badwater Greenstone, which is about 10,000 feet thick to the west of the anticline, thins to less than one half of that in its easternmost part.

How the gneiss anticline reacted to the post-Animikie orogenic forces is largely supposition, but some relations are clearly recorded. On a regional scale, the present structures suggest that the post-Animikie orogenic areas were subjected to a north-south compression. It seems likely that the strong upward and outward thrust of the gneiss occurred at this time. The synclinal reentrant of Michigamme Slate on the northeast flank suggests that the block also received some of its arcuate bend at that time. The cross faults on the convex (west) side of the bend are likely to be related to the bending; they possibly reflect nonsimultaneous movements of the gneiss along fractures formed approximately parallel to the east-west hingeline of the bend. As noted above, the cross faults postdate the outthrusting of the gneiss block; the present steep dip of the Sturgeon Quartzite was therefore probably attained before they formed. The fact that some of the cross faults displace Michigamme strata (pls. 2, 3) dates them as last-forming features of the gneiss uplift.

According to Higgins (1947, p. 479), the aggregate eastward movement on the cross faults is approximately 3.2 miles, assuming that the Sturgeon Quartzite belt was straight and its strike was northwest before faulting, but James (in James and others, 1961, p. 68) pointed out that the quartzite belt was probably not straight but had recurved—owing to the bend in the gneiss anticline noted above—and that the aggregate movement was probably much less. The striations in the fault zones, according to Higgins (1947, p. 479), pitch 2°–40° SW and suggest upward and eastward movement on the faults. Higgins regarded the faults as vertical tear faults resulting from a severe compression from the southwest, active after the gneiss block was in its uplifted position, and he has presented quartz petrofabric diagrams for the Sturgeon Quartzite and the gneiss to support his concept. However, James (James and others, 1961, p. 68–70), who regards the cross faults as being related to the upthrusting of the gneiss, has pointed out that the petrofabrics of quartz are little understood and subject to bias, and that the data presented by Higgins can be manipulated, by choice of axes, to support upthrusting as the cause of the measured fabrics; James therefore concluded that it is less hazardous to support any structural interpretations on other evidence.

To summarize briefly, the Carney Lake anticline owes its internal folding to a pre-Animikie orogeny. As an anticline it has had a long history as a positive tectonic element. It attained its present general outline in post-Vulcan and pre-Michigamme time, and this original structure was greatly accentuated and modified by faults during the post-Animikie or Penokean orogeny.

QUINNESEC (WISCONSIN COMPLEX) AREA

The outcrop area of Quinnesec Formation represents but a very small part of the north fringe of the general Wisconsin complex. The complex, like the Carney Lake Gneiss and other lower Precambrian terranes to the north, was basement upon which the Animikie rocks were deposited, and we assume that its pre-Animikie history was in general like that of these other terranes. By analogy, for example with the Dickinson Group of central Dickinson County and eastern

Iron County and with the Mona and Kitchi Schists of the Marquette district, we may infer that the Quinnesec terrane in pre-Animikie time was virtually as it now appears—that is, folded and foliated, more or less metamorphosed, and intruded by batholithic masses of pre-Animikie granite. Further, the distribution and character of the Animikie sediments suggest that the Wisconsin complex area was periodically positive during Animikie time as was the Carney Lake Gneiss area.

The present structure of the Quinnesec Formation area then is for the most part inherited from pre-Animikie time, but this structure was extensively modified by Animikie and post-Animikie intrusive bodies and by post-Animikie deformation and metamorphism. The general northwest strike of the beds and foliations and the tight folding noted at a few places could be inherited features. The late Animikie metagabbro bodies cut Quinnesec strata that were already well foliated, and, though generally massive and undeformed themselves, they contain abundant foliated schist inclusions. These deformed inclusions mean that a strong deformation preceded the intrusion of the metagabbro, and we presume that it was a pre-Animikie deformation, but there seems no way to prove or disprove this.

The foliation of the Quinnesec strata generally follows the bedding. It strikes northwestward to westward as a tight-fitting mantle on the north and east sides of the convex Hoskin Lake Granite lobe, and is in part truncated and partly shouldered aside by the metagabbro bodies. Minor tight plications in the felsic part of the Quinnesec north of Aurora and folds indicated in the mafic part by opposing pillow tops were likely premetagabbro structures, and thus probably pre-Animikie. The plicated felsic rocks were extremely deformed cataclastically before emplacement of the metagabbro. They retained this characteristic in the oligoclase-amphibolite zone of post-Animikie metamorphism, but lost it entirely in the greenschist zone where they have been extensively sericitized.

Another line of evidence which indicates a premetagabbro folded condition for the Quinnesec strata is the attitude of the metagabbro bodies discussed on page 71.

THE MENOMINEE TROUGH

By reference to the several map plates and sections and to figure 23 it is possible to visualize fairly well a three-dimensional picture of the present Menominee trough. In addition, much of the detailed structural data for certain of the mine areas, given previously by W. S. Bayley (1904) and by Van Hise and Leith (1911), is still valid and usable; therefore we shall limit our discussion to features not previously described and to comments on the origin of the major structures as we understand them.

LARGE-SCALE FOLDS

Three ages of folding have contributed to the present configuration of the Menominee trough: (1) in late-Randville and post-Randville time, (2) during post-Vulcan, and (3) in post-Animikie time.

POST-RANDVILLE FOLDING

The evidence for post-Randville-pre-Felch diastrophism has been summarized on page 41. This diastrophism was regional in scope. The distribution of pre-Vulcan remnants of the Randville strata suggests that the folding was very gentle, but the erosion that followed exposed the pre-Animikie rocks over very broad areas, and in these denuded areas the succeeding formations were deposited on the pre-Animikie basement.

In the Menominee district the post-Randville folding was apparently very gentle except locally, as near Iron Hill (pl. 2; fig. 11). From many factors previously considered we have inferred that diastrophic movements in late- and post-Randville time probably produced an east-trending shallow structural trough into which the Menominee sediments were deposited, but the original dimensions of this trough are unknown. What appear to us to be boulders of Sturgeon Quartzite in the basal conglomerate of the Felch Formation indicate a possible uplift of at least 1,000–1,500 feet in some nearby areas, probably in the Carney Lake Gneiss area and possibly in the Wisconsin complex area. This estimated uplift does not mean, however, that topographic highs of that magnitude existed, for erosion may have kept pace with the uplift as is indicated by the many beds of dolomite conglomerate in the Randville.

POST-VULCAN FOLDING

The post-Vulcan diastrophism was also regional in extent. Within the district it caused uplift of the Carney Lake anticline and sharp folding of the mantling formations, and erosion subsequent to the folding removed some of the Vulcan strata before the Michigamme Slate was deposited. In figure 24 we have attempted to show the geology of the district as it may have appeared in pre-Michigamme time, after the post-Vulcan folding and erosion. This paleogeology is highly conjectural, of course, but it explains some of the seemingly anomalous features of later post-Animikie structure.

The principal features are a west-pitching anticline which is centered on the Carney Lake Gneiss area and a west-pitching syncline which is centered in the Menominee trough area. The anticline is visualized as

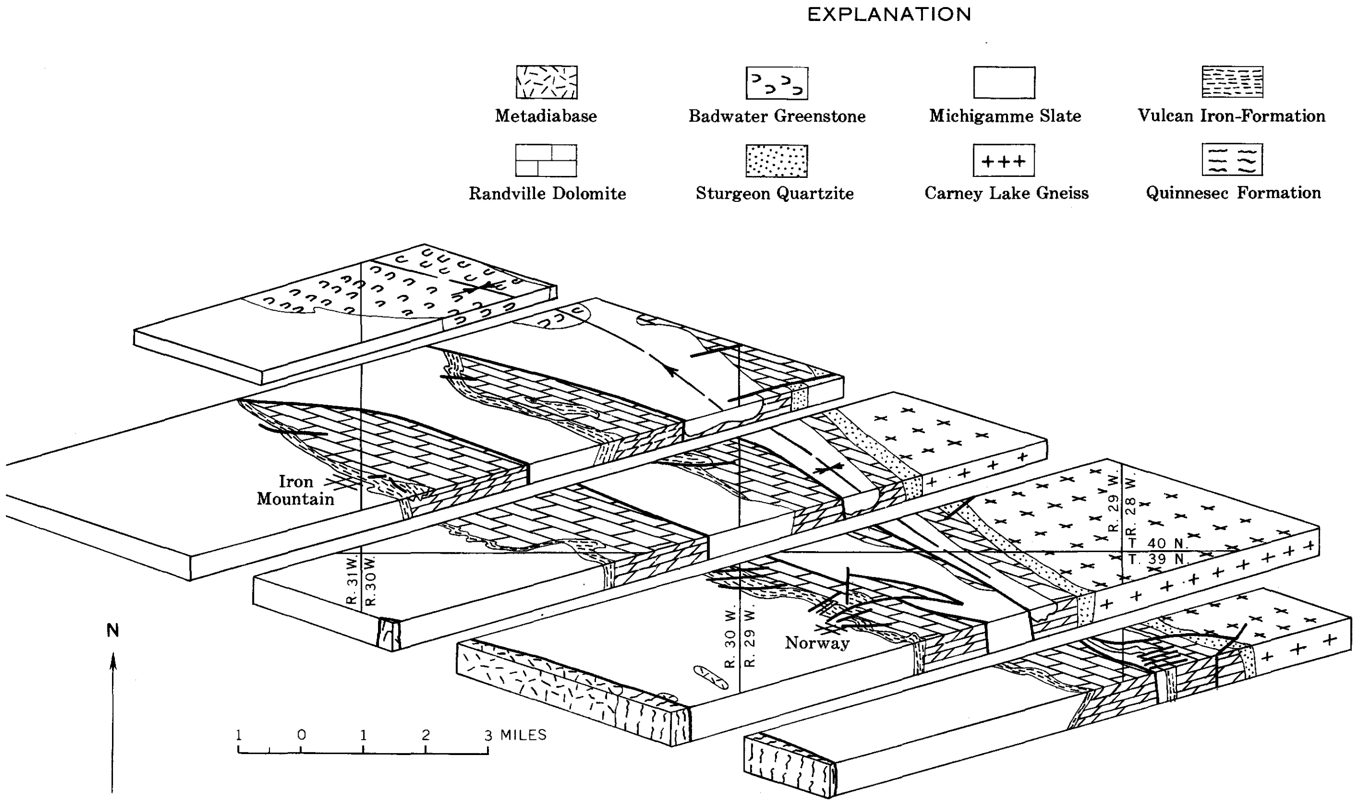

FIGURE 23.—Generalized structure of the Menominee iron-bearing district.

being broad and gentle, somewhat corrugated with secondary folds, and possibly faulted on the south flank. Pre-Michigamme erosion removed the Vulcan and Felch Formations from part of the crest of this anticline—that is, north of the north iron range where the Michigamme Slate now rests on the Randville Dolomite—and this erosion removed the Curry Iron-Bearing Member and part of the Brier Slate Member of the Vulcan, within the area outlined on figure 24 and labeled "Curry member removed." Secondary or drag folds related to the main anticline were narrow and steep sided but probably did not plunge steeply away from the anticline; 120–180 feet per mile is indicated in the area where the Curry Member was removed. These minor folds include the steep-plunging and often overturned folds that characterize especially the south iron range. The present attitude of these folds was probably caused by post-Animikie folding and later faulting on the range faults.

POST-ANIMIKIE FOLDING

The post-Animikie folding was likewise regional in scope, but extremely intense in contrast to the two previous gentle Animikie foldings. Large complicated synclinoria and anticlinoria and broad tracts of isoclinally folded beds characterize this orogenic period. Rock flowage on fold limbs, complicated drag folds, fold-faults, strong foliation and lineation, and mineral reconstitution are characteristic features.

For the northern Michigan area generally, the structural features suggest a general north-south compression, deflected and directed in a complicated way by the action and interaction of uplifted blocks of competent basement complex.

In the Menominee district the folds suggest the north-south compression in the western part, but a more northeasterly one in the eastern part where the trough narrows to a corridor between the basement massifs. Further, the overturned beds mantling the Carney Lake Gneiss and the similarly overturned beds and folds in the Menominee trough suggest that probably the uplift and mushrooming of the Carney Lake Gneiss largely controlled the folding within the trough; the Wisconsin complex, however, though uplifted during the folding, probably presented a more or less static buttress against which the folding took place.

Any interpretation of the trough folds requires a visualization of the terrane as it may have appeared before it was so deeply eroded as it is today.

Such a mental reconstruction has been tried for section *A–A′* (pl. 1); see figure 25. This inferred structure shows a reasonable fit to observed details

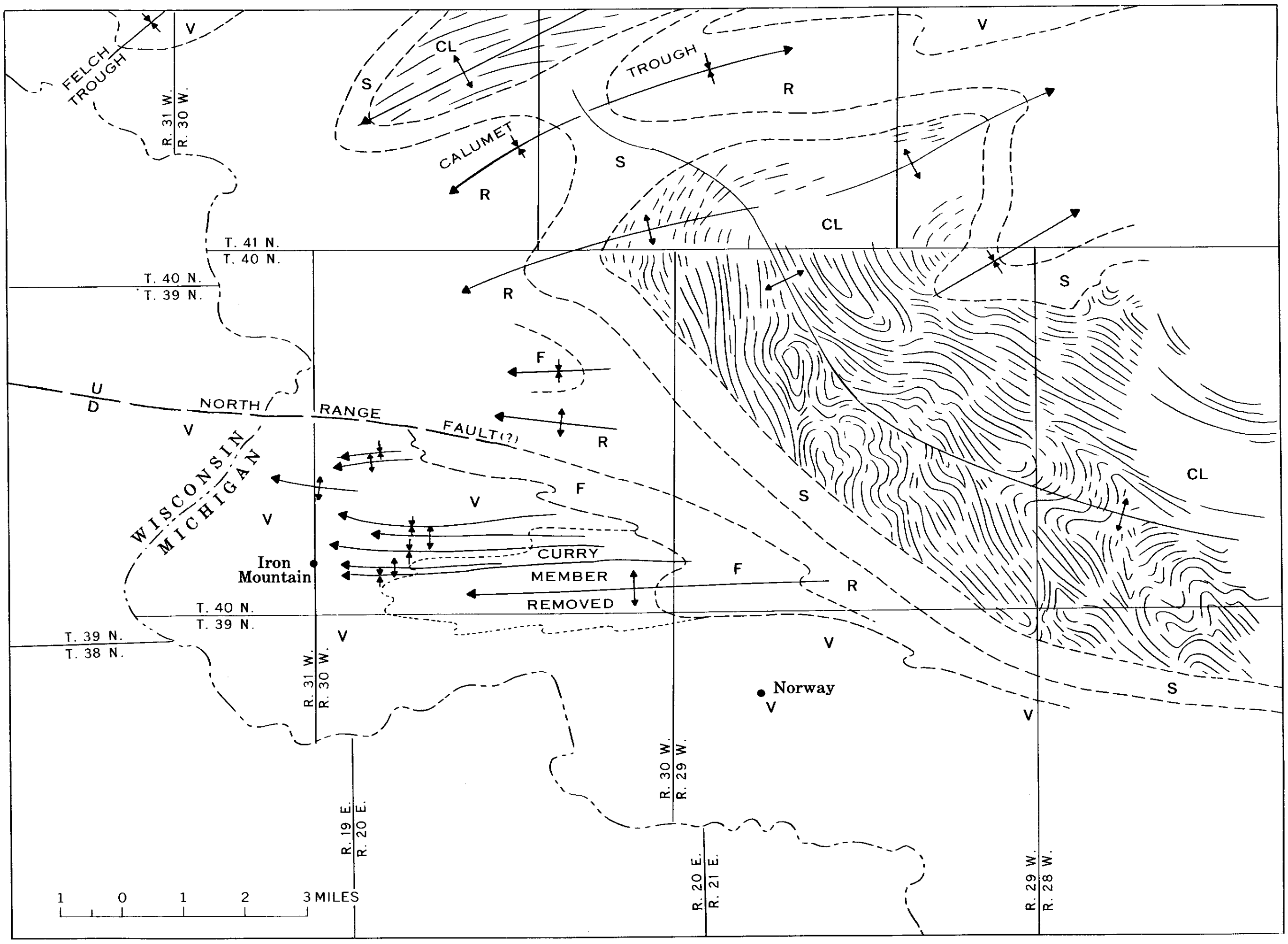

Figure 24.—Inferred structure at end of pre-Michigamme erosion. CL, Carney Lake Gneiss; S, Sturgeon Quartzite; R, Randville Dolomite; F, Felch Formation; V, Vulcan Iron-Formation.

such as drag folds, attitudes, and tops of beds; therefore, the reconstruction is believed to resemble closely the original. Only erosion need be invoked to produce the structure as we now see it. The major folds shown on the section, such as the Badwater syncline on the right end and the overturned anticline-syncline couplet on the left half are post-Animikie folds. The north range fault, which likely dates from the post-Vulcan folding (fig. 24), served as a locus for a post-Animikie anticline—the iron range moved up, whereas the previous movement on that fault was down on the iron range side (fig. 24). Some of the minor or drag folds are related to the post-Vulcan folding as previously noted (fig. 24). Those of the south iron range were overturned as a result of the post-Animikie folding and faulting. Some drag folds are probably related to the post-Animikie folding but cannot be separated from those folds formed earlier. Because of the map scale, only the largest drag folds are shown, but some feeling of the complicated detail actually observed may be derived from a sketch of the west wall of the Bradley open pit presented as figure 26, and from the photograph (fig. 27) which shows a series of plunging folds at the juncture of the Clifford and Traders open pits, sec. 20, T. 40 N., R. 30 W. (pl. 1).

Genetically related to the drag folding are axial-plane faults which formed generally by overextension of the inner short limbs of the folds after maximum folding, but which also cross incipient folds. In places the faults are not directly related to drag folds, but as a group they are systematic, and the apparent movement is largely consistent with the group. The east- or northeast-trending system of diagonal faults that is conspicuous in both iron ranges would appear to have originated in the folding process as described above.

The major east-southeast-trending faults seem to be younger than all the diagonal faults with the possible exception of those that bound the pie-shaped fault

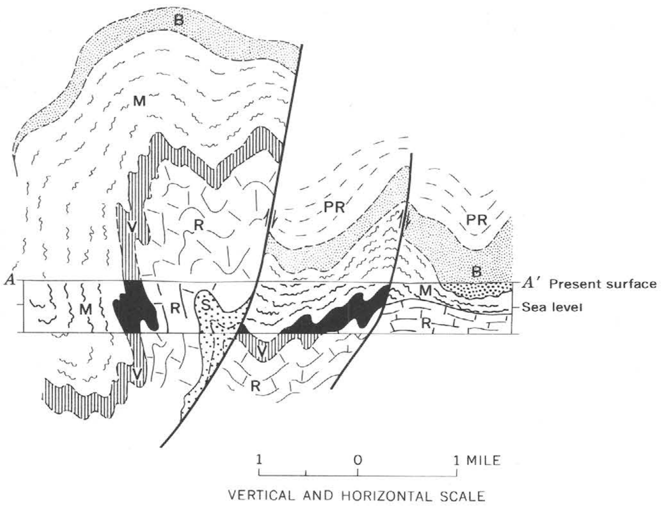

FIGURE 25.—Expanded version of section *A–A'*, plate 1 (in box), showing a hypothetical reconstruction of post-Animikie folds and faults. S, Sturgeon Quartzite; R, Randville Dolomite; black and vertical line pattern, Vulcan Iron-Formation; M, Michigamme Slate; B, Badwater Greenstone; PR, Paint River Group.

block at Norway (pl. 2), but this interpretation depends on the validity of certain assumptions. For example, the diagonal faults near Loretto (pl. 3) are assumed not to cross the south range fault, even though quite similar displacements have affected both north and south blocks. The two iron-range faults and the south fault (fig. 22) all show the same relative movement—south side up—and therefore seem to be related, but simultaneous movement is not inferred.

The south fault has dislocated the isofacies of post-Animikie regional metamorphism (pl. 1). It is thus younger than the metamorphism, but older than Upper Cambrian rocks, which it does not disturb (east of mapped area). Similarly, a large northeast-trending fault in Iron and Dickinson Counties, the Bush Lake fault, dislocates the isofacies of regional metamorphism (Bayley, 1959b). The timing on the above major faults indicates that they, and other seemingly related major east-trending faults such as the south range fault, are postorogenic.

The north range fault, it seems, is in a category by itself. We have inferred, from the distribution of the Vulcan Iron-Formation, that the fault was active during the post-Vulcan folding (see fig. 24). If so, the movement was north side up, whereas the post-Animikie movement on this fault was south side up; we infer, as depicted in figure 25, that upthrust on the fault largely controlled the anticline in the post-Vulcan formations. We believed that the north range fault moved early in the orogenic process, but no really definite time of movement can be assigned to it.

The amount of vertical movement on these faults can only be guessed. Diamond drilling in the area north of the north iron range suggests that the vertical movement on that fault is about 1,000 feet. We estimate 1,000–8,000 feet of displacement for the south iron range fault, and considerably more for the south complex fault.

The least understood fault structure in the district is that of the wedge-shaped graben near Norway (pl. 2). This graben is highly brecciated on each bounding fault and grabenlike in aspect, but is probably not a true graben in the sense that it is probably not a downdropped block, but a segment that lagged a little behind when the main block moved up on the main south range fault. In section (pl. 2, C–*C'*) the graben block reveals the kind of structure that has since been eroded from the top of the range. A south dip for

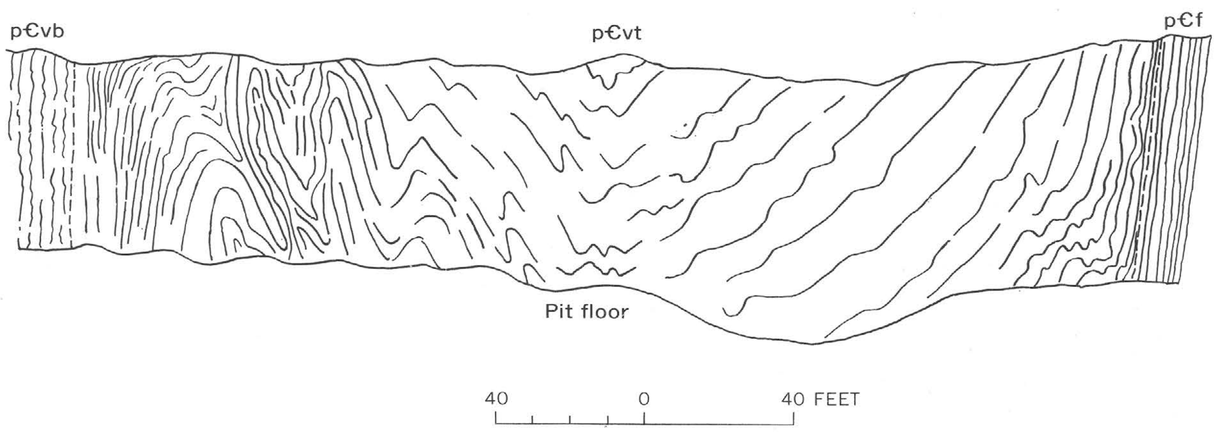

FIGURE 26.—Sketch of section of vertical west face of West Chapin mine open pit (Bradley pit) looking west (Oct. 1955). p€f, Felch Formation; p€vt, Traders Iron-Bearing Member of Vulcan Iron-Formation; p€vb, Brier Slate Member of Vulcan Iron-Formation.

FIGURE 27.—View looking northeast from southwest side of Clifford pit, sec. 20, T. 40 N., R. 30 W., showing three plunging anticlinal folds of Brier Slate Member with the Curry Iron-Bearing Member stripped off. Height of pit wall, approximately 60 feet. Upper flat surface is Precambrian peneplain.

the south range fault is indicated by the migration of that fault (by erosion) with respect to the lag block.

GEOLOGIC HISTORY

The geologic history of this area, insofar as we are able to determine, covers a period probably in excess of 3 billion years. It divides conveniently into three parts, (1) pre-Animikie, (2) Animikie, and (3) post-Animikie history.

PRE-ANIMIKIE HISTORY

The history starts with the deposition of the oldest known rocks, the volcanic and sedimentary rocks of the Quinnesec Formation, the scattered inclusions of volcanic and sedimentary rocks in the Carney Lake Gneiss, and probably the equivalent Dickinson Group rocks of central Dickinson County. The nature of these very ancient rocks suggests that they were deposited in water, probably in the sea, and that their depositional environment was not greatly different, if at all, from some present-day marine environments.

About 2.7 billion years ago these older rocks were strongly folded, in general on west or northwest axes, intruded by batholithic granite, and metamorphosed. A protracted period of erosion followed, which may have lasted hundreds of millions of years, during which the area was reduced to a plain of moderate relief near sea level.

ANIMIKIE HISTORY

A little more than 2 billion years ago the Animikie period began when, because of a general subsidence of the region, the sea slowly rolled over the old erosion surface and covered it with onlapping beach and shallow-water sand deposits (Sturgeon Quartzite). The sea advanced from the southeast, and when the shorelines had migrated far to the northwest, well beyond the limits of the district, all the district was covered by a warm shallow sea in which algae colonies flourished and carbonate rocks were deposited (Randville Dolomite). During Randville time the deposition of carbonate rocks was interrupted here and there by gentle warpings of the sea bottom which exposed areas of recently deposited carbonate rock to erosion, and which raised broad shallow tidal flats whereon newly deposited calcareous muds were subjected to desiccation. After a time, continued regional uplift, accompanied by local warping, caused the total withdrawal of the sea from the region, and thus ended Randville sedimentation. The withdrawal of the sea at this time, and the subsequent erosion, have resulted in a widely recognized hiatus in the Animikie record. How much time elapsed before the seas returned in Felch time, or whether the uplift in the Lake Superior region was in consequence of orogeny elsewhere, there is no way at present to tell.

The geologic record resumes with the transgression of the sea over the post-Randville erosion surface, which, because of the previous warping, was divided into shallow elongate basins that greatly restricted circulation. The direction from whence the sea came is yet unknown—from the east seems the most reasonable. As in Sturgeon time, the first deposits laid down by the transgressing sea were, at most places, sands and muds which were generally deposited alternately (Felch Formation); these alternations reflect slight oscillations in sea level. The Felch sediments must have been derived from local upwarps that remained out of water throughout Felch time. These islandlike areas were finally submerged near the end of Felch time by a final transgression of the sea which is recorded by the "Traders quartzite" of local usage. With these areas submerged and the source of clastic sediment largely cut off, chemical sedimentation became dominant, and the Vulcan Iron-Formation was deposited. Vulcan time was one of remarkable crustal stability, indeed virtual stagnation. To deposit, century after century, the minute and tenuous laminae of iron oxide and silica without any significant dilution by terrigenous clastic materials requires very special and constant conditions not only of the land areas but within the depositional basin.

James (1954) believes the Vulcan basin was partly barred from the open ocean by an offshore buckle, which, in the following epoch, was a center of volcanism. The barred basin is essential to the scheme in that it permits the formation of abnormal concentrations of iron and silica and at the same time provides the abnormal redox potentials in the bottom environment for their deposition. The land areas are believed to have been low lying and deeply weathered.

Near the middle and toward the end of Vulcan time, uplift, probably to the south of the district, caused extensive dilution, by clastic materials, of the iron being deposited (Brier and Loretto Slate Members). Vulcan deposition ended with uplift and warping of at least the marginal parts of the basin. In certain areas the Vulcan was folded, raised above sea level, and removed by erosion, and most of the debris was carried out of the district.

The last phase of Animikie deposition in this district started with a general rapid subsiding of the Animikie basin and a concomitant raising of the surrounding land areas. The types and thicknesses of clastic sediments deposited in this epoch indicate that the center of the basin (geosyncline) was northwest of the district. In the Michigamme Slate, the first to be deposited, there seems to be a change in the strata from shelf types—quartzite, dolomite, and slate—to geosynclinal types—graywacke-slate—as the district is traversed from southeast to northwest. Beyond the district, to the northwest, the Michigamme is composed largely of graywacke and slate. This situation suggests a Michigamme landmass to the south of the district.

The record of Animikie deposition, for this district, ends with the extrusion on the sea floor of a great lenticular mass of basaltic lava flows (Badwater Greenstone); however, younger Animikie formations present in Florence County, Wis. and Iron County, Mich., indicate that Animikie sedimentation continued long after the Badwater volcanism. Intrusive activity of basaltic lava, on a minor scale, also continued to the end of Animikie time.

POST-ANIMIKIE HISTORY

Folding, faulting, and uplift on a regional scale, and local metamorphism and intrusion of igneous rocks ended the Animikie era. Within the district the rocks were strongly folded and faulted, intruded by minor igneous bodies, and generally metamorphosed. A protracted epoch of erosion followed that probably lasted more than 1 billion years. Within this epoch, to the north and northwest, the Keweenawan Series was deposited, and fresh diabase dikes, thought to be related to Keweenawan volcanism, were intruded into most of the Animikie and older formations.

By approximately 500 million years ago the region had been eroded to a low-lying plain near sea level. In Late Cambrian time the sea again rolled across the region and deposited thick sandstone beds. These beds grade upward into carbonate rocks, probably of Ordovician age, but there the Paleozoic record ends. Except for regional uplift, the Paleozoic rocks have not been disturbed since they were deposited. They have been deeply eroded, however, and only remnants remain. The Mesozoic Era and Tertiary Period are not represented in any way. Probably during those ages the region was a very low lying shield area that eroded very slowly. Most of the Paleozoic rocks were removed by this erosion, but the Precambrian surface seems not to have been greatly changed by it.

In late Pleistocene time, continental glaciers moved over the district scouring and smoothing the rocks and removing any weathered rock or residual soils. Upon the melting of the glaciers, great quantities of sand and gravel were deposited from rivers of melt water. At many places the sands covered isolated blocks of ice, some of enormous size which, upon melting, left numerous deep holes (kettles) in the otherwise smooth sand deposits. Some of the largest kettles are now lakes, others were lakes and have been filled by vegetation, and still others are dry holes. The present streams formed as the glaciers melted. They still flow, for the most part, on the glacial deposits, into which the largest stream, the Menominee, has entrenched 150–200 feet.

ECONOMIC GEOLOGY

The Vulcan Iron-Formation and related ore deposits form the major mineral resource in this district. They were extensively but very selectively exploited in the past; in general, only certain iron-enriched parts, which constitute a small fraction of the total iron-formation, were mined. At present, there is no iron mining activity. Future prospects for renewed iron mining will depend primarily on demand. The techniques for upgrading the low-grade iron-formation to a saleable ore are known; it is largely the cost that will determine when these techniques will be used.

Nonmetallics, such as dolomite and sand and gravel, have been mined within the area for various purposes. Large quantities of these materials remain, and doubtless their production will continue in pace with the local demand. A yet untapped and nearly boundless resource of very pure silica may be found in the Sturgeon Quartzite. This material should be investigated to determine its suitability for such uses as glass, silicon alloys, and acid refractories.

IRON

DEFINITION OF THE TERM "IRON ORE"

The usage of the term "iron ore" in the Lake Superior region was defined by James and others (1961, p. 79) as follows:

The term "iron ore" has a dual meaning in the Lake Superior region in that it may be applied either to the material mined or to the material shipped to the furnaces. The fundamental usage, however, is not the commonly accepted one—that is, material that can be mined at a profit—but rather it applies to furnace feed. Inasmuch as about three fourths of the material mined is "direct shipping," the term "ore," used in the mining sense, is the same as the ore used in the metallurgical sense much of the time. But in 1958, for example, nearly half of the shipments was of material that had been treated so as to increase the iron content. This beneficiation product is referred to as "ore" in statistical records, and is graded to "ore specifications" just as is the direct-shipping ore, even though in the more usual sense the material treated is the ore and the product the concentrate.

* * * with the gradual increase in beneficiation and concentration, the duality in the definition of "ore" will stand out more clearly. At present the material mined for concentration is referred to as iron-formation, taconite, or low-grade ore.

The basic requirements for ore are set by the furnaces. Most of the acceptable material contains 51.50 percent or more iron and 5 to 12 percent silica. Other specifications are based on content of manganese, alumina, sulfur, and phosphorus, and on physical structure. A very small amount, about 0.6 percent of total shipments, is "siliceous ore"; this material, which typically contains 36 to 40 percent iron and 40 to 43 percent silica, is used only for special purposes and the demand is limited.

The economics of iron mining in the Lake Superior region have continued to change rapidly over the past few years; the main production at present is of concentrates from "taconite," or low-grade ores. The Vulcan Iron-Formation itself is mined for this purpose in the Felch district, a few miles north of the mapped area.

HISTORY OF IRON MINING

According to a historical sketch by N. H. Winchell (1895), iron ore was discovered in the Menominee district in 1848 by two explorers, J. W. Foster and S. W. Hill, but the first mining did not begin until 1870, when N. P. Saxton started digging pits and trenches on the site of the Breen mine. The first ore (53 tons) was shipped in 1873. All the major mines of the district had been located by 1878. Thereafter the new iron district was important among the ore producers from Lake Superior region until the last of its underground mines closed in 1945. The productive life span of the range was about 70 years, during which about 85 million tons of ore was mined and shipped, mainly

TABLE 35.—*Iron-ore production, Southern Dickinson County area, Menominee iron range*

[C, Chapin group; P, Penn group. (1892), date after which shipments were made as Penn group]

Mine	Opened	Abandoned	Ore shipped (tons)
Antoine (Clifford-Trader)	1895	1925	2,269,444
Aragon	1889	1931	11,160,975
Bradley	1937		[1] 486,786
Breen	1877	1907	75,425
Brier Hill (P)	1882	1883	14,981
Chapin	1880	1934	26,409,278
Cuff	1899	1942	83,306
Cundy	1896	1913	846,078
Curry (P)	1879	(1892)	[2] 416,928
Cyclops (P)	1878	(1892)	286,093
East Central (P)	([2])	([2])	([2])
East Vulcan	([2])	([2])	([2])
Eleanor (Appleton; Sturgeon River)	1887	1907	48,123
Emmett	1878	1884	131,940
Forest	1904	1904	11,988
Globe-Cornell	1880	1958	[1] 675,753
Half and Half	1889	1891	7,524
Hamilton (C)	1886	1892	[3] 96,072
Hersel	1890	1890	955
Indiana	1882	1920	244,527
Keel Ridge	1880	1899	93,101
Loretto	1893	1937	3,691,458
Ludington (C)	1880	1894	[3] 1,001,518
Millie (Hewitt) (C)	1881	1936	503,934
Munro	1903	1921	576,254
Norway (P)	1878	(1892)	[2] 1,291,352
Penn group of mines	1893	1945	21,629,154
Perry	1883	1883	3,138
Pewabic	1887	1918	9,359,339
Quinnesec	1878	1935	512,235
Saginaw (Perkins)	1879	1909	502,985
Stephenson	1879	1887	39,350
Verona	1900	1904	130,975
Vivian	1902	1913	482,187
Vulcan (P)	1877	(1892)	[2] 1,668,654
Walpole	1887	1891	19,089
West Chapin	1922	1936	144,760
Total			84,915,659

[1] Shipments to 1950. Bradley still open (1965).
[2] Included in Penn group total.
[3] Shipped as Chapin ore after dates given.

from underground mines. The annual production scarcely dropped below 1 million tons for the forty years between 1890 and 1930, and it exceeded 3 million tons for the peak production years, 1901 and 1906. The ore was shipped by rail to the port of Escanaba.

A tabulation of production by mines to 1950 is given in table 35; the figures are chiefly from publications by the Lake Superior Iron Ore Association (1938, 1952).

FUTURE EXPLORATION AND DEVELOPMENT

In view of the extensive exploration and mining that has been done, it is unlikely that any very large bodies of direct-shipping iron ore remain undiscovered in the Menominee range. Doubtless there are small ore bodies that have escaped detection, but it would not seem prudent to explore for or to develop them at this time; high-grade direct-shipping ores no longer command the premium they once did, and, indeed, they may be difficult to market at all. High-iron low-silica concentrates are now most favored by the steel mills; these concentrates are now being produced from low-grade iron-formation in great abundance, and no doubt will dominate the market in the near future. Any further development of the Menominee iron resources will

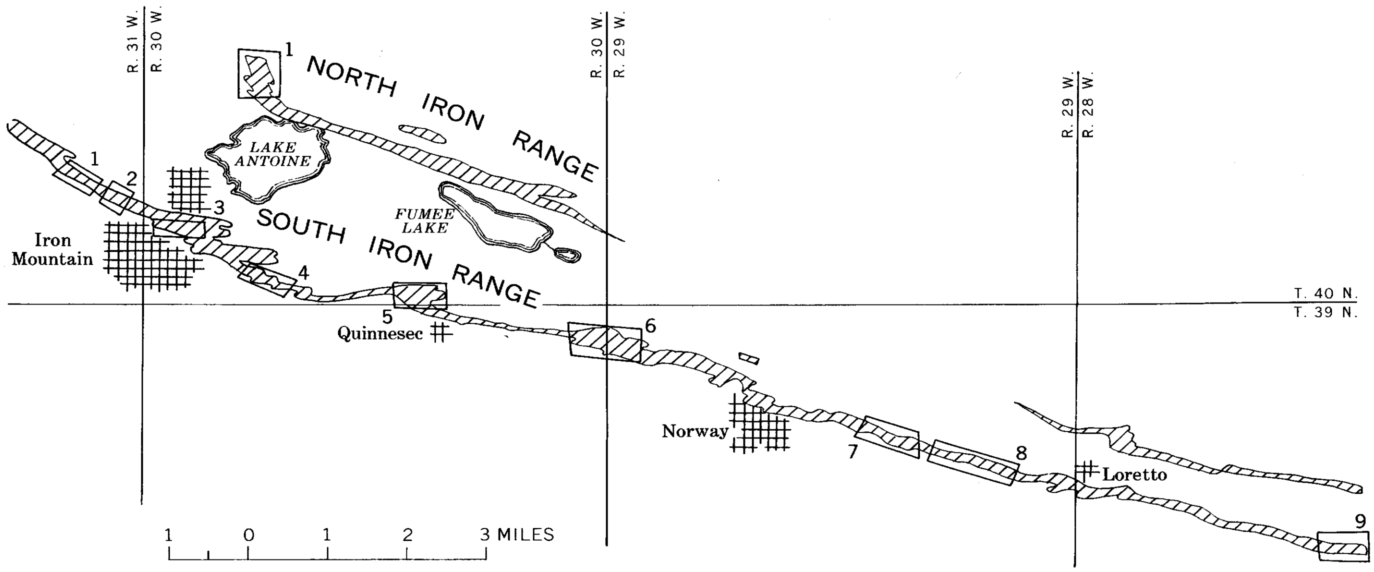

FIGURE 28.—Map of Menominee district showing distribution of iron-formation and the locations of specific areas of iron-formation discussed in the text.

depend on several factors: (1) the availability of large tonnages of low-grade iron-formation suitably situated for open pit mining, (2) the kinds of iron minerals, and (3) the sizes and distribution of the minerals in the iron-formation. Each of the three factors is critical: (1) the iron-formation must be available in very large quantities; (2) and (3) it must be amenable to beneficiation, and by some method economically commensurate with the total ore-to-market operation. We shall try here to evaluate, on admittedly insufficient information, only the geologic factors, which are: the availability of ore and the physical and chemical characteristics of the ore.

AVAILABILITY OF ORE

The only areas discussed here and indicated in figure 28 are those few that appear to fulfill the tonnage requirements for a long-term concentration operation—that is, where sufficient ore lies near the surface and can be mined by open pit methods. Reserves of iron-formation for each area are roughly estimated to a depth of 500 feet on the basis of 10 cubic feet per short ton.

NORTH IRON RANGE

On the north iron range, one area that may prove suitable for large-scale open pit mining is the area of the Clifford-Traders and Globe-Cornell pits. The estimated reserve of iron-formation is 70 million short tons.

This area is partly but not heavily covered by glacial deposits. These could easily be removed. A more reliable tonnage estimate will require the drilling of a series of short vertical holes in the broad northeast segment of this area. The quality of the natural ore for this area, as indicated by the analyses of ore shipments from the two abandoned open pits, is about 37 percent iron, 41 percent silica, and 0.01 percent phosphorus.

Though there are other places on the north iron range where iron is plentiful, they are generally undesirable because of deep glacial cover or abundant ground water. The Loretto-Appleton mine area, for example, may contain more than 60 million tons of iron-formation, but it is covered by 30–130 feet of glacial drift; thus, most of it is below the level of the adjacent Sturgeon River. The ore has been mined underground; probably the water does not present an insurmountable problem, but 5–8 million yards of stripping would be required to uncover the iron-formation.

SOUTH IRON RANGE

Area 1 (see fig. 28).—Traders Iron-Bearing Member. Approximately 150 feet thick. Shallow glacial cover; partial Cambrian sandstone cover. Length of belt, about 3,300 feet. Estimated reserve, 25 million tons. Average iron content, about 34 percent.

Area 2.—Bradley pit area. Heavy glacial cover east and west. East extension probably most desirable but hampered by existing roads and railroad. Length of belt, about 1,700 feet. Estimated reserve, about 15 million tons. Quality as indicated by Bradley ore: 38–44 percent iron natural, 35–42 percent silica, and 0.01 percent phosphorus.

Area 3.—Millie Hill. On the basis of tonnage and accessibility a very desirable area, but unfortunately close to dense residential area on the southwest. Approximate area, 2 million square feet having 100–200 feet of relief. Reserve estimate, about 83 million tons. Quality, as indicated by Millie ores and other sources: 28–38 percent iron natural, 40 percent silica, and 0.03 percent phosphorus. Magnetic oxide conversion tests made by the Bureau of Mineral Research, Michigan College of Mining and Technology, on 18 samples from the Millie pit area indicate that a very satisfactory magnetic concentration of this iron-formation can be made after a reducing roast and fine grinding, approximately 97 percent −325 mesh (Victor Kral, Ford Motor Co., written commun., 1958).

Area 4.—Keel Ridge mine area. Belt is about 3,500 feet long and about 150 feet wide. Covered by a few feet of soil at most places, but the soil is somewhat thicker in the western half. Only the Traders Iron-Bearing Member is present; it has an iron content of about 34 percent. Estimated reserves, about 25 million tons.

Area 5.—Vivian-Quinnesec mines area. Area extensively but thinly covered by glacial deposits, and, on the north edge, by Cambrian sandstone. Probably, because of the 100- to 400-foot separations between iron-formation belts, would require mining from several open pits having a possible aggregate surface area of about 800,000 square feet. Reserve estimate about 40 million tons, which, as previously noted, is for 500 feet of depth, which cannot be realistically applied to this area. Quality of ore indicated by Vivian and Quinnesec ore shipments for 1912 and 1935 respectively: 36 percent iron, 40 percent silica, and 0.02–0.34 percent phosphorus.

Area 6.—Munro mine area. This area has been extensively stripped and is otherwise only lightly covered by glacial deposits. Most of the exposures are on gentle to steep south-facing slopes. Cambrian sandstone overlaps on the north boundary. The exposed iron-formation is all Traders Member which dips generally steeply southwest, and which is cut by numerous east-west faults introducing slices of Felch Formation. It seems likely that some of the central part of the area will bottom at much less than 500 feet because of folding. The footwall contact can be expected to incline 50°–70° S. There is very roughly 1 million square feet underlain by easily accessible iron-formation; this area indicates 50 million tons of iron-formation, less an undetermined amount due to structural complications noted above. The ore is siliceous and probably contains about 34 percent iron.

Area 7.—Vulcan mine area. Area of the Traders Member is covered by as much as 80 feet of Cambrian sandstone. Hillside location is suited to stripping; the sandstone could be moved downhill on the Brier Slate Member, but several open stopes beneath the sandstone would have to be caved or otherwise dealt with before stripping commenced. There are roughly 35 million tons of 34-percent iron-formation in the belt.

The belt of Curry Iron-Bearing Member extends along strike for about 4,500 feet. The beds dip 60°–75° S. Soil cover is generally thin. Extensively mined underground, and some areas subsiding, especially in central part. Fifty million tons of iron-formation indicated; probably only one-half or one-third could be easily mined. Average iron content of Curry Member, about 31 percent.

Area 8.—This area, east of Vulcan, includes a belt about 6,000 feet long along which both iron-bearing members of the Vulcan are exposed or only lightly covered. The Traders and Curry Members are 100 and 150 feet thick, respectively, and dip about 85° S. They are separated by the Brier Slate Member, about 300 feet thick. As at other places, the iron-formations are presumed to contain an average of 31–34 percent iron. The Brier Member probably contains close to its average of 18 percent iron. The combined area of iron-formation is 1½ million square feet; the tonnage to 500 feet, 75 million tons. Open pit mining would, of course, be much hampered by the rib of Brier Slate Member between the pits. If methods could be devised to process the Brier Slate Member, which is also iron rich and represents a sizeable resource, the mining would be much simplified. There is roughly 800 million cubic feet of this ferruginous slate between the iron-formations to 500 feet depth, and this amount of slate probably contains more than 10 million tons of iron—nearly half as much as the combined iron-formations.

Area 9.—Breen mine area. This area extends to the east of the mapped area for some undetermined distance, but is at least 1,000 feet long. The three usual members of the Vulcan are present but not well exposed. The Curry Member here is approximately 38 percent iron, 40 percent silica, and 0.016 percent phosphorus. The Brier Slate Member is unusually iron rich, averaging about 26 percent. The quality of the Traders Member was not determined but is likely about 34 percent. If the Brier Member can be utilized, there is probably more than 20 million tons of concentrating material in this block.

MINERALOGY AND CHEMICAL COMPOSITION

The iron-formation was described on page 43. To summarize briefly, it is composed chiefly of finely divided crystalline silica and iron oxide, and some very minor constituents. Wide variations in the iron content of the iron-formation may be expected because of the various degrees of enrichment, but for the whole district it seems probable that the average iron-formation will contain about 32 percent iron, 40 percent silica, and 0.01–0.05 percent phosphorus.

As shown in figures 13–15 (p. 45), the iron oxide minerals are very fine grained and are very intimately mixed with the quartz matrix.

The iron occurs in two principal minerals: as hematite or specularite, and as magnetite or, if oxidized, martite. The proportions of these have not been determined. For the jasper beds that make up the bulk of the iron-formation, the iron minerals are chiefly confined to the granules and to a lesser extent to the quartz matrix. The granules average slightly less than 1 mm in diameter, and the amount of quartz matrix between granules varies widely from specimen to specimen, but commonly the distance between granules is about half their short dimension. Individual quartz grains or crystals range from about 0.02 to 0.34 mm in diameter. The average size (for 20 specimens) is 0.10 mm, about twice the size of the iron ore grains. The iron oxide minerals within the granules range from very fine (submicroscopic) dust to well-crystallized grains as much as 0.23 mm in diameter. The average size of iron ore mineral grains in 20 specimens examined is 0.04 mm, or slightly less than 325 mesh (Tyler). The following screen analysis of a sample of Traders iron-formation ground to pass a 3-mesh screen reflects clearly the small size of the quartz and iron-mineral grains and the intricate mixture between them.

Screen analysis of Traders Iron-Bearing Member −3 mesh

[From Dutton, 1942, p. 27]

Mesh (in.)	Weight percent	Percent Fe	Percent total Fe
+4	16.81	32.8	17.03
−4+6	19.82	31.4	19.22
−6+8	14.69	33.6	15.24
−8+10	11.68	31.4	11.33
−10+14	7.48	32.7	7.55
−14+20	6.11	32.7	6.17
−20+28	4.43	32.7	4.48
−28+35	3.32	32.8	3.36
−35+48	2.61	31.6	2.55
−48+65	2.21	29.8	2.04
−65+100	1.99	28.3	1.74
−100	8.85	34.0	9.29
Total or average	100.00	31.98	100.00

Note: The table shows that no appreciable fractionation between quartz and iron ore minerals was accomplished by the sieving. Probably mineral aggregates of nearly uniform composition dominate all size fractions.

Grain sizes determined for 22 specimens of Brier Slate are as follows: clastic quartz and feldspar grains average 0.05 mm; range, 0.11–0.01 mm. Martite or magnetite grains average 0.03 mm; range, 0.10–0.01. The rock is laminated, and the iron minerals are more or less segregated. Average analysis, 18 percent iron.

THE MENOMINEE DISTRICT IRON-FORMATIONS AS CONCENTRATING ORE

To release the iron minerals from the quartz will require fine grinding, probably to 90 percent −325 mesh. Such fine grinding is within the capabilities of present-day ore dressing, though expensive. However, the effectiveness of the various beneficiation techniques depends to a great extent on particle size; thus liberation at very fine sizes probably eliminates some beneficiation methods. For instance, the flotation technique is hampered by the presence of slimes produced by fine grinding. Thus the flotation method is not very selective in treating this type of low-grade iron-formation. According to M. E. Volin (written commun. 1962).

> Various schemes have been tried to separate the poor from the rich bands after first crushing or coarse grinding, such as heavy media, spirals and other gravity concentration methods. In general, it has been found difficult to reject any considerable amount of the highly siliceous portions without losing too much of the iron.

Of the many beneficiation processes tried on Menominee Group iron-formation, one in particular shows considerable promise. It is called "magnetic-oxide conversion," and the Institute of Mineral Research, Michigan College of Mining and Technology, has been conducting small-scale pilot-plant experiments using this process. In brief, the ore is ground to −20 mesh, after which it is roasted to render all of the iron minerals magnetic. It is then reground to about 96 percent −325 mesh and separated magnetically. Concentrates, by this method, are reported to average 65.18 percent iron and 9.03 percent silica. A 96.75-percent recovery was made from crude heads that averaged 31.56 percent iron (Victor Kral, Ford Motor Co., written commun., 1958).

From the recovery figure reported, the magnetic-oxide conversion method seems eminently suited to the Menominee ores. It will be expensive, compared to the concentration of some of the simpler ores of the region, but this expense may be more than offset by the good location and accessibility of the deposits.

DIRECT REDUCTION

Since a reducing roast seems a necessary part of the processing of the Menominee ores, it is natural to consider direct reduction, which represents but a greater degree of reduction—oxide ore to sponge iron. There are many patented variations of the direct reduction process, some of them in commercial operation, and many of these have been described in a review article by McAneny (1960). In general, the idea is to reduce the oxide ore to iron in a kiln, shaft furnace, retort, or other apparatus, by heat plus reductant gases and (or) reductant solids, such as coke dust; thereby the blast

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furnace is eliminated. Each of the many variations of the general process has advantages and disadvantages. A common disadvantage is the large amount of fuel required. Cheap fuel is a necessity. An advantage is the high quality of the product. Nearly all the gangue is removed from the ore; thus shipping costs are reduced to a minimum. M. E. Volin (written commun., 1962) has informed us that direct reduction experiments on Menominee iron-formation are being conducted at the Institute of Mineral Research.

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