Geology of the Marquette and Sands Quadrangles Marquette County Michigan

GEOLOGICAL SURVEY PROFESSIONAL PAPER 397

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

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By JACOB E. GAIR *and* ROBERT E. THADEN

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[Plates are in pocket]

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By Jacob E. Gair and Robert E. Thaden

ABSTRACT

The Marquette and Sands quadrangles lie astride the east end of the Marquette iron range, Michigan, and extend 6–7 miles north and south of the range. The northeastern part of the Marquette quadrangle extends into Lake Superior.

The area is underlain mainly by deformed and metamorphosed lower and middle Precambrian rocks formerly designated as Archean (basement complex) and Huronian rocks. Numerous upper Precambrian (Keweenawan) diabase dikes cut most of the older Precambrian rocks. In places, gently-dipping Cambrian sandstone lies unconformably on Precambrian rocks.

The rocks of the area give evidence of at least two waves of deformation and metamorphism prior to the beginning of middle Precambrian time and of a major wave of deformation and regional metamorphism at the close of middle Precambrian time. Since the beginning of late Precambrian time there has been little deformation and no regional metamorphism.

The major structural feature of the area is the west-trending Marquette synclinorium of middle Precambrian rocks, bounded on the north and south by structural uplifts of lower Precambrian basement rock. The synclinorium was formed at the close of middle Precambrian time, evidently along a structural axis established during early Precambrian time. The part of the synclinorium in the Marquette and Sands quadrangles contains several second-order folds and a number of faults related to the folding of the synclinorium.

The lower Precambrian rocks in the area comprise the Mona Schist, the Compeau Creek Gneiss, metadiabase, felsic porphyry, and possibly other intrusive rocks. These rocks are exposed north or south of the synclinorium and were formerly called the northern and southern complexes. The Mona Schist is the oldest formation in the area and consists of massive mafic metavolcanic rock, ellipsoidal greenstone, dark slate, layered or pseudolayered amphibole schist, and light slate. The Mona Schist has been intruded by the Compeau Creek Gneiss, and some remnants of the Mona are caught up in the gneiss. A major unconformity separates the Mona Schist and Compeau Creek Gneiss from the metasedimentary middle Precambrian rocks of Animikie age.

The formations in the Animikie Series, from the base upward, are the Enchantment Lake Formation, Mesnard Quartzite, Kona Dolomite, Wewe Slate, Ajibik Quartzite, and Siamo Slate. Within this sequence, the Ajibik Quartzite, which has a small amount of conglomerate at its base in at least one place, rests on the Wewe Slate disconformably or has a very slight angular discordance.

The Enchantment Lake Formation consists of a basal conglomerate, fine-grained graywacke, arkose, sericitic slate, quartz-sericite wacke, sericitic quartzite, and quartzite. The formation is probably lenticular and has a maximum thickness of 500 feet. It is correlated with the Fern Creek Formation of the Menominee Range.

The Mesnard Quartzite consists almost entirely of vitreous quartzite that weathers white to light pink. The Mesnard formerly included the rocks now assigned to the Enchantment Lake Formation. Crossbedding and ripple marks in the quartzite indicate deposition in shallow water. Locally the quartzite is brecciated. Small amounts of hematite occur in thin seams, along bedding, and as cement in breccia. The quartzite is 200–500 feet thick.

The Kona Dolomite consists mainly of massive to thin-bedded dolomite and quartzose dolomite, and of interlaminated dolomite and chert. A basal slate occurs in many places, and possibly everywhere, though its extent cannot be assessed because of poor exposures. White-weathering and red dolomitic quartzite and a few thin beds of a laminated orange and brown ferruginous quartz-feldspar siltite are locally noteworthy units in the formation. Algal structures are widespread and in a few places are associated with oolites. Both features are indicative of shallow-water deposition. The dolomite is silicified in many places. The Kona is 1,000–1,200 feet thick.

Wewe Slate is mostly either gray and thick-bedded (nonlaminated) or green and white and thinly laminated slate. Chlorite granules and some textural and chemical evidence of volcanic material are found in parts of the laminated green and white slate. Parts of the chloritic slate weather to a rusty-colored ferruginous material which in places has been test-pitted for iron. Because of poor exposures, the estimated thickness of 900 feet is highly conjectural.

Ajibik Quartzite consists of white-weathering to light-pink vitreous quartzite, both thick and thin bedded. Thin slate interbeds are seen in a few places. At the base of the formation in the SE¼ sec. 6, T. 47 N., R. 25 W., a bed of conglomerate, 1–2 feet thick, contains fragments of Wewe Slate, chert, and quartzite, and rests on Wewe Slate with little or no angular discordance. This conglomerate is believed to be the result of a mild disturbance and of slight uplift of the Wewe to sea level or a little higher between Wewe and Ajibik times. The Ajibik Quartzite is about 650 feet thick.

The Siamo Slate typically is dark gray and laminated and has thin interbeds of gray to nearly black vitreous quartzite. Parts of the Siamo tend to weather to a rusty-colored ferruginous slate which has been prospected for iron ore in places. Because the top of the Siamo is not exposed in the mapped area and the crumpled rock occurs in the core of the synclinorium, without marker beds, no estimate of thickness is possible.

Several different igneous rocks cut the lower and middle Precambrian rocks, and most of them have been metamorphosed. Metadiabase, quartz veins, and diabase cut both lower and middle Precambrian rocks; metagabbro, metapyroxenite, felsic porphyry, hornblende lamprophyre, syenite, pegmatite, and peridotite intrude only lower Precambrian rocks. Except for a small amount of metadiabase intruded by tonalitic offshoots from the Compeau Creek Gneiss and for felsic porphyry believed to be of the same age as the gneiss, the metamorphosed intrusive

rocks can be dated only as younger than their host rocks and older than the last regional metamorphism, which occurred during the middle to late Precambrian interval. Thus, the metamorphosed intrusive rocks in the Mona Schist and the Compeau Creek Gneiss, except as noted above, may be either of early or middle Precambrian age. Diabase is generally only slightly to moderately altered and therefore postdates the last regional metamorphism. The periodotite has been extensively serpentinized, intrudes only lower Precambrian rock, and has been considered, in previous reports, to be of early Precambrian age. However, the presence of considerable amounts of fresh olivine and pyroxene in the serpentinized peridotite suggests a late- or post-metamorphic age, and the peridotite is interpreted as having been emplaced late in the middle to late Precambrian metamorphic and orogenic interval.

Mineral explorations in the Marquette and Sands quadrangles have been concerned mainly with iron and copper, but the principal products to date have been sand and gravel, quartzite, and greenstone for aggregate, road rock, and breakwaters, and the Cambrian sandstone for structural stone.

Prospecting for iron has involved mainly the sinking of test pits in shear and breccia zones along which iron has been concentrated by ground water during weathering, or by hydrothermal solutions associated with the introduction of vein quartz. Such zones have been prospected in the Mona Schist, Mesnard Quartzite, Kona Dolomite, Wewe Slate, and Siamo Slate. Some iron ore evidently was taken prior to 1880 from an oxidized zone of shears and carbonate veins in chloritic slate of the Mona Schist at the Eureka mine in sec. 21, T. 48 N., R. 25 W. The likelihood of marketable concentrations of iron ore in occurrences such as those prospected in the Marquette and Sands quadrangles is remote.

Several test pits and shallow shafts for copper have been dug in the southeastern and south-central parts of the Marquette quadrangle and the northeastern part of the Sands quadrangle. Most of these pits explore concentrations of sulfides in or near fault or breccia zones, and, in one place, in a quartz vein. These sites are not favorable for large amounts of copper, but the nearby country rock, especially the slates in the lower part of the Kona Dolomite, should be tested for the possibility of low concentrations (but large total amounts) of copper.

INTRODUCTION

LOCATION AND ACCESS

The Marquette and Sands 7½-minute quadrangles are at the east end of the Marquette iron range, east-central Marquette County, in the northern peninsula of Michigan (fig. 1; pls. 1, 2), between 87°22′30″ and 87°30′ W., and between 46°22′30″ and 46°37′30″ N. The Marquette quadrangle lies north of and adjoins the Sands quadrangle along 46°30′ N. This latitude also nearly bisects the west-trending Marquette iron range. The eastern and north-central parts of the Marquette quadrangle extend into Lake Superior.

The city of Marquette, which occupies the central and east-central parts of the Marquette quadrangle, is one of the three major ports on Lake Superior for the shipment of iron ore.

The Duluth, South Shore and Atlantic and the Lake Superior and Ishpeming Railroads cross the Marquette quadrangle and connect the docks at Marquette with the iron mines to the west. The Chicago and North Western Railroad runs southeastward across the western part of the Sands quadrangle and connects Negaunee with cities along the west side of Lake Michigan.

TOPOGRAPHY AND DRAINAGE

The Marquette and Sands quadrangles are representative of the wide variations in topography formed by glaciation around the margins of Lake Superior.

The quadrangles divide naturally into three major topographic units: (1) steep-faced northwest-trending ridges which occupy the northern part of the Marquette quadrangle, (2) west-trending bedrock ridges and hills which coincide with the Marquette trough in the southern part of the Marquette quadrangle and northern part of the Sands quadrangle, and (3) sand plains which occupy much of the Sands quadrangle.

The steep rock ridges in the northern part of the Marquette quadrangle and steep-faced, partly dissected sand terraces in the west-central and south-central parts of the quadrangle trend roughly parallel to the Lake Superior shore; they probably represent the combined influence of joints or faults in the basement and of Pleistocene glaciation on topography and the shoreline. Along the east sides of Hogback and Sugarloaf Mountains in the northern part of the quadrangle, the land rises abruptly about 500 feet; elsewhere the ridges are much lower. The most rugged topography and greatest relief along the west-trending belt in the southern part of the Marquette quadrangle is at the east end, where the land rises rather abruptly from Lake Superior to heights of 400–500 feet above the lake.

West-southwestward along the belt into the Kona Hills, in the northwestern part of the Sands quadrangle, the surface rises gradually to an altitude of 840–960 feet above the lake. This belt gives way south and southeast to extensive flat to gently rolling sand plains. The sand plains occupy two levels, one between 490 and 530 feet, the other between 620 and 650 feet above Lake Superior. The lower plain, within the northeastern quarter of the Sands quadrangle, falls away eastward through an area of steep slopes and narrow steep-sided gullies toward Lake Superior and an adjoining low-lying sand plain. The higher of the two plains in the Sands quadrangle is separated from the lower in the central part of the quadrangle by a narrow north-trending, east-facing sand escarpment which owes most or all of its irregularities of form directly to glacial deposition. This escarpment turns and extends eastward from Strawberry Lake nearly to the east edge of the quadrangle. From there southward to near the southeast corner of the quadrangle, the higher sand

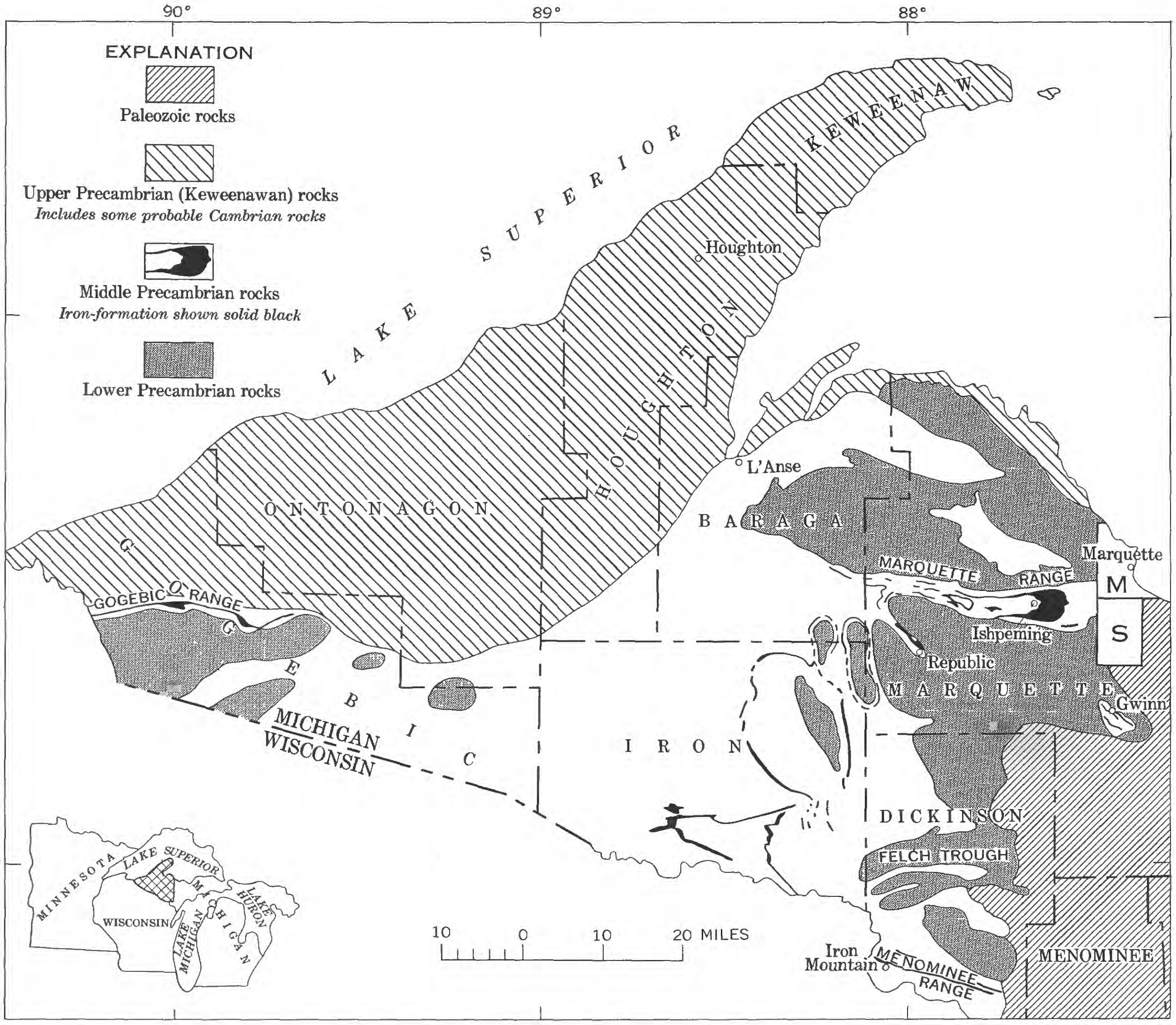

FIGURE 1.—Geologic sketch map of western part of northern peninsula of Michigan, showing location of Marquette (M) and Sands (S) quadrangles.

plain drops away eastward through a broad, very rough surfaced area which is a product both of glacial deposition and postglacial erosion. Scattered knobs and ridges of bedrock, or of bedrock partly covered by sand, stand above the plain in the southwestern part of the Sands quadrangle.

Although swamps are not as extensive in the area as in many nearby parts of northern Michigan, there are, at many different elevations, numerous small swamps, many of which are not shown on the topographic base map. They generally occupy low areas between bordering hills or small bare rock knobs.

The principal drainage of the Marquette quadrangle is provided by the Dead and Carp Rivers and their tributaries. These rivers head many miles to the west and trend irregularly eastward across the quadrangle to Lake Superior. The Carp River also flows for about 2 miles through the northern part of the Sands quadrangle.

Drainage in the eastern part of the Sands quadrangle is also into Lake Superior, through a number of creeks tributary to the Chocolay River. The southwestern part of the Sands quadrangle and the area south of the Kona Hills are drained into Lake Michigan by the East branch of the Escanaba River and its tributaries. The divide between drainage into Lake Superior and into Lake Michigan crosses the sand plain approximately along the path of the Chicago and North Western Railroad.

From the west edge of the Marquette quadrangle to Lake Superior, the Dead River falls about 400 feet and the Carp River falls about 670 feet. Several hydroelectric plants have been built along these rivers and supply part of the electric power needs of the area.

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PRESENT INVESTIGATION

The mapping of the Marquette and Sands quadrangles is the first step in a cooperative program of the U.S. Geological Survey and the Geological Survey Division of the Michigan Department of Conservation to remap the Marquette iron district on modern topographic base maps for publication at a scale of 1:24,000. The mapping carried out during the present program extends some 5 miles north and 3 miles south of the belt along the range mapped by Van Hise and Bayley (1897).

Approximately 25 man-months of fieldwork for the present geological study of the Marquette and Sands quadrangles were done during the four summers, 1957–1960. R. E. Thaden helped with the work during the summers of 1959 and 1960, and B. F. Jones during the summer of 1957.

Geologic mapping was done directly on topographic sheets enlarged to 1:12,000 from the standard 7½-minute (1:24,000) topographic maps of the Marquette and Sands quadrangles published by the U.S. Geological Survey. All known outcrops were plotted on the maps. Many outcrops as shown are in reality clusters of smaller outcrops too close together to be separated at the scale of the mapping. Isolated small outcrops and persistent thin dikes as little as 1 foot thick have been exaggerated in size sufficiently to be shown on the map.

Outcrops were located by pace-and-compass traverses from known points and, wherever possible, by direct reference to topographic features. Traverses in most of the area were 300–500 feet apart. Lighthouse Point in the Marquette quadrangle was mapped by planetable methods at a scale of 1:600. A nearly completely exposed section in a large roadcut and adjacent quarry about 1 mile southeast of the State prison in the Marquette quadrangle was mapped by tape and compass at a scale of 1:1,200.

Many specific locations given in the report are measured in feet north and east of the southwest corner, or north and west of the southeast corner, of the indicated section. Some locations are given relative to the quarter corner of a section—a point on the section line midway between the section corners.

Rock descriptions in this report are based both on field observations and on the microscopic examination of about 1,200 thin sections. Feldspar determinations are based on measurements of the extinction angle to (010) in sections perpendicular to X or Z, and by determinations of optic sign in conjunction with a comparison of indices of refraction with the index of refraction of Canada balsam. Amphiboles were identified by determinations of the extinction angle, $Z \wedge c$, and by measuring indices of refraction in standard immersion oils. Some of these measurements were facilitated by the use of a spindle stage devised by R. E. Wilcox of the U.S. Geological Survey (1959). Pyroxenes were identified by the measurement of indices of refraction in immersion oils and by the measurement of the optic angle on a universal stage.

Volume percentages of minerals generally were estimated visually, but in selected thin sections having proportions of minerals close to critical values for naming the rock, volume percentages were measured with a mechanical stage and point counter.

Twenty-one complete chemical analyses have been made in the laboratories of the U.S. Geological Survey.

ACKNOWLEDGMENTS

Blair F. Jones of the U.S. Geological Survey mapped a part of the Animikie belt in the southern part of the Marquette quadrangle in the summer of 1957 and collaborated in making a tape-and-compass traverse in the southeastern part of the Marquette quadrangle. The late Justin Zinn, professor of geology at Michigan State University, called our attention to several interesting and significant localities in the area and to several references in the lengthy literature of the Marquette range that otherwise might have been overlooked. He was also helpful in discussing problems of correlation and stratigraphic terminology.

Robert Reed of the Michigan Department of Conservation directed our attention to several test pits and other interesting localities that might otherwise have been overlooked. Harold Stonehouse, professor of geology at Michigan State University, pointed out an outcrop of diabase in the SE¼ sec. 6, T. 47N., R. 25 W. John Olson of Marquette provided transportation by motorboat to the islands in Lake Superior, within the Marquette quadrangle. Authorities at the Marquette branch of the Michigan State prison permitted passage through the prison grounds into parts of the area. Louis Vierling of Marquette showed us the location of the old Eureka iron mine, which now is little more than a small undercut reentrant in a streambank.

All rock analyses were performed by U.S. Geological Survey personnel, unless otherwise indicated.

PREVIOUS GEOLOGIC WORK

The Marquette iron range has been the subject of many previous geologic reports. The most comprehensive account to date is that of Van Hise and Bayley (1897), in the first part of which (p. 8–148) is an annotated bibliography containing 108 entries representing all prior published work on the Marquette area. Most of these earlier works describe observations at the east end as well as elsewhere along the range, and many give conclusions about the structure and stratigraphic sequence in the vicinity of Marquette. The most notable of these reports were by Houghton (1841), Foster and Whitney (1851), Kimball (1865), Brooks (1873, 1876), Wadsworth (1880), Rominger (1881, 1895), Williams (1891), and Van Hise (1891, 1892, 1893).

The fundamental concepts of the geology of the Marquette iron range were evolved during the last half of the 19th century and have been changed relatively little since the summary work of Van Hise and Bayley (1897). The major stratigraphic and structural features of the area as described by them were (1) a bipartite sequence of metasedimentary rocks of "Algonkian" age, which they called the "Marquette series" (following Van Hise, 1891) and which rests unconformably on a "basement complex" of "Archean" age, and (2) a west-plunging syclinorium along the axis of the Marquette range occupied by the folded Marquette series and bounded north and south by the older basement rocks.

Basement rocks north and south of the range were called, respectively, the northern complex and the southern complex. In the northern complex, greenstone and associated silicic metavolcanic rock, the oldest rocks in the area, were intruded by gneissoid granitic rock, and both groups of rock were cut by mafic dikes of different ages. The greenstone and silicic metavolcanic rock of the northern complex in the vicinity of Marquette were called the Mona Schist by Van Hise and Bayley. The granitic rocks were not named. They gave only cursory attention to the gneissoid granite and greenstone schist of the southern complex within the area of the present Marquette and Sands quandrangles.

Van Hise and Bayley (1897) followed Van Hise (1891) in dividing the metasedimentary rocks of the range into a lower Marquette Series and an unconformably overlying upper Marquette Series. Correlation of the metasedimentary rocks of the Marquette area with those in the type Huronian area of Canada, north of Lake Huron, had been suggested earlier (Hunt, 1861; Kimball, 1865; Brooks, 1873; Irving, 1883, 1885; Rominger, 1881; Van Hise, 1891, 1892, p. 184–186) and the name "Huronian" was widely used for these metasedimentary rocks prior to 1897. Van Hise and Bayley (1897), however, did not make formal use of "Huronian" for these rocks. The formations of the Lower Marquette Series, as named by them, were the Mesnard Quartzite, the Kona Dolomite, the Wewe Slate, the Ajibik Quartzite, the Siamo Slate, and the Negaunee Formation (table 1, in pocket). Their Upper Marquette Series consisted of the Ishpeming Formation (Goodrich Quartzite and Bijiki Schist), the Michigamme Formation, and the Clarksburg Formation. Only rocks of the lower series below the Negaunee Iron-Formation are present in the Marquette and Sands quadrangles. The Negaunee Iron-Formation and rocks of the upper series cross the axis of the west-plunging synclinorium west of the Marquette and Sands quadrangles.

Van Hise and Leith (1911, p. 251–283) summarized the work of the U.S. Geological Survey in the Marquette district, in general following Van Hise and Bayley (1897), but they formally applied the name "Huronian" to rocks of the Marquette Series. However, as a result of the discovery of an unconformity at the base of the Ajibik Quartzite by A. E. Seaman subsequent to 1897 (see Van Hise and Leith, 1911, pl. 19), and in accordance with the new subdivision of the Marquette Series and the Huronian proposed by a Canadian-American committee (Van Hise and others, 1905), the original lower Huronian of the Marquette area (Lower Marquette Series of former usage) was divided into two parts. The part beneath the Ajibik Quartzite was designated lower Huronian; the part from the base of the Ajibik to the top of the Negaunee formation was assigned to the middle Huronian (table 1). Since 1911, the validity of the original correlation of the metasedimentary rocks of the Marquette and other iron-mining districts of Michigan with the type Huronian of Canada has been disputed by a number of geologists, most recently by James (1958).

Leith, Lund, and Leith (1935) reviewed geologic features discovered after 1911 and revised the geologic map of the Lake Superior district of Van Hise and Leith (1911). Changes recognized by Leith, Lund, and Leith in the Marquette district (1935, p. 20) apply to areas west of the Marquette and Sands quadrangles (see table 1), and are not further considered here.

Between 1911 and 1966 the operating iron companies in the district have conducted intensive drilling programs and done detailed mapping, principally in the main producing areas around Negaunee and Ishpeming and westward along the range, and near Palmer, some 4 miles south of Negaunee. Few of the results of these studies have been published, and the limitation of such work to or near areas of iron formation has precluded such work from producing significant changes in geologic interpretation within the area of the Marquette and Sands quadrangles.

The work of Tyler and Twenhofel (1952), done under the auspices of the Jones and Laughlin Steel Corp., is undoubtedly the most significant of such studies; they advanced several interpretations radically different from those of Van Hise and his associates. Tyler and Twenhofel suggested that the lithologic units crossed time lines along a shoreline that transgressed generally from east to west or northeast to southwest, quartzite representing nearshore facies or stillstands of sea level. As a result, the Ajibik Quartzite, in the area east of Negaunee where it overlies the Mesnard, Kona, and Wewe sequence, may be a regressive facies of the Mesnard; west of Negaunee, the Ajibik resting directly on basement rocks may be merely a transgressive facies of the Mesnard. The major conclusion of Tyler and Twenhofel as affecting interpretation in the Marquette and Sands quadrangles is that there is no unconformity at the base of the Ajibik and hence no valid basis for subdivision of the original lower Huronian of the Marquette area into lower and middle Huronian units. Major map changes introduced by Tyler and Twenhofel did not extend into the area of the Marquette and Sands quadrangles or affect prior mapping there.

Topical studies of Precambrian rocks, some of which included the mapping of small areas, have been made in the Marquette and Sands quadrangles by Ayres and Higgins (1939), Brooke (1951), Buzas (1960), Creveling (1926), Jackson (1950), Krimmel (1941), Porturas (1945), Sahakian (1959), and Vehrs (1959). Of these workers, however, only Brooke has introduced any change in previous interpretations of structure or stratigraphy. He deduced an unconformity north of Enchantment Lake (formerly Mud Lake) in the Marquette quadrangle, a little above the base of the Mesnard of Van Hise and Bayley (1897), and concluded that the rocks between that unconformity and the underlying Archean greenstone constituted a subHuronian nonmarine series, possibly of Timiskaming age.

As a result of work by the U.S. Geological Survey since 1943, mainly in Iron and Dickinson Counties, Mich. (James, 1958, p. 27–30), the Precambrian rocks of the area are now assigned to the following informal subdivisions: lower Precambrian (formerly Archean), middle Precambrian (formerly Huronian), and upper Precambrian (from top of "Huronian" to top of Keweenawan rocks) (table 1). The correlation of middle Precambrian rocks with the type Huronian is considered invalid (James, 1958, p. 33–35), and such rocks in northern Michigan are now correlated with the Animikie Group of northern Minnesota and adjacent Ontario. Inasmuch as the base of the Animikie in the type area probably corresponds to the horizon at the base of the Ajibik Quartzite in the Marquette range, the assignment to the Animikie of the Mesnard, Kona, and Wewe rocks and their correlatives in Iron and Dickinson Counties required a downward expansion of the Animikie series in respect to the Animikie of the type area. The Chocolay and Menominee Groups (James, 1958, p. 35) correspond to the lower and middle Huronian of Van Hise and Leith (1911).

GEOLOGIC SETTING

The Marquette and Sands quadrangles contain parts of two major structure-time units, the Marquette synclinorium, consisting mainly of rocks of middle Precambrian age, and areas of older rocks to the north and south. The older rocks formed during early Precambrian time and are of two broad age groups and predominantly of two lithologies: (1) older mafic metavolcanic rocks, the Mona Schist, and (2) younger gneisic rocks, tonalite, granodiorite, and amphibolite, named the Compeau Creek Gneiss in this report. The volcanic rocks appear to have had a schistose structure prior to being intruded by the tonalite and granodiorite. Erosion which produced coarse conglomerate at the base of the middle Precambrian section must have resulted from substantial uplift (deformation?) at the end of early Precambrian time. Thus these rocks were probably deformed at least twice prior to middle Precambrian time. Metamorphism accompanying the intrusion of tonalite and granodiorite reached at least the level of the amphibolite facies in parts of the area, but the degree of metamorphism, if any, that accompanied the uplift at the close of early Precambrian time is unknown.

The rocks of middle Precambrian age in the Marquette synclinorium are mainly of metasedimentary origin. The original sediments were sand, mud (probably mainly detrital), mixtures of sand and detrital mud, shallow-water carbonate precipitates with intermixed and interbedded detritus, and interlaminated ferruginous and siliceous precipitates with some interlayered sand. The sequence of formations is shown in table 1. West of the area of the present report, a thick lens of mafic volcanic rock was deposited within the upper part of the middle Precambrian sequence, and thick mafic sills were intruded into the Negaunee Iron-Formation sometime before the end of middle Precambrian time. Relatively slight deformation and limited erosion occurred twice during middle Precambrian time, in the intervals between deposition of the Wewe Slate and Ajibik Quartzite and between deposition of the Negaunee Iron-Formation and the Goodrich Quartzite. The middle Precambrian was brought to a close by strong orogeny during which the Marquette synclinorium was formed. Metamorphism approximately concurrent with the orogeny increased in intensity westward from the Marquette-Sands area where it reached only the level of the lower greenschist facies.

After or possibly starting shortly before the end of the orogeny and metamorphism, some peridotite and perhaps some pegmatite of late middle or early late Precambrian age were emplaced. Nonmetamorphosed diabase dikes of late Precambrian age postdate the last orogeny and metamorphism in the area.

During Early or Middle Cambrian time, sandstone of probable continental origin was deposited widely in lower parts of the area, especially near the present shoreline of Lake Superior. Since Cambrian time the area has been gently uplifted, without orogeny.

LOWER PRECAMBRIAN METAVOLCANIC AND GNEISSIC ROCKS

GENERAL FEATURES

The lower Precambrian rocks are the oldest in the area and consist principally of older mafic metavolcanic rocks and younger felsic and amphibolitic gneisses. The bulk of the gneisses were formed by the intrusion of tonalite and granodiorite into country rock consisting at least in part if not entirely of the older metavolcanic rocks.

The metavolcanic rock, the Mona Schist, consists predominantly of ellipsoidal and other massive metabasalts and of layered amphibole schist. Chloritic, actinolitic, and quartz-sericite schists and slates and mafic meta agglomerate compose lesser parts of the formation.

The gneiss is named Compeau Creek Gneiss in the present report and consists of abundant light-colored foliated "granitic" rock—tonalite and granodiorite—and of less abundant amphibolite and dark varieties of tonalite and granodiorite, rich in hornblende, biotite, or chlorite.

The lower part of the Mona Schist is metamorphosed to the level of the lower greenschist facies (chlorite zone). The gneiss, on the other hand, reached the metamorphic level of the amphibolite facies. The upper part of the Mona Schist, between the lower part of the Mona and the gneiss, is at an intermediate metamorphic level between the greenschist and amphibolite facies; it is therefore thought to show thermal metamorphic effects resulting from the intrusion of the gneiss into the Mona Schist.

MONA SCHIST

DISTRIBUTION AND THICKNESS

The Mona Schist was named by Van Hise and Bayley (1895, p. 490–496) for extensive exposures in the Mona Hills in the central and south parts of secs. 27 and 28, T. 48 N., R. 25 W. These hills are not named on the present-day topographic map. The principal occurrence is in a westward-trending belt, some 2½–3½ miles wide, underlying the north limb of the Marquette synclinorium and extending across the width of the Marquette quadrangle. Other occurrences in the Marquette quadrangle are (1) at the core of the anticline in secs. 2 and 3, T. 47 N., R. 25 W., and secs. 35 and 36, T. 48 N., R. 25 W., (2) underlying the south limb of the eastward-plunging syncline in sec. 1, T. 47 N., R. 25 W., and sec. 6, T. 47 N., R. 24 W., and (3) outliers in secs. 1 and 12, T. 48 N., R. 26 W., north of the main belt of the Mona. These outliers seem to be surrounded by granitic rocks that intruded the Mona, but future work in the Negaunee quadrangle to the west may disclose continuity with the principal belt of Mona Schist.

Only a minimum thickness can be determined for the Mona. The true base and top of the formation are unknown in the area, inasmuch as the lowermost known part of the formation passes unconformably beneath Animikie metasedimentary rocks and the upper contact is against intrusive tonalite and related rocks. Widespread ellipsoidal structures in the lower part of the Mona indicate tops of flow layers persistently northward and thus rule out significant widening of that part of the Mona in the principal belt by folding. Because of the absence of marker beds in the Mona, the possibility of thickening of the section by imbricate faulting cannot be evaluated. Dips of foliation, and of layering in the hornblende schists where they can be measured on beds of ellipsoidal greenstone, are rather consistently between 70° N. and vertical. If the average dip is taken to be 75° N. and an assumption is made that there is no repetition by faulting, the minimum thickness of the Mona ranges from 13,000 to 21,000 feet.

DEFINITION AND SUBDIVISIONS

The most detailed study of the Mona Schist to date was by G. H. Williams (1891, p. 134–170). Williams distinguished between west-trending "banded greenstones" in the northern part of the main greenstone area near Marquette and massive aphanitic greenstone in the southern part of the area. The superposition of these rocks was not determined by him, for although he described spheroidal (pillow or ellipsoidal) structures in the aphanitic greenstones, the usefulness of such features in determining top direction was not known at that time. West-trending ferruginous slates lying between the "banded" and the massive aphanitic greenstones were named the Eureka series by Williams and were thought by him to be detrital rocks of Huronian age, in a tight syncline.

Williams' massive aphanitic greenstones correspond to approximately the lower half of the Mona Schist, referred to as the lower member in the present report. The thickness of the lower member ranges from 8,200 to 11,800 feet. Ellipsoidal structures are widely distributed in this part of the Mona and indicate submarine deposition for at least some of the greenstones. The lower part of the Mona has chloritic slate or schist and siliceous metavolcanic rock (felsite) interlayered with the massive greenstones.

The upper part of the Mona, characteristically a more or less distinctly layered amphibole schist, is named the Lighthouse Point Member in this report after Lighthouse Point in the easternmost part of the city of Marquette, where the best exposures and type locality of the member are found. The Lighthouse Point Member ranges in thickness from approximately 4,500 to 11,600 feet. The maximum thickness of the member near the west edge of the quadrangle, however, is highly questionable, because of the probability that the amphibole schist in the outlier north of the principal belt was brought to its present position by folding prior to or during engulfment by the granitic rock (W. P. Puffett, oral commun., 1966). The Lighthouse Point Member corresponds virtually to Williams' banded greenstones. Some massive greenstone and felsitic rock and very small amounts of mafic meta-agglomerate are intercalated with the layered amphibole schist in the Lighthouse Point Member; the lower part of the number consists mainly of chloritic slate. The rocks of Williams' Eureka Series are placed in the lower part of the Lighthouse Point Member in this report. They evidently are transitional between the massive metabasalt below and the layered metavolcanic rock above.

All or virtually all the Mona is of volcanic origin, and the bulk of it is mafic. Irregular and tabular intrusive bodies are common in the Mona, and most of these also are mafic.

The rocks of the Mona Schist are discussed in the following paragraphs in their estimated order of abundance.

METABASALT

Metabasalt composing the bulk of the lower member of the Mona is distributed in a belt about 2 miles wide, which immediately underlies the north limb of the synclinorium and extends across the width of the Marquette quadrangle. Ellipsoidal structures are common in this belt. Most of the greenstone in the core of the anticline near the southeast corner of the Marquette quadrangle is also metabasalt and is assigned to the lower member. Massive aphanitic metabasalt(?), some having pillow or probable pillow structures, in places is interbedded with amphibole schist of the Lighthouse Point Member. Tabular bodies of metabasalt(?) from 1 to 20 feet thick are common from the vicinity of Lighthouse Point westward into the central part of the Marquette quadrangle. In the western part of the quadrangle, metabasalt(?) forms an irregularly wedging or folded mass, 2,000 feet or more thick, between felsitic and chloritic rocks in the lower part of the Lighthouse Point Member and the layered amphibole schist in the upper part of the member. The rocks of this thick mass are best exposed in the S½ sec. 7, and the N½ sec. 18, T. 48 N., R. 25 W., and in the SE¼ sec. 12, T. 48 N., R. 26 W. Some 200–300 feet of massive porphyritic metabasalt forms a wedge in the amphibole schist of the Lighthouse Point Member along the Dead River, about 1,300 feet north of the S¼ cor. sec. 7, T. 48 N., R. 25 W.

Both massive and schistose ellipsoidal metabasalt are common in the lower member of the Mona Schist and resemble in detail those of much younger age in Iron County, Mich., described by Clements and Smyth (1899, p. 112–124), Gair and Wier (1956, p. 43), and Bayley (1959, p. 28–29). The pillows are more or less flattened bun forms and range in width from less than 1 foot to 11 feet; most are 1–3 feet wide. Elongation of the pillows on outcrop surfaces is generally eastward or a little south of eastward, parallel to the strike of the formation. The tops of the pillows invariably face northward.

Thin amygdaloidal zones in metabasalt and concentrations of amygdules in the outer parts of pillows have been found in a few places in the area. Amygdules are filled variously with chlorite, zoisite, and sphene, with calcite or calcite and chlorite, with chlorite and epidote, with sphene, rutile(?), and fibrous colorless amphibole, or with tremolite, epidote-clinozoisite, and leucoxene. Much of the titanium in some of these rocks is in the amygdules.

The mineralogy and textures of the metabasalt in the lower member of the Mona, although markedly variable in detail, are typical of basaltic rocks that have been thoroughly metamorphosed in the chlorite zone or the greenschist facies (Williams, 1891; James, 1955, p. 1468–1470; Gair and Wier, 1956, p. 43–44; Bayley, 1959, p. 31–32).

The metabasalt is dull shades of green, gray green, and brownish green. The principal minerals are chlorite or pale-green amphibole, sodic plagioclase, sericite, and epidote-clinozoisite, although in any particular sample any one of these minerals may be absent or present only in small amounts. Quartz, opaque minerals, leucoxene-sphene, and carbonate are minor minerals in most samples, and small amounts of biotite, rutile, or reddish-brown iron oxides occur in some of the metabasalts.

The amphibole of the metabasalt in the lower member of the Mona is tremolite-actinolite; most of it probably was derived from pyroxene. It is slightly to moderately pleochroic (X, colorless or pale straw yellow, to Z, pale green or brownish green), has maximum extinction, $Z \wedge c$, ranging from 19° to 22° in different samples, is biaxially negative with large optic angle, and has indices of refraction ranging from approximately 1.62

to 1.64 for N_X, 1.63 to 1.65 for N_Y, and 1.65 to 1.66 for N_Z. Grains vary from well-formed prisms as much as 0.75 mm long to shredlike fragments less than 0.05 mm. Fibrous uralite is common.

Chlorite from several samples of metabasalt in the lower member has brown or purple interference colors, has N_Y ranging from 1.1614 to 1.626, is biaxially positive, and has a small optic angle. It is probably the variety ripidolite (Winchell and Winchell, 1951, p. 381–383). Chlorite from one sample is bright green, has a purple interference color, has N_Y of 1.626, is optically negative, and is probably the variety diabantite, which carries somewhat more of the ferroantigorite molecule than does ripidolite.

Most of the metabasalt that is rich in actinolitic amphibole has only small amounts of chlorite. In the finer grained metabasalt having pronounced foliation structure, the primary fabric, although preserved in small patches, has been considerably disrupted by shearing, and chlorite generally is the dominant mafic mineral. The reasons for the preferential development of amphibole or chlorite are not well understood. Such preferential development may have been related to the availability of water or to differences in metamorphic equilibruim in more massive and less massive (sheared) volcanic rocks, as suggested by Wiseman (1934, p. 373–374).

Feldspar in the metabasalt is sodic plagioclase. Twinned and untwinned laths—generally less than 1.0 mm in length and pseudomorphous after primary calcic plagioclase—and groundmass aggregates of untwinned grains are characteristic. Variations in anorthite content (approximately An_5 to An_{20}) appear to have random geographic distribution and possibly represent slightly different equilibruim levels from place to place during the metamorphism of the original more calcic plagioclase. Aggregates of sericite or epidote, or mixtures of both, commonly containing some chlorite or carbonate, replace some or parts of some laths of primary plagioclase. In a few places, sphene is also one of the replacing minerals. Where there has been only partial replacement by these minerals, the existing plagioclase is sodic. The similarity of chemical analyses and normative compositions (CIPW) for typical basalt and for the metabasalt (tables 2 and 3) tends to rule out the possibility that either silica or sodium was introduced during metamorphism.

The small amount of quartz in the metabasalt may be either a result of the breakdown of calcic plagioclase or of the recrystallization of primary quartz. Derivation of quartz from calcic plagioclase is indicated by the association of wormy quartz with sodic plagioclase, sericite, chlorite, and epidote in some pseudomorphs of calcic plagioclase.

Biotite(?) or stilpnomelane(?) has been found in relatively few places in the metabasalt of the lower member of the Mona, in small amounts, as small flakes or plates in the groundmass, replacing plagioclase or amphibole, or intergrown along cleavage in chlorite. No attempt was made to identify this material conclusively. Stilpnomelane would be representative of the lower greenschist facies of metamorphism—the general metamorphic level in the area—but if the mineral is biotite

TABLE 2.—*Chemical analyses of mafic metavolcanic rocks and basalts*

[Nos. 1 and 3, standard rock analysis, Dorothy F. Powers. Nos. 2, 4, and 5, rapid rock analysis, Paul Elmore, I. H. Barlow, S. D. Botts, and G. W. Chloe]

	1	2	3	4	5	6	7	8	9	10	11
SiO_2	48.93	48.5	51.54	68.2	53.6	48.59	44.69	49.86	48.52	49.98	47.56
Al_2O_3	14.41	15.8	17.56	14.9	16.9	13.85	16.35	15.07	14.36	13.74	15.00
Fe_2O_3	4.05	3.6	1.36	.9	1.4	4.64	6.03	2.39	1.40	2.37	3.81
FeO	8.46	7.2	9.98	1.4	5.5	7.37	6.58	9.63	11.34	11.60	7.23
MgO	6.01	6.6	4.57	1.4	4.8	7.62	7.25	6.34	7.33	4.73	6.26
CaO	10.63	10.6	7.26	2.3	3.7	6.96	9.11	9.64	10.21	8.21	8.48
Na_2O	2.46	2.5	2.71	4.4	3.9	2.26	2.63	1.76	2.05	2.92	3.28
K_2O	.33	1.2	.32	3.9	3.5	2.62	.76	.43	.24	1.29	.91
H_2O+	2.78	2.3	2.75	1.3	3.4	2.53	3.30	3.18	2.88	1.22	3.02
H_2O-	.10		.14			1.14	.76	.34	.21		
CO_2	.30	.18	.49	1.0	1.8	.12	.04	Tr.	.00	------	1.69
TiO_2	1.04	.74	.89	.3	.97	2.02	2.18	.53	1.28	2.87	1.72
P_2O_5	.08	.06	.07	.15	.52	.23	.26	.20	.09	.78	.38
S	.10	------	.02	------	------	.02	.05	.15	.06	------	------
MnO	.21	.22	.26	.08	.14	.17	.17	.37	.22	.24	.22
CuO	------	------	------	------	------	.028	.060	------	------	------	------

1. Ellipsoidal metabasalt, JG–26A–58, 1,400N–4,200E, SW. cor. sec. 28, T. 48 N., R. 25 W.
2. Fine-grained massive metabasalt, JG–30–59, 4,700N–2,400E, SW. cor. sec. 18, T. 48 N., R. 25 W.
3. Layered amphibole schist, JG–103–58, Lighthouse Point, 4,100N–1,900E, SW. cor. sec. 24, T. 48 N., R. 25 W.
4. Feldspathized greenstone, JG–47–59, 3,580N–4,900E, SW. cor. sec. 7, T. 48 N., R. 25 W.
5. Feldspathized greenstone, JG–63–59, 1,800N–500E, SW. cor. sec. 6, T. 48 N., R. 25 W.
6. Chilled basalt from Keweenawan "Greenstone flow" (Cornwall, 1951, p. 185).
7. Chilled basalt from Keweenawan "Big Trap" (Cornwall, 1951, p. 188).
8. Composite metabasalt, Labrador (Baragar, 1960, p. 1634).
9. Howse Group basalt, Labrador (Baragar, 1960, p. 1634).
10. Average of 6 analyses of Oregon basalts (Washington, 1922, p. 797).
11. Average of 135 analyses of metabasalts (Fairbairn, 1934, p. 94).

TABLE 3.—*Normative compositions (CIPW) of mafic metavolcanic rocks and basalts*

[Nos. 1-3, mafic metavolcanic rocks, derived from analyses 1-3, table 2; nos. 4-7, basalts, derived from analyses 6, 7, 9, and 10 respectively, table 2]

Mineral	1	2	3	4	5	6	7
Quartz	3.42	-----	5.28	2.60	-----	-----	2.22
Orthoclase	1.67	7.23	1.67	15.57	4.45	1.11	7.78
Albite	20.96	20.96	23.06	18.86	22.53	17.29	24.63
Anorthite	27.24	28.36	34.75	19.74	30.30	31.41	20.29
Diopside	18.43	18.91	.46	11.90	10.56	15.16	12.28
Hypersthene	16.93	9.23	27.40	19.85	9.28	26.24	20.58
Olivine	-----	7.47	-----	-----	5.36	1.66	-----
Magnetite	5.80	5.34	2.09	6.73	8.82	2.09	3.48
Ilmenite	1.98	1.37	1.67	3.80	4.10	2.43	5.47
Apatite	.34	.34	.34	.34	.67	.34	2.02
Pyrite	-----	-----	-----	-----	-----	.24	-----

it probably represents local attainment of the upper greenschist facies (biotite zone).

Original basaltic textures (intergranular, intersertal, subophitic, with porphyritic varieties) have been extensively obliterated by shearing and recrystallization. The resulting secondary fabric varies from chaotic intergrowths to strongly alined metamorphic minerals. Ragged, shredded, and wormy grain boundaries are typical. In the more schistose metabasalt, platy minerals in particular, and to a lesser extent, tremolite-actinolite, tend to be well alined. Seams of granoblastic quartz follow shears in some of this metabasalt. Where pseudomorphs of original plagioclase remain, they tend to form small augen.

The porphyritic metabasalt in the Lighthouse Point Member in the SW¼ sec. 7, T. 48 N., R. 25 W., is characterized by a fine-grained schistose groundmass of alined deep-grass-green hornblende, granular sodic plagioclase, and epidote in which are set tabular alined aggregates of epidote, sericite, and chlorite, with or without some granular plagioclase. Primary textures have been completely obliterated in the groundmass, but the tabular aggregates, which are pseudomorphs of original prismatic plagioclase phenocrysts, averaging ½–1 inch in size, show the original porphyritic character of the rock.

Primary minerals, with the possible exceptions of rutile and opaque minerals, thus have been completely reconstituted in the metabasalt of the Mona. Except in the more highly metamorphosed basalt of the Lighthouse Point Member, relic primary basaltic textures remain in areas of pseudomorphous replacement and in relatively undisturbed "islands" between shears. Together with bulk chemical analyses (table 2) and existing saussuritic mineral assemblages typical of metamorphosed basalt or gabbro, the relic textures indicate the basaltic character of the premetamorphic rock. From this deduction, mainly, the essential primary minerals are in turn inferred to have been pyroxene and calcic plagioclase.

AMPHIBOLE SCHIST

DESCRIPTION

Amphibole schist composes most of the upper part of the Lighthouse Point Member. It is best exposed on Lighthouse Point in Marquette, but other excellent large exposures occur (1) along the stream that follows approximately the east section line of sec. 16, T. 48 N., R. 25 W., (the so-called "Brook Section" that was well described by Rominger, 1881, p. 25 and by Williams, 1891, p. 143–145), (2) east and west from the same stream in the S½ secs. 15 and 16, and (3) along the Dead River in the south-central part of sec. 7, T. 48 N., R. 25 W.

The schist is chemically similar to basalt (table 2, col. 3). Typically, it is dark green or gray green and weathers to deep brown, brownish green, and pale gray green or pale olive green. In many exposures this rock has a strong schistosity parallel to which are pronounced surfaces of parting that separate the rock into layers (figs. 2 and 3). Color variations that conform to the layers on weathered surfaces and thus accentuate the layered appearance represent mineralogical variations. Within in-

FIGURE 2.—Layered amphibole schist, Lighthouse Point, Marquette, Mich. Flat lensoid layers strike west and dip steeply north. Length of scale 12 inches.

FIGURE 3.—Layered amphibole schist, west end of Lighthouse Point, Marquette, Mich. Shows lamination within layers, lensing of layers, and boudinage structure. Length of scale 12 inches.

dividual layers, faint light-colored thin streaks and seams form laminae parallel to schistosity and layers, and thin light-colored veinlets cut across schistosity and layers.

The layers appear to result from two conditions. In one, there is little or no color difference and apparently no gross mineral variation from layer to layer. The surfaces between layers evidently represent either original surfaces of sedimentation or of relatively strong shearing or combinations of bedding and shearing surfaces. The other condition resulting in the layers is a variation in the proportions of hornblende and groundmass minerals from layer to layer as well as from lamina to lamina within the layers. In a few places, also, grass-green and blue-green hornblendes alternate from layer to layer. These mineral variations are generally accompanied by textural variations, particularly between alined hornblende and granoblastic mosaics of plagioclase and quartz.

Individual layers consist generally of flat lenses, 1–3 inches thick and 10–30 feet long. These tabular lenses commonly are interleaved near their margins in such a way as to maintain a continuing layer of uniform thickness from lens to lens. The layers of amphibole schist are strikingly uniform in appearance and produce a characteristic striped or banded pattern on weathered outcrop surfaces (fig. 2). Blocks typically break away from the bedrock with slabby forms. The trend of the layers is rather uniformly west to slightly north of west across most of the area, and dips are vertical to steeply north.

In places the amphibole schist is unlayered or is in beds or lenses several feet thick, without internal lamination.

Hornblende forms about 50–75 percent of most of the schists. Much of it is in prisms or lensoid aggregates of parallel prisms, 0.2–1.0 mm long that extinguish uniformly under crossed nicols. The grains, individually and within seams or lenticular aggregates, are strongly alined parallel to schistosity and layering. This alinement can be seen parallel to traces of foliation surfaces in the field but is not commonly seen on the foliation surfaces themselves; hence lineation caused by alined hornblende needles, which is evident from uniform extinction in thin section, is not readily measured in many places in the field.

The hornblende generally is moderately to strongly pleochroic, having the maximum absorption parallel to Z, ranging from dull grass green to blue green. Pale-green hornblende is common in some layers, but such layers make up only a small part of the schist. Within the belt of amphibole schist, there appears to be no systematic distribution of grass-green hornblende as compared with blue-green hornblende. These types in places alternate from layer to layer and may occur within the limits of a thin section, or rarely in single crystals zoned from blue-green cores to grass-green rims. They probably represent small compositional differences. The blue-green hornblende in one sample has $N_X=1.656$, $N_Y=1.669$, $N_Z=1.675$, $2V_X=80°$, and $Z \wedge c=16°$.

In addition to hornblende, the schist generally contains some 15–35 percent of one of the following minerals or combinations of them: plagioclase (An_{20-38}), plagioclase-quartz, plagioclase-quartz-epidote-carbonate, plagioclase-epidote-carbonate, plagioclase-chlorite, epidote, or epidote-quartz. Any of these minerals (other than hornblende), however, may be absent or present only in small amounts. They generally form a groundmass for lenses and seams of strongly alined hornblende. Such hornblende aggregates typically are 0.2–0.5 mm thick; groundmass areas between them range from less than 0.05 to about 0.3 mm across. The quartz and plagioclase grains generally are equant or slightly elongated, are less than 0.05 mm in size, and form intergrown granoblastic mosaics. The plagioclase of the mosaics is untwinned, but some isolated (relic?) twinned laths of plagioclase also occur in the schists. Epidote-clinozoisite is fine to medium grained, granular, and rod shaped. Biotite is abundant in some layers, and magnetite, sulfides, sphene, and leucoxene are sporadic minor minerals.

Veinlets containing carbonates, epidote, and quartz alone or in various combinations are very common.

Lepidoblastic, nematoblastic, and microaugen fabrics are common and result from intergrowth of plagioclase and quartz with strongly oriented chlorite, biotite, and hornblende. Scattered, simply twinned stubby laths of sodic plagioclase are the only textural forms possibly retained from the primary fabric.

ORIGIN OF THE LAYERED STRUCTURE

Williams (1891, p. 154–158), and all geologists who studied the layered amphibole schists before him, concluded that the layering was a result of some type of sedimentation. The chemical and general mineralogical similarity of the schists to the massive greenstones of the lower part of the Mona, however, led Williams to consider the possibility that the schists were modified from an originally massive volcanic rock. His inability to devise a satisfactory hypothesis for the transformation forced him to conclude that the schists are "highly metamorphosed diabase tuffs."

Eskola (1932), Schmidt (1932, p. 183), Turner (1941), and other geologists have described processes of metamorphic differentiation whereby originally massive unlayered rocks have been converted into layered bodies, or the layering of originally layered rocks has been accentuated. Thin streaky lenses and laminae (generally 0.2–0.5 mm thick) of light- and dark-colored minerals within layers of the amphibole schist probably were at least in part formed by metamorphic differentiation. The influence of metamorphic differentiation in forming the thick layers cannot be ruled out, but because of the absence of a clear transition along the strike from the layered schist into unlayered rock of similar composition, there is little evidence directly suggesting such a process.

The lenticular form of the layers has led to some speculation (H. L. James, oral commun., 1958) that they may be greatly stretched and flattened pillows, presumably resulting from unusual compressive shearing attending the intrusion of granitic material into the Mona along the north side of the area of amphibole schist. However, the possibility that pillows of varied size and shape were squeezed and stretched into uniform flat lenses, 1–3 inches thick and 10–30 feet in diameter, throughout a section of 2,000 feet or more thick and for a distance of 5–6 miles along the strike seems extremely remote. Furthermore, it seems unlikely that any vestige of original pillow forms would survive the shearing and recrystallization accompanying such deformation or survive the transfers of material necessary for the development of the thin dark- and light-colored laminae from pillows of uniform composition.

The layered amphibole schist is therefore interpreted as metamorphosed mafic tuff, following Williams. The present layering is considered to represent primary tabular layers and lenses of mafic volcanic ash now modified by recrystallization, by bedding slippage and low-angle shearing across bedding, by shearing and some secondly segregation within layers, and very locally by boudinage.

The unlayered amphibole schist may have been fine-grained mafic lava or thick beds of (subaerial?) ash without internal layering.

CHLORITIC AND ACTINOLITIC SLATES AND SCHISTS

Chloritic slate and schist are widely distributed within the Mona but are generally not well exposed. Their most extensive occurrence is in the lower part of the Lighthouse Point Member, where, in a poorly defined belt across the Marquette quadrangle, they range in thickness from about 2,800 feet on the west to about 300 feet on the east. These chloritic rocks probably correspond to the Eureka Series of Williams (1891). The best exposures in this belt are on the southwest shore of Lighthouse Point, a short distance north of the breakwater and near the old Eureka mine in the north-central part of sec. 21, T. 48 N., R. 25 W. Mr. Louis Vierling, a longtime resident of Marquette, pointed out the southeast undercut bank of the creek, 3,900 N–3000 E, of SW. cor. of sec. 21, as the location of the Eureka mine; this does not agree with the map of Van Hise and Bayley (1897, atlas, sheet 36), but does agree with records of the Cleveland Cliffs Iron Co.

Thin chloritic slate and schist are interbedded with the dominant massive metabasalt of the lower member of the Mona, particularly in secs. 22 and 26–29, T. 48 N., R. 25 W.

The Mona under the south limb of the east-plunging syncline in sec. 1, T. 47 N., 25 W., and sec. 6, T. 47 N., R. 24 W., is best exposed in the Harvey quarry in the NW¼ sec. 6. T. 47 N., R. 24 W. The Mona there is mainly chloritic schist (some of which is also rich in quartz and muscovite) and actinolitic schist, but in contrast to chloritic schist in the lower member of the Mona north of the synclinorium, most of this schist shows layering. In this respect, it resembles some of the amphibole schist of the Lighthouse Point Member, but the evidence as to which part of the Mona it belongs is inconclusive.

DESCRIPTION AND ORIGIN

The chloritic schist and slate under the north limb of the synclinorium typically are dull shades of green or brownish green and vary in fissility from strongly schistose to slaty. Major constituents are chlorite and quartz, or chlorite, sericite, and quartz. Rounded chloritic granules in some of these rocks may have been amygdules, or detritus from underlying greenstone. At one place the schist consists almost entirely of chlorite studded with brownish cloudy epidote and clinozoisite. Swirly and sinuous "lines" of dusty brownish opaque material in this rock probably represent original flow structure or the outlines of flattened shards of a glassy volcanic rock. Carbonate, leucoxene, and red- or yellow-brown iron oxides generally are minor constituents, and biotite is known at one locality. The typical fabric of these rocks results from an intimate mixture of fine-size granules of quartz and strongly oriented tiny flakes of chlorite and sericite. Irregular-shaped or roughly lenticular zones commonly are rich in either chlorite or sericite to the near exclusion of the other.

The origin of most of the chloritic schist and slate is not clear. The abundance of chlorite, the general lack of detrital textures, and the proximity to known metavolcanic rocks suggest a volcanic origin. The relative abundance of quartz, on the other hand, suggests a sedimentary origin. Quite possibly these rocks originated as water-laid tuffs or mixtures of volcanic ash and chemically precipitated silica.

The Mona in the Harvey quarry, immediately underlying Animikie rocks on the south limb of the east-plunging syncline, is a medium-grained chlorite-quartz-muscovite schist having crinkled foliation surfaces transected by fracture cleavage. Granoblastic-textured quartz in seams and lenses parallel to foliation makes up about half the crinkled schist and suggests a metasedimentary origin.

Southward in the quarry from this chlorite-quartz-muscovite schist, the rock becomes more mafic and somewhat finer grained. Most is quartz-chlorite schist, quartz-plagioclase-chlorite schist, and strongly foliated actinolitic schist, in which small amounts of carbonate, sericite, leucoxene, epidote, and iron oxides are common. These schists are cut by some thin dikes of greenstone and granitic rock. Sericitic quartz-chlorite schist in the central part of the Mona outcrop in the quarry, and actinolitic schist near the south end of the quarry are distinctly layered, a result of alternating concentrations of quartz and chlorite or of quartz and amphibole, plagioclase, and epidote-carbonate. This sequence in the Mona suggests a change from dominantly sedimentary to dominantly tuffaceous rocks.

A laminated quartz-actinolite rock of probable mixed sedimentary and volcanic origin is unique in the area and deserves special mention even though it is a minor occurrence. It is found in one small outcrop about 1,150W, SE. cor. sec. 35, T. 48 N., R. 25 W., in the belt of Mona that forms the core of the anticline in secs. 2 and 3, T. 47 N., R. 25 W. It is interlayered with greenschist, massive porphyritic, biotitic felsite, and chlorite schist. In the laminated rock, quartzitic layers 1–5 cm thick alternate with thin seams and lenses ($<$1–10 mm) of actinolitic amphibole or sodic plagioclase and actinolite, with or without carbonate and iron oxides.

In layers consisting mainly of actinolitic amphibole, N_X of the amphibole is close to 1.640, N_Y is 1.660–1.670, N_Z is 1.672–1.673, and $2V_X$ is 74°. In plagioclase-rich layers, fibrous actinolite grows along cleavage and twin planes in the plagioclase and apparently formed by replacement of plagioclase. The fibrous actinolite has N_X 1.630, N_Y 1.641, and N_Z 1.650, and evidently is poorer in iron than the actinolitic amphibole in the amphibole-rich layers. The sodic plagioclase and actinolite that partly replaces it probably are pseudomorphous after earlier more calcic plagioclase. The plagioclase grains are mainly stubby laths having simple twinning, although one blocky grain with two sets of lamellar twins at about right angles to one another was seen. Common grain sizes range from about 0.05 to 0.25 mm, but a few laths are 0.6–0.8 mm in size. Because grain boundaries are generally somewhat jagged, the plagioclase grains look as though they might have had a pyroclastic origin. Carbonate is intergrown with and appears to replace both sodic plagioclase and actinolite.

The iron oxides associated with both the amphibole-rich and plagioclase-actinolite layers are magnetite and secondary iron oxides formed by the oxidation of magnetite, probably during weathering. The magnetite evidently formed, at least in its present form, during regional metamorphism.

The quartzitic layers have small isolated crystals of amphibole scattered through them. Quartz in these layers is medium grained, recrystallized, and elongated parallel to layering.

Although the origin of this laminated rock is not clear, the relatively large size of recrystallized quartz grains and the layering suggest a detrital origin. The abundant amphibole or twinned plagioclase in some layers, on the other hand, indicates the probability of igneous origin. The rock therefore probably consists of alternating layers of detrital quartz and pyroclastic material.

HEMATITE-RICH ZONES

The chloritic slate, especially in the lower part of the Lighthouse Point Member, commonly has secondary seams and lenticles along foliation, fractions of a millimeter in thickness, and thin crosscutting veinlets of quartz and carbonate. The carbonate is partly replaced by reddish-brown iron oxide. Where such replacement is extensive, the iron oxide tends also to infringe on neighboring chlorite. A hematite-rich shear zone in chloritic slate in the southeast bank of a creek 3,900N–3,000E from SW. cor. sec. 21 is about 5 feet wide. Earthy hematite evidently was concentrated in this shear zone by the oxidation of chlorite or of chlorite and carbonate. This seam has been prospected and is probably at the actual site of the so-called Eureka mine. Some reddish iron oxide also occurs along shear surfaces in the exposure of chlorite schist and slate near Lighthouse Point, on the shore some 600 feet southwest of the lighthouse. The concentration of earthly iron ore in the shear zones and in the carbonate-bearing seams and veinlets probably are near-surface features related to weathering and the movement of meteoric water.

FELSITE

Felsite is a comparatively minor component of the Mona Schist in the Marquette quadrangle. In the lower member of the Mona, felsite slate and prophyritic felsite schist are each known from a single location, and massive porphyritic felsite from two places. The slate occurs near the center of sec. 22, and the massive felsite occurs in the NW¼NW¼ sec. 29 and in the outcrop 1,150 W, SE. cor. sec. 35, T. 48 N., R. 25 W. The porphyritic felsite schist forms a layer in the northern part of the Harvey quarry, south of the Animikie-Mona contact in the NW¼ sec. 6, T. 47 N., R. 24 W. Similar rocks are more widely exposed in the lower part of the Lighthouse Point Member. Thin, comformable or slightly crosscutting tabular bodies of porphyritic felsite occur widely in the hornblende schist of the Lighthouse Point Member. They range from a few inches to about 15 feet in thickness. The crosscutting bodies are intrusive, but conformable felsite may be of intrusive or volcanic origin.

The main varieties of felsite are in the lower part of the Lighthouse Point Member and are fissile, or flinty and nonfissile, poorly laminated fine-grained quartz-sericite rocks, with or without opalescent or granular "quartz eyes," and massive to schistose medium-grained fragmental quartz-feldspar rock. The fissile and flinty quartz-sericite rocks occur principally between the SW ¼ sec. 16 and the NE¼ sec. 23, T. 48 N., R. 25 W. Fissile quartz-sericite slate is interlayered with quartz-chlorite slate in two large outcrops, one in the SW¼SW¼, the other in the SE¼SW¼ sec. 16, T. 48 N., R. 25 W. The coarser grained fragmental rocks are exposed mainly near the SW. cor. sec. 16 and in the SE¼SE¼ sec. 17, T. 48 N., R. 25 W.

DESCRIPTION

The felsitic rocks typically are gray or pale gray green where fresh; they weather to pale gray, brownish gray, rusty pink, or cream. The fine-grained fissile and flinty rocks (most grains less than 0.05 mm) consist of granoblastic mosaics of quartz through which either strongly or weakly oriented flakes of sericite are interspersed. Small amounts of feldspar and chlorite are common in such mixtures. The feldspar occurs in untwinned grains in the quartzose mosaics and as small, generally twinned phenocrysts or fragments studded through the groundmass mosaic. Recrystalized phenocrysts of quartz now are generally lenticles along foliation surfaces in the slate, whereas phenocrysts of quartz or feldspar in the nonfissile rock are subeuhedral crystals or angular fragments. The feldspar phenocrysts in one sample of flinty quartz-sericite rock have been identified as sodic oligoclase (An_{13}). Small amounts of brownish iron oxide, leucoxene, and biotite occur in some of these rocks. Quartz-sericite slate exposed along the shore near the north end of the breakwater in the SW¼ sec. 24, T. 48 N., R. 25 W., contains a small amount of tourmaline.

The medium-grained felsite that is well exposed in the SE¼ sec. 17 is a chaotic mixture of angular fragments of sodic plagioclase and quartz, between 0.5 and 1.0 mm in size, separated by pronounced shear seams that are deflected around individual crystal fragments. Granular quartz and strongly oriented flakes of chlorite and biotite, all less than 0.04 mm in diameter, occupy the shear seams. The fabric of this rock, allowing for mineralogical differences, is similar to that of a greenstone tuff shown by Bayley (1959, fig. 12).

ORIGIN

Laminated fine-grained flinty quartz-sericite rock from near the west end of Ridge Street in Marquette was called novaculite by Williams (1891, p. 152–153) in deference to local practice of the time, but was recognized by him as silicic volcanic rock and compared with silicic tuffs from a number of European localities. An analysis of the flinty light-colored laminated rock from Ridge Street (probably the outcrop at 4,200N–1,600E, SW cor. sec. 23, T. 48 N., R. 25 W.) was given by Williams (1891, p. 152) and is reproduced here (table 4). It indicates a rhyolitic composition for the flinty quartz-sericite rock. Textural as well as chemical reasons for this interpretation are given in detail by Williams. We follow Williams in interpreting these and similar rocks in the lower part of the Lighthouse Point Member as silicic tuffs.

TABLE 4.—*Chemical analysis of laminated siliceous metavolcanic rock*

[From Williams, 1891, p. 152. Analyst, W. F. Hillebrand]

SiO_2	76.99	K_2O	3.65
Al_2O_3	13.92	Na_2O	.56
Fe_2O_3	.45	Li_2O	Tr
FeO	.77	H_2O	2.35
MnO	Tr	P_2O_5	Tr
CaO	.32		
MgO	1.12	Total	100.13

The chemical analysis of the quartz-sericite rock from Ridge Street compares rather closely with analyses of four samples of metarhyolite from the Hemlock Formation in Iron County, Mich. (Gair and Wier, 1956, table 8). Although the quartz-sericite rock from Ridge Street in Marquette has substantially less K_2O than three of the four analyzed samples of the metarhyolite from Iron County, its K_2O content, like that of metarhyolite sample 4 of Gair and Wier, is close to the average K_2O for eight rhyolite samples listed by Clarke (1920). The relatively low Na_2O content of the felsite from Marquette compares with the Na_2O content in all four samples of metarhyolite of Gair and Wier (1956, table 8).

MAFIC META-AGGLOMERATE

Distinctly fragmental mafic volcanic rock is a rare component of the Lighthouse Point Member in the Marquette quadrangle, but its presence tends to substantiate the pyroclastic origin of the associated amphibole schist.

Several layers of nonschistose medium- to fine-grained fragmental rock, from a few inches to about 3 feet in thickness, have been found in two places in the SW¼ of sec. 16, and one layer of schistose crystal tuff, 3–4 feet thick, has been found at one place in the NW¼SW¼ sec. 15, T. 48 N., R. 25 W. The meta-agglomerate in sec. 16 is interlayered with amphibole schist, whereas that in sec. 15 is interlayered with a schistose greenstone mapped as metabasalt(?).

The nonschistose meta-agglomerate in sec. 16 is gray green with a faint maroon tinge; it weathers to shades of brown. Angular fragments in the rock are deep maroon and range from about 0.5 to 8.0 mm in size. The fragments now consist of fine-grained aggregates of muscovite-sericite through which are thin stringers of magnetite partly oxidized to hematite. The original nature of these fragments is unknown. They "float" in a granoblastic groundmass, mainly of tremolite, blebby feldspar, and granular magnetite and minor carbonate, biotite, and dusty opaque material. In one layer of meta-agglomerate or an altered vitrophyre, aggregates of muscovite-sericite, calcite, and tremolite are pseudomorphs of original fragments and euhedral crystals of plagioclase and are set in a base mainly made up of fine-grained sericite and iron oxides. The pseudomorphs of fragments and phenocrysts compose roughly 75 percent of the rock.

The schistose crystal tuff consists of abundant fragments and euhedra of plagioclase, 0.5–1.0 mm in size, in a strongly schistose groundmass of alined blue-green hornblende, granular quartz, plagioclase, and magnetite. Some hornblende crystals are as much as 1.0 mm long but are alined with and appear to be part of the schistose groundmass, in contrast to equant feldspar of similar size, around which the schistosity commonly bends. Individual alined hornblende crystals commonly are lenticular, as if mechanically ground off at the ends. Except for the abundant feldspar grains studded through the fabric, the mineralogy and fabric of this rock are similar to those of the typical layered amphibole schist.

COMPARISONS OF CHEMICAL ANALYSES, NORMS, AND MODES METAMORPHISM

Three chemical analyses of mafic metavolcanic rocks of the Mona help to show that the parents of these rocks were tholeiitic basalts and that their metamorphism was essentially isochemical. Differences and similarities between these metabasalts, feldspathized metabasalts, and basalts and metabasalts from other areas are best seen by comparing individual oxide components in the chemical analyses (table 2). Modes of the chemically analyzed rocks from the Mona Schist are shown in table 5. The metabasalt of the Marquette area is seen to be quite similar to basalts and metabasalts from other areas. This similarity is made especially evident by comparing analyses 1 and 2 with analyses 6, 7, 9, and 10 in table 2. Analysis 3 in table 2, however, is relatively high in SiO_2 and Al_2O_3, and low in CaO and MgO. The rock represented by analysis 3 possibly was affected more by metasomatism during metamorphism than the rocks represented by analyses 1 and 2, or, being of tuffaceous origin, it may have had a small admixture of aluminous and siliceous sediments. Standard (CIPW) norms derived from chemical analyses 1–3 of table 2 are compared with the norms of nonmetamorphosed basalts (table 3). Although the norms of the metavolcanic rocks do not even approximate the present mineralogy of the rocks, they provide an additional and standard basis for a chemical comparison of metamorphosed and nonmetamorphosed rocks of approximately equivalent type. Epinorms and mesonorms come closer to the existing mineralogy of low-grade and medium-grade metamorphic rocks than do CIPW norms, by providing for water-bearing normative minerals (Barth, 1955, 1959). A comparison of metamorphic norms for the metamorphic rocks with CIPW norms for the nonmetamorphosed rocks, however, would be unsatisfactory here because the two types of norms are based on different standards. The similarity of the norms of the two metabasalts (norms 1 and 2) is quite evident, and their similarity with norms 4–7 of table 3 indicate their essentially isochemical metamorphism.

TABLE 5.—*Modes of five chemically analyzed samples of mafic metavolcanic rocks, Mona Schist, Marquette quadrangle*

[For chemical analyses of corresponding rocks, see analyses 1–5, table 2]

Mineral	1	2	3	4	5
Quartz	1.0	1.7	------	10.2	5.3
Quartz and sodic plagioclase [1]	------	------	34.4	------	------
Sodic plagioclase	10.0	9.3	------	------	------
Sericitized sodic plagioclase	------	------	------	53.2	54.5
Potassium-feldspar	------	.6	------	5.5	7.5
Epidote minerals	26.3	32.0	4.2	------	------
Chlorite	16.3	.6	4.7	------	16.8
Biotite	------	------	3.2	------	Tr
Hornblende	------	------	50.5	------	------
Actinolitic amphibole	44.6	55.3	------	------	------
Carbonate	1.0	.6	1.2	8.2	12.9
Magnetite	------	------	.7	------	------
Unidentified opaque	Tr	Tr	------	.7	1.2
Leucoxene	.7	Tr	.7	1.5	------
Apatite	Tr	Tr	Tr	Tr	1.2
Rutile	------	------	------	Tr	.3

[1] Quartz and sodic plagioclase form fine-grained intergrowths which are counted together.

The norm for No. 3, a layered amphibolitic schist from the Lighthouse Point Member, is relatively low in diopside and slightly high in anorthite. Also revealed in the CIPW computation for No. 3, but not shown in table 3, is an abnormally high proportion (approximately 12:11) of FeO to MgO in the diopside and hypersthene. The low diopside is nominally a result of relatively high Al_2O_3 and low CaO, which, in the CIPW calculation, results in most of the available CaO being assigned to anorthite. The high FeO content of the diopside and hypersthene is a result both of the comparatively low MgO in No. 3 and the comparatively low Fe_2O_3 content of the rock, which necessarily reduces the amount of FeO assigned to magnetite. These abnormalities of No. 3 in tables 2 and 3 do not appear to be systematic, as would be expected if the analysis and norm represented merely a more siliceous parent or differentiated magma or lava than the other analyses and norms. The evidence, however, is insufficient to indicate the possible cause or causes of the abnormalities.

Two of the three analyzed mafic rocks from the Mona have normative quartz, and all three have MgO/CaO ratios between 0.79 and 0.88; an origin from tholeiitic magma is thus indicated, by a comparison with ratios computed from analyses given by Turner and Verhoogen (1951) in table 16, p. 180, using molecular proportions of MgO and CaO. This conclusion is also supported by the Na_2O content of the two analyzed samples of metabasalt, which compares closely with Na_2O in typical tholeiitic basalt. The fine-grained metabasalt (No. 2, tables 2 and 3), because of having almost 1 percent more K_2O than the other two analyzed rocks, is deficient in normative quartz and has normative olivine, but its MgO/CaO ratio is typical of tholeiite, rather than of oceanic (olivine) basalt. Turner and Verhoogen (1951, p. 179–181) have emphasized that distinctions between olivine basalts and tholeiitic basalts are not always sufficiently clear cut for a determination based on one criterion. The weight of evidence, however, indicates derivation of the mafic volcanic rocks of the Mona from a tholeiitic magma.

In several places in the southern part of the area of Compeau Creek Gneiss in the Marquette quadrangle, remnants of greenstone forming poorly defined septa in the gneiss appear to have been secondarily feldspathized. Feldspathization probably occurred during the emplacement of the Compeau Creek Gneiss. Chemical analyses of two samples of feldspathized greenstone are shown in table 2 (Nos. 4 and 5), and the corresponding modes are listed in table 5 (Nos. 4 and 5). The potassium-feldspar and the relatively large amounts of quartz and sodic plagioclase in the feldspathized greenstone are attributed here to metasomatism that occurred during emplacement of the gneiss. Carbonate and sericite appear to be later than the feldspars and quartz and are interpreted as products of low-grade metamorphism that occurred subsequent to the formation of the gneiss.

METAMORPHISM

Studies by the U.S. Geological Survey since 1943 have identified at least two episodes of regional metamorphism in northern Michigan. The most recent was associated with a major orogeny that occurred after deposition of the youngest middle Precambrian (Animikie) rocks and prior to eruptions of upper Precambrian (Keweenawan) igneous rocks. One or more earlier episodes of metamorphism occurred before the deposition of middle Precambrian rocks.

The pattern of regional metamorphism given by James (1955, pl. 1) represents changes that took place during the last major orogeny, in the post-Animikie, pre-Keweenawan interval. The metamorphism reached the level of the chlorite zone in the eastern part of the Marquette iron range, as is well shown in the lower member of the Mona Schist by characteristic assemblages of sodic plagioclase, epidote minerals, sericite, quartz, and pale-green actinolitic amphibole or chlorite. James' map shows an increase in metamorphic grade westward along the range, the isograds trending obliquely across the axis of the range. Notwithstanding the nominal low-grade metamorphism in the Marquette area at the eastern end of the range, the Lighthouse Point Member along the northern third of the belt of Mona Schist is characterized by mineral assemblages rich in blue-green hornblende, oligoclase-andesine (An_{20-38}), quartz, and epidote.

The change from assemblages rich in chlorite or pale-green actinolitic amphibole in the lower member of the Mona to assemblages rich in blue-green hornblende in the upper part of the Mona might represent only premetamorphic compositional differences or a real increase in metamorphic grade from the greenschist to the epidote amphibolite or almandine amphibolite facies (for a discussion of the characteristics of these facies, see Fyfe and others, 1958, p. 217–218.) Several geologists have considered green and blue-green (aluminous) or other dark hornblendes—characterized by moderate to high indices of refraction ($N_Y > 1.651$)—and oligoclase or more calcic plagioclase to be indicators of a higher metamorphic grade than pale-green actinolitic hornblende or chlorite in association with albitic plagioclase, quartz, and sericite. (See Eskola, 1925, p. 43–44, 73–78; Wiseman, 1934, p. 365–369, 382–384; Turner, 1948, p. 90, 94; James, 1955, table 2, p. 1468–1470; Fyfe and other, 1958, p. 217–218.) James specifically related blue-green hornblende to the biotite and the garnet zones and green hornblende to the upper part of the garnet zone and to the staurolite and sillimanite zones of regional metamorphism. Although Wiseman considered deep-green hornblende to be characteristic of the garnet and higher zones, he also found some in low-grade (chlorite-biotite) zones in the southwest Highlands of Scotland. Eskola (1925, p. 44) also reported some dark-green hornblende from a low-grade metamorphic zone in Karelia, in apparent contradiction of his general observations. Turner (in Williams and other, 1954, p. 220; and in Fyfe and others, 1958, p. 218) pointed out the difficulty or impossibility of optically distinguishing deep-green actinolite from dark aluminous hornblende, and therefore in the later publication proposed eliminating use of the epidote amphibolite facies and distinguishing only between a greenschist and an almandine amphibolite facies on the basis of anorthite contents of less or more than 10 percent in plagioclase.

Plagioclase in the layered amphibole schist of the Lighthouse Point Member occurs mainly in fine-grained granoblastic mosaic aggregates charged with abundant inclusions. It has been possible only to check the plagioclase for its range in indices of refraction, all of which measure greater than 1.540 and less than 1.557. These indices indicate oligoclase-andesine (An_{20-38}), which in turn suggests that the dark amphibole is a true aluminous hornblende. The metamorphic grade of the Lighthouse Point Member is clearly higher than that of the greenschist facies, and the member is therefore more highly metamorphosed than the lower member of the Mona. The Lighthouse Point Member is classified as an amphibolite (Fyfe and others, 1958, p. 161, 218). Evidence discussed below indicates that the Lighthouse Point Member was metamorphosed prior to the regional metamorphism of the post-Animikie, pre-Keweenawan interval.

The evidence of earlier metamorphism of the Lighthouse Point Member is based on structural, textural, and mineralogic features in the amphibole schist and in crosscutting bodies of felsic porphyry and metadiabase. Dikes of metadiabase cut the amphibole schist on Lighthouse Point. The fabric of the host rocks is strongly schistose. In marked contrast to this, the fabric of the crosscutting metadiabase dikes is not sheared parallel to the schistosity of the host rocks and commonly retains the original diabasic texture and suggests that the schistosity of the host rocks was developed prior to intrusion of the metadiabase. Tabular intrusive bodies of porphyritic felsite in the Lighthouse Point Member tend to conform to the schistosity and layering of the amphibole schist, but locally the felsite cuts these structures at low angles and thus was probably emplaced during or after deformation of the schist. Such bodies of felsite are crosscut by the metadiabase on Lighthouse Point, thus strengthening the conclusion that the intrusion of the metadiabase postdates the deformation of the amphibole schist. The strongly alined blue-green hornblende is the major constructional element of the rock schistosity and therefore probably formed simultaneously with the schistosity. Thus, the blue-green hornblende antedated the intrusion and metamorphism of the metadiabase. Mineral assemblages in the metadiabase indicate metamorphism in the greenschist facies or the chlorite zone. This degree of metamorphism conforms to that in the lower member of the Mona and in Animikie rocks to the south and to that in premetamorphic mafic dikes cutting amphibolite-bearing Compeau Creek Gneiss to the northwest. The metamorphism of the metadiabase is assigned to the regional metamorphism of the post-Animikie, pre-Keweenawan interval, whereas the metamorphism of at least the northern part of the Mona Schist is interpreted here as an earlier event.

The most likely cause of the higher metamorphic grade and apparently greater deformation in the northern part of the belt of Mona Schist, compared with the southern part of the Mona, was the igneous activity connected with the formation of the Compeau Creek Gneiss. Beds of amphibolite in the gneiss indicate temperatures there well above the level of the greenschist facies. Apparently during emplacement of the gneiss, the metamorphic grade of the southern part of the Mona Schist was not raised above the grade of the greenschist facies, which grade it again attained during the post-Animikie, pre-Keweenawan metamorphism. During the later time of low-grade regional metamorphism, mafic dikes that postdated the earlier metamorphism were altered, but relatively little change may have occurred in the lower part of the Mona Schist if that rock had already undergone low-grade metamorphism. During the later metamorphism the earlier formed assemblages rich in blue-green hornblende evidently remained virtually intact, although small amounts of chlorite, quartz, and sodic plagioclase in the amphibole schist may be retrogressive minerals formed at that time.

Brown or green biotite is associated with deep-green or blue-green hornblende in some places in the amphibole schist. Most such occurrences are close to intrusions of upper Precambrian (Keweenawan) diabase and may be a product of contact metamorphism.

COMPEAU CREEK GNEISS

DEFINITION, DISTRIBUTION, TRENDS, AND AGE

The name Compeau Creek Gneiss is here used for lower Precambrian rocks that consist largely of foliated light-colored tonalite and granodiorite, dark chloritic, biotitic, and hornblendic varieties of these rocks, and amphibolite. Quartz monzonitic and monzonitic compositions are not common, and granite is rare. Many definitions of granite in the literature are based on ratios of approximately 65 : 35 of alkali feldspar to sodium-calcium feldspar, the alkali feldspar ranging from orthoclase to albite or oligoclase (An_{20}). Inasmuch as the natural break in feldspar solid-solution series occurs between potassium-feldspar and plagioclase, the classification of granitic rocks in this report is based on a ratio of potassium-feldspar to plagioclase of 2 : 1. The gneiss is named after Compeau Creek, the type locality, along and near which the rock is extensively exposed in the northern part of the Marquette quadrangle. Similar gneiss is also exposed along the south side of the Marquette synclinorium in the northern part of the Sands quadrangle and in the drainage of the East Branch of the Escanaba River in the southwestern part of the Sands quadrangle. Many characteristic features of the gneiss, and some not seen elsewhere, are well exposed in a wave-washed outcrop one-quarter mile south of Wetmore Landing in sec. 29, T. 49 N., R. 25 W., about 4 miles northwest of Marquette (pl. 3).

Some of the gneiss near the southern margin of the synclinorium in the northern part of the Sands quadrangle is exceptionally rich in quartz and probably corresponds to the Palmer Gneiss of earlier workers (Van Hise and Bayley, 1897, p. 210–218; Van Hise and Leith, 1911, p. 255–256). The probable origin of the Palmer Gneiss by granulation and the introduction of secondary quartz into "normal" gneiss (Van Hise and Bayley, 1897, p. 216–218), however, preclude correlations with it in a strict stratigraphic sense. Compeau Creek Gneiss is silicified to a cherty rock and is designated as silicified gneiss at several places in the southwestern part of the Sands quadrangle. Septa of greenstone which are interpreted as inclusions or remnants of Mona Schist occur widely in the gneiss in the northern half of the Marquette quadrangle.

Inasmuch as there is no surface continuity between the gneisses north and south of the Marquette synclinorium, there is no certainty that they form a continuous body within the earth's crust. Measurements of the gneisses with a scintillation detector at several places by Robert Reed of the Michigan Department of Conservation and by us gave readings of 0.024–0.025 milliroentgen per hour north of the synclinorium and 0.035–0.045 milliroentgen per hour south of the synclinorium. The differences in radioactivity might indicate discrete bodies of different age. The gneisses are correlated here, however, and given the same name because of their lithologic similarities, their structural trends, and because they are of early Precambrian age and form part of the basement of the Marquette synclinorium.

The Compeau Creek Gneiss has rather persistent east-to-east-southeast-trending foliation and layering. Folds are not abundant; the only ones detected in the layered gneiss are sporadically distributed within the limits of individual outcrops and typically have amplitudes and limb lengths of a few feet or less. Sets of folded subparallel partings mapped as "bedding" in the southern half of the gneiss outcrop south of Wetmore Landing (pl. 3) may be relic bedding from country rock predating the gneiss. The most notable departure from typical trends in the gneiss is in an outcrop in the SW¼ sec. 6, T. 48 N., R. 25 W., where tightly folded, steeply dipping amphibolite trends from southeast to nearly south for at least 100 feet along the strike.

The Compeau Creek Gneiss in the Marquette and Sands quadrangles, is part of the northern and southern complexes, respectively, of Van Hise and Bayley (1897, p. 149–151, 169–176, 190–192, 209–218), which were essentially geographic subdivisions of a basement complex of "Archean" age. Such rocks that were formerly called "Archean" lie unconformably beneath the Animikie Series and are now designated early Precambrian in age without connotation as to correlation with Precambrian rocks elsewhere (James, 1958, p. 31–33). The pre-Animikie age of the bulk of the Compeau Creek Gneiss is indicated by the presence of boulders of gneissic rocks in some lowermost rocks of Animikie age and by the absence of granitic rock intrusive into the Animikie rocks. Small amounts of post-Animikie granite are known in the southern complex (James, 1958, p. 39); therefore, minor occurrences of weakly foliated or nonfoliated granitic rock now mapped as part of the Compeau Creek Gneiss may be really of post-Animikie age. The light-colored dikes in the outcrop south of Wetmore Landing (pl. 3) may be such post-Animikie intrusive rocks. Abundant bodies of pegmatite in the southwestern part of the Sands quadrangle, too small to be mapped, clearly are younger than the gneissic host rock and may be of post-Animikie age.

It is unknown whether the Compeau Creek Gneiss is correlative with gneiss of pre-Dickinson or post-Dickinson age in the west-central or southwestern parts of the southern complex (James, 1958, p. 31–33), or whether the gneiss is of yet another pre-Animikie age. On the basis of potassium–argon ages of about 1,800 million years for pegmatite intrusive into Animikie rocks in Dickinson County, Mich. (James and others, 1961, p. 58), the pre-Animikie rocks, including the Compeau Creek Gneiss, are probably at least 2 billion years old.

DESCRIPTION

Modes of the Compeau Creek Gneiss, in volume percentages, are listed in tables 6 and 7, and seven chemical analyses of gneiss are given in table 8. Table 9 shows the compositional basis for the rock names used in the modal tables and text. The modes show that compositions are somewhat gradational and that the indicated boundaries are in general arbitrary. Epinorms and mesonorms, derived from the chemical analyses of the gneiss by the method of Barth (1955, 1959), are listed in table 10. Both epinorms and mesonorms are shown for light-

TABLE 6.—*Modes and averaged modes (volume percent) of light-colored gneiss [1] Compeau Creek Gneiss, Marquette and Sands quadrangles*

[Averaged modes show arithmetic mean and range, if any (in parentheses), for each mineral present in greater than trace amounts. (?), identification uncertain]

Mineral	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18
Sodic plagioclase [2]	61.2 (47–73)	67	64	53.5	58.5	50.6 (40–60)	----	58	58.5 (51–71)	46 (43–51)	----	43.5	43	47	51	20	58	21.5
Feldspar-sericite [3]	----	----	----	----	----	----	52 (44.5–63.5)	----	----	----	43 (34.5–48)	----	----	----	--	---	----	----
Quartz	26 (10.5–38.5)	14	26	28.5	27	37.5 (28–51)	42.6 (32.5–55.5)	17.5	22.2 (4–35.5)	32.5 (26.5–36)	39.2 (32.5–45.5)	31.5	28.5	9.5	5	34	6	43.5
Potassium-feldspar [4]	3.8 (0–8.5)	Tr	3	4	2.5	1.4 (0–6)	1 (0–4.5)	1	12.8 (10–19.5)	14 (7–19.5)	14.4 (11–16.5)	23	28	31	44	[5] 45	5	29
Sericite [6]	2.5 (0–6.5)	4	2	1.5	1	7 (1–15.5)	----	Tr	1.3 (0–2.5)	2.6 (0–7)	----	.5	Tr	----	--	---	11	1.5
Chlorite [7]	3.6 (Tr–8)	6	4	2	2	1.8 (Tr–6)	.5 (0–1.5)	19	2.5 [8] (Tr–3)	1 (0–4)	2 (Tr–3.5)	----	1	5.5	--	---	15	Tr
Biotite	.2 (0–2)	----	----	2	----	1 (0–9.5)	.2 (0–2)	----	----	3.3 (0–6)	---- (0–Tr)	----	.5	4.5	--	---	.5	4.5
Hornblende	----	----	----	5	4.5	----	----	----	Tr [8] (0–6)	----	----	----	----	----	--	---	----	----
Epidote	.6 (0–4)	5.5	----	2.5	3	Tr (0–3.5)	---- (0–Tr)	1.5	(0–Tr)	Tr (0–0.5)	----	Tr	----	----	--	---	----	----
Carbonate	.6 (0–3.5)	1.5	----	Tr	2.5	1 (0–5.5)	.7 (0–5)	----	.8 (0–2.5)	Tr (0–1)	.5 (0–2)	Tr	----	----	--	Tr	1.5	Tr
Sphene-leucoxene	.2 (0–1)	1	Tr	Tr	.5	Tr (0–0.5)	Tr (0–0.5)	2	Tr (0–0.5)	Tr (0–0.5)	Tr (Tr–1)	----	----	----	--	---	1	Tr
Magnetite	(0–Tr)	1	----	----	Tr	----	.2 (0–2)	----	Tr (0–0.5)	----	Tr (0–0.5)	Tr	----	----	--	---	1	----
Hematite [9]	Tr (0–1)		----	.5	.5	Tr (0–1)	.2 (0–0.5)	Tr	Tr (0–0.5)	Tr (0–0.5)	.5 (Tr–1.5)	Tr	----	----	--	---	----	Tr
Unidentified opaque	Tr (0–1)	----	1	----	----	(0–Tr)	----	1	Tr (0–0.5)	----	----	----	----	1.5	--	---	----	Tr
Pyrite	----	----	----	Tr	----	Tr (0–0.5)	----	----	----	----	----	----	----	----	--	---	----	----
Zircon	(0–Tr)	----	Tr	----	Tr	Tr	Tr	----	(0–Tr)	(0–Tr)	(0–Tr)	Tr	----	----	--	---	Tr	Tr
Apatite	Tr (0–.5)	----	----	Tr	Tr	(0–Tr)	(0–Tr)	Tr	(0–Tr)	(0–Tr)	Tr (0–0.5)	Tr	----	----	--	---	.5	----
Rutile	(0–Tr)	----	Tr	----	----	(0–Tr)	(0–Tr)	----	(0–Tr)	----	(0–Tr) (?)	----	----	----	--	---	----	----

[1] Light-colored gneiss contains a total of less than 10 percent chlorite, biotite, and hornblende.
[2] Most is slightly to moderately sericitized; some is slightly saussuritized; some stained by orange-colored iron oxide.
[3] Feldspar extensively sericitized; all identifiable feldspar is plagioclase; stained by orange-colored iron oxide in places; in few samples also includes small amounts of quartz, carbonate, chlorite, saussurite, or leucoxene.
[4] Most is microcline.
[5] Includes 5 percent granophyre.
[6] No determination of amount apart from feldspar in rocks containing "feldspar-sericite" or containing scattered separate small flakes of sericite in feldspar.
[7] Includes interleaved biotite or muscovite in few samples.
[8] High value is rough estimate from intergrown chlorite and hornblende composing 9 percent of mode in one sample.
[9] Includes some yellow-brown iron oxide.

1. Averaged mode, 14 samples of tonalite gneiss; Marquette quadrangle, north of Marquette synclinorium.
2. Tonalite gneiss (for chemical analysis, see No. 1, table 8); Marquette quadrangle, 400N–1,280E, SW. cor. sec. 9, T. 48 N., R. 25 W.
3. Tonalite gneiss (for chemical analysis, see No. 2, table 8); Marquette quadrangle, 50N–4,780E, SW. cor. sec. 4, T. 48 N., R. 25 W.
4. Tonalite gneiss (for chemical analysis see No. 4, table 8); Marquette quadrangle, 4,120N–1,750E, SW. cor. sec. 32, T. 49 N., R. 25 W.
5. Tonalite gneiss (for chemical analysis see No. 3, table 8); Marquette quadrangle, 3,760N–540E, SW. cor. sec. 33, T. 49 N., R. 25 W.
6. Averaged mode, 11 samples of tonalite gneiss; Sands quadrangle, south of Marquette synclinorium.
7. Averaged mode, 10 samples of extensively sericitized tonalite gneiss; Sands quadrangle, south of Marquette synclinorium.
8. Chlorite tonalite gneiss; Marquette quadrangle, 4,500N–2,500E, SW. cor. sec. 5, T. 48 N., R. 25 W.
9. Averaged mode, 5 samples of granodiorite gneiss; Marquette quadrangle, north of Marquette synclinorium.
10. Averaged mode, 6 samples of granodiorite gneiss; Sands quadrangle, south of Marquette synclinorium.
11. Averaged mode, 4 samples of extensively sericitized granodiorite gneiss; Sands quadrangle, south of Marquette synclinorium.
12. Quartz monzonite gneiss; Marquette quadrangle, 1,500N–1,700E, SW. cor. sec. 30, T. 49 N., R. 25 W.
13. Quartz monzonite offshoot of quartz monzonite layer in gneiss; cuts amphibolite layer; Marquette quadrangle, 1,500N–4,360E, SW. cor. sec. 31, T. 49 N., R. 25 W.
14. Monzonite gneiss; Marquette quadrangle, 3,040N–2,200E, SW. cor. sec. 7, T. 48 N., R. 25 W.
15. Monzonite offshoot of monzonite layer in gneiss; cuts amphibolite layer; Marquette quadrangle, 20N–4,500E, SW. cor. sec. 3, T. 48 N., R. 25 W.
16. Porphyritic granite gneiss; Marquette quadrangle, 3,120N–2,440E, SW. cor. sec. 17, T. 48 N., R. 25 W.
17. Chlorite diorite gneiss; Marquette quadrangle, 1,800N–540E, SW. cor. sec. 6, T. 48 N., R. 25 W.
18. Biotitic quartz monzonite gneiss; Sands quadrangle, 3,750N–640E, SW. cor. sec. 17, T. 46 N., R. 25 W.

TABLE 7.—*Modes and averaged modes (volume percent) of dark-colored gneiss* [1], *Compeau Creek Gneiss, Marquette and Sands quadrangles*

[Average modes show arithmetic mean and range, if any (in parentheses), for each mineral present in greater than trace amounts. [E, estimated; (?), identification uncertain]]

Mineral	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18
Sodic plagioclase [2]	30.7 (22.5–39)	47.5	42.8 (38.5–50)	----	----	----	30.5	37 (36–38)	----	36.5 (34.5–38.5)	----	41	----	69	----	57.7 (57–58.5)	39.5	----
Saussurite-plagioclase [3]	----	----	----	49	47.5 (30.5–64.5)	36	----	----	52	----	47	----	47.5	----	73	----	----	48.8 (48–50)
Quartz	4 (Tr–8)	Tr	2.3 (Tr–5.5)	6	1.7 (0–8)	5	12.5	19.2 (17.5–21)	11	27 (25.5–28.5)	38	41.5	22	1	5.5	21.5 (15–28)	33.5	32 (28–35.5)
Potassium-feldspar [4]	(0–Tr)	1	----	.5	1.2 (0–4)	3.5	[5] 2(?)E	----	1.5	(0–Tr) (?)	----	4	----	5	Tr	.2 (0–.5)	----	(Tr–2) .8
Hornblende	61.5 (60–63)	48.5	50.3 (38–57)	37	47.4 (30–64)	51.5	48.5	40 (38–42)	28	28.5 (19.5–27.5)	8.5	----	----	15	15	----	----	----
Biotite	----	----	1.3 (0–4)	----	----	.5	----	1.8 (Tr–3.5)	Tr	9	----	10.5	15.5	----	----	1 (0–2)	16	13.5 (11.5–16)
Epidote	1 (0–2)	.5E	.3 (Tr–1)	1	----	----	1.5	1.2 (.5–2)	1.5	2.2 (2–2.5)	[6] 1.5	Tr	[7] 8	2	2.5	1.5 (Tr–3)	Tr	3.5 (0–6)
Chlorite	1.2 (.5–2)	1.5	Tr	5.5	1 (0–4)	.5	----	Tr	4	.3 (Tr.–5)	3	Tr	3.5	3.5	1	12 (8–16)	1	0–Tr
Sericite	----	----	.5E (0–1.5E)	----	----	----	----	----	----	----	----	----	----	Tr	----	3.7 (2.5–5)	6	----
Carbonate	Tr	Tr	.5	----	.2 (0–.5)	.5	3	----	.5	----	----	----	Tr	2.5	1	1.2 (Tr–2.5)	3.5	.3 (Tr–1)
Sphene-leucoxene	Tr	1	1.2 (.5–2)	.5	.3 (0–.5)	2	1	.7 (.5–1)	.5	Tr	1.5	Tr	.5E	Tr	1	.2 (Tr–.5)	.5	.3 (Tr–.5)
Magnetite	----	----	.3 (0–1)	----	(0–Tr)	.5	----	----	----	1 (0–2)	----	3	----	1	----	.5 (0–1)	----	----
Hematite	(0–Tr)	Tr	Tr	Tr	(0–Tr)	Tr	Tr	(0–Tr)	Tr	Tr	Tr	Tr	----	1E	Tr	.2 (Tr–.5)	----	Tr
Unidentified opaque.	1 (0–2)	Tr	(0–Tr)	----	.3 (0–1)	----	1	(0–Tr)	Tr	.2E (0–.5E)	.5	----	2.5	----	----	----	----	----
Pyrite	----	----	(0–Tr)	----	----	----	----	----	----	----	----	----	----	----	1	----	----	(0–Tr)
Zircon	(0–Tr)	----	(0–Tr)	Tr	(0–Tr)	----	----	----	----	----	----	Tr	Tr	----	Tr	(0–Tr)	Tr	Tr
Apatite	.2 (0–.5)	Tr	Tr	Tr	(0–Tr)	----	Tr	Tr	Tr	Tr	Tr	Tr	Tr	Tr	Tr	.2 (0–.5)	Tr	(0–Tr)

[1] Dark-colored gneiss contains a total of more than 10 percent chlorite, biotite, and hornblende.
[2] Most is slightly to moderately sericitized; some is slightly saussuritized.
[3] Mainly epidote; epidote and sodic plagioclase; or epidote, sodic plagioclase, and sericite, commonly also containing small amounts of chlorite.
[4] Most is microcline.
[5] Probable potassium-feldspar, deeply colored by hematite.
[6] Large plates along grain boundaries of plagioclase or hornblende or enclosed mainly by hornblende or quartz; does not include fine-grained epidote in saussurite.
[7] Mainly in darker layers associated with biotite, quartz, chlorite, and opaque mineral; does not include epidote in saussurite.

1. Averaged mode, 2 samples of amphibolite; Marquette quadrangle, secs. 3 and 6, T. 48 N., R. 25 W.
2. Amphibolite (for chemical analysis, see No. 6, table 8); Marquette quadrangle, 4,140N–60W, SE. cor. sec. 36, T. 49 N., R. 26 W.
3. Averaged mode, 3 samples of amphibolite; Sands quadrangle, sec. 10, T. 47 N., R. 25 W., and sec. 18, T. 46 N., R. 25 W.
4. Amphibolite with extensively saussuritized plagioclase (for chemical analysis, see No. 7, table 8); Marquette quadrangle, 820N–1,350E, SW cor. sec. 19, T. 49 N., R. 25 W.
5. Averaged mode, 6 amphibolites with extensively saussuritized plagioclase; Marquette quadrangle, secs. 3 and 6, T. 48 N., R. 25 W., and secs. 30 and 31, T. 49 N., R. 25 W.
6. Amphibolite with extensively saussuritized plagioclase; Sands quadrangle, 1,450N–900 E., SW. cor. sec. 10, T. 47 N., R. 25 W.
7. Quartz amphibolite (for chemical analysis, see No. 5, table 8); Marquette quadrangle, 1,250N–1,300E, SW. cor. sec. 6, T. 48 N., R. 25 W.
8. Averaged mode, 2 samples of quartz amphibolite; Sands quadrangle, secs. 17 and 18, T. 46 N., R. 25 W.
9. Hornblende tonalite gneiss containing extensively saussuritized plagioclase; Marquette quadrangle, 4,100N–880W, SE. cor. sec. 25, T. 49 N., R. 26 W.
10. Averaged mode, 2 samples of hornblende tonalite gneiss; Sands quardangle, sec. 12, T. 46 N., R. 26 W., and sec. 21, T. 46 N., R. 25 W.
11. Hornblendic tonalite gneiss containing extensively saussuritized plagioclase; Sands quadrangle, 1,280N–1,050E, SW. cor. sec. 10, T. 47 N., R. 25 W.
12. Biotite tonalite gneiss; Sands quadrangle, 3,750N–640E, SW. cor. sec. 17, T. 46 N., R. 25 W.
13. Biotite tonalite gneiss containing extensively saussuritized plagioclase; Sands quadrangle, 1180N–300E, SW. cor. sec. 31, T. 47 N., R. 25 W.
14. Hornblende diorite gneiss; Marquette quadrangle, 950N–4,750E, SW. cor. secs. 30, T. 49 N., R. 25 W.
15. Hornblende diorite gneiss containing extensively saussuritized plagioclase; Marquette quadrangle, 380N–1,400E, SW. cor. sec. 19, T. 49 N., R. 25 W.
16. Averaged mode, 2 samples of chlorite tonalite gneiss; Sands quadrangle, south of Marquette synclinorium.
17. Biotite tonalite gneiss; Sands quadrangle, sec. 24, T. 46 N., R 26 W.
18. Averaged mode, 3 samples of extensively sericitized biotite tonalite gneiss; Sands quadrangle, south of Marquette synclinorium.

colored gneiss, because the gneiss, which evidently formed at medium grades of metamorphism (amphibolite facies) in earlier Precambrian time, has partly retrogressed in places, under the influence of low-grade metamorphism of late middle Precambrian age (p. 6). Epinorms and mesonorms are compared with modes (volume percent) representing the same samples. The lack of close agreement probably results from inherent defects in the arbitrary assignments of molecules during the calculation of the norm. The fact that the modes of Nos. 1 and 2 in table 10 are more nearly in agreement with the epinorms than with the mesonorms reflects the response of the rock to low-grade metamorphism. The mode of No. 3, however, is not very clearly correlative with either the epinorm or the mesonorm. The mode of No. 4 clearly indicates the amphibolite facies of metamorphism, so the mode is compared only with the mesonorm.

TABLE 8.—*Chemical analyses of Compeau Creek Gneiss, Marquette quadrangle*

[Nos. 1, 3, 6, and 7, rapid rock analyses. Nos. 2, 4, and 5, standard rock analyses, Dorothy F. Powers. Nos. 3 and 7, Cl and F contents, standard rock analyses, Vertie C. Smith]

	1	2	3	4	5	6	7
SiO_2	66.8	72.24	68.4	68.17	49.22	50.8	55.2
Al_2O_3	16.1	15.16	15.6	15.33	12.58	15.4	14.5
Fe_2O_3	1.7	.61	1.2	1.23	3.52	3.4	1.9
FeO	1.6	.54	1.4	1.26	12.89	6.4	4.2
MgO	1.8	.68	1.7	1.39	6.20	7.2	6.2
CaO	2.6	.81	2.1	3.34	7.87	8.2	6.8
Na_2O	5.4	5.70	5.4	4.79	2.53	3.8	3.8
K_2O	1.6	2.67	2.0	2.22	.61	1.6	1.6
H_2O^+	1.6	.44	1.4	.81	2.06	1.8	1.4
H_2O^-		.08		.17	.06		
TiO_2	.42	.18	.37	.37	1.67	1.2	.86
P_2O_5	.13	.06	.12	.12	.23	.32	.42
MnO	.08	.04	.04	.07	.26	.20	.12
CO_2	.30	.21	.16	.44	.02	<.05	<.05
Cl	------	.01	.02	.01	.01	------	.01
F	------	.02	.03	.04	.04	------	.10
SO_3	------	------	------	------	------	------	1.9
BaO	------	------	------	------	------	------	1.0

1. Tonalite gneiss, JG–33–59, 400N–1,280E, SW. cor. sec 9, T. 48 N., R. 25 W.
2. Tonalite gneiss, JG–55–59, 50N–4,780E, SW. cor. sec. 4, T. 48 N., R. 25W.
3. Tonalite gneiss, RT–35–59, 3,760N–540E, SW. cor. sec. 33, T. 49 N., R. 25 W.
4. Tonalite gneiss, JG–84–59, 4,120N–1,750E, SW. cor. sec. 32, T. 49 N., R. 25 W.
5. Quartz amphibolite, JG–60–59, 1,250N–1,300E, SW. cor. sec. 6, T. 48 N., R. 25 W.
6. Amphibolite, JG–78A–59, 4,140N–60W, SE. cor. sec. 36, T. 49 N., R. 26 W.
7. Amphibolite, RT–45–59, 820N–1,350E, SW. cor. sec. 19, T. 49 N., R. 25 W.

TABLE 9.—*Compositional basis for rock names of Compeau Creek Gneiss*

Ratio of hornblende (or biotite, chlorite) to total rock	>⅞ Plagioclase <⅛ Potassium-feldspar		⅔–⅞ Plagioclase ⅛–⅓ Potassium-feldspar	⅓–⅔ Plagioclase ⅛–⅔ Potassium-feldspar		<⅓ Plagioclase >⅔ Potassium-feldspar
	<10 percent quartz	>10 percent quartz	>10 percent quartz	<10 percent quartz	>10 percent quartz	>10 percent quartz
<⅒	Diorite	Tonalite	Granodiorite	Monzonite	Quartz monzonite	Granite.
>⅒ to <⅓	Hornblende diorite (or chlorite diorite).	Hornblende tonalite (or biotite tonalite or chlorite tonalite).	Hornblende granodiorite.			
>⅓ to <⅔	Amphibolite	Quartz amphibolite.				

TABLE 10.—*Comparison of normative and modal compositions, Compeau Creek Gneiss*

[E, epinorm; M, mesonorm; mode given as volume percent, from tables 6, and 7 (Nos. 1–3 here are Nos. 2–4 in table 6, and No. 4 here is No. 7 in table 7). Norms of Nos. 1–here are from analyses 1, 2, 4, and 5 respectively, table 8]

Mineral component	1			2			3			4	
	E	M	Mode	E	M	Mode	E	M	Mode	M	Mode
Quartz	22.6	23.1	14	25.1	25.3	26	23.7	23.8	28.5	2.8	12.5
Orthoclase	9.5	4.3	Tr	13.7	14.2	3	13.0	8.7	4.0	-------	2
Albite	49.0	49.0	[1] 67	51.5	51.5	[1] 64	43.5	43.5	[1] 53.5	24.0	30.5
Anorthite	-------	10.5	-------	-------	3.5	-------	-------	14.0	-------	-------	-------
Corundum	-------	-------	-------	-------	1.7	-------	-------	-------	-------	2.1	-------
Zoisite (epidote)	9.6	-------	5.5	3.2	-------	-------	12.8	-------	2.5	-------	1.5
Hornblende	-------	-------	-------	-------	-------	-------	-------	.75	5	50.3	48.5
Biotite	-------	8.3	-------	-------	2.9	-------	-------	6.9	2	6.4	-------
Muscovite (sericite)	-------	-------	4	3.2	-------	2	-------	-------	1.5	-------	-------
Chlorite [2]	6.2	-------	6	2.5	-------	4	2.3	-------	2	-------	-------
Chlorite [3]	.5	-------		-------	-------		2.7	-------		-------	-------
Carbonate	-------	-------	1.5	-------	-------	-------	-------	-------	Tr	-------	3
Hypersthene	-------	-------	-------	-------	-------	-------	-------	-------	-------	5.4	-------
Titanite	-------	.9	[4] 1	-------	.3	[4] Tr	-------	.9	[4] Tr	3.6	[4] 1
Ilmenite	.6	-------	-------	.2	-------	-------	.6	-------	-------	-------	-------
Magnetite	1.8	1.8	1	.6	.6	-------	1.0	1.0	-------	3.9	-------
Hematite	-------	-------		-------	-------	-------	-------	-------	.5	-------	Tr
Unidentified opaque	-------	-------	-------	-------	-------	1	-------	-------	-------	-------	1
Apatite	.3	.3	-------	Tr	Tr	-------	.3	.3	Tr	.5	Tr

[1] Sodic plagioclase, close to Ab_{90}.
[2] Amesite, $(Mg, Fe)_2 Al_2SiO_5 (OH)_4$ (see Barth, 1955, p. 351–352).
[3] Antigorite, $(Mg, Fe)_3 Si_2O_5 (OH)_4$ (see Barth, 1955, p. 351–352).
[4] Includes both titanite (sphene) and leucoxene.

The lighter colored gneiss resembles granite gneiss or foliated granite megascopically. It was called granite in the monographs and earlier studies, and, even though true granite is rare in the Compeau Creek Gneiss of the present area, it is convenient to use "granitic" collectively for the lighter colored quartz-feldspar gneiss when no specific composition is involved.

On dark, weathered, or lichen-covered outcrops, tonalite and granodiorite that are rich in biotite or hornblende generally cannot be readily distinguished from light-colored varieties of these rocks, so relationships between them are obscure in most outcrops. In places, however, the darker varieties appear to form crude layers or poorly defined schlieren, ranging from a few inches to probably no more than a few scores of feet in thickness, within or alternating with the lighter colored varieties. Darker gneiss layers commonly fray out into light-colored gneiss. Much of the gneiss within half a mile of the northern edge of the Marquette quadrangle is rich in hornblende (see modes of Nos. 7, 9, and 15, table 7). In that area, light-colored feldspathic rock forms seams, layers, and crosscutting veins in the darker gneiss, and migmatite is common.

The bulk of the rock in the outcrop south of Wetmore Landing (pl. 3) is typical light-colored foliated tonalite gneiss. The septum of hornblende tonalite in the south part of the outcrop has a saccharoidal texture and is only weakly foliated parallel to the walls of the septum. The septum of dark tonalite clearly cuts across the foliation of the gneiss and probably represents either an intrusion into the gneiss or an irregular zone of recrystallized gneiss.

Amphibolite, though only abundant in a few places, is widespread in the gneiss in both the Marquette and Sands quadrangles. Layers of amphibolite typically range in thickness from about 1 inch to several feet and alternate with generally thicker layers of tonalite, granodiorite, or quartz monzonite. Beds of amphibolite commonly fray out into light-colored rock. In places the layering of amphibolite is emphasized by thin seams of feldspar or feldspar-quartz within the darker layers. Offshoots from the "granitic" layers commonly cut the amphibolite. Several such offshoots that have been examined in thin section are quartz monzonite or monzonite (for example, see modes of Nos. 15 and 17, table 6), but the compositions of the layers from which the offshoots came have not been determined.

Light-colored gneiss and gneiss consisting of interlayered dark- and light-colored rock are cut by pegmatite, which is discussed in the following section on intrusive rocks. Where dark gneiss alone is cut by coarse-grained "granitic" rock, it is uncertain whether the "granitic" rock is of the age of the light-colored gneiss or of the later pegmatite.

The "granitic" gneiss is pale salmon colored or light gray, but on weathered surfaces typicaly is pink, orange brown, or dull gray; it is medium grained (most grains between 0.1 and 2 mm) and generally nonporphyritic. Tonalite gneiss or granodiorite gneiss rich in hornblende, biotite, or chlorite is greenish or mottled green and salmon colored, and on weathered surfaces is shades of gray or gray green and may show pinkish seams parallel to foliation. Amphibolite weathers to brownish green, commonly streaked by light-colored seams, or has a "salt and pepper" appearance.

The gneiss is more or less strongly foliated and jointed in virtually all outcrops. On outcrop surfaces of the light-colored gneiss, foliation is marked by discontinuous traces of shear surfaces along which small amounts of chlorite are common, and in the darker gneiss, it is marked by similar shears as well as by streaky "granitic" seams, or by alined chlorite, biotite, or hornblende. The texture of the light-colored feldspar-quartz gneiss is mainly granitoid, having some elongation of strained quartz grains parallel to foliation. This type of fabric in many places is modified by shears along which sericite, chlorite, or finely granular or elongated quartz, and dusty hematite or leucoxene are aggregated in seams or pods. In some of the gneiss granular epidote is strung out along shear seams. Plagioclase crystals generally are not elongated parallel to foliation, although in places they have been sheared into slices.

The preferred alinement of hornblende varies from strong to weak in the hornblende-rich gneiss, and the hornblende may be evenly distributed through the rock or concentrated in layers that alternate with layers of plagioclase and quartz or saussuritized plagioclase. Hornblende crystals generally form granoblastic mosaic intergrowths with other hornblende, with plagioclase, or with plagioclase and quartz. Hornblende in amphibolite generally has some preferred alinement of prisms parallel to the layering of the rock.

In the biotite-rich gneiss, well-alined flakes or books of biotite, with or without accompanying granular epidote or quartz, are generally concentrated in thin layers that alternate with layers rich in feldspar and quartz. Less commonly, biotite occupies thin seams between slightly elongated and well-alined feldspar and quartz and fills interstices at the ends of these grains.

Augenlike metacrysts of feldspar have grown adjacent to and out from foliation surfaces in some of the dark gneiss.

The dominant sodic plagioclase of the tonalitic and granodioritic gneisses is everywhere clouded by sericite, dusty hematite, chlorite, fine-grained epidote, or by saussuritic mixtures of these minerals. Saussuritized plagioclase is particularly common in the amphibolite. In the exceptionally quartz-rich gneiss in the northern part of the Sands quadrangle, plagioclase has been almost completely sericitized. The sodic plagioclase of the Compeau Creek Gneiss generally cannot be accurately identified because of clouding by alteration products and a lack of twinning in many grains. Indices of refraction of the plagioclase range from somewhat less than to slightly more than the index of refraction of balsam and so give the approximate composition.

In light-colored gneiss having foliated granitoid fabric, potassium-feldspar, mainly microcline, commonly is intergrown randomly with plagioclase and quartz. Generally the potassium-feldspar partly rims plagioclase or occurs in relatively small equant grains interstitial to the plagioclase. Such potassium-feldspar appears to have formed only a little later than the sodic plagioclase and approximately at the same time as the quartz. In some strongly foliated light-colored gneiss, microcline commonly is intergrown with quartz in quartzose seams, subparallel with foliation, and is virtually absent from intervening areas rich in sodic plagioclase. The quartz-microcline seams are clearly later than the sodic plagioclase, but whether they are merely the last products of crystallization, deposited along shear seams as the gneiss was emplaced, or a product of later activity is not known. Coarse crystals of microcline occur in places in the gneiss, and indeed many of the pegmatitelike blobs in the gneiss in the southwestern part of the Sands quadrangle are single large crystals of microcline with a little included (graphic?) quartz. Such crystals in amphibolite are clearly porphyroblasts, but whether they are porphyroblasts or phenocrysts in the light-colored gneiss cannot be determined without knowing the absolute ages of the large crystals and groundmass minerals. Potassium-feldspar in the exceptionally quartz-rich gneiss along the south margin of the Marquette synclinorium in the northern part of the Sands quadrangle generally is not associated with the patches and seams of quartz but is irregularly distributed within areas of sericitized plagioclase.

Horneblende of the gneisses is strongly pleochroic from pale yellow or pale yellow green in X to brownish green or grass green in Y to deep grass green in Z. The maximum extinction angle ($Z \wedge c$) is 17°–20°, and the optic angle is large and negative. Indices of refraction of horneblende from three samples are N_X 1.667–1.670, N_Y 1.686–1.690, and N_Z 1.690–1.693. Indices of refraction of hornblende from two other samples are N_X 1.648–1.649, N_Y 1.660–1.664, and N_Z 1.668–1.669. According to Tröger (1956, table 187, p. 77), these two ranges of data indicate that the proportions of Mg: (Fe^{++}, Mn, Ti) in the variable part of the horneblende formula are 40 : 60 and 70 : 30, respectively.

The light-colored gneiss that is exceptionally rich in quartz along the south side of the Marquette synclinorium (see modes of Nos. 6 and 7, table 6) probably corresponds to the Palmer Gneiss described by Van Hise and Bayley (1897, p. 210–218) and Van Hise and Leith (1911, p. 255–256) in the southern complex west of the present area. Megascopically the rock varies from a typical "granitic" gneiss that contains irregularly shaped quartz "seams" to a sheared rock that is similar to an arkose in appearance. Quartz generally forms wormy interconnected "grains" and stringers having highly irregular form but vague subparallel alinement. On weathered surfaces, quartz aggregates separated by anastomosing shears commonly have a clastic aspect, but in thin section the aggregates are seen to be more or less interconnected and show no evidence of detrital fabric. The textures of these rocks commonly differ from typical granitoid texture in that the quartz and feldspar grains tend to be segregated and interconnected with their own kind and not ingrown grain by grain. The formation of this rock during post-middle Precambrian shearing of "normal" tonalite gneiss is suggested by: (1) the textural features described above, (2) the relatively high quartz content but the otherwise typical tonalitic composition (allowing for sericitization of feldspar), (3) the extensive sericitization of plagioclase, and (4) proximity of most of this rock to the contact with overlying middle Precambrian rocks, along which shearing and faulting occured during post-middle Precambrian folding. Quartz in the gneiss at the begining of post-middle Precambrian deformation is believed to have been granulated, dissolved, and reprecipitated, and thereby concentrated in zones of greatest deformation.

In the southwestern part of the Sands quadrangle, particularly in the NE¼ sec. 7, T. 46 N., R. 25 W., zones of massive deep salmon-colored fine- to medium-grained tonalite, ranging from less than 1 foot to several feet thick, occur more or less conformably in strongly foliated lighter colored medium- to coarse-grained tonalite gneiss. In thin section, the rock is seen to have a weak foliation and a pronounced cataclastic fabric. Fragmental grains of quartz and feldspar of different sizes and shapes are surrounded by finely ground quartz and dusty hematite. Much finely ground quartz has recrystallized to chertlike quartz, and this rehealing undoubtedly accounts largely for the massive character of the fragmented rock. The rock has been deformed almost entirely by fragmentation, in contrast to the enclosing foliated tonalite gneiss in which the well-alined fabric formed by slippage along grain boundaries and recrystallization-elongation of quartz. In outcrop the deep salmon-colored tonalite appears to be of later age than the enclosing gneiss because of its more massive character and because very locally it truncates foliation of the gneiss at small angles. In places, light-gray foliated gneiss contains metacrysts (augen) of salmon-colored feldspar and frays out into the deep salmon-colored tonalite. The salmon-colored massive tonalite, while in a semicrystalline state, probably intruded the gneiss along foliation surfaces and was fragmented either during intrusion or subsequently.

QUARTZ VEINS AND SPATIALLY ASSOCIATED SILICIFIED ZONES

Quartz veins occur widely in the gneiss. In several places they are abundant and have a persistent trend. The age or ages of the quartz veins are unknown, although the veins in the outcrop south of Wetmore Landing (pl. 3) are at least of two ages. Most veins are less than 3 inches thick, but some are several feet thick. The thick veins are related to faults or other notable fracture zones in the gneiss. In one such fracture zone, extensive silicification of the gneiss to a chertlike rock preceded and possibly accompanied the emplacement of the quartz veins. Several other small areas of similar silicified rock are found in the gneiss. The locations of the silicified zones and the most prominent quartz veins are shown in figure 4 and by letter symbol on plates 1 and 2. Quartz veins in the gneiss adjacent to the fault between the gneiss and silicified Kona Dolomite in the SW¼ sec. 2, T. 47 N., R. 25 W., are discussed along with the silicification of the dolomite (p. 44) and are not discussed further here or indicated in figure 4. The two areas having the most conspicuous vein quartz are the one with spatially associated cherty rock mentioned above, located in sec. 1, T. 46 N., R. 26 W., and a zone along an east-trending fault in the south third of sec. 5 and SE¼SE¼ sec. 6, T. 48 N., R. 25 W.

The vein quartz in secs. 5 and 6 is mainly white and occurs irregularly along the line of the fault in a zone 50–100 feet wide. The quartz is generally massive and as much as 1 foot thick but in places is quite porous. Locally it is pink or has aggregates of reddish-brown or yellowish-brown iron oxide in drusy voids. The tonalite adjacent to this zone is strongly veined with quartz.

The belt of quartz-rich rock in sec. 1, T. 46 N., R. 26 W., extends along the top of the ridge across the central part of the section. The rock is mainly mottled whitish, pink, red-brown, yellow-brown, and greenish cherty

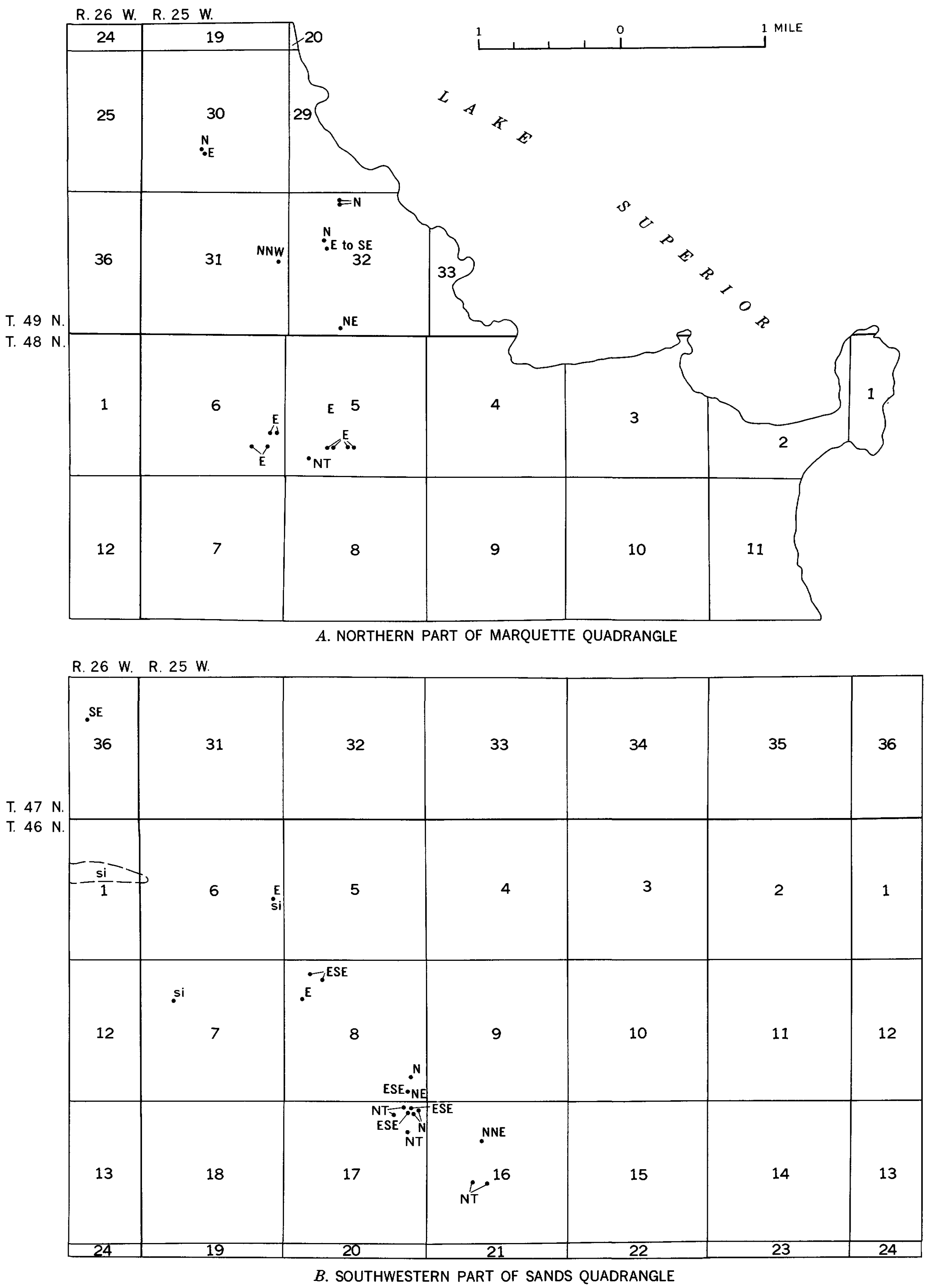

FIGURE 4.——Locations of silicified zones (si) and conspicuous quartz veins in Compeau Creek Gneiss. Trends of most dikes shown. NT indicates no trend evident.

quartz cut by white veins of coarser grained quartz. Near the east end of the ridge in sec. 1, fragments of salmon-colored feldspar and tonalite are abundant in the quartzose rock. A small outcrop of gneiss on the east edge of sec. 1 is cut by many thin quartz veins. Vehrs (1959) interpreted this quartzose rock as a pregranite quartzite that had been partly granitized. However, fragments of feldspar and gneissic country rock clearly "float" in the quartz rock, and abundant quartz veins at the east edge of sec. 1 are obviously intrusive into the gneiss. We therefore believe that the belt of quartzose rock marks an east-west zone of faulting or strong cataclasis in the gneiss along which silica was introduced and that silicification of much of the finely comminuted gneiss resulted. The mottled colors of much of the quartzose rock result from fine-grained intermixed chlorite, sericite, and reddish iron oxide. We believe these effects resulted from the silicification of parts of the ground-up gneiss in which there were concentrations of these differently colored substances. The white quartz veins probably developed late in the period of silica emplacement, but possibly came at some later time, after fracturing of the cherty silicified rock.

Tourmaline has been found in several veins between the south-central and the southwestern parts of the Sands quadrangle, and iron oxides occur in a few of the veins; otherwise the quartz appears to be without associated minerals. Some veins in the southwestern part of the Sands quadrangle are probably related genetically to pegmatite.

ORIGIN

The origin of the Compeau Creek Gneiss is obscure. Data collected and relationships observed during the present study shed some light on the problem but do not fully explain it. Sahakian (1959) concluded that a representative part of the gneiss that he studied in sec. 36, T. 47 N., R. 26 W., formed from a country rock of mafic Keewatin volcanic flows (Mona Schist?) that were metamorphosed to hornblende schist during the pre-Animikie Laurentian disturbance and were intruded by granite during the later pre-Animikie Algoman revolution to form banded gneiss. The gneiss was further metamorphosed and intruded by pegmatite and other granitic rock during the post-Animikie Killarney revolution. This interpretation is plausible and is not contradicted by available evidence, but enough is not yet known about the absolute ages of these rocks to relate these events to specific disturbances and revolutions.

The origin of at least some of the gneiss by the intrusion of felsic magma or solutions into the Mona Schist is indicated by several lines of evidence. Dikes of "granitic" rock, interpreted here as offshoots of the Compeau Creek Gneiss, intrude the Mona in the northern part of the main belt of greenstone and south of the east-plunging syncline in the Marquette quadrangle; "granitic" rock cuts dark layers in the gneiss or is intimately associated with such layers to form migmatite; and "granitic" gneiss carries isolated lenses of Mona-like greenschist, the foliation of which is conformable with that in the gneiss. These septa of schistose greenstone generally are not more than a few feet thick and 100 feet long, and some have been seen to fray out into the gneiss. They are widely distributed in the gneiss north of the principal belt of Mona Schist as shown on the map in figure 5 and in plate 1.

It is by no means certain, however, that all greenschist septa in the gneiss are of like origin. They may be remnants of mafic country rock that predated the gneiss, lenses of altered mafic intrusive rock, or schlieren segregated in the original magmatic part of the gneiss. The septa, except one in the NE¼ sec. 36, T. 49 N., R. 26 W., and one in the SW¼ sec. 19, T. 49 N., R. 25 W., are rich in chlorite and appear to be of low metamorphic grade, despite the fact that their gneissic environment is at the amphibolite level. This discrepancy in metamorphic grade between the gneiss and the included septa could be readily explained if the septa were postgneiss intrusive bodies but is more difficult to explain if the septa are of pregneissic age. Possibly a high water content helped to preserve typical low metamorphic mineral assemblages in remnants of pregneiss rock while the gneiss was being metamorphosed at the amphibolite level.

The largest greenschist septum, which is well exposed below the power dam in the channel of the Dead River in sec. 10, T. 48 N., R. 25 W., and which has been mapped westward from there for some 2,000 feet (pl. 1), appears to antedate the gneiss and therefore to be a remnant of mafic country rock (Mona Schist?). This septum is a chlorite-plagioclase-quartz-carbonate schist. Plagioclase and quartz form granoblastic intergrowths between flakes and anastomosing folia of chlorite. The carbonate is irregularly distributed in relatively large equant grains and aggregates, clearly later than the foliation of the schist. In a few narrow zones parallel to foliation there are abundant highly altered porphyroblasts, aggregates of quartz, carbonate, and clay(?) pseudomorphous after feldspar and transecting the foliation of the schist. These porphyroblasts indicate postfoliation feldspathization of the chlorite schist and are probably related to the regional emplacement of felsic material when the gneiss formed. A septum of chlorite schist in the gneiss in the west-central part of sec. 11, T. 48 N., R. 25 W., has blocky feldspar crystals which also lie astride foliation surfaces and appear to be porphyroblasts.

Septa of schistose greenstone at 3,900N–1,750E, SW. cor. sec. 6, T. 48 N., R. 25 W., and at 950N–4,650E, SW. cor. of sec. 31, T. 49 N., R. 25 W., are exposed within a few feet of crosscutting dikes of massive metadiabase, and almost certainly antedate the dikes. These septa could be either conformable mafic intrusive bodies of earlier age than the dikes or remnants of pregneiss country rock.

From these facts the following conclusions are drawn: (1) At least part of the present area of gneiss was once occupied by Mona Schist, and at least some of the septa of schistose greenstone in the gneiss are relics of the Mona Schist; the Mona therefore predates the present gneiss. (This latter part of the first conclusion accords with Williams, 1891, pp. 146–147, and with Van Hise and Bayley, 1897, pp. 150–151, 153, 170–171.) (2) A substantial part of the gneiss must have been some type of siliceous igneous material, the intrusion of which was spatially controlled by one or more types of preexisting layered or foliated country rock. The transection of dark layered gneiss, particularly amphibolite, by "granitic" rock shows that the gneiss comprises rocks of at least two ages and two origins. It is unknown whether the dark layered gneiss was derived from Mona Schist or from other pregneiss country rock.

Two offshoots of light-colored layers in the gneiss have been identified as quartz monzonite and monzonite (modes of Nos. 13 and 15, table 6). Foliated monzonite (mode of No. 14, table 6) has been identified from one place. Not enough is known of the distribution of these

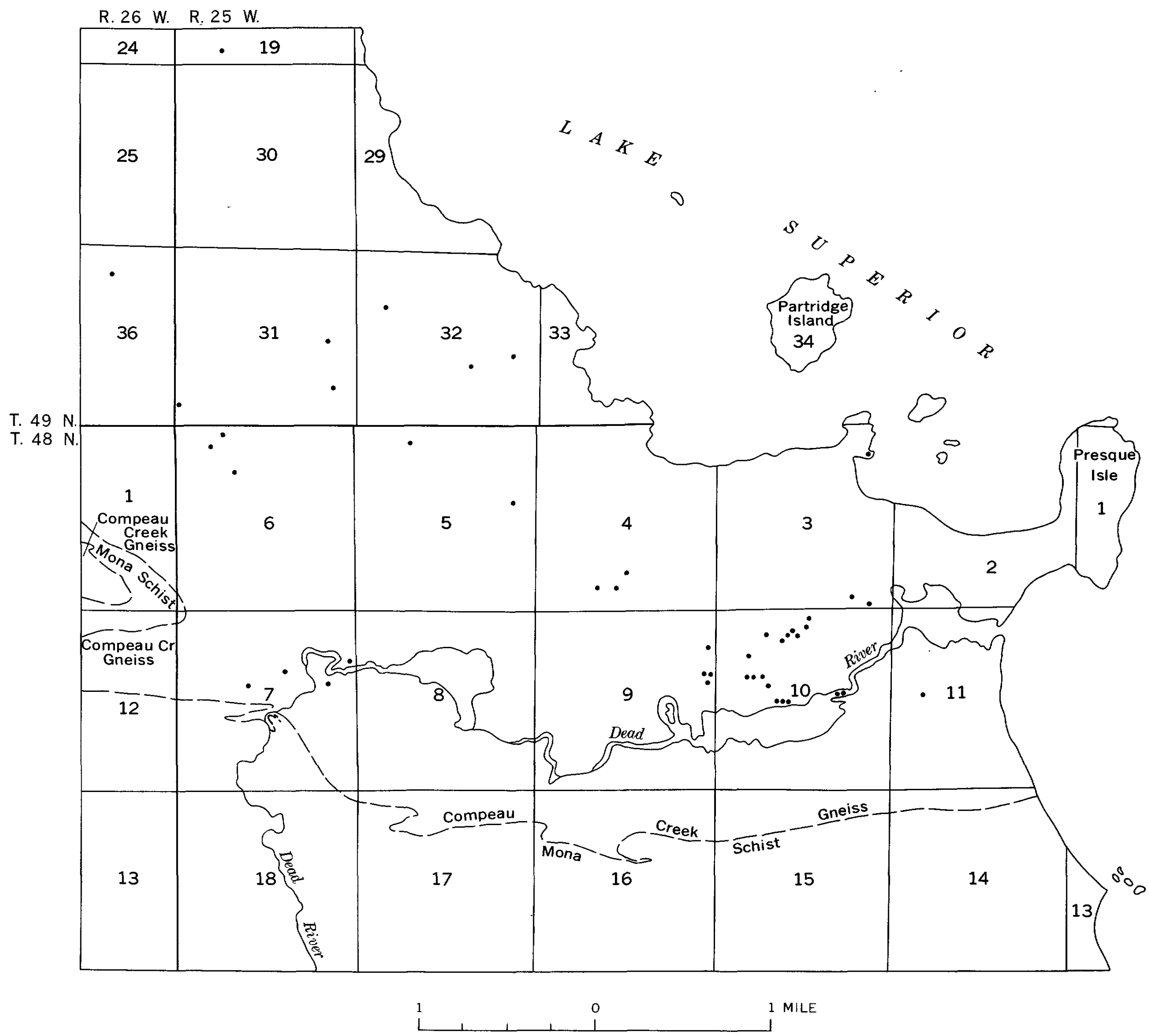

FIGURE 5.—Locations of septa of schistose greenstone (black dots) in Compeau Creek Gneiss, north half of Marquette quadrangle.

types of rock, or their relationships to the tonalite and granodiorite gneisses, however, to attempt to explain their significance.

METAMORPHISM

The association in the amphibolite of deep-green hornblende that has indices of refraction mainly between 1.650 and 1.693 and aggregates of sodic plagioclase and saussurite indicates partial retrograde metamorphism of the Compeau Creek Gneiss from the level of the amphibolite facies (see data for garnet zone, Wiseman, 1934, p. 380–386, and for amphibolite, Sugi, 1935, p. 129). The saussurite, an indicator of metamorphism below the level of the amphibolite facies, in association with sodic plagioclase suggests an originally more calcic composition of the plagioclase and therefore tends to substantiate the higher original metamorphic grade indicated by the hornblende. The association of saussurite and sodic plagioclase is taken as evidence of retrograde metamorphism to the greenschist facies from the level of the amphibolite facies. Chlorite commonly replaces hornblende or biotite on grain surfaces or along cleavages and almost certainly is a result of retrograde metamorphism, probably to chlorite grade, as is the universal clouding of sodic plagioclase by sericite or saussurite in the light-colored gneiss.

Biotite can be metamorphically cogenetic with hornblende (Turner, 1948, p. 78, fig. 17), so the biotite-rich gneiss is considered to be a product of metamorphism in the amphibolite facies. Where minor biotite is intergrown with hornblende, it is uncertain whether it was cogenetic with hornblende or an alteration product of the hornblende, developed below the level of the amphibolite facies (Turner, 1948, p. 91, fig. 24; p. 95, fig. 26).

It is evident that during the pre-Animikie emplacement of "granitic" material to form the present gneiss, most of the invaded country rock was metamorphosed to the level of the amphibolite facies. Later, almost certainly during the widespread orogeny and metamorphism of the post-Animikie, pre-Keweenawan interval, metamorphism near the eastern end of the Marquette range was at a level below that of the amphibolite facies, and some retrograde metamorphism occurred in the less stable minerals of the Compeau Creek Gneiss.

MIDDLE PRECAMBRIAN METASEDIMENTARY ROCKS—ANIMIKIE SERIES

Layered rocks of middle Precambrian age in northern Michigan have been correlated with the Animikie Group of northeastern Minnesota and adjacent Ontario by James (1958). Formerly these rocks in Michigan had been correlated with the Huronian rocks on the north shore of Lake Huron. James further divided the rocks of the Animikie Series into the Chocolay Group at the base, the Menomine Group, the Baraga Group, and the Paint River Group at the top. Of these, only rocks of the Chocolay and part of the Menominee Groups occur in the Marquette and Sands quadrangles.

CHOCOLAY GROUP

The Chocolay Group contains the oldest rocks of the Animikie Series and was defined by James (1958, p. 35) as equivalent to the lower Huronian of earlier reports. It was named for Mount Chocolay (not named on present-day topographic maps) and Chocolay Junction on the Duluth, South Shore and Atlantic Railroad, close to the southeast corner of the Marquette quadrangle, near which are extensive exposures of lower Animikie rocks. The Chocolay Group in the Marquette area consists of the Enchantment Lake Formation, overlain in conformable succession by the Mesnard Quartzite, the Kona Dolomite, and the Wewe Slate.

ENCHANTMENT LAKE FORMATION

DEFINITION, DISTRIBUTION, AND THICKNESS

The Enchantment Lake Formation is at the base of the Chocolay Group and the Animikie Series. The formation is probably discontinuous and lenticular, although this is not known with certainty, because in places it has been thinned by faults, and in areas where it appears to be absent, it may be faulted out or merely covered by talus from the overlying Mesnard Quartzite. It is here named for Enchantment Lake in secs. 29 and 32, T. 48 N., R. 25 W., and is best exposed at the type locality in sec. 29, a short distance north of the lake, and in scattered outcrops from one-half mile northwest to 1½ miles east of the lake. North and northwest of Enchantment Lake, the formation is roughly gradational from coarse congolmerate at the bottom through interbedded graywacke slate, wacke, arkose, and subarkose to sericite slate and sericitic quartzite at the top. The formation is also well exposed near the west end of the west-plunging anticline in the N½ sec. 3 and near the SW. cor. sec. 9, T. 47 N., R. 25 W. The maximum thickness is approximately 500 feet north of Enchantment Lake where the top and bottom of the formation are sharply defined.

ROCK NAMES AND PETROGRAPHY

The names of most rocks in the Enchantment Lake Formation are based on compositions and relative amounts of detritus and matrix. Although most matrix minerals are recrystallized and the rocks therefore are metamorphic, names of the corresponding sedimentary rock, such as conglomerate, arkose, graywacke, or wacke are used without the awkward prefix "meta" where no specific metamorphic names such as quartzite or slate are in general use. Definitions for graywacke, wacke, arkose, and subarkose, as used in this report, are generally modified somewhat from those given by Pettijohn (1949, p. 243–261; 1957a, p. 301–305, 322–331), and Gilbert (in Williams and others, 1954, p. 289–292, 297–304):

GRAYWACKE—Dark, poorly sorted rock containing sharply angular to subrounded detrital quartz or quartz and feldspar "floating" in a matrix consisting of chlorite, sericite, cherty quartz, reddish-brown iron oxide, leucoxene, and black opaque material. Detritus in some of the graywacke includes rock fragments, but they are not abundant. The matrix composes 15–85 percent of the rock (see fig. 6).

WACKE—Light-colored, poorly sorted rock containing sharply angular to subrounded detrital quartz "floating" in a matrix consisting mainly of sericite or of sericite and cherty quartz. The detritus in a few places includes substantial amounts of feldspar and small amounts of rock fragments. The matrix composes 15–85 percent of the rock (see fig. 7).

ARKOSE—Light-colored to grayish-salmon-colored moderately well sorted rock consisting mainly of subrounded detrital quartz and feldspar. Feldspar composes 25–75 percent and matrix minerals less than 15 percent of the rock.

SUBARKOSE—Feldspathic quartzite, moderately well sorted, consisting mainly of subrounded detrital quartz. Feldspar composes 5–25 percent and matrix minerals less than 15 percent of the rock.

Modes and averaged modes of typical rocks of the Enchantment Lake Formation, and grain sizes of nonmatrix detritus are shown in table 11. Sericite slate, quartzite, and conglomerate are not listed in the table.

The significant difference between graywacke and wacke, as used here, is in the color and composition of the matrix. Feldspar has been considered an essential or varietal constituent of graywacke by many authors (Fischer, 1934; Krynine, 1945; Pettijohn, 1949, p. 244–245, 255), but Pettijohn (1954; 1957a, p. 303–314) modified his earlier classification and, although recognizing that feldspar commonly is abundant in graywacke, he does not (1957a) isolate a variety of graywacke low in both feldspar and rock fragments—the subgraywacke of his earlier classification or the low-rank graywacke of Krynine (1945). Furthermore, he does not (1957a) distinguish between detrital quartz and feldspar grains, but groups them together in order to contrast them with the amount of matrix, as a measure of sorting. In graywacke and wacke of the Enchantment Lake Formation, feldspar exceeds rock fragments, but some of the graywacke and most of the wacke, nonetheless, have only small amounts of feldspar (both sodic plagioclase and microcline). In this report, no formal distinction is made in type of graywacke or wacke on the basis of the feldspar/quartz ratio. In some of the following description, though, feldspar-poor and feldspar-rich graywackes and wackes are indicated.

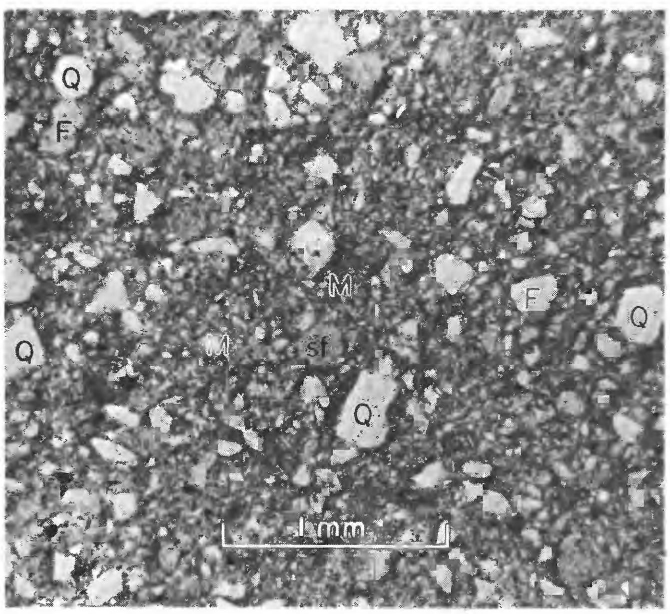

FIGURE 6.—Photomicrograph of graywacke from Enchantment Lake Formation, 1,100N–3,250E, SW. cor. sec. 29, T. 48 N., R. 25 W. Detritus of quartz (Q), feldspar (F), and sericitized feldspar (sf) "floats" in matrix (M) of sericite, chlorite, and fine-grained quartz. Plane light.

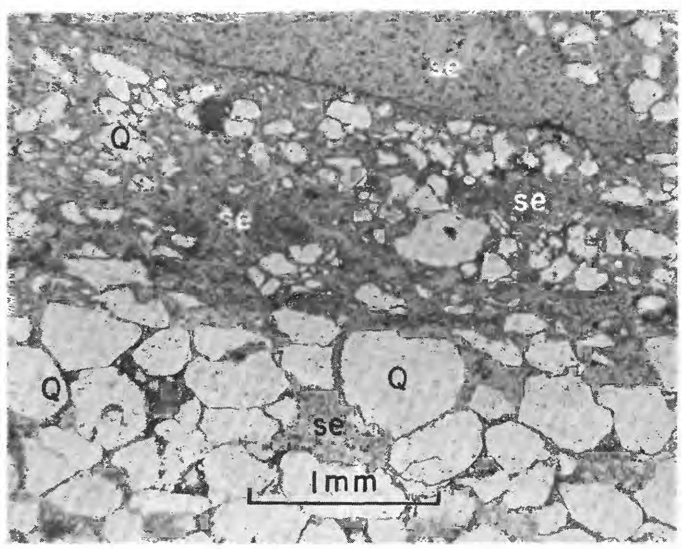

FIGURE 7.—Photomicrograph of quartz wacke and interlayered sericite slate (in upper part of photograph), Enchantment Lake Formation, 700N–3,050E, SW. cor. sec. 29, T. 48 N., R. 25 W. Detrital quartz (Q) in central part of photograph "floats" in abundant matrix consisting mainly of sericite (se). Plane light.

TABLE 11.—*Modes and averaged modes, in volume percent, and grain sizes, Enchantment Lake Formation, Marquette quadrangle*

[Range of modes for each component shown in parentheses]

	1	2	3	4	5	6	7	8	9	10
Quartz detritus	19.5	48	17	80.5	49	22	15.5	43	83	42
	(10–45)	(45–51)	(16–18)	(74–84.5)	------	(14–30)	------	(38–48)	------	------
Feldspar detritus	14	3	14	14.6	41	3.[illegible]	10	1.5	9	54
	(1–24)	(Tr– 9.5)	(14–14)	(11–18)	------	(3–4)	------	(1–2)	------	------
Rock fragments [1]	.5	2	1	.4	------	2.5	8	1.7	------	------
	(0– 2)	(Tr– 3.5)	(0.5–1.5)	(0–0.5)	------	(1–[illegible])	------	(0.5–3)	------	------
Matrix [2]	64	47.5	67	4.2	10	71.5	65	53.2	8	[3]4
	(44–83.5)	(38.5–52.5)	(66–68)	(1.5–7)	------	(63–80)	------	(46–60.5)	------	------
Carbonate [4]	.5	------	------	------	------	------	------	------	------	------
	(0– 9.5)	------	------	------	------	------	------	------	------	------
Opaque [5]	.5	------	.75	------	------	.5	1	------	------	------
	(0– 2.5)	------	(0–1.5)	------	------	(0–1)	------	------	------	------
Grain size, detritus range in mm [6]	0.01– 0.5	0.05–0.60	0.015–0.12	0.05–3.5	0.07–0.2	0.05–1	0.03–0.2	0.06–0.27	0.5–3	0.06–0.2

[1] Includes clastic chert, sericite rock, "granitic" rock, and possible mafic metavolcanic rock.
[2] Generally includes some fine-grained detritus and dusty iron oxides; in graywacke, in addition, a mixture of chlorite, sericite, chert, and leucoxene; in wacke, mainly sericite and chert.
[3] Almost entirely chlorite.
[4] Replaces parts of matrix.
[5] Most is magnetite.
[6] Does not refer to fine-size detritus in matrix.

1. Averaged mode, 9 samples of graywacke; middle part of formation, Enchantment Lake area.
2. Averaged mode, 4 samples of quartz wacke; upper part of formation, Enchantment Lake area.
3. Averaged mode, 2 samples of quartz-feldspar wacke; upper part of formation, Enchantment Lake area.
4. Averaged mode, 4 samples of subarkose; middle part of formation, NW¼ sec. 34, T. 48 N., R. 25 W., and NE¼ sec. 3, T. 47 N., R. 25 W.
5. Arkose; middle part of formation, NW¼ sec. 34, T. 48 N., R. 25 W.
6. Averaged mode, 2 samples of quartz wacke; middle part of formation, NE¼NE¼ sec. 33, T. 48 N., R. 25 W., and NE¼NE¼ sec. 3, T. 47 N., R. 25 W.
7. Quartz-feldspar wacke; middle part of formation, NE¼NE¼ sec. 33, T. 48 N., R. 25 W.
8. Averaged mode, 2 samples of quartz wacke; upper part of formation, SW¼SW¼ sec. 28, and NW¼NW¼ sec. 34, T. 48 N., R. 25 W.
9. Subarkose; lower part of formation, NW¼NE¼ sec. 3, T. 47 N., R. 25 W.
10. Arkose; lower part of formation, SE¼NW¼ sec. 1, T. 47 N., R. 25 W.

Arkose and subarkose, as defined above, differ from graywacke and quartz-feldspar wacke mainly in containing more detrital grains and less matrix material. Detrital grains typically are in contact with one another, quartz grains may be cemented by silica overgrowths, and the matrix forms thin seams between detrital grains. Thus, the sorting of the arkose and subarkose is better than that of the graywacke and wacke. Arkose has a higher feldspar/quartz ratio than most graywacke of the area. Feldspars are both sodic plagioclase and microcline. The matrix of some arkose and subarkose contains chlorite, but generally the matrix is mainly sericite or sericite and fine-grained quartz, with some fine-grained feldspar.

Quartzite is red or reddish gray, occurs in beds less than 6 inches thick, and generally has granoblastic texture and small amounts of interstitial sericite and dusty hematite.

Sericite slate is gray, buff, or pale olive green. It consists largely of sericite, but commonly also has small amounts of hematite, and leucoxene.

LITHOLOGIC DISTRIBUTION

The Enchantment Lake rocks can be accurately assigned to the lower, middle, or upper parts of the formation in two places: north of Enchantment Lake and in the N½ sec. 3, T. 47 N., R. 25 W. The formation in those areas is delimited by directly adjacent outcrops of underlying and overlying formations. In several other places, rocks can be assigned either to the basal or the upper part of the formation on the basis of proximity to the corresponding contact plus sufficiently varied lithologies to permit an approximate correlation with the well-delimited sections. Where outcrops of the Enchantment Lake Formation are not bordered by exposures of either underlying or overlying rocks, the assignment to the lower, middle, or upper part of the formation is made by analogy with the lithology of the two delimited sections. The recurrence of most lithologic types in the section makes the assignment of a rock to a given part of the formation impractical in areas where only one type is exposed. Evidence of stratigraphic thinning indicates that the formation is probably lenticular, and most of the rocks of the type section probably lens out or change facies along strike. Thus a pebbly conglomerate close to outcrops of overlying Mesnard Quartzite in one place may actually be at the base of the Enchantment Lake Formation.

In the type area the lower part of the formation consists of conglomerate, the middle part consists of graywacke, wacke, arkose, and subarkose, and the upper part contains sericite slate, wacke, and quartzite. Rocks in other parts of the area are compared below with the type section, from bottom to top.

The distribution of 115 samples of the formation by type of rock and position in the formation is given in the following table. The table indicates the more abundant rock types exposed in the area, even though the sampling undoubtedly does not show accurately the relative amounts of the different types. The table also shows the varied lithology of the lower, middle, and upper parts of the formation.

Distribution of 115 samples of Enchantment Lake Formation by rock type and position in the formation

	Lower part	Middle part	Upper part
Conglomerate (most have composition of graywacke or wacke)	6	1	0
Graywacke (slaty and massive)	0	16	0
Wacke	3	8	31
Arkose	1	6	0
Subarkose	2	12	1
Quartz-sericite slate	1	2	2
Sericite slate	0	8	5
Chlorite slate	0	1	0
Sericitic quartzite	0	0	1
Quartzite	0	3	5
Total samples collected	13	57	45

LOWER PART OF FORMATION

Enchantment Lake area.—The base of the Enchantment Lake section consists of conglomerate that strikes eastward and dips vertically, the contact between it and the greenstone basement to the north being exposed in several places in the vicinity of 1,300N–2,000E, S.W. cor. sec. 29 (Old State Road or Mud Lake locality of Irving, *in* Williams, 1891, p. 21 and Van Hise and Bayley, 1897, p. 223, 232). Not more than 20 feet of conglomerate is exposed across the strike in any single outcrop, and the total thickness of this rock in the type section is probably not greater than 50 feet.

The conglomerate is a greenish-gray rock in which angular to subrounded boulders and smaller fragments of pink or salmon-colored "granitic" rock are mixed with cobbles and smaller fragments of greenstone and white vein quartz and with coarse angular grains of salmon-colored feldspar (fig. 8). The largest boulder of "granitic" rock observed is about 1.5 feet in maximum diameter. The fragments are surrounded by a greenish fine- to medium-grained, more or less foliated matrix, rich in chlorite, sericite, carbonate, and quartz. Feldspar, leucoxene, magnetite, biotite, sulfide, rutile, apatite, and reddish iron oxide occur in the matrix in minor or trace amounts.

Other areas.—A vertical bed of conglomerate about 4 feet thick is at the base of the exposed section of Enchantment Lake Formation in the NE¼ sec. 33, T. 48N., R. 25 W., but in a similar position, a short distance eastward in the NW¼ sec. 34, 5 feet of similar conglomerate is underlain by 30 feet of gray subarkose. Presumably these rocks in secs. 33 and 34 are at or near the bottom of the formation, but their exact stratigraphic position in respect to beds in the Enchantment Lake section is not known.

The conglomerate is mottled bluish black and tan or pale olive green. Platy or irregular-shaped angular bluish-black hematite-rich fragments as much as 3 inches across are mixed with pieces of vein quartz, chert, jasper, sericite rock, and probable metavolcanic rock (fig. 9). The matrix is generally tan or pale olive green and consists mainly of sericite and fine- to medium-grained clastic chert and quartz. Feldspar, magnetite,

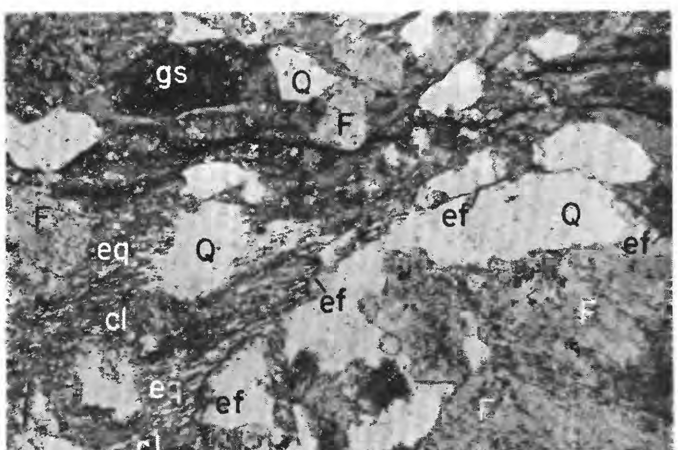

FIGURE 8.—Photomicrograph of conglomerate with composition of graywacke, Enchantment Lake Formation, 1,300N–1,850E, SW. cor., sec. 29, T. 48 N., R. 25 W. Grains of detrital quartz (Q), feldspar (F), and greenstone (gs), and pebble-size fragment of tonalite (ef marks the edge of the fragment) set in matrix of chlorite (cl), and thin elongated aggregates of fine-grained quartz (eq). Plane light.

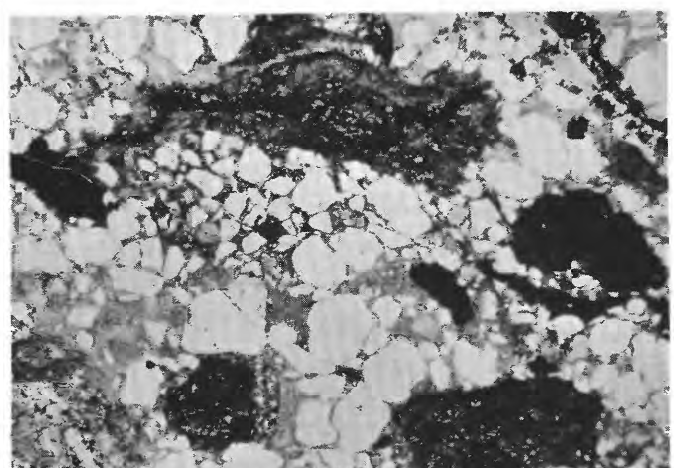

FIGURE 9.—Photomicrograph of conglomerate, Enchantment Lake Formation, 4,500N–150E. SW. cor., sec. 34, T. 48 N., R. 25 W. Coarse fragments of dark hematite-rich material set in matrix consisting mainly of sericite and medium-size detrital quartz. Plane light.

hematite, and leucoxene occur in small amounts. Fragments of probable metavolcanic rock are generally tan or reddish and consist of fine-grained intergrown quartz, feldspar, and sericite, and scattered magnetite, hematite, and dusty leucoxene(?). The dark pieces of ferruginous rock are relatively abundant, and their presence at the base of the Animikie section is particularly interesting because of their resemblance to iron-formation. An indication of the probable origin of this material is given in the same conglomerate by some fragments of metavolcanic(?) rock that contain seams, veins, and patches of hematite. One such fragment has subparallel tabular seams of hematite cut by cross fractures filled with quartz. The hematite, therefore, is probably of pre-Animikie hydrothermal origin. Weathering and erosion of such rock veined by hematite could have supplied tabular pieces of hematite to the conglomerate.

The subarkose beneath the conglomerate in the NW¼ sec. 34 contains abundant clastic quartz and small amounts of clastic chert and feldspar, loosely packed or locally "floating" in a sericite matrix. Hence locally the subarkose grades to wacke. Geothitic iron oxide, magnetite, leucoxene, and tourmaline exist in small or trace amounts.

Nearly all exposures of the Enchantment Lake Formation near the SW. cor. sec. 9, T. 47 N., R. 25 W., consists of salmon-colored conglomerate in proximity to underlying "granitic" gneiss. A few readings of strike and dip and the several reentrants of conglomerate into the area of gneiss indicate that trends follow the edge of the area of gneiss and the conglomerate forms roughly a westward- to southwestward- to south-facing dip slope. The number of measurable dips, however, is insufficient for determining thickness. Cobbles and smaller fragments of salmon-colored "granitic" rock and white vein quartz are very abundant; pieces of greenschist and jaspery chert are not common. The matrix contains angular and subangular medium- to coarse-grained feldspar and quartz. Thus the conglomerate is arkosic.

At 250N–300E, SW. cor. sec. 9, a small exposure of vitreous fine-grained salmon-colored arkose lies between conglomerate and gneiss. Southeastward from this point, in sec. 16, most exposures of the Enchantment Lake Formation are near exposures of gneiss and consist of fine-grained pink or salmon-colored feldspathic quartzite or arkose and some scattered quartz pebbles. Southeastward from the SW. cor. sec. 9, the conglomerate therefore either has lensed out or is very thin and not exposed.

Pink, salmon-colored, or brick-red arkose and feldspathic quartzite similar to that characterizing the basal part of the Enchantment Lake Formation in the NW¼ NW¼ sec. 16, T. 47 N., R. 25 W., occurs on the north flank of the uplift of Compeau Creek Gneiss in the NW¼ sec. 9, borders the gneiss in the SW¼SW¼ sec. 4, and occurs within 50 feet of Mona greenstone in the NW¼ NE¼ sec. 3, T. 47 N., R. 25 W.

The poorly exposed Enchantment Lake rocks about 1,300 feet north of the center of sec. 1, T, 47 N., R. 25 W., in the Marquette quadrangle, are apparently somewhat infolded with underlying Mona Schist and are strongly deformed, recrystallized, and silicified. At the base of the formation there, cherty chloritic slate containing abundant irregular pods of secondary quartz and thin quartz veins is probably not more than 10 feet thick and is overlain by fine-grained salmon-colored arkose which is probably not more than 20 feet thick. The arkose has a granoblastic fabric, typically with uniformly intergrown equant quartz and feldspar. Locally there are thin quartzitic zones and isolated coarse quartz grains in the arkose.

MIDDLE PART OF FORMATION

Enchantment Lake Area.—The middle part of the formation in the type area is best exposed in three long narrow east-trending outcrops in sec. 29, approximately centered at 1,050N–700E, 1,000N–2,200E, and 1,050N–2,800E from the SW. cor. Most of the rock in these outcrops is fine-grained, schistose to slaty, and dark-gray or greenish-gray graywacke. Graywacke and wacke are interbedded, with some alternations; because of weathering and fine grain size, they are not readily distinguished in the field. The west outcrop consists mainly of fine-grained feldspar-poor graywacke slate in which there are some irregularly and widely spaced angular fragments of pink and gray-salmon-colored medium-grained arkose. These fragments range in size from about 1 to 8 inches. Some are roughly tabular, and all are probably derived from nearby beds and are probably analogous in origin to fragments in intraformational conglomerate. The upward succession and corresponding thickness in the central outcrop are: greenish-gray feldspar-poor graywacke slate, 35–40 feet; grayish feldspar-poor graywacke slate, 25 feet; gray-salmon-colored quartz-carbonate wacke, 10 feet; grayish feldspar-poor graywacke slate containing considerable magnetite, 15–20 feet; and gray-salmon-colored feldspar-rich graywacke, 2–3 feet. Near the east end of this outcrop a lens of carbonate-rich rock, probably not more than 30 feet long and 10 feet thick, is interbedded with the graywacke. Whether this rock is a sedimentary facies of the formation or a vein is not known. The east outcrop is mainly uniformly gray-green feldspar-rich graywacke slate having some interbedded feldspar-poor graywacke slate and thin nonslaty feldspar-rich graywacke of the same color and a thin lens of gray-salmon-colored feldspar-rich graywacke bounded by thin layers of fissile green (chlorite?) slate. By projection along strike, the dominant graywacke of the east outcrop appears to underlie graywacke and wacke of the central and west outcrops.

Medium- to coarse-grained light-gray and speckled salmon-colored arkose and subarkose form a small outcrop immediately south of the west outcrop mentioned above, and are probably correlative with similar rock to the east in secs. 33 and 34; they pass south of the central outcrop and north of the outcrops on the north shore of Enchantment Lake.

The greatest thickness across any one outcrop is 110 feet, and the approximate total thickness of rocks in the middle part of the formation in the type area is 400 feet.

Evidently because of the repetition of facies combined with virtually uniform dips across the section of graywacke and associated rocks north of Enchantment Lake, Van Hise and Bayley (1897, p. 233) thought that these rocks are repeated by isoclinal folding. Bedding in these strongly cleaved rocks is obscure, probably in part because it nearly parallels the steeply dipping cleavage. However, no indication was found immediately north of Enchantment Lake that the upper and lower contact or beds of the formation were folded at a lesser scale than that of the major folding of the area; therefore, although the middle part of the formation may have undetected small-scale drag folds, we believe that the repetition of facies is a depositional characteristic.

Other areas.—Graywacke is virtually missing outside the type area. A small amount of feldspar-poor graywacke occurs in the western part of the outcrop area in the NE¼ sec. 33, T. 48 N., R. 25 W, but none is found in the principal exposures farther east in the section and in the NW¼ sec. 34. The rocks there are mainly interbedded subarkose, arkose, wacke, and sericite slate, containing some thin beds of conglomerate and ferruginous quartzite. Some of these undoubtedly are the graywacke of Van Hise and Bayley (1897, p. 234). Several beds of buff-greenish-gray feldspathic quartz-pebble conglomerate, each about 10 inches thick, are interbedded with subarkose and arkose about 20 feet above the bluish basal conglomerate and 100 feet below the Mesnard Quartzite in the NE¼NE¼ sec. 33. The arkose strongly resembles the rock in the middle part of the formation in the type section, which crops out immediately south of the graywacke slate in the SW¼SW¼ sec. 29. The thickness of the Enchantment Lake Formation in secs. 33 and 34, between the base of the Mesnard Quartzite and the base of the bluish conglomerate, ranges from 120 to 300 feet, and rocks in the middle part of the formation are roughly 80–200 feet thick.

Some of the best exposures of the middle part of the formation occur on the south limb of the west-plunging anticline in the N½ sec. 3, T. 47 N., R. 25 W., where the rocks are buff or gray quartz-rich wacke and subarkose, buff or pale-olive-green sericite slate, and pale-gray to reddish quartzite and sericitic quartzite. Beds of wacke, subarkose, and quartzite typically are 1–4 inches thick and are separated by thicker layers of sericite slate. These rocks are poor in feldspar and nonchloritic.

A small outcrop of feldspathic quartzite or subarkose in the SW¼SE¼ sec. 4, T. 47 N., R. 25 W., is probably in the middle part of the formation.

UPPER PART OF FORMATION

Enchantment Lake area.—The upper part of the formation in the type area is exposed on the north side of two outcrops of Mesnard Quartzite, in the SW¼SE¼ sec. 29, and 150 feet northeast of the eastern outcrop of Mesnard. It is mainly light-gray to buff wacke, commonly somewhat slaty. About 60–75 percent of the rock is detrital quartz; most of the rest is sericite; detrital chert, yellow-brown iron oxide, rutile, zircon, and sphene or leucoxene generally occur in minor or trace amounts. This rock is interbedded with fissile buff to pale-olive-green sericite slate and thin quartzite. Thin seams of detrital quartz occur in the slate. The maximum exposed thickness of this unit in the type area is approximately 55 feet.

Other areas.—East and southeast of the type area, the upper part of the formation consists of wacke or sericite slate similar to that in the type area, or of interbedded wacke and slate. Exposures occur adjacent to Mesnard Quartzite near the SW. cor. sec. 28, in the NE¼ sec. 33, and in the NW¼ sec. 34, T. 48 N., R. 25 W., in the Harvey quarry, and near the W¼ cor. sec. 1, in the north-central and northwest parts and the SE¼ sec. 2, in the north-central part of sec. 3, and near the center of the SE¼ sec. 4, T. 47 N., R. 25 W. Less than 10 feet of the rock is exposed in most outcrops. The maximum exposed thickness is approximately 60 feet near the SW. cor. sec. 28.

CORRELATION AND AGE RELATIONS

The distinctive rocks of the Enchantment Lake Formation were recognized in the type section and in several other places during early work in the area, but were mapped as the lower part of the Mesnard Quartzite (Van Hise and Bayley, 1897, p. 223–231, 232–236, 238–240; Van Hise and Leith, 1911, p. 254–255, pl. 17; Leith and others, 1935, pl. 1). Van Hise and Leith, however, hesitated at placing these rocks with the Mesnard. They pointed out lithologic similarities between the conglomerate near Enchantment Lake (Mud Lake at that time) and parts of the "pre-Huronian" Kitchi Schist, north of Ishpeming, some 12 miles west of Enchantment Lake. They stated that if the rocks between the vitreous Mesnard Quartzite and the greenstone basement near Enchantment Lake could have been mapped without reference to district-wide considerations the rocks probably would have been shown as a distinct sedimentary unit, unconforamble above the Mona Schist and unconformable below the Mesnard Quartzite.

W. A. Seaman (in Snelgrove and others, 1944, p. 11) correlated the conglomerate and associated rocks of the Enchantment Lake area with the Kitchi Schist and postulated an unconformity between the so-called Kitchi and the Mesnard in the Enchantment Lake area. This postulated unconformity evidently was an eastward projection of the known great unconformity between the type Kitchi and "Huronian" rock (the Ajibik Quartzite) north of Ishpeming.

The Mesnard Quartzite has been considered a direct correlative of the Sturgeon Quartzite of Iron and Dickinson Counties, Mich., for more than 60 years and continues to be so accepted by the U.S. Geological Survey (James, 1958, p. 35). Pettijohn (1943a) described arkosic and graywacke conglomerates and associated slate, argillite, and sericitic quartzite at the base of the Sturgeon Quartzite from several localities bordering the southern part of the southern complex in the Menominee and Calumet iron-bearing districts of Dickinson County. At Fern Creek, near Norway, Mich., he recognized tillite and associated argillite containing ice-rafted cobbles overlying arkosic basal conglomerate. This conglomerate in turn overlies granitic basement rocks. He proposed the name Fern Creek Formation for these rocks of evident glacial association and all beds between them and the granitic basement.

Trow (1948) extended the Fern Creek Formation to include all the beds of conglomerate, slate, and impure quartzite lying between pure vitreous Sturgeon Quartzite and the "Archean" basement gneiss in the Menominee district. He found tillite only in the previously defined Fern Creek Formation but thought that the other units were related to glaciation largely as lacustrine, alluvial, and fluviatile outwash deposits. Trow's Fern Creek section has a maximum thickness of 260 feet.

Both Pettijohn (1943a, p. 396) and Trow (1948) mentioned the probability of a correlation between the sub-Sturgeon conglomerate and associated rocks in Dickinson County and the lower Mesnard conglomerate and associated rocks described by Van Hise and Bayley in the Marquette area.

Brooke (1951) considered the basal metasedimentary rocks north of Enchantment Lake to be a "sub-Huronian" unit, possibly correlative with the Fern Creek Formation. As had Trow in the Menominee district, Brooke postulated an erosional interval after the deposition of the basal metasedimentary rocks and prior to deposition of the overlying vitreous (Mesnard) quartzite. Brooke's evidence for an erosional interval was the marked lithologic difference between the Mesnard and the "sub-Huronian" metasedimentary rock and some small discrepancies in strike between them.

The Enchantment Lake Formation has a strong lithologic similarity to the Fern Creek Formation of the Menominee district and is here correlated with the Fern Creek, both for that reason and because of the analogous stratigraphic positions of the two formations, conformable or virtually so beneath quartzite and markedly unconformable on basement rocks. No objective evidence has been found in the Marquette area for an erosional interval between the Enchantment Lake and the Mesnard, evidence such as differences in strikes and dips of adjacent parts of the two formations, or pebbles of the Enchantment Lake Formation in Mesnard Quartzite. On the contrary, an increase in quartzitic and sericite slate interbeds, and somewhat better sorted wacke in the upper part of the formation as compared to the middle and lower parts, may represent a lithologic gradation between the two formations. The Enchantment Lake and Fern Creek rocks clearly parallel the course of the lower Animikie Mesnard and Sturgeon Quartzites and truncate underlying basement rocks. Because of this relationship, correlation of the Enchantment Lake and Fern Creek rocks with the Kitchi of the Ishpeming area would mean that the lower part of the Animikie section over a wide area northeast and northwest of Ishpeming had been eroded just down to the level of the basal part of the Animikie, following which the Ajibik Quartzite was deposited. The unlikelihood of such a precise erosional base level and (1) the fact that the reported lithologic similarities between the Enchantment Lake Formation and the Kitchi are gross ones and are not evident in detail and (2) the lack of evidence that the Kitchi of the Ishpeming area unconformably overlies the Mona Schist indicate that the Enchantment Lake Formation and the Kitchi Schist are not correlative.

Because of the equivalence of the Enchantment Lake and the Fern Creek rocks and because of the apparent lack of an erosional interval between the Enchantment Lake Formation and the Mesnard Quartzite in the Marquette area, the Enchantment Lake Formation is assigned to the base of the Chocolay Group in the Animikie Series.

ORIGIN

The Enchantment Lake Formation probably originated on an initial seaward slope of a newly forming depositional basin at the beginning of Animikie time, or on the front slope of a delta expanding into such a basin. In such an environment, rapid deposition (from turbid media, submarine slumping, mudflows, or landslides) and alternating transitions from shallow marine foreset to shallow-water or subaerial topset beds could occur. Such alternations are common in modern deltas (Shepard and others, 1960). The Enchantment Lake sediments probably were not transported great distances by running water, and their transport was probably not preceded by long weathering. The poorly sorted sediments were probably deposited below the depth of wave action but probably in water sufficiently shallow that from time to time the top of the poorly sorted pile was encroached upon by thin lenses of fairly well sorted arkosic or quartzose detritus. Poor sorting and an absence of graded bedding may be due in part to submarine mudflows, according to Crowell (1957). The above interpretation is based on the following considerations:

1. Boulders of "granitic" rock at the base of the formation in the type section are "foreign," and have been transported either from north of the main belt of Mona Schist, a distance of several miles, or from the south, an unknown distance along the present steeply dipping (folded) erosion surface at the base of the Animikie.
2. In the middle part of the formation, graywacke and wacke, representing "poured-in" types of sediment (Fischer, 1934; Pettijohn, 1943b, p. 956; 1957a, p. 313) are interlayered with arkose, a rock with much better size sorting and one generally attributed to a different depositional environment than graywacke (Pettijohn, 1957a, p. 328–330). Furthermore, in the middle part of the formation, "poured-in" wacke is interlayered with texturally and compositionally well sorted sericite slate and quartzite, and arkose and subarkose are interlayered with thin conglomerate, wacke, and sericite slate.
3. Graywacke is in thin deposits, does not have graded bedding, and is not associated with mafic volcanic rocks, in contrast to typical graywacke (Pettijohn, 1943b, p. 954–956; 1957a, p. 312–314).
4. Graywacke and wacke contain angular to subrounded clastic grains of fine size (mainly <0.4 mm), but many times coarser than matrix grains.
5. The quartz-rich wacke and sericite slate in the upper part of the formation are better sorted than are the rocks in the basal and middle parts, and some of the wacke in the upper part of the formation has internal laminations resulting from varying concentrations of quartz along bedding.
6. The abundant "clay paste" [1] matrix of the graywacke and wacke, studded with markedly coarser detritus, signifies rapid deposition from a turbid medium (Kuenen and Migliorini, 1950; Pettijohn, 1957a, p. 312). However, the lack of graded bedding, a typical product of turbidity currents "discharging" into deep-water basins, and the alternation of graywacke or wacke with thin lenses of fairly well sorted medium-grained arkose or quartzite suggest a shallow-water environment.
7. Heterogeneous mixtures of sediments such as conglomerate, sandstone, siltstone, shale, arenite, and graywacke, in thin layers, have been attributed to deltaic deposition by Krynine (1940).
8. Submarine slumping and landslides are common at relatively shallow depths on sea bottoms with gradients of as little as 1°, as noted by Shepard on the foreset slope of the Mississippi delta (Shepard and others, 1960, p. 60). Shepard found that valleys in the foreset slope, probably caused by submarine landslides, die out at depths of less than 200 feet. If submarine landslides or mudflows come to rest within a short distance of their starting points, still on the seaward slope, settling of their heterogeneous sediments without the development of grading or internal lamination seems likely (Kuenen and Migliorini, 1950, p. 121), and such deposits indeed are characteristic of the foreset slope of the Mississippi delta.

MESNARD QUARTZITE

DEFINITION, DISTRIBUTION, AND THICKNESS

The Mesnard Quartzite was named by Van Hise and Bayley (1895, p. 517) from Mount Mesnard (in 1959 locally and informally given the name "Mount Marquette" as part of a tourist development) in the southeastern part of the Marquette quadrangle. The formation as originally defined and mapped included both the Enchantment Lake Formation and the Mesnard Quartzite of the present report. The name of the formation is here restricted to thick and massive vitreous quartzite, which is generally colorless to light gray and weathers white to shades of pink or light red. Some thin beds of gray slate, quartz-pebble conglomerate, and wacke are interlayered in the vitreous quartzite. The base of the redefined Mesnard Quartzite is placed where massive white-weathering vitreous quartzite first ap-

[1] Although this term is derived from the terms "clay matrix" and "paste," used by Pettijohn to describe the interstitial material of nonmetamorphosed graywacke (1943b, p. 943–944; 1957a, p. 303), it is no less useful for metamorphic graywacke or wacke, in which the matrix is now largely sericite, chlorite, and chert, or simply sericite.

pears above quartz wacke or sericite slate of the Enchantment Lake Formation. The top of the formation is at the horizon where vitreous quartzite gives way to slate that in turn is overlain by dolomite or where quartzite passes almost directly into dolomite.

The largest exposures are on Mount Mesnard and westward from there for about 1 mile along the north limb of the Marquette synclinorium and in the N½ secs. 2 and 3, T. 47 N., R. 25 W., on the south flank of the uplift of Mona Schist. The westernmost exposure on the north limb of the Marquette synclinorium is on the north shore of Enchantment Lake. The formation is projected westward from there between outcrops of Kona Dolomite and Mona Schist toward outcrops of Mesnard Quartzite in the adjoining Negaunee quadrangle. The westernmost exposures in the southern part of the synclinorium are in the NW¼ sec. 9, T. 47 N., R. 25 W. The southwestern extension of the Mesnard into secs. 8 and 17, T. 47 N., R. 25 W., is inferred from the gently arched structure of the overlying Kona Dolomite in secs. 7, 8, 17, and 18, and from the domelike fold of the underlying Enchantment Lake conglomerate and the Compeau Creek Gneiss in secs. 8, 9, and 16. Quartzite and cherty rock that evidently were mapped as Mesnard Quartzite along the south limb of the synclinorium in the SW¼ sec. 18 and the adjoining part of sec. 13, T. 47 N., R. 26 W., by Van Hise and Bayley (1897, atlas sheets 34 and 37) and by Van Hise and Leith (1911, pl. 17) are interpreted here as quartzite interbeds in silicified dolomite, and as massive silicified dolomite and vein quartz (Gair and others, 1961b). These are assigned to the Kona Dolomite.

The maximum closely delimited thickness of Mesnard Quartzite is 500 feet, measured across the part of the formation that extends eastward from Enchantment Lake for 1–2 miles. The quartzite there dips steeply and in several places is bounded closely by outcrops of underlying and overlying rocks. East of Mount Mesnard, the inferred thickness is between 600 and 700 feet, but because there is no control on the lower contact, the influence of the fold mapped on Mount Mesnard cannot be evaluated. The minimum closely delimited thickness is 200 feet and also occurs along the north limb of the synclinorium in the NW¼ sec. 34; the disparity between the maximum and minimum thicknesses may be a result of thinning by undetected bedding faults.

DESCRIPTION

Nearly all exposures of the Mesnard consist dominantly or entirely of white, pink, or light-gray quartzite. Veinlets of white quartz or white quartz and specular hematite cut the quartzite in many places. Small amounts of red or bluish-black hematite are concentrated in cracks, in scattered spots, and locally along bedding. Spots of specular hematite are generally ⅛–¼ inch across but in one place are 2 inches in diameter. In a few places, quartzite is colored red by hematite in the matrix, and in places brecciated quartzite has angular white fragments set in a matrix of reddish ferruginous quartzite. Good examples of such ferruginous brecciated quartzite occur on the top of the nob in the NW¼NE¼ sec. 1, T. 47 N., R. 25 W. (The east-west ridge along which this nob occurs was called Mount Chocolay on former maps.) Several old test pits have been found on Mount Mesnard near 2,750N–500E, SW. cor. sec. 35, across a ferruginous zone of shears and breccia about 100 feet wide. The shears and breccia are evidently of no great extent, and the average iron content in them is probably less than 10 percent.

Bedding may be massive and difficult to see in smoothly glaciated outcrops, or it may be conspicuous. Tabular layers separated by parting surfaces range from about 1 inch to about 2 feet in thickness, and within these, alternating dark ferruginous and white laminations, ⅛–¼ inch thick, are common. Bedding is also marked in places by thin coarsely granular or pebbly zones. Ripple marks and crossbedding are widespread and indicate deposition in shallow water; they also indicate top directions. Crossbedding may best be seen on the north slope of Mount Mesnard, particularly in the SE¼NW¼ sec. 35; along the north slope of the ridge to the west, in the NE¼NW¼ sec. 34, and along the prominent ridge in the NW¼NW¼ sec. 2, T. 47 N., R. 25 W. The quartzite is strongly jointed, and closely-spaced strike joints are apt to be confused with bedding.

Most of the vitreous quartzite contains more than 90 percent quartz. Quartz grains interlock in granoblastic mosaics as a result of secondary overgrowths or have subangular to round detrital outlines and are separated by thin selvages of sericite, chlorite, cherty quartz, finely granular hematite, or combinations of these minerals. Mixed sericite and cherty quartz form the most common matrix material. Leucoxene, feldspar, zircon, tourmaline, rutile, and finely granular opaque material occur in minor amounts in some of the quartzite. Nonmatrix quartz grains range in size from 0.06 to 3 mm, but most are between 0.1 and 0.6 mm. The coarsest grains tend to be concentrated parallel to bedding in seams or lenses not more than a few grains thick. In granoblastic quartz mosaic fabrics, rings of dusty hematite or of other dusty material may outline the original subrounded to rounded detrital shapes that are now largely obscured by secondary quartz overgrowths. Small amounts of finely granular and dusty reddish-brown iron oxide, which accentuate bedding in places, are localized in the matrix of those beds, particularly in

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triangular-shaped areas between detrital grains. The iron oxide generally is intergrown with sericite and fine-grained quartz. Such iron oxide probably collected with other matrix material during deposition and tended to coat the detrital grains. Slightly ferruginous chlorite-bearing quartzite has been found at 3,500N–3,800E, 3,000N–4,900E, and 2,400N–2,400E, SW. cor. sec. 34, T. 48 N., R. 25 W., and at 1,600S–1,900E and 1,800S–400E, NW. cors. secs. 2 and 4, respectively, T. 47 N., R., 25 W. Small amounts of reddish-brown oxide in the chlorite-bearing quartzite probably formed by weathering of the chlorite.

Pebbly zones ranging from a few inches to 2 feet in thickness have been found in the quartzite at the following locations:

Marquette quadrangle:
- T. 48 N., R. 25 W., sec. 33, from SW. cor., 4,250N–4,700E and 3,700N–4,900E [1]
- T. 47 N., R. 25 W., sec. 2, from NW. cor., 1,900S–4,100E, 1,850S–4,450E, and 1,900S–4,700E

Sands quadrangle:
- T. 47 N., R. 25 W., sec. 2, from SW. cor., 2,450N–3,800E

Pebbles consist of white to reddish vein quartz, cherty quartz, and quartzite and range in diameter from less than 1 to 3 inches. Except in the 6-inch conglomerate bed mentioned in the footnote, they are set in a matrix of medium-grained white or reddish quartzite. Some, and perhaps all, of these zones were observed by Van Hise and Bayley (1897, p. 234–238). Conglomerate noted by Van Hise and Bayley (1897, p. 238) between Mesnard Quartzite and the underlying granitic basement, north of the quarter post between secs. 1 and 2, T. 47 N., R. 25 W., is here interpreted as tectonic breccia developed from Mesnard Quartzite in the slippage zone between the Mesnard and basement rock.

Slate interbeds in the Mesnard Quartzite range from 1 to 30 feet in thickness and are exposed in relatively few places. The slate typically is gray or green, fissile, and weathers grayish brown, buff, or pale yellow. The main constituents are sericite or sericite and chlorite. Fine-grained chert and detrital quartz generally form less than 10 percent of the rock content. Small amounts of fine-grained reddish-brown iron oxide are ubiquitous, and traces of magnetite, rutile, and tourmaline occur in some of the slate. Interbeds of vitreous quartzite, 1–2 inches thick, are present in one of the slate beds.

ORIGIN AND SOURCE OF SEDIMENTS

Most beds of the Mesnard Quartzite are well sorted, and original quartz grains were rounded to subrounded. Together with the widespread crossbedding and ripple marks, the good sorting and grain shapes suggest stable conditions of sedimentation under which detritus was winnowed and reworked within the zone of wave action and nearshore currents. Slate interbeds probably represent local short-term increases in the depth of the bottom to below the base of wave action.

Crossbeds have been used to indicate the source direction of sediments in northern Michigan by Trow (1948) and Pettijohn (1957b). By rotating the attitudes of pairs of crossbeds around their lines of strike on a stereographic net so that the topset bed of each pair is brought to a horizontal position, the foreset bed presumably is brought close to the attitude it had when it was deposited. The updip direction then should also be the direction from which sediments came. This method has been used on 13 pairs of crossbeds from the Mesnard, and the results indicate that the source of sediments was to the west-northwest (fig. 10). This conclusion agrees with the findings of Trow (1948), for the Sturgeon Quartzite in the Menominee range, and of Pettijohn (1957b), for middle Precambrian quartzites over a wide area of the southern Canadian Shield in the upper Great Lakes region. A granitic land area, an unknown distance west-northwest of the map area, was probably the source of Mesnard sediments.

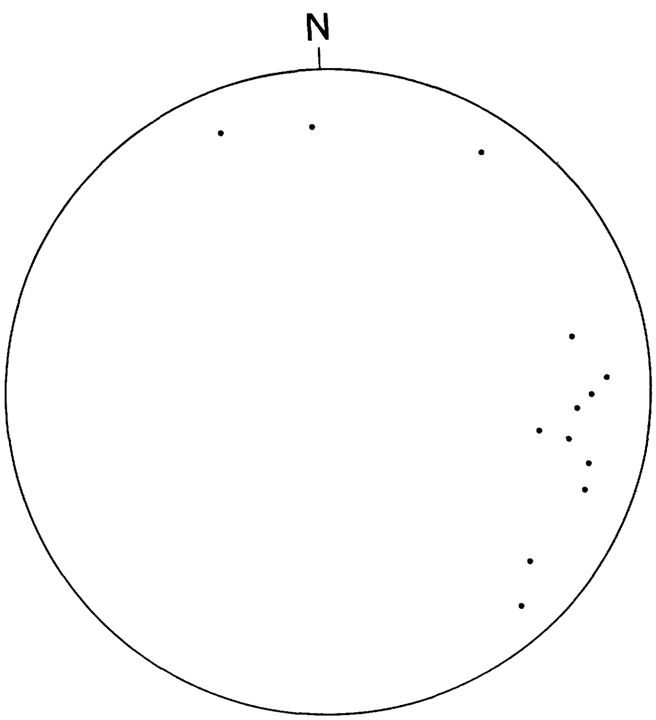

FIGURE 10.—Orientation diagram of 13 sets of crossbeds, Mesnard Quartzite, Marquette quadrangle. Dip lines of crossbeds reoriented to prefolding position and projected into lower hemisphere of Schmidt equal-area net.

[1] A 6-inch conglomerate bed in which quartz pebbles "float" in a red slate that in turn is interbedded with vitreous quartzite.

CORRELATION AND AGE RELATIONS

The Mesnard Quartzite is considered to be a direct correlative of the middle Precambrian Sturgeon Quartzite of Iron and Dickinson Counties (James, 1958, p. 35). Lithologic similarity, both of the formations and of the entire stratigraphic sections within which they occur, and similar stratigraphic position in respect to basement rocks below and the distinctive dolomite formation above constitute evidence in support of this correlation. The Mesnard, therefore, is of early Animikie age, but because of the downward extension of the Animikie in northern Michigan by James (1958), compared with the type Animikie, the Mesnard is at a lower horizon than the base of the type Animikie in the Thunder Bay area, Ontario.

KONA DOLOMITE

COMPONENTS AND THEIR GENERAL DISTRIBUTION

The Kona Dolomite was named by Van Hise and Bayley (1895, p. 523) from the Kona Hills in secs. 12 and 13, T. 47 N., R. 26 W. Dolomite and interlaminated chert-dolomite are the characteristic rocks of the unit, but quartzose dolomite, quartzite, and slate are widespread and make up a substantial part of the formation. Transitional lithologies between these types are not common.

A thin, generally poorly exposed slate unit occurs probably everywhere between vitreous Mesnard Quartzite and the characteristic dolomite or chert-dolomite of the Kona (pls. 1, 2, 4). This slate resembles slate interbedded with dolomite higher in the formation, and in places it contains thin interbeds of dolomite. Hence, it has been mapped with the Kona on plates 1 and 2, although in a few places it is sufficiently thick and well exposed to be mapped separately.

Algal structures occur widely in the Kona and are indicated by letter symbol on plates 1, 2, and 4. A distinctive fine-grained laminated chocolate-brown and orange-colored siltite forms one or two beds at several localities. Parts of the Kona have been extensively silicified and intruded by vein quartz (Gair and others, 1961b). The largest exposures of silicified Kona occur in the eastern part of the two quadrangles (pl. 4), but small masses, 100 feet or less in surface dimensions, are present in many places in the western part of the area. Lovering's (1962) usage of the term "jasperoid" is adopted here for chertlike silicified (replaced) dolomite, and the term "chert" is restricted to rock of primary or inferred primary origin.

UPPER LIMIT AND THICKNESS

The top of the Kona is in general poorly defined. It is closely delimited only in the SE¼ sec. 13, T. 47 N., R. 26 W., where dolomite is conformably overlain in a nearly continuous exposure by slate that has quartzite interbeds but no dolomite. The dolomite and slate are separated by a narrow zone of massive chert or jasperoid that probably represents a zone of fold-slippage along the upper contact of the Kona rather than a true fault.

The thickness of the Kona can best be determined across the north-trending belt between Little Pelesier Lake and Carp River Lake in the north-central part of the Sands quadrangle. There the formation appears to be uncomplicated by faults or second-order folds, and the top and bottom are fairly closely delimited. A width of nearly continuous outcrop of 2,000 feet and an average dip of about 30° indicate a minimum thickness of 1,000 feet. The greatest mapped width of the Kona here is about 2,500 feet and represents a maximum thickness of 1,200 feet. Along the north limb of the synclinorium, in the Marquette quadrangle, where exposures of dolomite are extensive, the top of the Kona is not well defined and there is some second-order folding of the formation. Therefore, no attempt has been made in this study to determine the thickness of the Kona in that area.

DESCRIPTION

DOLOMITE AND LAMINATED CHERT-DOLOMITE

Dolomite and laminated chert-dolomite are well exposed (1) in a broad gently folded belt crossing the Kona and Ragged Hills on the south side of the synclinorium, (2) in the north-trending belt between Carp River Lake and Little Pelesier Lake, (3) in the core of the synclinorium in the central part of secs. 2 and 3, T. 47 N., R. 25 W., and (4) along the north limb, both on the ridge west of Mount Mesnard and in the area of second-order folding south of Enchantment Lake.

Beds of dolomite range from about 2 inches to several feet in thickness in rock having no chert laminations. The thicker, outwardly massive beds generally are internally laminated, but partings are not formed on these laminae. Laminae are planar, wavy, or lenticular, may pinch and swell, and commonly are disrupted by microfaults, some of which may represent slumping after partial consolidation. Many such minor features are seen only on polished surfaces.

In laminated chert-dolomite, ¼- to 2-inch layers of dolomite generally alternate with conformable paper-thin to ½-inch chert layers. Chert laminae are tabular, form flat lenses, or are wavy. Short cusps of chert along some bedding surfaces may be a product of boudinage. Chert laminae may have been deposited alternately with dolomite laminae or they may be jasperoid replacements of selected dolomite beds. They are interpreted as primary where the laminae are persistent, and are quite regular in thickness and strictly conformable with the alternating dolomite laminae. The laminated chert-dolomite described here is thus interpreted as a product of primary carbonate-silica deposition.

Oolites are found in some of the dolomite and probably were originally more common than at present. The oolites now seen have been all but obliterated by recrystallization and are generally marked only or mainly by dusty hematite rings. Some have quartz rims around carbonate cores and may be partly silicified carbonate oolites.

The weathered dolomite is mainly pale yellowish tan but can be rusty brown, maroon, buff, or nearly white. Most fresh dolomite is pale yellowish tan, salmon pink, or salmon brown. Chert is pale yellowish tan, salmon, or orange brown. Outcrops of massive dolomite are characterized by somewhat rounded surfaces caused by weathering back along bedding and cross fractures. Slight weathered-out ridges tend to follow laminations. Laminated chert-dolomite, on the other hand, typically weathers with pronounced lamellar ridges along resistant chert laminae.

The index of refraction, N_O, of carbonate from several differently colored dolomite beds in the southeastern part of the Marquette quadrangle ranges from 1.679 to 1.682 and indicates that, regardless of rock color, the carbonate is virtually pure dolomite. X-ray tests of several samples has confirmed this identification. Variations in color are caused largely by different amounts of iron oxides, which form tiny inclusions within and between crystals of dolomite.

After deposition, both dolomite and chert recrystallized to mosaic fabrics, probably during regional metamorphism in the post-Animikie-pre-Keweenawan interval. Grains of recrystallized chert and dolomite are generally less than 0.05 and 0.3 mm in diameter, respectively. After recrystallization, layers of dolomite were cut by veinlets and were partly replaced by lenses, knots, or irregularly shaped masses of deep-red dolomite, heavily charged with tiny particles of hematite. Similar red dolomite in places forms the groundmass for dolomite breccia fragments. In a few places, coarse red euhedral metacrysts of dolomite are spotted through fine- to medium-grained lighter colored crystalline dolomite. These metacrysts, with their heavy charge of dusty hematite, are analogous to the postmetamorphic veins and knots and probably formed after the general metamorphic recrystallization of the rock. Small irregular seams and patches of light-colored coarse dolomite commonly cut the fine- to medium-grained granoblastic dolomite mosaics or are disposed along bedding. We cannot determine whether crystallization of coarse dolomite in these areas occurred at the same time as crystallization of the granoblastic mosaics or later. Thin veinlets and cross-joint fillings of quartz, quartz-carbonate, quartz-carbonate-microcline, or quartz-carbonate-tourmaline are minor postmetamorphic features in the Kona.

QUARTZOSE DOLOMITE

Quartz-bearing dolomite containing 5–50 percent quartz are common in the Kona in all parts of the area. It is generally dark reddish orange, salmon, or maroon, has fine- to medium-grained granoblastic texture, and shows considerable minor variation in grain size within the limits of a thin section. Oolitic structure is common in parts of the rock. Most of the quartz is of detrital origin, but some also is in overgrowths on detrital quartz (fig. 11) and in small irregular blebs and veinlets. Both dolomite and quartz have been extensively recrystallized. This recrystallization is well illustrated in some thin sections showing grains of recrystallized dolomite and quartz crossing the borders of relic oolites (fig. 12). Detrital quartz and recrystallized dolomite grains are generally less than 0.2 mm in size. Either because of irregular overgrowths in quartz or corrosion of detrital or overgrowth quartz by dolomite (Carozzi, 1960, p. 259), or because of both, adjoining quartz and dolomite commonly have interlocking boundaries. In some places, dolomite crystals adjoining overgrowth quartz are coarser than in the surrounding rock and evidently were affected by the solutions that brought in the overgrowth silica during metamorphic recrystallization. In places the detritus consists both of quartz grains and rock fragments of fine-grained dolomite, the frag-

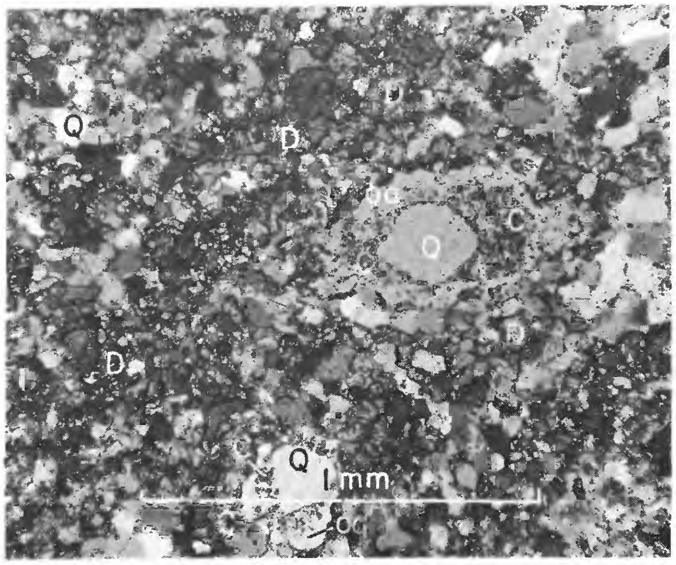

FIGURE 11.—Photomicrograph of quartzose dolomite, 4,750N–250E, SW. cor sec. 31, T. 48 N., R. 25 W. Intergrown mosaic of dolomite (D) and quartz. Detrital quartz (Q) enlarged by overgrowths of quartz (oq). Small inclusions of carbonate (C) in large overgrowth aprons outline two comparatively large grains of detrital quartz. Nicols partly crossed.

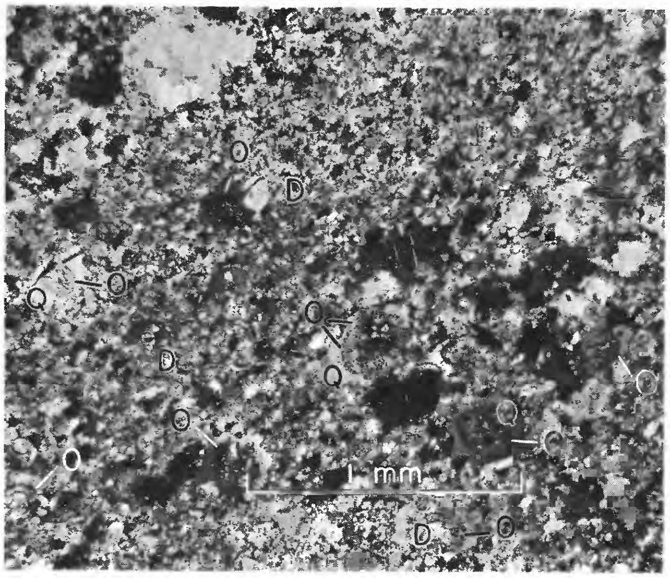

FIGURE 12.—Photomicrograph of quartzose oolitic dolomite, 2,650N–1,300E, SW. cor. sec. 34, T. 48 N., R. 25 W. Minute grains of iron oxide or carbonate outline relic oolites (O), now overgrown by recrystallized dolomite (D) or quartz (Q). Nicols partly crossed.

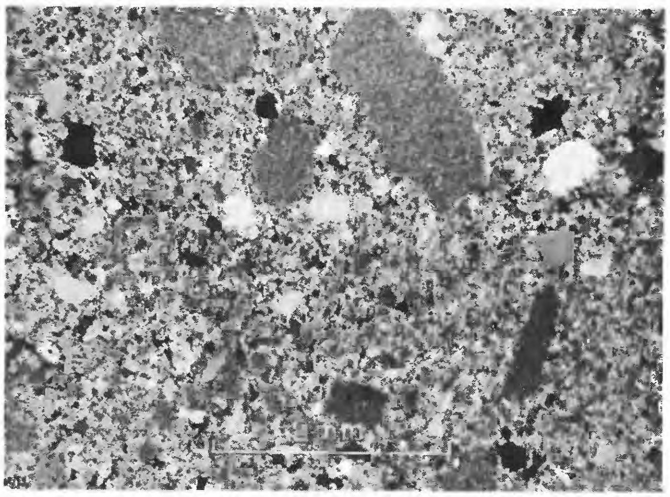

FIGURE 13.—Photomicrograph of quartzose dolomite with detrital fragments of a fine-grained dolomite, 2,950–450E, SW. cor. sec. 34, T. 48 N., R. 25 W. Nicols partly crossed.

ments probably being analogous in origin to intraformational conglomerate (fig. 13).

Small amounts (as much as approximately 2 percent) of dusty or finely granular red or yellow-brown iron oxides are nearly ubiquitous, and sericite in amounts as great as 5 percent is common in the quartzose dolomite. Traces of a finely granular dark opaque mineral (magnetite?) are found in most samples of this rock. Chlorite, microcline, and biotite occur in a relatively small fraction of the quartzose dolomite. The biotite occurs near intrusions of metadiabase and so is probably a relic of contact metamorphism that preceded the post-Animikie, pre-Keweenawan regional metamorphism.

QUARTZITE

Several beds of impure quartzite occur in the lower part of the Kona in the NE¼NE¼ sec. 1, in the SW¼ sec. 2, in the west-central part of sec. 3, T. 47 N., R. 25 W., and in the NE¼ sec. 32, T. 48 N., R. 25 W. The greatest known thickness of any of these beds is 80 feet, near the southeastern end of the large roadcut on U.S. Highway 41, about on the line between R. 24 W. and R. 25 W. (pl. 5). Beds of quartzite, generally not more than 20 feet thick, occur farther west in the dolomite, particularly in the hill just east of Little Pelesier Lake and in several localities in the Kona and Ragged Hills, north and east of Grace Lake. The stratigraphic position of these beds in the formation is uncertain.

Characteristically the quartzite consists of detrital quartz with 10 percent or less interstitial dolomite, iron oxides, sericite, recrystallized chert, or combinations of

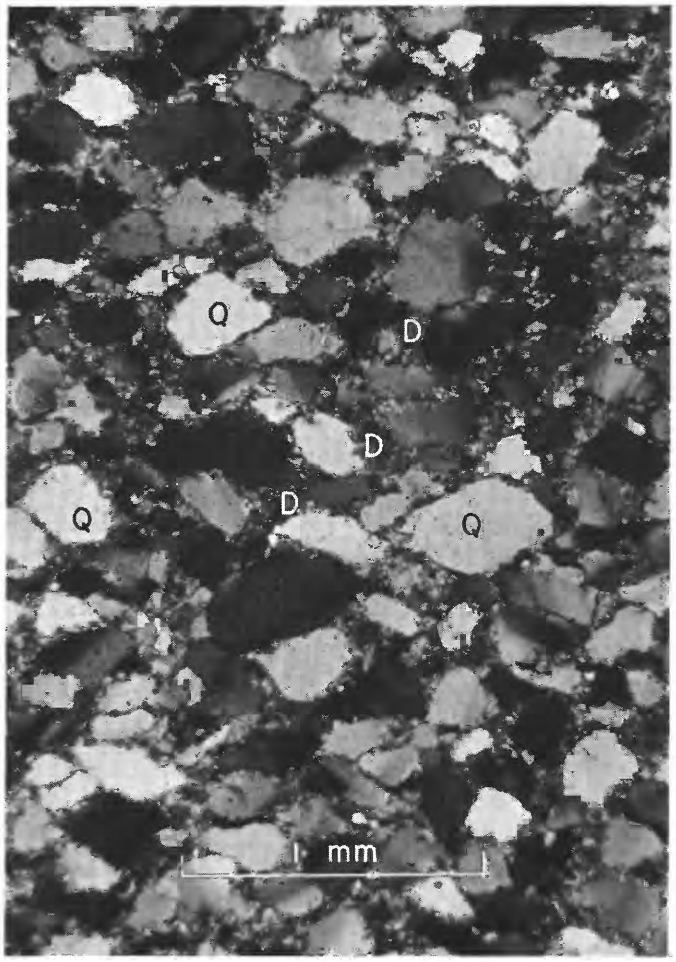

FIGURE 14.—Photomicrograph of dolomitic quartzite, approximately 300S, SE. cor. sec. 36, T. 48 N., R. 25 W. Loosely packed detrital quartz (Q) separated by matrix seams of dolomite (D) stained reddish brown by iron oxide. Small amount of marginal overgrowth on some quartz grains. Crossed nicols.

these (fig. 14). Minor leucoxene, chlorite, epidote, and zircon occur in some of the quartzite. Interstitial dolomite is generally less than 5 percent of the rock, but in places is forms 10–35 percent, and the weathered dolomitic quartzite it somewhat friable. Interstitial dolomite is everywhere stained reddish or yellowish brown by iron oxides. Depending on the degree of oxidation of interstitial dolomite, which probably is a near-surface feature, and on the amounts and kinds of interstitial iron oxides, the color of the quartzite ranges from light pink to red to dark reddish brown. The generally reddish-brown colors of the Kona quartzite help to distinguish it from Mesnard Quartzite.

A few 5- to 10-foot beds of white quartzite that occur in sequence with slate and dolomite on the hill east of Little Pelesier Lake and in several places in the Ragged Hills are lithologically indistinguishable from Mesnard Quartzite.

Minor amounts of quartzite have more than 10 percent matrix chert and sericite and grade toward cherty quartz wacke. At one place such quartzite contains lenses of sericite slate about 0.5 mm thick, but in general slate and quartzite occur in separate distinct beds and are not closely interlayered or intergradational.

SLATE

A well-defined belt of slate occurs at the base of the Kona on Mount Mesnard, extending from about 1,000 feet west to 800 feet east of the surge tank (pl. 4). The slate is also well exposed elsewhere on the north limb of the synclinorium in the NE¼ sec. 33 and the NW¼ sec. 34. In the core of the synclinorium, a sharply defined belt of the slate passes around the north, northeast, east, and southeast sides of Buschell Lake (Copper Lake of earlier reports) in sec. 2, T. 47 N., R. 25 W. (pl. 4). Several old copper prospects are located along this slate belt from southeast to south of Buschell Lake. The basal slate is also well exposed in and adjacent to the large roadcut along U.S. Highway 41, near the NE. cor. sec. 1, T. 47 N., R. 25 W. (pl. 5); the slate, however, does not crop out well and cannot be traced westward from there. About half a mile west of the roadcut, near the base of the Kona on the south flank and toward the west end of the east-plunging syncline, slate and sericitic quartzite have copper mineralization and have been extensively prospected. These places, as well as other localities in secs. 2, 3, 7, 8, and 9, T. 47 N., R. 25 W., and in sec. 13, T. 47 N., R. 26 W., where slate has been found in the Kona, are indicated by letter symbol on plates 1 and 2. The maximum exposed thickness of slate in the lower part of the Kona is 180–200 feet in the roadcut on U.S. Highway 41 (pl. 5) and in the belt north and northeast of Buschell Lake. The slate, which is 180–200 feet thick along U.S. Highway 41 on the north limb of the east-plunging syncline, is only about 50 feet thick on the south limb, just north of the Harvey quarry (pl. 5). Probably part of the slate on the south limb has been faulted out of the exposed section. Some undetected imbricate faults and isoclinal folds may also exist within the slate unit on the north limb, adding to the apparent thickness of the slate there.

Most of the slate beds in the Kona are gray, maroon, or pale green, and some are tan or buff. They consist generally of fine-grained clastic and cherty quartz mixed with abundant sericite and small amounts of dusty and finely granular iron oxides. Sericite typically ranges from 60 to 90 percent, quartz from 10 to 40 percent, and iron oxides from less than 1 to about 3 percent of the rock. Rutile, leucoxene, chlorite, microcline, tourmaline, and zoisite are minor minerals in some of the slate. Thin silty beds consisting of fine-grained sericitic quartzite, ferruginous wacke, quartzose dolomite, dolomitic quartzite, or dolomitic wacke are commonly interspersed through the quartz-sericite slate. In places, the slate consists almost entirely of sericite or of sericite containing thin carbonate or carbonate-quartz interbeds. Carbonate-bearing layers generally weather to shades of brown.

Sericite-rich parts of the slate generally have a strong cleavage caused by the alinement of the micaceous flakes. Quartz usually is equant and “floats” in the sericite. Most thin silty layers have granoblastic mosaic textures or weakly foliated graywackelike texture.

Thin veinlets and nodules of medium- to coarse-grained vein quartz or vein quartz and carbonate are very abundant in some parts of the slate. Nodules commonly lie astride, and are somewhat elongated parallel to the cleavage and therefore formed during or later than deformation. This feature is particularly well displayed in the slate of the lower part of the Kona in the roadcut on U.S. Higway 41 near the NE. cor. sec. 1, T. 47 N., R. 25 W. (pls. 4, 5).

LAMINATED SILTITE

A distinctive nonfissile fine-grained siltite with chocolate-brown and orange-colored layers forms one or two beds in the Kona, not more than 10 feet thick. These beds can be used as horizon markers for short distances. In plates 1 and 2 the principal occurrences are shown by letter symbol in the NE¼ sec. 31, the NE¼ sec. 32, the NE¼ sec. 33, and the NW¼ sec. 34, T. 48 N., R. 25 W., in the NE¼ sec. 36, T. 48 N., R. 26 W., in the SE¼ sec. 7, T. 47 N., R. 25 W., and in the SE¼ sec. 13, T. 47 N., R. 26 W.

This siltite consists of dominant quartz and microcline and also contains minor hematite and sericite, and, in some places, dolomite. One thin section also contains minor rutile, sphene(?), and zircon(?). The rock has a fine-grained detrital texture partly modified to a granoblastic mosaic fabric. Sericite flakes and carbonate grains are spread among granular quartz and iron-stained granular microcline. Tiny granules of reddish-brown iron oxide are erratically distributed in loosely packed aggregates but are more highly concentrated in the chocolate-brown layers. The fine-grained iron-stained microcline was identified only by X-ray (analyst, P. D. Blackman).

ALGAL STRUCTURES

Certain arching, swirly, or somewhat bulbous laminated features in the dolomite or chert-dolomite (figs. 15–17) are similar to structures both in ancient and modern carbonate deposits that are generally acknowledged to be stromatolitic growths representing algal colonies (Walcott, 1914; Twenhofel, 1919; Rezak, 1957; James and others, 1961, p. 35).

Some workers have attributed these features to tectonic folding or to preconsolidation slumping, but the similarity to forms built by blue-green algae in modern times almost conclusively indicates an organic origin. Wherever three-dimensional exposures are found, these features can be seen to be gently to strongly domed or slightly bulbous, rather than arched along linear axes as they appear to be in some two-dimensional exposures. The algal domes or bulbs commonly rise from a bedding surface. They may be widely spaced along a bed or closely packed. Where they are closely packed, the sides of the adjoining arches meet in sharp downward-pointing V's (fig. 16; Twenhofel, 1919, figs. 1, 2). Domes typically are circular to moderately oval in plan (fig. 15*B*) and tend to be wider than they are high. Generally circular or oval outlines may be somewhat irregular in detail. Domes (colonies) range in width from about 1 inch to 6 feet. Although the smallest and the largest domes seen were not at the same locality, domes as small as about 2 inches wide do occur near domes as much as 1 foot wide. The conditions that governed the variations in the sizes of these domes are unknown; possibly the colonies were at different stages of growth when the growth in any given bed was halted.

Attempts by different workers to classify stromatolitic structures on the basis of biologic affinities have not been successful, inasmuch as the stromatolites represent products of the reaction of algae on their environment, rather than parts of the algae themselves. As Riedel (1953) and

FIGURE 15.—*A* (left) Algal structure in laminated chert-dolomite, 2,850N, SW. cor. sec. 34, T. 48 N., R. 25 W. Cross-sectional view. Overall length of scale approximately 7.5 inches. *B* (above), Same occurrence as *A*. View looking onto surface of steeply dipping bedding from right side of *A*.

Rezak (1957, p. 130–131) have emphasized, classification therefore should be arbitrarily based on differences in the forms of the stromatolitic growths. Most, if not all, of the algal structures in the Kona are similar in form to those of the genus *Collenia* Walcott (Walcott, 1914; Twenhofel, 1919, p. 346; Rezak, 1957, p. 131). Twenhofel named the species *Collenia kona* from a single conspicuous occurrence, probably the one in the woods above the southeast end of the large roadcut on U.S. Highway 41, in the NW¼NW¼ sec. 6, T. 47 N., R. 24 W. From Twenhofel's description, it is not apparent that the species is based on characteristics uniquely different from species of *Collenia* described by Walcott, and the species name evidently has only geographic significance. Forms considerably different in detail from that at Twenhofel's locality have been found at many other places in the Kona during the present study. Many of these resemble species of *Collenia* illustrated by Walcott (1914), but no attempt is made here to evaluate them for the designation of species, either according to Walcott's or Rezak's criteria.

In the laminated chert-dolomite domes, the alternation of chert and dolomite laminae is entirely similar to that in layers without algal structure. Probably chert laminae in both types of rock are also similar in genetic relationship to the carbonate, interpreted here as a primary precipitate because the chert laminae are thin, persistent, quite regular in thickness, and conformable with adjacent arching dolomite laminae. This alternation of carbonate and chert would not necessarily require that algae be responsible first for the precipitation of carbonate and later of silica, although the influence of some types of ancient algae in precipitating silica has been pointed out by Clarke (1916) and by Grout and Broderick (1919). The growth of carbonate stromatolites has long been attributed to the influence of blue-green algae in removing carbon dioxide from solution during photosynthesis; this removal increases the alkalinity of the water and causes precipitation of calcium carbonate. Black (1933, p. 167, 186), Young (1935, p. 158), Cloud and Barnes (1948, p. 98, 100), and Rezak (1957, p. 147) have pointed out, however, that in addition to the influence of algae in precipitating calcium carbonate, modern stromatolites grow through the ability of algal filaments to bind whatever sediments are delivered to the site of deposition. Perhaps the arching laminated chert-dolomite algal structures of the Kona therefore represent merely the binding action of algae on carbonate mud and silica gel alternately precipitated through no action of the algae.

Modern stromatolitic deposits are forming in intertidal zones and on tidal mudflats (Black, 1933; Rezak, 1957). Oolites also are generally considered products of shallow-water environments and have been noted in association with stromatolites by Kalkowski (1908), Bradley (1929), and Young (1933). One thin section made from stromatolitic Kona Dolomite shows oolites between stromatolites. Stromatolitic parts of the Kona therefore were almost certainly shallow-water deposits.

SILICIFIED DOLOMITE AND SLATE

DESCRIPTION

Silicification of the Kona has occurred widely and resulted in replacement of dolomite or slate by jasperoid

FIGURE 16.—Algal structure in laminated dolomite, approximately 400S, SE. cor. sec. 36, T. 48 N., R. 25 W. Overall length of scale approximately 7.5 inches. Probably same locality as that shown by Twenhofel (1919, figs. 1, 2).

FIGURE 17.—Algal structure in laminated dolomite. Specimen from 850N–3,400E, SW. cor. sec. 7, T. 47 N., R. 25 W. Rock also has small irregular nodules of chert which stand up in relief on the weathered right side of specimen.

and vein quartz (figs. 18, 19). Outcrops having substantial silicification are indicated on plates 1, 2, and 4. (See Gair and others, 1961b, fig. 178.1.) Silicification in the eastern part of the area occurred close to the basal contact of the Kona or close to faults that removed the Mesnard and placed the Kona against basement rocks (pls. 2, 4). In the western part of the map area (pl. 2), there is no evidence that silicification was concentrated in the lower part of the formation. In the Kona and Ragged Hills the stratigraphic distance to the top or bottom of the formation at most places is difficult or impossible to determine owing to gentle undulatory folding and to an absence of marker beds. In the belt of Kona Dolomite between Carp River and Little Pelesier Lakes, however, patches that underwent silicification are about evenly distributed from the bottom to the top of the Kona.

Silicified dolomite, or jasperoid, typically is very fine grained and compact and is flesh colored, pale tan, salmon colored, or shades of red; in some places it is bluish black. Associated coarser grained vein quartz generally is white or reddish brown. Banded red and white jasperoid is a common form of the silicified dolomite in many places, most notably on Mount Mesnard, near 2,400 N–950E, SW. cor. sec. 35, T. 48 N., R. 25 W. There and elsewhere in silicified parts of the Kona, a jasperoid breccia of whitish angular fragments in a deep red matrix is common. Such breccia was well illustrated by Van Hise and Bayley (1897, pl. 8, facing p. 250).

Silica replaces individual beds of dolomite which range from an inch or less to several feet in thickness (fig. 18) and masses of thin beds many feet in aggregate thickness. Original thin beds are commonly preserved as differently colored layers of jasperoid, the coloring depending largely on the types and amounts of minor iron oxides. Streaky tan or buff dolomite zones occur within some beds of jasperoid and are interpreted by us as relics of original beds of dolomite. In places such

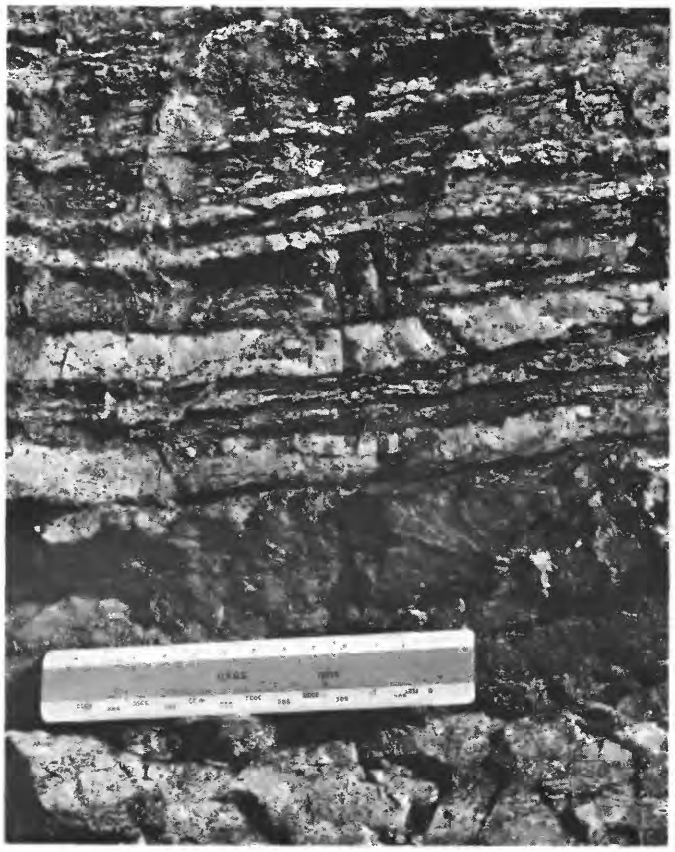

FIGURE 18.—Partly silicified laminated chert-dolomite, immediately south of surge tank on Mount Mesnard. Rock in approximately lower two-fifths of photograph almost completely silicified. Some silicification of dolomite along cross fractures in upper part of photograph. Primary or secondary origin of siliceous laminae (lighter colored layers) in upper part of photograph is uncertain. Scale is approximately 7.5 inches long.

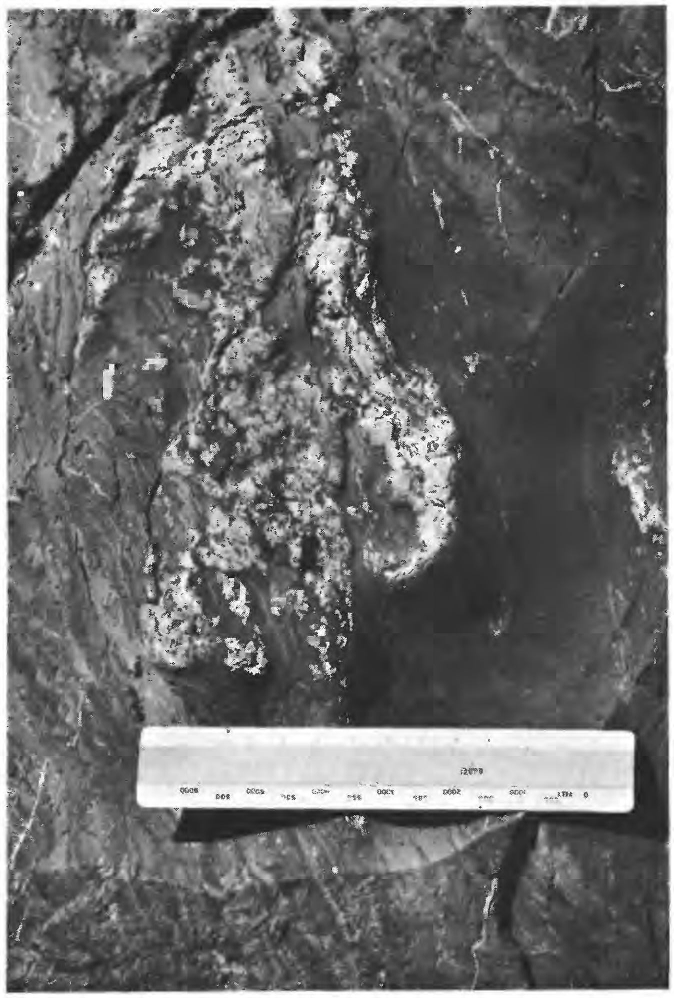

FIGURE 19.—Partly silicified slate member of Kona Dolomite, immediately north of fault and approximately 150 feet east-southeast of bench mark 1089, in south central part of sec. 3, T. 47 N., R. 25 W. Irregular knots of secondary quartz in laminated maroon slate. Scale is approximately 7.5 inches long.

streaky dolomite zones coalesce along the strike, and the jasperoid bed gradually gives way to a dolomite bed. In a few places jasperoid or vein quartz is in sharp contact with and appears to surround blocks of dolomite.

The jasperoid consists of granoblastic mosaics of very fine grained quartz. Dusty iron oxides are irregularly concentrated in the mosaics. Knots and masses of vein quartz have grains of greatly variable size, ranging from a fraction of a millimeter to several millimeters. Grains or small blebby aggregates of relatively coarse (vein) quartz occur widely in the jasperoid. Under magnification of 500–800 times in thin section, small loose clusters of very fine grained carbonate crystals may be seen in the jasperoid and probably are relics of presilicification dolomite.

Small lenticular knots and irregular-shaped masses of vein quartz are the most characteristic features of silicified slate in the Kona (fig. 19), but part of a large continuous mass of vein quartz and white, reddish-brown, and greenish jasperoid extending about 1,000 feet along Migisy Bluff in the SW¼ sec. 2, T. 47 N., R. 25 W., is also probably in part silicified slate. This conclusion is based on the interpretation that streaky sericitic and chloritic zones in the jasperoid were derived from presilicification slate.

South of, but immediately adjacent to, massive white vein quartz forming part of the large silicified outcrop in the SW¼ sec. 2, deposits of secondary silica also occur in tonalitic basement rock (at 1,450N–2,250E, SW cor. sec. 2), in the form of veins of white quartz as much as 4 inches thick. Some of these veins wedge out upward in the tonalite.

GEOLOGIC FACTORS CONTROLLING SILICIFICATION

The principal geologic factors controlling silicification appear to have been the availablity of silica-bearing solutions, the composition of the original rock, and the presence of channelways near sites of silicification.

Where laminated chert-dolomite passes along strike into banded light-colored and cherty rock, it seems clear that the layers of dolomite gave rise to layers of reddish jasperoid through a particle-by-particle (molecular or ionic?) selective exchange between carbonate and silica, whereas original layers of colorless chert remained virtually intact to form the light-colored laminae. North and northeast of Grace Lake (sec. 13, T. 47 N., R. 26 W.), interbedded quartzite and jasperoid are extensively associated with laminated jasperoid that is almost certainly silicified dolomite. For this reason the interbedded quartzite and jasperoid probably represent the selective silicification of dolomite layers of original interbedded quartzite and dolomite (Gair and others, 1961b).

Dolomitic rock probably was more readily silicified than slate; in fact, the evidence is not clear whether slate was replaced in any sense of chemical exchange, or merely displaced by silica.

Silicified dolomite in many partly silicified outcrops is sharply limited by bedding surfaces or is localized along cross fractures (fig. 18). These avenues were the principal routes for the movement of silica close to sites of replacement, and even where silica replaced entire beds of dolomite grain by grain, the access provided by bedding surfaces and cross fractures must have accelerated the process enormously. Breccia zones, although they have a more limited distribution than bedding surfaces or cross joints, also were important passageways for silica-bearing solutions. The widespread association of breccia and secondary silica indicates that in many places silicification was almost entirely localized along or adjacent to breccia zones.

The most extensively silicified parts of the Kona are in the eastern part of the area, mainly along or close to the contact of the Kona with the Mesnard (west-central part of sec. 35, T. 48 N., R. 25 W.; north-central part of sec. 1, T. 47 N., R. 25 W.) or to faults (SW¼ sec. 2, SE¼ sec. 3, T. 47 N., R. 25 W.) along which the Kona was moved against gneissic basement rocks (Gair and others, 1961b). Because of these occurrences of silicification, it seems that slippage zones along the Kona-Mesnard contact and the indicated faults provided major access routes for silica-bearing solutions. The age of much, if not all, of the silicification in the eastern part of the area thus was late or postdeformational, but the failure to find fragments of silicified dolomite in Cambrian rock of the area precludes placing a younger time limit on the silicification.

That silica-bearing solutions were available is of course obvious, but the source of the solutions is unknown. Silica could have moved downward from an overlying erosion surface, according to Leith's concept (1925), but the thorough bed-by-bed replacement of thick masses of dolomite, the lenticular knots of coarse-grained quartz along foliation in Kona slate, and the upward-wedging quartz veins in basement rock a few feet from massive white vein quartz that replaces the Kona probably constitute better evidence for upward or lateral movements of (heated) solutions in the eastern part of the area. The ultimate source or sources of the silica in this area remain unknown, and the question can only be asked: Did silica come from nearby rock, for example from leaching of underlying tonalite or Mesnard Quartzite, from a deep and obscure igneous source, or by some devious artesian route from the leaching of iron-formation, 10–15 miles to the west?

In the western part of the area, except for isolated silicified breccias, silicification is not clearly or generally associated with faults or other definable zones of rock movement. Much of the silicification there might be of the type suggested by Leith, that is related to the surface representing Cambrian to Pleistocene erosion.

ORIGIN AND CORRELATION

Algal structures, oolites, and scattered grains of detrital quartz indicate deposition of the Kona in generally shallow water. By analogy with carbonate forming in modern seas and with widely accepted ideas on the formation of nonfossiliferous limestone and silica gels, the original carbonate of the Kona probably was precipitated from sea water or precipitated alternately with silica (chert). All the carbonate identified is nearly pure dolomite, and the recrystallized fabric offers no sign of replacement of one carbonate by another. Therefore, whether the original Kona carbonate was calcite or dolomite remains unknown.

The Kona rests conformably on the Mesnard Quartzite and is considered to be directly correlative with the Randville Dolomite of Dickinson and Iron Counties, Mich. (James, 1958, p. 35). The Kona, therefore, is of early Animikie age in the middle Precambrian rock sequence.

WEWE SLATE

NAME, DISTRIBUTION, AND THICKNESS

The Wewe Slate was named by Van Hise and Bayley (1895, p. 530–540) from the Wewe hills in secs. 22 and 23, T. 47 N., R. 26 W., west of the Sands quadrangle. Subsequently (Van Hise and Leith, 1911, pl. 17), the type Wewe rocks of that area were mapped as Mesnard Quartzite.

Besides the dominant slate, the formation contains sericitic quartzite, wacke, and conglomerate. Most of the nonslaty rocks, however, occur west of the area of this report. In the Marquette and Sands quadrangles the Wewe is almost entirely massive (nonlaminated) and laminated slates. The most extensive outcrops occur in the axial part of the synclinorium, along the Carp River valley between Carp River Lake in the SW¼ sec. 5 and the mouth of Morgan Creek in the north-central part of sec. 4, T. 47 N., R. 25 W. Other good exposures are in the axial part of the synclinorium, north of the center of sec. 5, in the SE¼ sec. 6, and in the NW¼ sec. 8, T. 47 N., R. 25 W., on the north flank of the synclinorium in the SW¼ sec. 31, T. 48 N., R. 25 W., north of the center of sec. 12, and near the SE. cor., sec. 13, T. 47 N., R. 26 W.

The maximum mapped (inferred) width of the Wewe is 3,000 feet on the north limb of the synclinorium in sec. 31, T. 48 N., R. 25 W., but this belt contains only a few small outcrops. Attitudes are obscure in the area of the most extensive single outcrop along the Carp River in the NW¼ sec. 4, T. 47 N., R. 25 W., but the inferred structural trend there indicates that only a small thickness of slate is exposed. The most reliable determination of thickness can be made north and northeast of Carp River Lake in the S½ sec. 5 where, assuming that the average dip is 40°, a thickness of 650 feet of Wewe is exposed across closely spaced outcrops. This thickness perhaps represents about two-thirds of the total thickness (approximately 900 feet) of the formation in that belt.

DESCRIPTION AND ORIGIN

Wewe Slate is light to dark gray, greenish gray, dull green, or salmon colored and generally is fissile. It weathers to dark gray, dull gray green, rusty gray or brown, and salmon. Rust-colored Wewe Slate has been prospected in the SW¼ sec. 31, T. 48 N., R. 25 W. Laminations generally are green and white, gray and rusty brown, or reddish brown and white.

The dominant minerals in most of the Wewe Slate are quartz, chlorite, and sericite. Dusty reddish-brown iron oxide and tiny needles of rutile occur widely in the formation but generally in small amounts. Leucoxene and magnetite are minor constituents in parts of the slate. Leucoxene and rutile together, however, compose about 10 percent of one thin section. Quartz occurs as fine-grained detritus or irregular cherty mosaics, generally scattered through a micaceous-chloritic ma-

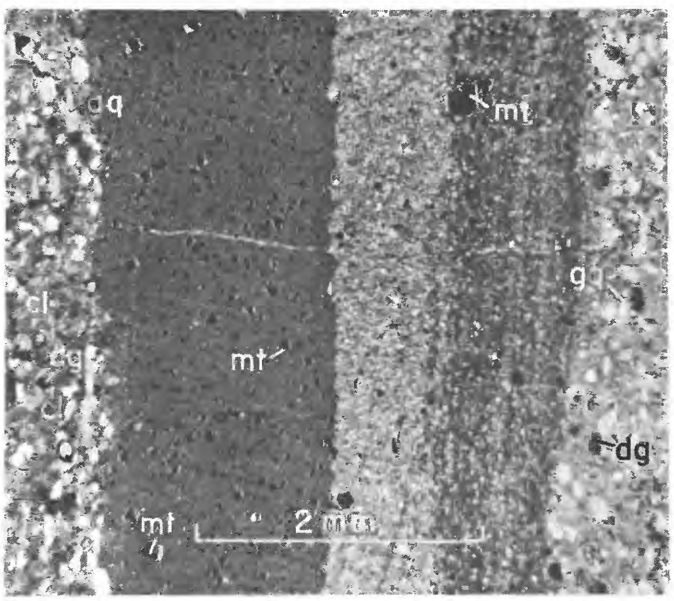

FIGURE 20.—Photomicrograph of laminated Wewe Slate, approximately 1,750 feet south of N¼ cor. sec. 5, T. 47 N., R. 25 W. Coarser grained layers contain granular aggregates of quartz (gq), other dark granules (dg) of unknown composition, and fine-grained detrital quartz (Q) and feldspar(?) set in matrix rich in chlorite (cl). Darkest layer contains much chlorite, very fine grained quartz, and fine-grained sphene(?). Grains of magnetite (mt) most abundant in darkest layer. Plane light.

trix, which tends to be foliated. Laminations are a result of varying concentrations of quartz, chlorite, or iron oxide (fig. 20). Ferric iron oxide generally occurs in reddish laminated slate and is derived mainly from the oxidation of chlorite, probably during weathering.

Although the Wewe Slate has been considered to be of detrital origin and although the bulk of the rock does appear to be so, the slate in at least three localities bears some indication of volcanic or partly volcanic origin. At 600N–800E, SW cor. sec. 31, T. 48 N., R. 25 W., a laminated greenish sericite-chlorite-chert slate containing much leucoxene and rutile has several textural features suggestive of volcanic origin. In some laminae, greenish chloritic material is somewhat aggregated into ovoid, lenticular, and angular granules which may have been fragments of mafic ash, and chert has vague oolitic or amygdular forms. Aggregates of sericite have vague lathlike forms, some of which crisscross as in diabasic texture, and some scattered quartz grains about 0.1 mm in diameter have concave or angular reentrants such as are common in partly resorbed quartz phenocrysts of lavas. At 250N–4,000E, SW. cor. sec. 31, some Wewe Slate contains distinct alined sericite-rich augen or lenticles in a foliated ferruginous quartz-sericite groundmass. The lenticles or augen strongly resemble amygdules. Sericite-rich zones also extend along cracks crossing foliation and along a thin quartz veinlet. Evidently sericitization of favored zones, including preexisting lenticles (amygdules), occurred during or after metamorphism. At this locality, also, other amygdule-like features contain oxidized or partly oxidized chlorite rimmed by sericite, and some extensively silicified slate has tabular and wedge-shaped quartz fragments (pseudomorphs?) of pyroclastic aspect. Laminated green and white cherty slate approximately 1,750 feet south of the N¼ cor. sec. 5, T. 47 N., R. 25 W. (fig. 21), contains considerable leucoxene and rutile, lath-shaped sericite aggregates, quartz pods that resemble amygdules (fig. 22), and granular structures of obscure origin, possibly ash fragments modified by recrystallization and deformation. Some light-colored chert layers, ¼–½ inch thick, in the laminated green and white slate in sec. 5 have wavy surfaces against thicker (4–5 inches) greenish layers. Thin seams of reddish-purple iron oxide tend to be concentrated along the surfaces of the chert layers. Abrupt terminations of chert layers suggest some preconsolidation slumping.

The slate from sec. 5 has been analyzed chemically and has a composition (table 12) that is somewhat abnormal for argillaceous rock of entirely detrital origin (see Pettijohn, 1957a, p. 342–346; 357–368, table 61, figs. 87, 88; Shaw, 1956, tables 2, 9, 10). Particularly noteworthy are the relatively large amounts of FeO, total iron oxide, MgO, and TiO_2, and the relatively small amount of K_2O, which, in conjunction with the textural features just cited, are probably best explained as a result of the mixing of mafic volcanic material with detritus.

TABLE 12.—*Chemical analysis of Wewe Slate, Marquette quadrangle*

[3,400N–2,800E, SW. cor. sec. 5, T. 47 N., R. 25 W. Rapid rock analysis. Paul Elmore, I. H. Barlow, S. D. Botts, and G. W. Chloe]

SiO_2	59.0	K_2O	1.0
Al_2O_3	12.8	H_2O	4.7
Fe_2O_3	1.2	TiO_2	1.0
FeO	10.1	P_2O_5	.11
Mgo	5.4	MnO	.22
CaO	.92	CO_2	1.4
Na_2O	1.5	Total	99.35

CORRELATION AND AGE RELATIONS

The Wewe Slate overlies the Kona Dolomite conformably and is overlain disconformably or with slight angular unconformity by the Ajibik Quartzite. Its stratigraphic position is equivalent either to the upper part of the Randville Dolomite or to the early part of the hiatus between the Randville and Felch Formations in central Dickinson County (James and others, 1961, p. 33–37).

MENOMINEE GROUP

The representative formations of the Menominee Group in the Marquette and Sands quadrangles are the Ajibik Quartzite and the Siamo Slate. These formations together correspond stratigraphically to the Felch Formation in the type area of the group in Dickinson County (James and others, 1961, p. 36–39). The upper formation of the group, the Negaunee Iron-Formation, occurs in normal sequence west of the mapped area.

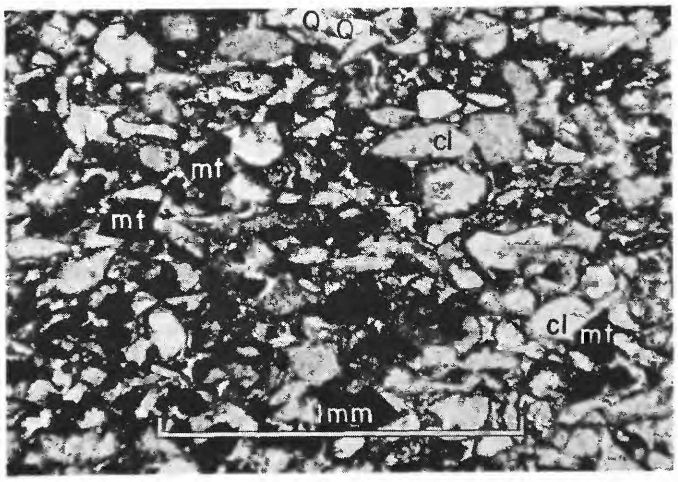

FIGURE 21.—Photomicrograph of granular structure in laminated Wewe Slate, approximately 1,750 feet south of N¼ cor. sec. 5, T. 47 N., R. 25 W. This layer contains irregular granules of chlorite (cl) separated by opaque seams of iron oxide, scattered euhedra of magnetite (mt), and a few grains of clastic quartz (Q). Plane light.

AJIBIK QUARTZITE

Name, distribution, and thickness.—The Ajibik Quartzite was named by Van Hise and Bayley (1895, p. 540–554) for the Ajibik Hills northeast of Palmer, about 2½ miles west of the Sands quadrangle, and west-southwest of Grace Lake in the SE¼ sec. 13, T. 47 N., R. 26 W. Most of the formation in the Marquette and Sands quadrangles consists of massive quartzite. Slate, thin-bedded quartzite, siltite, and conglomerate are minor components. Massive quartzite is well exposed on the north limb of the synclinorium near the N¼ cor. sec. 6, in the axial part of the synclinorium, in the NW¼ sec. 5, the SE¼ sec. 6, and the north-central part of sec. 7, T. 47 N., R. 25 W. Thin slaty phases are interbedded with quartzite near the N¼ cor. sec. 6, and on the south side of County Highway 480 at 3,700N–2,000E, SW cor. sec. 7, T. 47 N., R. 25 W. Some 60 feet of slate that is probably interbedded with quartzite is exposed in the steep north-facing slope at the west end of the large outcrop of quartzite, at 1,100N–2,950E, SW cor. of sec. 6. The small amount of exposed slate is probably not representative of the amount actually present in the formation. Thin-bedded quartzite, siltite, and conglomerate occur along the north-facing cliff just south of the Carp River in the vicinity of 1,000N–3,500E, SW. cor. sec. 6, T. 47 N., R. 25 W., and a thin pebbly zone occurs in the long narrow outcrop of Ajibik along the north side of the low east-trending ridge in the NW¼ SW¼ sec. 5, T. 47 N., R. 25 W. Silty quartzitic layers have also been noted in the large outcrop of quartzite in the north-central part of sec. 6, and conglomerate, or pseudoconglomerate (breccia), has been found in quartzite in the SW¼NW¼ sec. 5. The conglomerate in the SE¼ sec. 6 occurs at the base of the formation.

The thickness of the Ajibik is best measured across the south-trending belt in the SE¼ sec. 6 and the N½ sec. 7, T. 47 N., R. 25 W., where the mapped width is approximately 1,500 feet and the average dip is 25° W. The calculated thickness is approximately 650 feet, which is very close to the thickness calculated by Van Hise and Bayley (1895, p. 553). The width of the belt of Ajibik in the NW¼ sec. 5 is approximately 2,500 feet, but as a result of second-order folds there, bedding to a considerable extent trends in the nominal cross-strike direction. The Ajibik belt on the north limb in the NE¼ sec. 6 is fairly well delimited, is about 600 feet wide, and dips about 70° S. The Ajibik beds, however, may be somewhat thinned there by an unmapped extension of the east-southeastward-trending fault that enters the north-central part of sec. 6.

Description.—The Ajibik Quartzite in most places in the area is lithologically indistinguishable from Mesnard Quartzite. It is a vitreous medium-grained clastic rock, mainly of light-gray, flesh, or tan color. It is massive without discernible bedding, or forms distinct beds, several inches to about 2 feet in thickness. Immediately above the basal conglomerate in the SE¼ sec. 6, on the south side of the Carp River, quartzite, siltite, and a little argillaceous rock form platy beds, ½–2 inches thick, upward through some 30 feet of the section, stratigraphically, to the base of massive thick-bedded quartzite. The quartzite generally his minor sericite and iron oxide (not more than 2 percent) and weathers white, light pink, brown, gray, or deep reddish brown. In places, yellowish or reddish-brown spots ¼–½ inch in diameter stain light-colored Ajibik, whereas similar discolorations are rare in the Mesnard. The stains are therefore a fairly reliable indicator of the Ajibik. Although the Ajibik is generally a little more ferruginous than the Mesnard, the variations in iron content in both quartzites are within virtually the same range and so are not distinctive. In at least one place, iron oxide has been formed by the oxidation of matrix chlorite, but because it is generally evenly distributed in the matrix and does not corrode quartz grains or replace other minerals, the iron of the Ajibik appears to have been deposited with the rock. Siltite resembles the common medium-grained quartzite of the Ajibik in all respects but that of grain size.

Slate in the Ajibik is light gray or gray green, weathers light tan, grayish brown, or red, and is highly sericitic. It is seen only in thin seams or lenses, less than 6

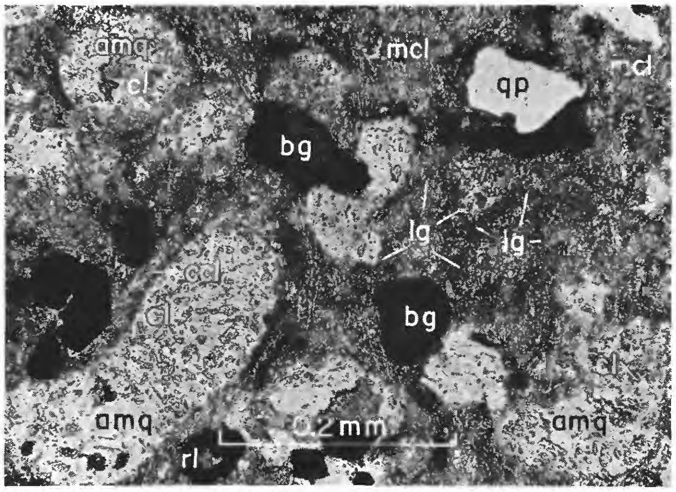

FIGURE 22.—Photomicrograph of granular structure in Wewe Slate, approximately 1,750 feet south of N¼ cor. sec. 5, T. 47 N., R. 25 W. Enlargement of upper part of layer on right side of figure 20. Shows amygdulelike aggregates of secondary quartz (amq) and chlorite (cl), brownish granules (bg) of unknown composition, ringlike structure of leucoxene (rl), and possible quartz phenocryst (qp) rimmed by leucoxene(?). Large elongate granule in lower left part of photograph has narrow rim of chert and chlorite (ccl). Bulk of matrix is chlorite (mcl). Two vague granules outlined by brownish leucoxene(?) (lg) ringing chlorite lie immediately to right of center of photograph. Plane light.

inches thick, except for the one area mentioned, in the SE¼ sec. 6, which possibly is Siamo Slate faulted in from across the Carp River to the north.

The conglomerate at the base of the Ajibik in the SE¼ sec. 6 is 1–2 feet thick, and consists mainly of angular fragments of dark-gray Wewe Slate and light-colored chert in a quartzitic and slate-fragment matrix. Some conglomeratic fragments occur in the thin-bedded quartzite for the first few feet above the bed of conglomerate. The conglomerate also contains some rounded pebbles and small cobbles of quartzite or vein quartz. The conglomerate weathers to a grayish rusty brown, and on weathered surfaces the conglomeratic nature of the rock is quite obscure. Within the limits of this outcrop, the bed of conglomerate appears to be parallel to beds of the underlying Wewe Slate, but irregularities in the surface with about 6 inches of relief between the conglomerate and the Wewe indicate some erosion of the Wewe prior to deposition of the Ajibik.

Relationships to underlying formations of the Animikie Series.—The relationship of the Ajibik to underlying rocks has been of considerable significance in the growth of stratigraphic concepts for the entire Lake Superior region as well as for the Marquette iron range.

Van Hise and Bayley (1895, p. 550; 1897, p. 294, 307) noted the conglomerate, which had been discovered by A. E. Seaman, at the base of the Ajibik in the SE¼ sec. 6, but they thought that it represented only slight local uplift and erosion and so placed the Ajibik in the middle of their Lower Marquette Series. They believed that from Mesnard through Ajibik time the sea transgressed westward across the area; consequently the lower part of the Ajibik might have been deposited near shore at the same time that the upper part of the Wewe was being deposited in deeper water to the east (1897, p. 295, 304). Later this idea was abandoned when Seaman's discovery of conglomerate at the base of the Ajibik east of Teal Lake near Negaunee became the main reason for subdividing the Lower Marquette Series into two parts, the lower and middle Huronian, which were correlated with the type Huronian section in Ontario (Van Hise and others, 1905, p. 91). Evidently the conglomerate in the SE¼ sec. 6 was part of the evidence for the unconformity separating lower Huronian and middle Huronian rocks, although specific mention of the conglomerate at this locality does not appear in published literature after the 1897 report by Van Hise and Bayley.

Tyler and Twenhofel (1952) decided that all conglomeratic rocks at the base of the Ajibik east of Teal Lake were tectonic features, boudinage or breccia, and expanded Van Hise and Bayley's concept of the possible time equivalence of the lower part of the Ajibik and the upper part of the Wewe. Tyler and Twenhofel considered the Mesnard, Kona and Wewe as virtually time-equivalent rocks, in effect, different facies of the same formation, which were deposited during the time that the Ajibik west of Teal Lake was being deposited. They placed the Ajibik east of Teal Lake and in the Marquette and Sands quadrangles conformably above their Mesnard, Kona and Wewe unit, evidently as a regressive depositional facies of the Ajibik to the west. Inasmuch as conglomerate beds marking unconformities may or may not cross time lines, and as there are no other possible time markers in these layered rocks, the conclusions of Tyler and Twenhofel apparently were based on (1) theoretical considerations and (2) an assumption that different lithologic units adjacent to one another in a general "on-strike" direction grade into one another. Structural details and careful mapping in the Marquette and Sands quadrangles and in the Palmer quadrangle, west of the Sands quadrangle, generally refute the idea of lithologic gradations of the magnitude they assumed.

Mapping in the Marquette and Sands quadrangles has established only a slight disconformity or very low angle unconformity at the base of the Ajibik. Whether this discontinuity is area wide cannot be asserted on the basis of our work so far. The solution to this problem must at least await detailed mapping in the quadrangles to the west, particularly in the critical belt extending 3 miles eastward from Teal Lake.

Correlation and age relations.—In the absence of a definite angular discordance between the Wewe and the Ajibik in the SE¼ sec. 6, the platy fragments of Wewe Slate and chert alone in the conglomerate at the base of the Ajibik might be construed as being no more significant than an intraformational conglomerate. The rounded quartzose pebbles and small cobbles, however, indicate a definite erosional interval in the eastern part of the Marquette Range between Wewe and Ajibik time. This erosional interval obviously was induced by only a slight crustal disturbance and probably corresponds to the definite hiatus between Palms Quartzite and Bad River Dolomite in the Gogebic Range (Van Hise and Leith, 1911, p. 228, 230; Aldrich, 1929, p. 79, 84) and to the inferred interval between the Felch Formation and Randville Dolomite in Dickinson County (James and others, 1961, p. 37, 39). The Ajibik, therefore, is here considered equivalent to the Palms Quartzite and to the lower part of the Felch Formation. These rocks are of early middle Animikie age (James, 1958, table 1).

SIAMO SLATE

Name, distribution, and thickness.—The Siamo Slate was named by Van Hise and Bayley (1895, p. 554) for the Siamo Hills (not named on latest topographic maps of U.S. Geological Survey) near the west end of Teal Lake. The formation is believed to lie conformably on the Ajibik Quartzite and is mapped from the top of massive light-colored quartzite to the stratigraphically lowest beds of Negaunee Iron-Formation. The iron-formation, however, occurs only west of the area of this report, so the Siamo is the youngest formation mapped; it occupies the axial part of the Marquette synclinorium in the western adjoining parts of the Marquette and Sands quadrangles.

The formation is poorly exposed in most of the area, and the contact between it and the Ajibik must be projected close to outcrops of Ajibik from a few well-controlled locations, such as in the north-central parts of secs. 6 and 7, T. 47 N., R. 25 W. and the SE¼ sec. 6, T. 47 N., R. 25 W., and in the north-central part of sec. 12, T. 47 N., R. 26 W. The best exposure is near the northwest corner of the Sands quadrangle.

The sparse outcrops, the presence of small folds in the axial part of the synclinorium, the lack of marker beds, and the absence of the upper contact of the Siamo in the area studied preclude a determination of thickness. Van Hise and Bayley (1895, p. 560–561) estimated the thickness of the Siamo to be between about 600 and 1,250 feet.

Description.—A full description of the Siamo must await completion of mapping in the quadrangles west of the present area. The Siamo in the Marquette and Sands quadrangles is mainly laminated and nonlaminated dark argillaceous slate having subordinate interbedded dark quartz-rich rocks. The slate commonly weather to rusty brown; as a result, they may superficially resemble iron-formation, especially where outcrops have been broken into piles of oxidized rubble by weathering and by glacial erosion. Such rubbly ferruginous slate is found in several test pits near 4,100N–4,500E and at the dirt road at 4,150N–3400E, SW. cor. sec. 6, T. 47 N., R. 25 W.

The slates have strongly foliated or matted fabrics of sericite or sericite-chlorite and fine-grained quartz (detritus, metachert?) (fig. 23). Leucoxene, rutile, and pyrite are minor constituents in some of the slate. Small granules and dust-size particles of reddish and yellow-brown iron oxide typically compose ½–3 percent of the content of the exposed slate, although locally considerably more iron oxide is present. It is difficult to determine how much of the iron oxide was derived by weathering of chlorite and pyrite or introduced during weathering and how much is primary. Future studies of subsurface Siamo in the area of mining to the west should reveal this. In a few places, hematite granules are aggregated in spots pseudomorphs of an earlier mineral are thus suggested.

Thin laminations and lenses (mostly 1 mm to about 1 cm thick) and thicker beds (several inches to as much as 8 feet thick) of the subordinate quartz-rich rocks alternate with the slate. They consist mainly of angular to subrounded detrital quartz (0.1–0.4 mm) and argillaceous or metachert matrix material. Fragments of metachert and a sericite slate(?) compose as much as 2 percent of the detritus. These interbeds have detrital textures or detrital textures modified by metamorphism, especially by quartz overgrowths. Depending on the proportion of larger fragments to matrix, as well as on the composition of the matrix, the interbeds can be classified as sericitic quartzite, sericitic-chloritic quartzite, wacke, metachert quartz wacke, or graywacke (nonfeldspathic). Like the wacke and graywacke of the Enchantment Lake Formation, such rocks in the Siamo are characterized by medium- or fine-grained detritus "floating" in a matrix of finer grained sericite, sericite-chlorite, metachert, or mixtures of these substances. These rocks are generally dark gray, greenish gray, or nearly black where fresh, but they become light pink, red, or shades of brown or reddish brown from weathering, particularly from the oxidation of chlorite.

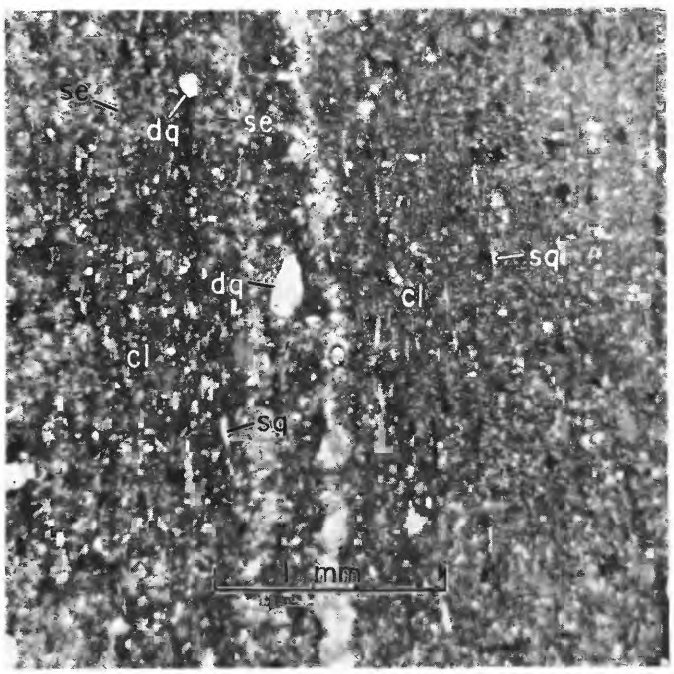

FIGURE 23.—Photomicrograph of Siamo Slate, 550N–350E, SW. cor. of Marquette quadrangle. Rock consists of fine-grained well-alined chlorite (cl) mixed with very fine grained cherty quartz, thin layers of fine-grained detrital quartz (Q), thin alined lenticles of secondary quartz (sq), scattered isolated detrital grains of quartz (dq), thin alined blades of sericite (se), and scattered granules of leucoxene and pyrite (opaque). Plane light.

LOWER(?) PRECAMBRIAN INTRUSIVE ROCKS

Sills and dikes of felsic porphyry and granitoid dikes cut the Mona Schist and are probably offshoots of felsic components of the Compeau Creek Gneiss. The gneiss, which is interpreted as being composed of a mixture of intrusive material and pregneiss country rock, has not been included with "intrusive rocks."

FELSIC PORPHYRY AND RELATED GRANITOID ROCK

General features.—Thin sills or dikes of felsic porphyry and rock of granitic appearance follow the schistosity layering or cut it at generally low angles in the Lighthouse Point Member of the Mona Schist. The best exposures of felsic porphyry in the area are on Lighthouse Point (pl. 6). Most of the sills or dikes are less than 15 feet thick, and many are less than 3 feet thick. Strictly conformable bodies are presumed to be intrusive because they resemble those that crosscut or pass from a conformable to a crosscutting relationship. Some of the conformable bodies, however, may well be silicic metavolcanic rock intercalated with the amphibole schist. Tonalitic rock in thin dikes in the S½ secs. 15 and 16, T. 48 N., R. 25 W. (some of this granitoid rock was described in the "Brook section" of Williams, 1891, p. 143–145) is mostly granitoid to granitoid porphyritic in texture; some, however, also is similar to typical felsic porphyry (fig. 24), and both may have come from a common magmatic source. Several granitic-looking dikes cut the Mona Schist in the south part of the Harvey quarry, NW¼ sec. 6, T. 47 N., R. 24 W, and a tonalite dike cuts the greenstone in a small quarry overgrown with brush near the center of sec. 1, T. 47 N., R. 25 W. These rocks my be equivalent to the felsic porphry in age and parent magma.

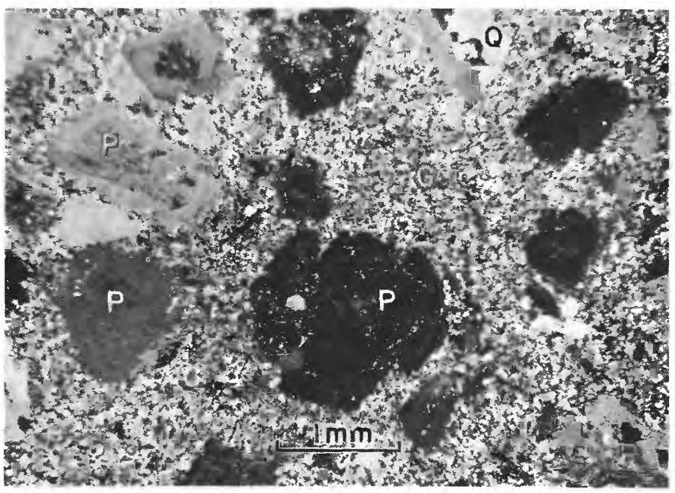

FIGURE 24.—Photomicrograph of felsic porphyry, 2,050N, SW. cor. sec. 15, T. 48 N., R. 25 W. Phenocrysts of sodic plagioclase (P) and quartz (Q) set in fine-grained groundmass (G). Some plagioclase is zoned and some is twinned. Quartz phenocrysts either were aggregates in magma or have recrystallized since the time of intrusion. The groundmass is mainly quartz and albite but contains minor sericite, hornblende, and chlorite. Crossed nicols.

Description.—The felsic porphyry and granitoid-textured dikes are gray, pale greenish gray, or mottled gray and salmon, and weather to pink. The porphyry is characterized by medium- to coarse-grained (0.5–3.5 mm) phenocrysts of feldspar and quartz set in a slightly to moderately foliated groundmass of fine-grained quartz, feldspar, and sericite (fig. 24). The ratio of phenocrysts to groundmass ranges from about 1:6 to 2:3. Feldspar phenocrysts are mainly sodic plagioclase (approximately An_{10-20}) and typically compose 20–30 percent of the rock, whereas phenocrysts of potassium-feldspar and quartz are generally less than 5 percent. At 3,100N–2,400E, SW. cor. sec. 17, T. 48 N., R. 25 W., however, a porphyritic granitoid-textured light-colored rock intrusive into the Mona Schist has 40 percent microcline and only 20 percent plagioclase. Amphibole, chlorite, and carbonate are commonly concentrated along shears and make up not more than a few percent of the material. Between shears (foliation surfaces) the fabric of the groundmass is a granoblastic mosaic. Foliation surfaces typically bend around phenocrysts to produce a microaugen structure.

Most of the felsic porphyry and granitoid-textured dike rocks in the area have a composition close to that of tonalite or granodiorite, as is shown both by the composition of the phenocrysts and by two chemical analyses (table 13, Nos. 3, 4).

TABLE 13.—*Chemical analyses of intrusive rocks, Marquette quadrangle*

[Nos. 1 and 2, standard rock analyses, Dorothy F. Powers. Nos. 3–8, rapid rock analyses, Paul Elmore, I. H. Barlow, S. D. Botts, and G. W. Chloe. Nos. 7 and 8, F and Cl contents, standard rock analyses, Vertie C. Smith]

	1	2	3	4	5	6	7	8
SiO_2	49.72	49.71	72.3	66.3	50.2	51.9	57.2	62.0
Al_2O_3	10.11	13.05	15.4	15.8	13.5	15.1	14.5	19.1
Fe_2O_3	1.88	3.00	.6	1.0	1.2	2.3	1.8	1.7
FeO	10.25	6.80	.52	1.5	8.2	5.2	4.1	.6
MgO	9.90	9.58	.51	1.5	9.0	8.5	6.1	.43
CaO	10.10	8.98	1.2	2.9	7.0	4.2	5.0	1.2
Na_2O	2.58	2.84	5.7	5.2	4.1	3.3	3.7	5.7
K_2O	.30	1.45	2.2	2.3	1.3	4.0	3.0	6.7
H_2O+	2.73	2.41	.88	1.5	2.7	3.5	2.5	.79
H_2O-	.13	.15						
TiO_2	1.06	1.14	.16	.47	1.1	.87	.76	.59
P_2O_5	.07	.09	.04	.19	.35	.38	.55	.05
MnO	.21	.17	.04	.07	.18	.17	.12	.04
CO_2	.94	.47	.05	1.4	.87	.31	.30	.65
S	.02	------	------	------	------	------	------	------
F	------	------	------	------	------	------	.13	.16
Cl	------	------	------	------	------	------	.01	.02

1. Metadiabase, JG–68–58, 1,300N–700W, SE. cor. sec. 24, T. 48 N., R. 26 W.
2. Metagabbro, JG–60–58, Picnic Rocks at 2,400N–1,900E, SW. cor. sec. 13, T. 48 N., R. 25 W.
3. Felsic porphyry, JG–12–59, 2,050N, SW. cor. sec. 15, T. 48 N., R. 25 W.
4. Felsic porphyry, JG–26–59, 4,900N–3,500E, SW. cor. sec. 17, T. 48 N., R. 25 W.
5. Hornblende lamprophyre, BJ–122B–57, 1,750N–2,100E, SW. cor. sec. 30, T. 48 N., R. 25 W.
6. Hornblende lamprophyre, BJ–124G–57, 1,500N–2,200E, SW. cor. sec. 30, T. 48 N., R. 25 W.
7. Porphyritic hornblende syenite, JG–70A–59, 4,200N–2,200E, SW. cor. sec. 5, T. 48 N., R. 25 W.
8. Syenite, RT–9–59, 550N–1,800E, SW. cor. sec. 3, T. 48 N., R. 25 W.

Origin and age relations.—The felsic porphyry cannot be related to other silicic igneous rock of the area with certainty. The texture of the porphyritic granitoid rock in secs. 15 and 16, T. 48 N., R. 25 W., is intermediate between that of the "granitic" rock of the Compeau Creek Gneiss and that of the typical felsic porphyry. The felsic porphyry may therefore be a textural variety of the tonalitic dikes, which almost certainly were emplaced during formation of the Compeau Creek Gneiss.

On Lighthouse Point, felsic porphyry is truncated by metadiabase (pl. 6), which was probably emplaced after the development of foliation and blue-green hornblende in the mafic metavolcanic country rock. This metamorphism of the mafic country rock on Lighthouse Point has been attributed (see p. 17) to deformation and a rise in temperature associated with the formation of the gneiss. Accordingly, the felsic porphyry is interpreted here as lower Precambrian synkinematic intrusions related to the formation of the gneiss.

LOWER OR MIDDLE PRECAMBRIAN INTRUSIVE ROCKS

The intrusive rocks described here include metadiabase and metagabbro, metapyroxenite, hornblende lamprophyre, syenites, pegmatite, aplite, and quartz veins. The pegmatite, aplite, and quartz veins are less certainly of early or middle Precambrian age than the other intrusive rocks. The intrusive rocks form discrete bodies that either crosscut or are conformable with the structure of surrounding country rock. The crosscutting relationships of several types of intrusive rock, both to the country rock and to one another, are best exposed on Lighthouse Point (pl. 6).

METADIABASE AND METAGABBRO

DISTRIBUTION

Metadiabase and metagabbro are widely exposed in the areas underlain by lower Precambrian rocks. They form dikes in the Compeau Creek Gneiss, and dikes, sills, and irregularly shaped bodies in the Mona Schist. Relatively few intrusive bodies of altered mafic rock occur in the areas of Animikie metasedimentary rocks. Most of these are in the Kona Dolomite, in the following locations: the NW¼NW¼ sec. 33, NE¼SW¼ and SW¼ NW¼ sec. 34, T. 48 N., R. 25 W.; the SE¼NW¼ sec. 3, SW¼ sec. 7, and SW¼NW¼ sec. 8, T. 47 N., R. 25 W. Metadiabase cuts the Enchantment Lake Formation at 4,100N–4,000E., SW. cor. sec. 3 and near the SW. cor. sec. 9, T. 47 N., R. 25 W., and cuts the Wewe Slate in SE¼NW¼ sec. 8.

A small dike, rich in carbonate minerals and leucoxene, branches from a dike of metadiabase in the SW¼ SW¼ sec. 9, T. 48 N., R. 25 W., and a pod about 50 feet long, rich in carbonate and leucoxene and containing faint relic plagioclase laths, is located in the Enchantment Lake Formation about 1,600 feet south and 150 feet west of the center of sec. 29, T. 48 N., R. 25 W. These bodies may represent metadiabase or diabase that has been almost completely replaced by carbonate minerals.

A thin highly schistose chloritic dike (fine-grained metadiabase?) appears to cut both gneiss and silicified dolomite and to cross the fault between them in the SW¼ sec. 2, T. 47 N., R. 25 W. The fault and the silicification are probably no older than the post-Animikie, pre-Keweenawan interval (p. 44), so the dike evidently was emplaced after late Animikie time. The emplacement of the dike after Animikie time is difficult to explain in view of the evident metamorphic condition of the dike rock. Possibly the dike was of Keweenawan age and its alteration was influenced by shearing during emplacement.

DESCRIPTION

Bodies of altered mafic intrusive rock range in size from dikes less than 1 foot thick to dikes and less regularly shaped masses as much as 400 feet thick. The plug of metagabbro forming Picnic Rocks in the central part of sec. 13, T. 48 N., R. 25 W., is probably at least 2,000 feet in diameter. The rocks generally are dull shades of green and weather to dark gray, greenish gray, and nearly black. Much of the coarse-grained metagabbro at Picnic Rocks, as well as metagabbro and metadiabase in several other places, is mottled dark-green and salmon colors, the latter color resulting from tiny inclusions of hematite in partly altered plagioclase. Mineralogically and chemically (table 13, Nos. 1 and 2) most of the metadiabase and metagabbro is similar to the metabasalt of the Mona Schist described above (p. 8–11; table 2), and mineralogically, chemically, and texturally they are virtually identical with altered mafic instrusive rocks from elsewhere in northern Michigan (James, 1955, p. 1468–1470, fig. 2 of pl. 4; Gair and Wier, 1956, p. 61; Bayley, 1959, p. 71). The common replacements of the original minerals, many of which form pseudomorphs, are tremolite-actinolite and saussuritic mixtures of sodic plagioclase (An_{5-15}), epidote, quartz, sericite, and chlorite (N_Y=1.629 in one sample). The indices of refaction, N_X–N_Y–N_Z, for actinolite-tremolite from several samples are 1.637–1.653–1.661, 1.621–1.633–1.652, 1.618–1.633–1.651, 1.630–1.640–1.656, 1.635–1.652–1.657, and 1.635–1.646–1.656. Ilmenite is widely altered to leucoxene, and in one sample, grains of banded sphene (identified by X-ray; analyst, Fred Hildebrand, U.S. Geological Survey) and hematite are interpreted as an alteration of a primary intergrowth of ilmenite and magnetite.

Thin metamorphosed mafic sills with offshooting dikes in the outcrop south of Wetmore Landing (pl. 3) are hornblende-rich rocks.

Aside from a few thin dikes that are foliated parallel to their trends, the metamorphosed mafic dikes are massive and retain recognizable parts of igneous fabrics. Textures range from fine-grained intergranular and diabasic in the thinner dikes to coarse grained subophitic, gabbroic, and poikilitic in the thicker masses but have been modified to some extent by recrystallization.

The metagabbro that forms Picnic Rocks is somewhat unusual in that it forms a relatively large pluglike mass, is coarser grained and more feldspathic, and probably contained some primary amphibole. It is rich in amphibole ($N_X1.641–N_Y1.654–N_Z1.662$) and partly epidotized plagioclase (approximately An_{10-12}) and has 1–2 percent microcline. Sphene, carbonate, magnetite, and apatite are minor minerals. Small veins and clots of syenite in the metagabbro probably are differentiates of the gabbroic magma. Some idiomorphic amphibole crystals are molded against plagioclase; this fact suggests that such amphibole originated by crystallization from a melt rather than by replacement of pyroxene. Pyroxene seen at the core of one amphibole crystal, however, indicates that at least some of the amphibole formed from pyroxene during metamorphism. Individual grains of amphibole typically have mottled pleochroism, possibly because of failure to reach an equilibrium composition during metamorphism.

Near the southwest edge of the body, adjacent to the parking lot of the Lake Shore Engineering Co., the intrusive rock is in contact with fine-grained greenstone and has considerable pyrite along the contact. Parts of the metagabbro in that vicinity are highly feldspathic (An_{30-43}), have coarse euhedra of amphibole that have mottled pleochroism, and are cut by some stringers of pegmatitic metagabbro. Some biotite is intergrown with (replaces?) the amphibole, and a little microcline-perthite, apatite, magnetite, carbonate, and hematite are present in the groundmass. An unusual feature is the presence of sphene-epidote reaction rims around ilmenite(?) against plagioclase. The euhedral form of the amphibole molded against plagioclase here also suggests that the amphibole originated by crystallization from a melt rather than by replacement of pyroxene during metamorphism.

The one available chemical analysis, of a sample from the largest island at Picnic Rocks (table 13, No. 2), indicates a gabbroic rather than dioritic composition, however, from which the primary dark mineral is deduced to have been pyroxene. Perhaps the rock near the southwest edge of the body, and some of that on the Picnic Rocks, crystallized from magma rich in residual fluids which yielded primary amphibole and intermediate plagioclase that were little affected by subsequent metamorphism.

METAMORPHISM AND AGE RELATIONS

Mafic intrusive rocks are sensitive to temperature changes. Most of the metadiabase and metagabbro is almost entirely reconstituted, mineralogically, and most is at least partly changed, texturally, from original diabase and gabbro. Mineral assemblages are generally typical of the chlorite zone or the lower part of the greenschist facies. All these rocks must have been emplaced prior to the last major metamorphism of the area, which occurred during the post-Animikie–pre-Keweenawan interval and which produced the present mineral assemblages. Locally, as where amphibolite of the Compeau Creek Gneiss or metavolcanic rock containing blue-green hornblende in the Lighthouse Point member of the Mona Schist is cut by the altered mafic dikes, the metamorphic grade of the host rocks appears to be higher than that of the dikes. (Small amounts of blue-green hornblende in metadiabase on Lighthouse Point are attributed to the contact thermal effect of nearby diabase of Keweenawan age.) Thus, these dikes appear to postdate higher grade metamorphism of the host rocks, evidence of which was not entirely destroyed during the post-Animikie, pre-Keweenawan metamorphism. Metadiabase and metagabbro in the lower part of the Mona, however, have the same low metamorphic grade as their host rocks, and so cannot be dated in this manner.

The metadiabase and metagabbro are probably of at least three different ages: (1) Metagabbro that appears to be cut by and to have contributed inclusions to tonalitic dike rock in two places (1,300N–3,900E, SW. cor. sec. 16, and 1,300N–200E, SW. cor sec. 15, T. 48 N., R. 25 W.) is evidently older than the Compeau Creek Gneiss. (2) Two dikes of metadiabase in the west-central part of the Marquette quadrangle appear to cut both Mona Schist and Compeau Creek Gneiss and so are younger than the gneiss. Metadiabase dikes cutting the Lighthouse Point Member of the Mona Schist (p. 17; pl. 6) probably also are younger than the gneiss, inasmuch as they cut intrusive porphyritic felsite which is thought to be correlative with the gneiss (p. 51). (3) Mafic metaintrusive rocks in the gneiss south of Wetmore Landing (pl. 3) appear to be intermediate in age between the first two. They cut the foliation of the gneiss and thus are younger than the foliation, but they are rich in hornblende. This composition indicates that they were metamorphosed at the amphibolite level either at the same time as the gneiss or some later time, but prior to the metamorphism of other known mafic dikes in the gneiss or on Lighthouse Point. The intrusion of most of the metadiabase, however, cannot yet be more accurately dated than the time interval between the development of the respective host rocks and the last metamorphism.

METAPYROXENITE

Three isolated occurrences of metapyroxenite have been found in the main belt of Mona Schist in the Marquette quadrangle. They are in the SE¼SE¼ sec. 24, T. 48 N., R. 26 W., in the west-central part of sec. 19 T. 48 N., R. 25 W., and a short distance northeast of the center of sec. 21, T. 48 N., R. 25 W. The occurrences in secs. 24 and 19 are located only by letter symbol on plate 1; the metapyroxenite in sec. 21 is shown only in outcrop.

The bulk of the metapyroxenite now consists of amphibole (75–95 percent). Data on the amphibole from one sample are N_X1.622–N_Y1.634–N_Z1.654; 2V, 70°–78° (−); $Z \wedge c$, 14°; pleochroism, pale straw yellow (X) to light grass green (Z). The primary mafic mineral is assumed to have been pyroxene because amphibole pseudomorphs of pyroxene were found in one sample and because the bulk of the minerals other than amphibole in the rock are partly altered plagioclase, epidote, chlorite, carbonate, and leucoxene, which indicate that the rock has been metamorphosed. Other minor minerals are pyrite, quartz, and hematite. Grain sizes in the metapyroxenite are 0.4–0.8 mm, and the fabric is panidiomorphic to hypidiomorphic granular.

The relationship of the metapyroxenite to other mafic intrusive rocks has not been seen in the field, except that the metapyroxenite in sec. 24 lies immediately adjacent to metadiabase. Presumably this rock was a pyroxene-rich differentiate of magmas that formed the abundant mafic intrusive rocks in the Mona Schist. The metapyroxenite is presumed to be about the same age as nearby metadiabase.

HORNBLENDE LAMPROPHYRE

Dikes and sills of hornblende lamprophyre, 1–15 feet thick and of very limited extent, cut the Mona Schist and have north, northwest, and west trends at the following localities in T. 48 N., R. 25 W.: SW¼SW¼ sec. 14, SE¼SW¼ sec. 22, NE¼NE¼ sec. 23, NW¼NW¼ sec. 23, S. central part of sec. 23, and E½SW¼ sec. 30.

The partial alteration of apparent primary mineral assemblages to metamorphic assemblages indicates that the lamprophyre was emplaced prior to the metamorphism of the post-Animikie, pre-Keweenawan interval. The lamprophyre consists largely of sodic plagioclase and hornblende, or potassium-feldspar (othoclase?), sodic plagioclase, and hornblende, apparently of premetamorphic origin, and so is classified as spessartite or vogesite. Ayres and Higgins (1939) called the dike in the NW¼NW¼ sec. 23 spessartite with thin marginal selvages of hornblendite and malachite. Lamprophyre dikes mentioned by Ayres and Higgins, east from the NW¼NW¼ sec. 23 and on Middle Island Point, have not been identified as such, except for the dike in the NE¼NE¼ sec. 23, but are probably among the dikes mapped as metadiabase during the present work.

NAME AND DESCRIPTION

The term "lamprophyre" is a widely used field term for any dark-colored dikes, particularly if they cut granitic rock and are associated with aplite dikes. Classically, however, the term has been used for dark porphyritic dikes in which the phenocrysts are well-formed mafic minerals (pyroxene, amphibole, or biotite). The mineral of the phenocrysts commonly occurs also in the groundmass, and the groundmass may have a sugary texture. The term "hornblende lamprophyre" is used here for dark-green intrusive bodies characterized by euhedral phenocrysts of hornblende set in a groundmass mainly of subhedral and euhedral plagioclase and hornblende. The lamprophyre resembles metadiabase and can be distinguished in the field only where the porphyritic character is well defined and the phenocrysts are relatively large. Minor minerals in the groundmass are sphene, magnetite(?), apatite, and quartz. A small amount of microcline was seen in one sample, and reddish potassium-feldspar is present in the groundmass of one sample. Actinolitic amphibole, chlorite, carbonate, epidote, leucoxene, quartz, and dusty hematite are common metamorphic minerals in the lamprophyre.

Phenocrysts generally compose 20–50 percent of the lamprophyre and range in size from 1 to 3 mm. Feldspar and hornblende crystals in the groundmass, on the other hand, generally range from 0.05 to 0.7 mm in size.

Pleochroism of most of the hornblende is pale yellow to olive or brownish green. Euhedral phenocrysts of pale-green slightly pleochroic actinolitic(?) amphibole in some of the lamprophyre are probably pseudomorphic replacements of darker amphibole, as indicated by (1) replacement rims of pale-green amphibole around brownish-green hornblende in some lamprophyre and (2) aggregates of pale-green amphibole, the aggregates as units being pseudomorphic after an earlier amphibole.

The groundmass typically has subdiabasic or bostonitic texture of obvious igneous origin, or, if partly recrystallized retains remnants of igneous texture. The molding of tabular and lathlike feldspar around amphibole in an igneous crystallization fabric shows that the amphibole shapes did not develop during metamorphism.

The plagioclase is now mainly An_{5-10}, but although it is in the shape of primary crystals of igneous origin, it has been more or less modified (saussuritized), as shown by inclusions of epidote, sericite, chlorite, and carbonate. The primary plagioclase, therefore, was probably somewhat more calcic than the present feldspar.

There is a considerable range in chemical composition of the lamprophyre (table 13, Nos. 5, 6; Ayres and Higgins, 1939, table 1, p. 569). Silica, iron oxides, MgO, and CaO are in the gabbro-diorite range, but the K_2O content of No. 6 in table 13 is abnormally high. The high K_2O content is attributed to potassium-feldspar in the groundmass. The sample has considerable reddish groundmass feldspar which is shown to be potassium-feldspar (orthoclase?) by the yellow-staining technique (sodium cobaltinitrite applied to an uncovered thin section that has been etched in fumes of hydrofluoric acid). Ayres and Higgins (1939, p. 576), however, assigned excess K_2O of their dark-green spessartite to the amphibole.

ORIGIN AND AGE RELATIONS

The abnormally high K_2O content of some of the hornblende lamprophyre and the unknown spatial relationships of lamprophyre to other dike rocks make the origin and age of the lamprophyre obscure.

The metagabbro at Picnic Rocks has relatively a high K_2O content (table 13, No. 2), and near the southwest edge of the mass, in the parking lot of the Lake Shore Engineering Co., much of the metagabbro has large euhedral hornblende crystals, is coarsely lamprophyric in aspect, and is associated with seams of pegmatitic metagabbro. A concentration of residual fluids in parts of the Picnic Rocks mafic magma probably accounts for the relatively high K_2O content, the hornblende euhedra, and the mafic pegmatite seams. The dikes of hornblende lamprophyre, which are somewhat similar to the metagabbro at Picnic Rocks in containing hornblende euhedra and high K_2O may also have formed from a mafic magma in which a concentration of volatiles occurred during differentiation.

The lamprophyre cannot be dated on geologic evidence more closely than the long interval between the formation of the Mona Schist and the latter part of Animikie time. Furthermore, the lamprophyre dikes may be of one or several ages. Those with potassium-feldspar ($Na_2O:K_2O$ about 1:2–1:4) for example, may be of different age and source than those with less K_2O ($Na_2O:K_2O$ about 1:1).

SYENITES

Steep dikes of syenite and porphyritic hornblende syenite, 1–50 feet thick, cut Compeau Creek Gneiss in the following places in the north-central part of the Marquette quadrangle:

T. 49 N., R. 25 W.:
1. North central part of sec. 34, north side of Partridge Island (hornblende syenite).

T. 48 N., R. 25 W.:
2. NW¼ SW¼ sec. 3 (hornblende syenite).
3. SE¼ SW¼ sec. 3 (syenite).
4. W½ NE¼ sec. 5 (syenite).
5. SW¼ SW¼ sec. 5 (syenite).
6. NE¼ NW¼ sec. 5 (hornblende syenite).
7. NE¼ NE¼ sec. 7 (syenite).
8. Central part of sec. 8 on the east shore of reservoir (hornblende syenite).

The dikes all trend approximately north and cut the Compeau Creek Gneiss. Compared with dikes of other rock in the area that have similar thicknesses, they persist for only short distances. Several dikes of porphyritic hornblende syenite at locality 2 are cut by a dike of metadiabase. In the metagabbro at Picnic Rocks, syenite veins and clots range from less than an inch in width to several inches in width.

The syenite is salmon colored or dull red, weathers dark gray, and is difficult to distinguish from granitic rock on weathered surfaces.

The main constituent of the syenite is feldspar, both sodic plagioclase and potassium-feldspar. Staining of uncovered thin sections with sodium cobaltinitrite indicates that potassium-feldspar composes about half the syenite and much of the groundmass in the hornblende syenite and occurs both as discrete crystals and as partial replacements of plagioclase.

Minor minerals of the syenite are sericite, chlorite, carbonate, quartz, sphene, rutile, and leucoxene. The sericite, leucoxene, and carbonate—and possibly some of the quartz—are secondary minerals probably related to regional metamorphism. The chlorite may be primary or may be pseudomorphic after primary biotite. The chemical analysis of one sample (table 13, No. 8) is typical for syenite (see Clarke, 1920, p. 435). Radiation from the dike in the NE¼NE¼ sec. 7 was four to five times the background radiation; a scintillation detector registered 0.120 milliroentgen per hour over the syenite dike as compared with 0.024–0.025 milliroentgen per hour over the bordering tonalitic rock. The radioactive material in the syenite has not been identified. The small syenite masses in the metagabbro at Picnic Rocks are probably feldspathic differentiates of the gabbroic magma. The origin of the larger syenite bodies is unknown.

The porphyritic hornblende syenite is a reddish-green rock characterized by euhedral phenocrysts of hornblende set in a groundmass mainly of subhedral and euhedral plagioclase and hornblende and anhedral potassium-feldspar (fig. 25). Minor minerals in the groundmass are apatite and a dark opaque mineral (magnetite?). Epidote, quartz, sericite, and dusty hematite are common metamorphic minerals in the hornblende syenite.

Phenocrysts of hornblende form from less than 10 to about 45 percent of the hornblende syenite. Indices of refraction of hornblende from one sample in the NE¼NW¼ sec. 5 are N_X=1.631, N_Y=1.650, and N_Z=1.659.

The hornblende syenite is not greatly different in appearance from some border phases of the metagabbro at Picnic Rocks and appears to be intermediate mineralogically and chemically (table 13, No. 7) between the felsic syenite and the hornblende lamprophyre. The syenite, porphyritic hornblende syenite, and the hornblende lamprophyre may be closely related as differentiates of gabbroic magma.

PEGMATITE

Pegmatite is rare in the Marquette quadrangle and in the northern half of the Sands quadrangle, but in the southwestern part of the Sands quadrangle it forms abundant thin dikes, conformable and crosscutting lenses, and ill-defined blotchy masses in the gneiss. A few of these masses are as much as 10 feet thick, but most are less than 3 feet thick and many less than 6 inches thick. In some places pegmatitic rock appears to grade into "granitic" gneiss and may only be a coarser grained phase of that rock. In general, however, the pegmatites form discrete bodies that clearly crosscut gneissic (foliated) "granitic" rock, amphibolite, interlayered amphibolite and "granitic" rock, or layered amphibolite containing crosscutting "granitic" rock. In places where amphibolite containing crosscutting "granitic" rock is in turn cut by pegmatite, the distinction in the ages of the pegmatite and "granitic" rock is particularly clear. The locations of the larger pegmatites are indicated by letter symbol on plate 2.

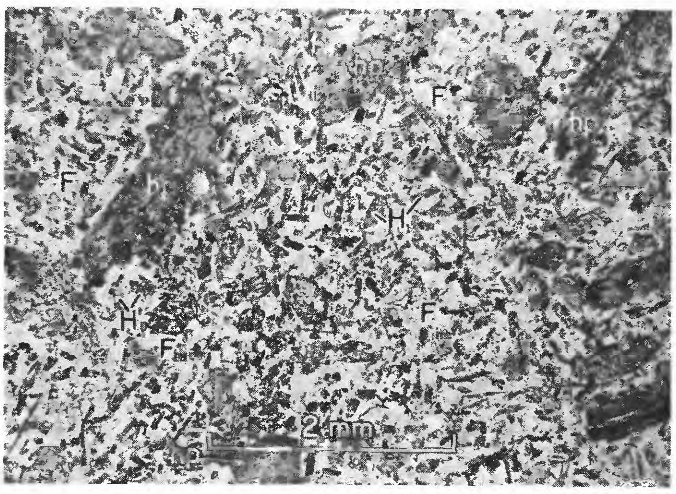

FIGURE 25.—Photomicrograph of porphyritic hornblende syenite, 4,200N–2,200E, SW cor. sec. 5, T. 48 N., R. 25 W. Euhedral phenocrysts or hornblende (hp) set in groundmass mainly of feldspar (sodic plagioclase and potassium-feldspar) (F) and subhedral to euhedral hornblende (H). Plane light.

Most of the pegmatite is poor in quartz. The thin dikes, which contain only a small part of all the pegmatite in the area, are relatively quartzose, whereas the abundant pegmatitic clots and lenses have little quartz. Only a few samples of the pegmatite have been collected, however, so we cannot generalize on the composition of the feldspar. The feldspar in one east-trending dike, 1 foot thick, is mainly albite, whereas several pegmatitic clots consist mainly of microcline.

Even the approximate age or ages of the pegmatite cannot be determined by geologic evidence alone. A pegmatite cutting the southern complex at Republic, Mich., 25 miles west of the southwestern part of the Sands quadrangle, has been dated by the potassium-argon and strontium-rubidium methods by L. T. Aldrich (written commun., 1962) as being between 1,760 and 1,830 million years old. To judge by the ages determined by Aldrich in other parts of northern Michigan, the age of the pegmatite at Republic is late Animikie, or at the youngest, post-Animikie and pre-Keweenawan. Age determinations of the pegmatites in the southwestern part of the Sands quadrangle by radioactive decay methods will be necessary before more can be known of their time relations to each other and to the pegmatite at Republic.

APLITE

Light-colored quartz-feldspar dikes, mostly less than 1 foot wide and of even, fine to medium grain size, occur in many places in the Compeau Creek Gneiss. The dikes generally cannot be followed for more than 20–30 feet on weathered outcrop surfaces, and they are not indicated on plates 1 and 2. Most have allotriomorphic to hypidiomorphic granular texture. The few that have been sampled are tonalitic in composition. Most of the aplite dikes cut foliation and hornblende-rich zones in the gneiss, and so are postgneiss in age. At the large outcrop immediately south of Wetmore Landing in sec. 29, T. 49 N., R. 25 W. (pl. 3), some sharply defined aplite (early light-colored tonalite) dikes less than 3 inches wide are crossed by the foliation of the gneiss. These dikes and poorly defined zones of similar aplite along small shears in the gneiss of the same outcrop probably were reconstituted from the gneiss, along cracks or along the small shears, respectively. Northwest of the top of Sugarloaf Mountain, at 4,750N–

2,000E, SW. cor. sec. 32, T. 49 N., R. 25 W., dikes of aplite, 6 inches thick are cut by quartz veins 2–3 inches thick; both the aplite and quartz are cut by a diabase dike. On geological evidence, therefore, the aplite can only be dated between the age of the gneiss (lower Precambrian) and the age of the diabase (upper Precambrian).

INTRUSIVE ROCK OF POSE-ANIMIKIE, PRE-KEWEENAWAN AGE(?)

SERPENTINIZED PERIDOTITE

LOCATION AND SIZE

Serpentinized peridotite forms approximately the northern two-thirds of Presque Isle, located near the NE. cor. T. 48 N., R. 25 W. The peridotite is overlain unconformably by Cambrian (Jacobsville) sandstone and is not seen in contact with any other rock. Van Hise and Bayley (1897, p. 184) mentioned that diabase cuts peridotite (presumably on Presque Isle), but they did not give a locality; we have not found definite diabase or metadiabase cutting the peridotite on Presque Isle, but some light-colored highly altered rock on the west side of Presque Isle, near 400W–3,500N, SE. cor. sec. 2. T. 48 N., R. 25 W., may be strongly weathered metadiabase. The possible maximum size of the peridotite body is roughly delimited by granitic or mafic metavolcanic rocks to the east (islands half a mile away), west, and south, but not to the north. From the distinctive shape of Presque Isle, however, the peridotite body is presumed to have approximately the same size and shape as the northern two-thirds of the "island."

DESCRIPTION

The serpentinized peridotite is a dark-green rock which weathers to black, dark green, light green, reddish brown, or yellowish brown. It is characteristically cut by joints that form a blocky pattern. Conspicuous flat-lying joints occur on the north and northeast sides of Presque Isle.

Wadsworth (1884) described lherzolitic and wehrlitic varieties of peridotite from Presque Isle that consisted of olivine, enstatite, and diallage, as well as serpentine and minor minerals. Tests made by Wadsworth for chromite were negative. Creveling (1926) listed olivine, hypersthene, and augite from one sample, plus the varieties of serpentine, antigorite and chrysotile and the related bowlingite. Five thin sections of the peridotite collected during this study are wehrlitic. No orthorhombic pyroxene has been seen. The peridotite now has approximately 10 percent monoclinic pyroxene, 15 percent olivine or less, a few percent magnetite(?), and generally less than 1–2 percent secondary biotite, hematite, goethite, dolomite, and quartz; the remainder is platy and fibrous serpentine. Whether some reddish serpentinelike material is bowlingite or partly oxidized antigorite is unknown. Much of the serpentine appears to be pseudomorphous after olivine, and the original texture, having typical abundant irregular cracks cutting the serpentinized olivine, is still clearly visible in thin sections. Pyroxene crystals show little evidence of alteration to serpentine or other secondary products, so the peridotite originally contained about the same amount of pyroxene as now.

Olivine composed approximately 85 percent of the original rock. During the serpentinization of the olivine, tiny opaque particles (magnetite?) collected in numerous lines which appear to lie both along crystallographic planes of the original olivine and along or adjacent to cracks and grain boundaries.

The texture of much of the original rock was allotriomorphic granular and medium grained, resembling that of typical dunite, but locally it was porphyritic because of large tabular crystals of pyroxene. The pyroxene phenocrysts commonly poikilitically enclose euhedra of olivine or antigorite pseudomorphs of olivine.

Thin, generally flat lying veins of light-colored dolomite, red chert and pale-green serpentine are very common in the peridotite near the north edge of Presque Isle. Flat-lying joints, dipping about 10° N., are also very abundant there and have been followed by many of the veins. Old test pits near the pronounced cove east of the northernmost part of the road explored a thin vein consisting of galena, pyrite, pyrrhotite, chalcopyrite, and traces of the nickel minerals violarite and millerite (Snelgrove and others, 1944, p. 65). A semiquantitative spectrographic analysis of one sample of the peridotite shows relatively large trace amounts of chromium (0.7 percent) and nickel (0.15 percent). Snelgrove, Seaman, and Ayres (1944, p. 64–65) found amounts of nickel ranging from 0.02 to 0.14 percent in the peridotite, and 0.32 percent in the sulfide vein material from the north end of Presque Isle.

Parts of the serpentinized peridotite lying just below the Cambrian sandstone along the west side of Presque Isle are extensively altered to yellow-brown iron oxide and are cut by flat-lying and reticulated networks of thin white to tan carbonate veins. These vein networks appear to be related to the weathered surface at the base of the overlying sandstone, whereas the veins on the northeast side of Presque Isle appear to be in joints formed tectonically or by cooling of the ultramafic magma. The veins on the west side of the "island" therefore probably were deposited by meteoric solutions, and those on the northeast side by hydrothermal solutions. Thin veins of orange-colored serpentine cut the serpentinized peridotite in places. Fragments of red chert and highly altered (oxidized) peridotite occur in lenses of basal conglomerate in the unconformably overlying sandstone. Much of the oxidation of the serpentinized peridotite, and also the abundant carbonate and chert veins associated with the oxidized rocks, are therefore interpreted as related to a pre-Paleozoic erosion surface.

METAMORPHISM AND AGE RELATIONS

Wadsworth (1884) concluded that the peridotite at Presque Isle was intrusive into the Paleozoic sandstone because the sandstone near the contact with the peridotite is discolored, is more indurated, and contains vein and chalcedonic quartz and because he found no fragments of peridotite in the sandstone. Perhaps he overlooked the fragments of oxidized peridotite in the similar-colored sandstone, or these may not then have been exposed along the actively eroding west shore of Presque Isle. Van Hise and Bayley (1897, p. 184) reported (1) fragments of peridotite in the sandstone and (2) diabase cutting peridotite but not sandstone, and they accordingly dated the peridotite as older than the sandstone. We also consider the peridotite to be older than the sandstone and attribute the features noted by Wadsworth to the circulation of ground water along the contact.

The restriction of the peridotite, as well as of the ultramafic rocks northwest of Ishpeming, to areas of pre-Animikie rocks has led some geologists to conclude that all these ultramafic rocks are of pre-Animikie age. On the basis of presently known field relationships, this conclusion may or may not be valid.

Although the peridotite is extensively serpentinized, the presence of considerable olivine (normally unstable during metamorphism) and pyroxene indicates that the rock has not been thoroughly metamorphosed. This conclusion in turn strongly suggests that the rock does not predate the metamorphism of the post-Animikie, pre-Keweenawan orogenic interval. The peridotite body apparently is thicker than any of the masses of metagabbro or metadiabase in the Marquette-Sands area, so if it were premetamorphic in age its size might have been a factor in shielding some of the minerals from destruction during metamorphism. The virtually complete alteration of comparably large bodies of mafic rock elsewhere in northern Michigan during the post-Animikie metamorphism and orogeny (see Gair and Wier, 1956, p. 60–61, 66; Bayley, 1959, p. 65–75), however, leads to the expectation that if the peridotite were of premetamorphic age the percentage of unaltered olivine should be virtually nil and the percentage of unaltered pyroxene much less than it is. On the other hand, the extensive serpentinization does indicate a high degree of alteration. This may have occurred after the post-Animikie, pre-Keweenawan metamorphism, entirely by autometamorphism (deuteric alteration). Because of the worldwide association of orogeny and ultramafic intrusion, however, it seems likely that the Presque Isle peridotite was intruded during, perhaps toward the close of, the post-Animikie, pre-Keweenawan orogeny and metamorphism, and was partially serpentinized at that time.

UPPER PRECAMBRIAN INTRUSIVE ROCK—KEWEENAWAN SERIES

DIABASE

GENERAL RELATIONSHIPS

Diabase dikes are abundant in the northern half of the Marquette quadrangle but less so to the south. Diabase cuts dikes of metadiabase and all formations up to and including the Kona Dolomite. The dikes generally are negatively polarized magnetically and only slightly metamorphosed. They trend mainly east and east-northeast, range in thickness from a few inches to 120 feet, and, from outcrops alone, appear to be continuous for as much as 3 miles. Aeromagnetic surveys (pl. 7; Case and Gair, 1965), however, show that some extend for much greater distances. Generally the thicker dikes are the more persistent, but some that are less than 10 feet thick are continuous for at least 1 mile. On Lighthouse Point (pl. 6) two thick diabase dikes pinch out opposite each other within a remarkably short distance. They probably are connected below, or were connected above, the present erosion surface.

DESCRIPTION

There are two types of diabase in the area. Most is medium grained (0.1–2.0 mm), dark gray, and commonly only slightly altered. Coarser grained diabase of this type generally has a faint amber cast, whereas the finer grained of this type is black. The dark-gray diabase locally has been considerably altered, so its mafic material is all secondary. In places the dark-gray diabase is porphyritic or glomeroporphyritic, and on freshly broken surfaces the phenocrysts have the look of fish scales. The second type of diabase is greenish with a reddish or salmon-colored groundmass; it is more highly altered than most of the dark-gray diabase and is much less common. This diabase forms several dikes, one of which is comparatively wide and extends eastward to Lake Superior from the bottom of the south side of Sugarloaf Mountain in sec. 32, T. 49 N., R. 25 W. The other reddish-green dikes are small. One is an offshoot of the large dike, one is in the hill just north of the

S¼ cor. sec. 33, T. 49 N., R. 25 W., and one is in the NW¼ sec. 5, T. 48 N., R. 25 W., south of Compeau Creek. On the knob 1,100 feet southeast of the top of Sugarloaf Mountain, dark-gray diabase and reddish-green diabase intersect, but because contacts between the two rocks are hidden, the relative ages of the two are uncertain.

The major constituents of the slightly altered dark-gray diabase are plagioclase and pyroxene. Plagioclase ranges from An_{48} to An_{70} in different samples. In porphyritic diabase the phenocrysts are well within, or near the calcic end of, the labradorite range, whereas smaller laths in the groundmass are about An_{50}. The pyroxene is augite and pigeonite(?). In several samples the intermediate index of refraction, N_Y, of the pyroxene is between 1.670 and 1.695, and the optic angle measured on a universal stage is 43°–48° (+). N_Z is 1.699 in one sample in which N_Y is between 1.670 and 1.678. The interference figures of some of the pyroxene, however, indicate optic angles of less than 20°, so apparently some of the diabase contains augite and some contains pigeonite.

Plagioclase and pyroxene are intergrown in the typical subophitic and intergranular fabrics of diabase. In chilled zones or in dikes less than about 1 foot thick, the diabase generally has a fine-grained black groundmass (nearly opaque in thin section) in which occur small granules of pyroxene and randomly directed small laths and microlites of plagioclase. In one sample there is evidence of two generations of pyroxene. A long thin prism of pyroxene has later granules of pyroxene clustered along it.

Small amounts of olivine occur in a few of the dikes, as relics amidst replacing chlorite or antigorite. Many dikes, however, have sharply defined, medium-sized granular aggregates of chlorite, antigorite, or a reddish platy mineral (bowlingite?, iddingsite?) that are probably pseudomorphs of olivine. In places, such aggregates are poikilitically enclosed in fresh pyroxene, and in one sample they have the shape of euhedral olivine. Olivine probably never composed more than 5 percent of the diabase.

Magnetite is the principal minor (1 percent or less) primary constituent of the dark-gray diabase. It is ubiquitous and appears to have crystallized relatively late. The surfaces of some unaltered pyroxene crystals are coated with magnetite granules, and in one thin section magnetite occurs along cracks in plagioclase. Pyrite and granophyre are other minor constituents in some of the diabase. The granophyre consists of graphic intergrowths of potassium-feldspar and quartz.

Chlorite (including some antigorite), sericite, biotite, amphibole, carbonate, and colored iron oxide are common secondary minerals in the diabase; in most places they form less than 15 percent of the rock. They are mainly interstitial to and tend to rim and partly replace the primary minerals irregularly through the rock. Sericite occurs, especially in partly altered plagioclase phenocrysts, and may be associated with carbonate. The nature and distribution of the altered material indicates that the alteration was deuteric.

In a few places, parts of some dark-gray diabase dikes are more highly altered than most of the diabase. Primary pyroxene has been almost entirely converted to chlorite, amphibole, and biotite, but the plagioclase remains largely unaltered. The schistose chloritic dike in the SW¼ sec. 2, T. 47 N., R. 25 W. (see p. 51) may be similarly altered diabase.

The reddish-green diabase differs mineralogically from the dark-gray diabase in being generally more altered and in having abundant interstitial granophyre. About half of the primary pyroxene in this rock has been changed to secondary mafic minerals. The granophyre consists of graphic and wormy intergrowths of quartz and potassium-feldspar, and is invariably colored pale red by inclusions of dusty hematite.

AGE AND ORIGIN

The two types of diabase are the youngest Precambrian rocks in the area. Their relative age is indicated by (1) their mappable crosscutting relationship to the metamorphosed rocks (except the Wewe and Siamo Slates and the Ajibik Quartzite), (2) their crosscutting relationship to structures produced during the late middle Precambrian orogeny, and (3) their nonmetamorphosed condition, which indicates primary crystallization after the post-Animikie, pre-Keweenawan regional metamorphism. The diabase is dated as Keweenawan by analogy with similar rocks in the Keweenaw Peninsula of Michigan.

The reddish-green or granophyric diabase probably represents local accumulations of K_2O and SiO_2 in the reservoirs from which the diabase magma came. Inasmuch as the diabase crystallized mostly if not entirely after emplacement, there could have been little differentiation in the magma reservoir as a result of crystallization. Concentrations of K_2O and SiO_2 must have been a result of some other process of differentiation.

The relative ages of the gray diabase and the red granophyric diabase are unknown. The greater alteration of the granophyric diabase may be a result of its greater age and may mean that it was emplaced before the completion of the post-Animikie metamorphism, or it may be a result of more volatiles in its relatively potash- and silica-rich magma. This latter cause of greater alteration is suggested in one of the normal diabase dikes, the contact zone of which is highly altered and has 1–2 percent quartz and a little granophyre.

CAMBRIAN ROCKS

JACOBSVILLE SANDSTONE

The Jacobsville Sandstone was named by Lane and Seaman (1907, p. 692) after the town of Jacobsville on the Keweenaw Peninsula. The formation has been most comprehensively studied by Hamblin (1958, p. 16–62). Hultman (1953) gives many details concerning the Jacobsville in the Marquette area.

In the area of the present report, exposures occur only in the Marquette quadrangle (pl. 1) within approximately 3 miles of the shore of Lake Superior. Northwest from Presque Isle, outcrops are along the shore, but from Marquette south and southeast they are both along the shore and inland, mainly along the Carp River and adjacent to Quarry Pond (sec. 26, T. 48 N., R. 25 W.). Areas shown west of the center of sec. 21 and in the N½ sec. 22, T. 48 N., R. 25 W., are largely heavy concentrations of fragments in the soil. The outcrop farthest inland is on Morgan Creek in the southwestern part of sec. 33, T. 48 N., R. 25 W. The highest outcrop is on the south flank of Mount Mesnard in the NW¼SW¼ sec. 35, at altitudes between 960 and 980 feet.

In the Marquette quadrangle, the Jacobsville is mainly a reddish to reddish-brown friable feldspathic lenticular sandstone with intercalated lenses of red or gray conglomerate and reddish shale. Bleaching to light gray, pale gray green, and white along bedding, cracks, and other relatively permeable parts of the rock is common. Dips generally do not exceed 15° and crossbedding is widespread. According to Hamblin the Jacobsville is poor in feldspar (less than 15 percent), except close to the basal contact with underlying granitic rock. Accurate measurement of the amount of feldspar in the Jacobsville of the Marquette area has not been made during the present study. Pebbles in conglomeratic lenses are mainly quartzite, cherty sericitic slate, chlorite-sericite slate, dolomite, iron-formation, or ferruginous slate and, at Presque Isle, weathered peridotite and red chert. In many places, flakes of muscovite as much as several millimeters in diameter occur in the standstone. Many additional descriptive details may be found in the reports by Hamblin (1958) and Hultman (1953).

Some 40–50 feet of the Jacobsville is exposed in places along the shore near the north edge of the Marquette quadrangle and along the Carp River. A thickness of at least 60 feet of sandstone was quarried at the present Quarry Pond, according to local residents. Hamblin reports 300 feet of Jacobsville in one shore exposure between Marquette and Big Bay, 25 miles to the northwest. The presence of flat-lying Jacobsville at altitudes between 960 and 980 feet on the south flank of Mount Mesnard, some 370 feet above the gently dipping outcrops on the lake shore, may indicate a minimum thickness of 300–400 feet. Hamblin, however, has reported great variations in thickness in short distances and attributes this, as well as the widespread crossbedding and cut-and-fill structures, to fluviatile deposition. The distribution of the sandstone south of Marquette indicates that the topography at the time the Jacobsville was deposited had at least several hundred feet of relief. Hamblin believes that much of the present topography adjacent to the south shore of Lake Superior has been exhumed and is a near replica of early or middle Cambrian topography.

Although some earlier workers thought that the Jacobsville is of Keweenawan age, stratigraphic considerations outside the Marquette area have led Hamblin (1958, p. 62) to conclude that the Jacobsville represents continental deposition of Early and Middle Cambrian age.

STRUCTURE

GENERAL FEATURES

The west-plunging Marquette synclinorium and the basement rocks to the north and south form the major structural units in the area. Structural elements mapped in the basement rocks are: small folds in layered gneiss and mafic metavolcanic rock, foliation, joints, and lineations (axes of minor folds; crenulations on bedding and foliation). Structural elements mapped in the middle Precambrian metasedimentary rocks of the synclinorium are: large and small folds, flow cleavage (foliation), joints, and lineations (fold axes, crenulations, intersections of bedding and cleavage).

Large folds in the synclinorium are evidenced by the map pattern, by minor folds on the limbs of the larger ones, and by comparisons of bedding-cleavage relationships from one limb to the other. Structural details near the east end of the synclinorium are shown on plate 4. Small folds, having distances between limbs (wave lengths) of generally less than 20 feet, both in the lower and middle Precambrian rocks, are shown on plates 1 and 2 only by structural symbols indicating fold axes or divergent attitudes of bedding. Foliation results from preferred mineral alinements, which may be strong and comprise a large proportion of all mineral grains in the more schistose or slaty rocks or may be slight and involve a small percentage of mineral grains along relatively wide spaced shear surfaces in some granitoid gneisses, mafic rocks, dolomite, and quartzite. Much of the quartzite and dolomite are without cleavage, and some amphibolite in the gneiss and the more massive metabasalt in the Mona Schist have no foliation.

Faults have been observed directly only in a few places and for only a few feet. For the most part the existence and positions of the faults are inferred from (1) anomalous changes in thickness of formations; (2) juxtaposition of formations normally separated by one or more formations; (3) offsets in contacts of formations or dikes; (4) unusual shear zones or breccias, generally marked by secondary quartz, in conjunction with (1), (2), or (3); and (5) long, straight, narrow clefts in the Compeau Creek Gneiss where such clefts are alined with faults inferred from the other criteria.

Little is known of relationships between structures in the basement and in the middle Precambrian rocks of the synclinorium, although trends of foliation and gneissic layering in many places in the basement rocks are virtually parallel with the trend of the major axis of the synclinorium. The synclinorium formed during the orogeny at the end of middle Precambrian time, whereas structural trends in the basement were formed prior to middle Precambrian time, as indicated by (1) discordant top directions in middle Precambrian and lower Precambrian rocks, and (2) development of foliation and gneissic layering in basement rocks before the intrusion of diabase that was subsequently metamorphosed during the orogeny at the end of middle Precambrian time. Deformation strong enough to have formed the main trends in the basement cannot be related to the comparatively minor disturbances at the end of early and middle Animikie times, and so it must have occurred prior to Animikie (middle Precambrian) time. Whether the late middle Precambrian deformation utilized lower Precambrian movement surfaces in the basement, whether similarly oriented forces caused the earlier and later deformations, or whether some combination of these and other factors caused the relationship is unknown.

CONTACT BETWEEN ROCKS OF ANIMIKIE AND PRE-ANIMIKIE AGE

The contact between Animikie and older rocks is in places a fault (see p. 67–68) and in places an erosional-depositional unconformity, in part modified by post-depositional tectonic action. Where the base of the Animikie section is an uncomformity, it is characterized by conglomerate or arkose and generally by structural discordance between Animikie and older rocks. Only conglomerate of the Animikie has been seen in physical contact with pre-Animikie rocks, in the S½ sec. 29, T. 48 N., R. 25 W., and in the SE¼ sec. 8, the SW¼ sec. 9, and the NW¼ sec. 16, T. 47 N., R. 25 W. The rather tightly folded contact between Animikie and pre-Animikie rocks in the east-central part of sec. 2, T. 47 N., R. 25 W. (pl. 4) is characterized by sheared arkosic grit and sheared gneiss which generally are difficult to distinguish from one another.

The structural discordance between the Animikie and older rocks is most apparent in two places: in sec. 29, T. 48 N., R. 25 W., and in sec. 9, T. 47 N., R. 25 W. The most notable discordance occurs at the base of the Enchantment Lake Formation, in its type area in sec. 29. There, weakly to moderately well cleaved conglomerate at the base of the Animikie section is in sharp vertical contact with a slaty greenstone phase of the Mona Schist. The slaty greenstone is between 50 and 100 feet thick and grades northward away from the contact into fairly massive schistose greenstone. The conformable stratigraphic sequence from the conglomerate southward to the lower part of the Mesnard Quartzite on the north shore of Enchantment Lake indicates that top directions in the vertically dipping Animikie rocks are to the south. Layers of Mona greenstone dip 50°–70° N., and widely distributed pillow structures indicate that top directions are to the north through a thickness of 1½ to about 3 miles of Mona Schist. The pillow structures having recognizable tops that occur nearest to the exposed unconformity are about 3,000 feet to the west and nearly on strike with the unconformity, in the NE¼SE¼ sec. 30. Foliation in the slaty greenstone and cleavage in the conglomerate are nearly parallel to one another and to the contact between the two rocks and commonly are strongly lineated in a downdip direction. Thus, nearly vertical movements are indicated in the contact zone between the steeply dipping Animikie and pre-Animikie rocks, and top directions point away from one another and demonstrate a "back-to-back" structure.

The deformation along the north margin of the Marquette synclinorium that produced the "back-to-back" structure might have been mainly rotation about the major fold axis of the synclinorium, or some combination of steeply inclined faults and shears. If the deformation were a rotation about the axis of the synclinorium, however, the unfolding of the conglomerate and overlying Animikie rocks to their original attitudes would also have "unfolded" a thickness of 1½–3 miles of Mona Schist and placed this great mass of greenstone in an upside-down position at the beginning of Animikie time.

The alternative explanation of a combination of faulting and shearing movements along the north margin of the synclinorium seems to be much more feasible and does not require uniform rotation of a great thickness of the Mona Schist first into an upside-down position in pre-Animikie time followed by a rotation back to a

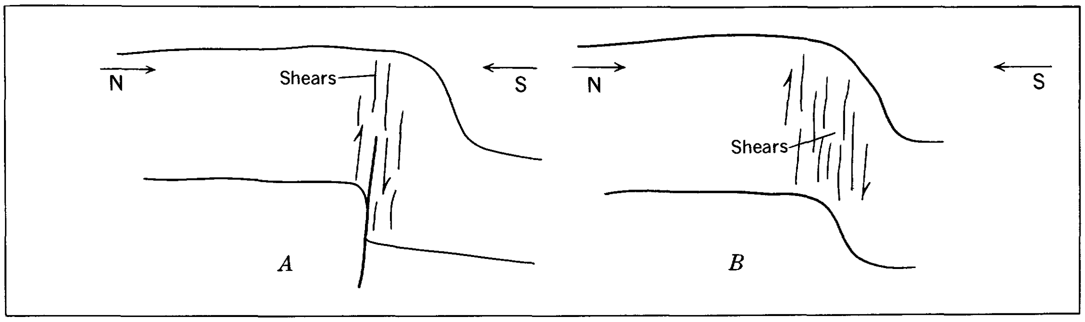

FIGURE 26.—Two possible mechanisms by which downdropping of pre-Animikie basement rocks might have been accomplished along the north side of Marquette synclinorium under influence of regional compression. *A*, Fault-flexure; high-angle reverse fault; some offsets along closely spaced shears. *B*, Shear fold; displacements along closely spaced subparallel shears.

steeply dipping upright position since the beginning of Animikie time. Instead, the Mona Schist could have been folded to about its present attitude prior to Animikie time and could have maintained approximately that attitude since then, except in a narrow zone of shearing or faulting along the margin of the synclinorium.

Steep normal or reverse faults, or a shear fold resulting from movements like those on such faults (fig 26), might have been as important along the north flank as faults were along the south flank (p. 67–68) in causing relative downdropping of the basement beneath the synclinorium. As the basement rocks were relatively downdropped, the Animikie rocks were folded to their present steeply inclined attitudes along the north flank. The folding of Animikie rocks in the synclinorium suggests north-south regional compression and crustal shortening, but almost vertical movement dominated near the contact. The probability of regional compression across the synclinorium suggests that marginal faults in the basement are steeply inclined reverse faults. Faulting and shear folding of basement rocks combined with the sliding or dragging of conglomerate along the unconformity that exists between the conglomerate and the greenstone could have brought the Animike rocks into their present position in respect to the Mona Schist. Such movements are somewhat analogous to those of juxtaposed shear folding and flexure folding (bedding-plane slip) that normally occur during the folding of alternating competent and incompetent rocks such as sandstone and shale.

Hypothetical mechanisms to explain the "back-to-back" structure are shown diagrammatically in figure 27. In figure 27*C*, as the unconformity is folded by shearing of the greenstone, a steeply inclined fault develops along one part of the shear zone, the slippage surface between conglomerate and greenstone passes into the fault, and the conglomerate is further steepened by being dragged along the fault. In the situation shown in figure 27*D*, as the upper surface of the greenstone is shearfolded downward to the south, the conglomerate bends accordantly with the surface of the greenstone and at the same time is dragged relatively downward to the south along the contact. In either process, once the conglomerate is steeply tilted, thinning and stretching by movements along steeply inclined cleavage could occur. The steeply plunging lineations on cleavage surfaces within the conglomerate suggest such movements. Furthermore, the few exposures of conglomerate at the base of the Animikie section along the north limb might be a result of shearing out of the conglomerate in places.

The mechanisms suggested in figure 27 are somewhat similar to one shown by James and others (1961, p. 72) to explain overturning and apparent draping of Animikie layered rocks against faulted blocks of basement rock. A somewhat similar mechanism of steep reverse or normal basement faults has been used to explain the formation of some monoclines in the Colorado Plateau (Baker, 1935; Baltz, 1962; Kelley, 1951).

At the second place where the structural discordance between Animikie and pre-Animikie rocks is well shown, in the SW¼ sec. 9, T. 47 N., R. 25 W., and in nearby parts of adjoining sections, foliation and streaky layering in the gneiss trend a little south of east, have steep dips, and are truncated by the contact between the Enchantment Lake Formation and the gneiss. The Enchantment Lake rocks strike parallel to the trend of the contact and dip 20°–35° W. and SW.

FOLDS

North of Marquette synclinorium.—North of the Marquette synclinorium the Mona Schist trends generally between S.70°E. and east; it appears to dip continuously at 40°–80°N. and to have top directions persistently northward, as shown by pillow structures. Layering in the gneiss trends between S.70°E. and S.80°E. and dips vertically or steeply north in most

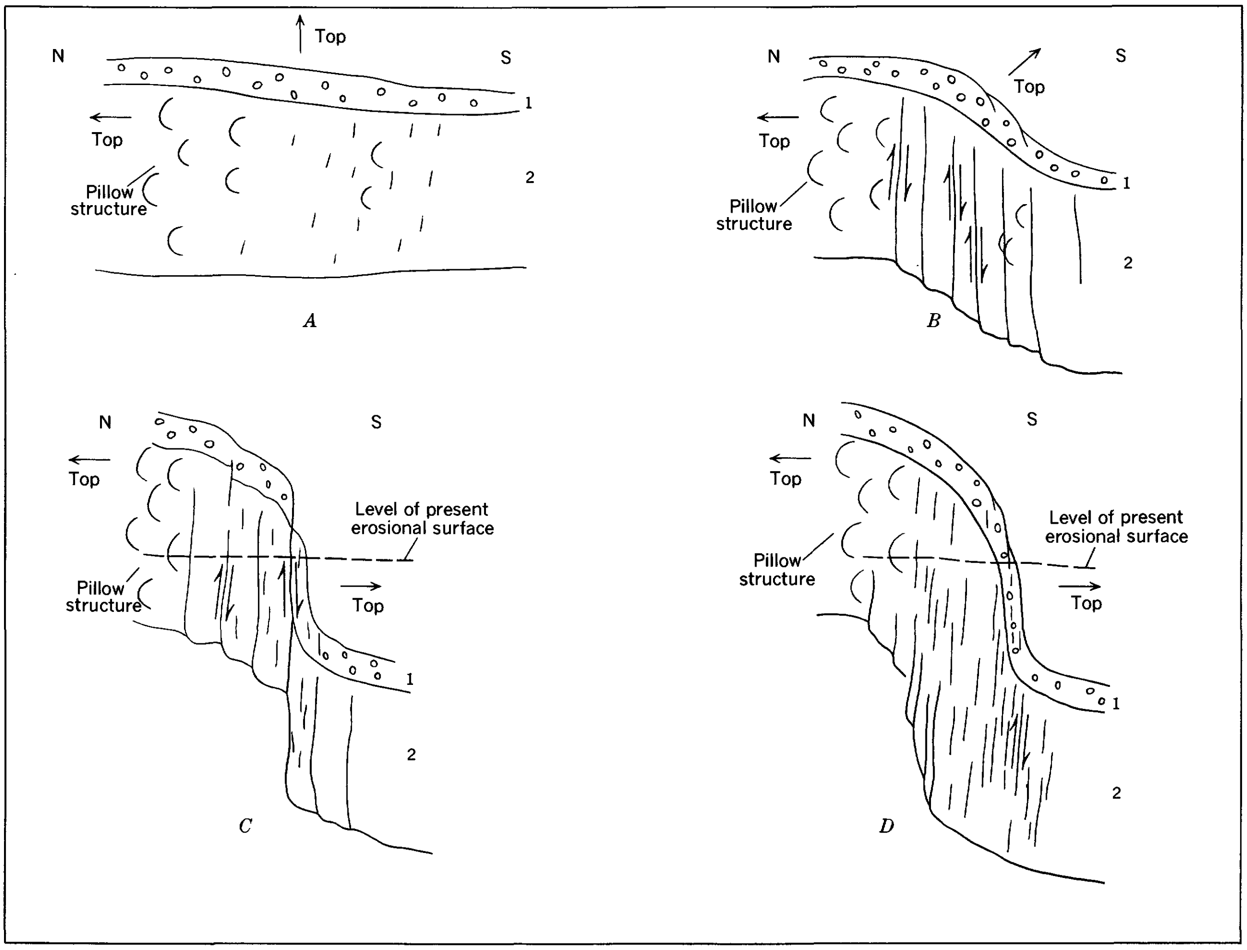

FIGURE 27.—Hypothetical mechanisms to explain "back-to-back" structure north side of Marquette synclinorium. *A*, Early Animikie time. *B*, Late Animikie time, early stage of deformation. *C* or *D*, Late Animikie time, late stage of deformation.

places. No large folds have definitely been identified in the Mona Schist or the Compeau Creek Gneiss. The areas of layered amphibole schist in secs. 1 and 12, T. 48 N., R. 26 W., appear to be the limbs of a large fold mapped by W. P. Puffett in the Negaunce quadrangle to the west (oral commun., 1966). The most conspicuous small folds seen north of the synclinorium, both in amphibole schist near the north edge of sec. 12, T. 48 N., R. 26 W., and in layered amphibolitic gneiss near the center of the SW¼ sec. 6, T. 48 N., R. 25 W., occur in or adjacent to this large fold.

South of Marquette synclinorium.—Small folds in layered gneiss in the southwestern part of the Sands quadrangle, near the SW. cor. sec. 31, T. 47 N., R. 25 W., at the NW. cor. sec. 8, T. 46 N., R. 25 W., and in the SE¼ sec. 16, T. 46 N., R. 25 W., plunge steeply southeast. South of the center of sec. 18, T. 46 N., R. 25 W., a small fold plunges gently northwest. The prevailing northwest trend of fold axes as well as similar trends of foliation indicate a pronounced northwest "grain" in the basement in this part of the Sands quadrangle.

The existence of a large northwest-trending infold of middle Precambrian rocks in the central part of the Sands quadrangle is suggested by a strong positive northwest-trending magnetic anomaly in that area (p. 69; Case and Gair, 1965).

Along the south side of the synclinorium near its east end, in the SE¼ sec. 2 and the adjoining part of sec. 1, T. 47 N., R. 25 W., the contact between lower and middle Precambrian rocks is in several tight west-plunging folds and indicates folding of the basement gneiss conformably with the rocks of Animikie age, at least close to the contact (pl. 4). Small folds with west- and northwest-plunging axes occur in the SW¼ sec. 10 and in the SW¼ sec. 9, T. 47 N., R. 25 W. These folds may be related to the large upwarps in the basement in secs. 4 and 9, T. 47 N., R. 25 W.

Folds within synclinorium.—The eastern end of the Marquette synclinorium, east from sec. 5, T. 47 N., R. 25 W., to Lake Superior, is occupied by several large folds of late Animikie age, including an anticline and a syncline that plunge west and an east-plunging syncline (pl. 4; Gair and others, 1961a). These folds are all related to the formation of the synclinorium. In sec. 5 the sinuous contacts of the Kona, Wewe, and Ajibik Formations indicate rather uniform small-scale folding, which probably extends westward in the Siamo Slate, along the core of the synclinorium, to the west edge of the mapped area and beyond. Kona Dolomite and Mesnard Quartzite south of the synclinorial axis in sec. 2, T. 47 N., R. 25 W., are tightly folded, and at least some of the folds continue into basement rocks (pl. 4).

The structure now mapped as an east-plunging syncline in the N½ sec. 1, T. 47 N., R. 25 W., was interpreted by Van Hise and Bayley (1897) as a canoe-shaped fold, whose east end was west of the lakeshore. Detailed mapping (pls. 4, 5) however, revealed no evidence of closure at the east end, or of a westward plunge of the fold axis west of the lakeshore (Gair and others, 1961a). In an effort to determine whether this syncline is open to the east and whether it signifies a major reversal of plunge for the Marquette synclinorium and a repetition of the Animikie section to the east, north-south aeromagnetic traverses were flown by the U.S. Geological Survey east of the present area in the summer of 1961 (Case and Gair, 1965, sheet 3). A continuation of east-trending structures to the vicinity of Munising, 30–35 miles east of Marquette, was indicated by a linear pair of low positive anomalies, but no evidence was found of a recurrence of the middle Animikie iron-formation east of Marquette.

The swing of Animikie rocks to the south in secs. 4 and 9, T. 47 N., R. 25 W., results from a major upwarping of basement rocks in the eastern parts of both sections and in the SW¼ sec. 9. This arching of the basement, although mainly along axes trending and plunging west-southwest to west-northwest, probably has an offshoot trending north-northwest which is, in effect, a crossfold extending diagonally beneath the synclinorium and reflected by the observed second-order folding of the Kona and the inferred folding of the Mesnard in the NW¼ sec. 32, T. 48 N., R. 25 W.

The broad belt of gently arched dolomite in secs. 7, 8, 17, and 18, T. 47 N., R. 25 W., and secs. 12 and 13, T. 47 N., R. 26 W., reflects the arching of the basement in sec. 9 to the east. The dolomite in this belt has many gentle, slightly plunging folds, and in many places along or adjacent to fold axes it is virtually flat lying (pl. 2, section D–D′). This anticlinal area is separated from the more tightly folded synclinorial core to the north by a west-trending fault.

FOLIATION AND CLEAVAGE

Attitudes (poles) of foliation in the gneiss north and south of the synclinorium, and of foliation and layering in the Mona Schist north of the synclinorium, have been plotted on a Schmidt equal-area net; the plotted points are contoured and compared in figures 28 *A–D*. Foliation south of the synclinorium has been plotted in two figures to bring out differences between the northern and southwestern parts of the area.

Most of the Mona Schist and Compeau Creek Gneiss contain a pervasive foliation, which in the two formational units is conformable on a regional scale. The trend of foliation is the dominant trend of the region and lies mainly between S.60°E. and S.80°E. The dips are mainly steeply north to vertical north of the synclinorium, and steeply north, vertical, or steeply south immediately south of the synclinorium. Farther south, in the southwestern part of the Sands quadrangle, southeast-striking, southwest-dipping foliation is prevalent.

The differences in the attitude of foliation from north of the synclinorium to the southern part of the Sands quadrangle may be due to unknown causes related to the origin of the foliation in early Precambrian time or may reflect buckling of the basement during late middle Precambrian orogeny. The positive magnetic anomaly that trends northwest across the central part of the Sands quadrangle (see p. 69) may represent a synclinal trough of middle Precambrian metasediments. Buckling of the basement rocks related to such infolding of Animikie rocks might explain the differences in the attitude of foliation from the northern to the southwestern parts of the quadrangle.

Cleavage in the metasedimentary rocks of Animikie age in the synclinorium, measured principally in slate and dolomite, trends mainly S.70°E. to east and dips steeply north (fig. 29). The cleavage is remarkably similar in orientation to foliation in the basement rocks north of the synclinorium (figs. 28*A*, *B*) and in the northern third of the Sands quadrangle, south of the synclinorium (fig. 28*C*.) The virtual duplication of orientation patterns for foliation of pre-Animikie age and cleavage of late Animikie age indicates a recurrence of the major pre-Animikie deformation plan at the end of Animikie time.

JOINTS

Joints occur everywhere in the Compeau Creek Gneiss, in much of the Mona Schist, and in most exposures of middle Precambrian rocks.

By far the most conspicuous joints in the Mona Schist are in the Lighthouse Point Member, where most are perpendicular to layering and strike both perpendicular and parallel to the trend of layers (fig. 2., pl. 6). These joints are here classified as cross joints and longitudinal joints, respectively. Joints oblique to layering or to the direction of strike (shear joints) are uncommon in the

Lighthouse Point Member. Blocky jointing is common in the more massive mafic metavolcanic rock and in mafic intrusive rocks. In tabular intrusive bodies, most joints are approximately perpendicular and parallel to bounding surfaces. Most such joints are probably primary cooling fractures.

About 650 joints have been measured in the Compeau Creek Gneiss north and south of the synclinorium, and 105 joints have been measured in rocks of Animikie age, mainly the Mesnard Quartzite and Kona Dolomite. The poles of these joints have been plotted on the lower hemisphere of a Schmidt equal-area net; the plotted

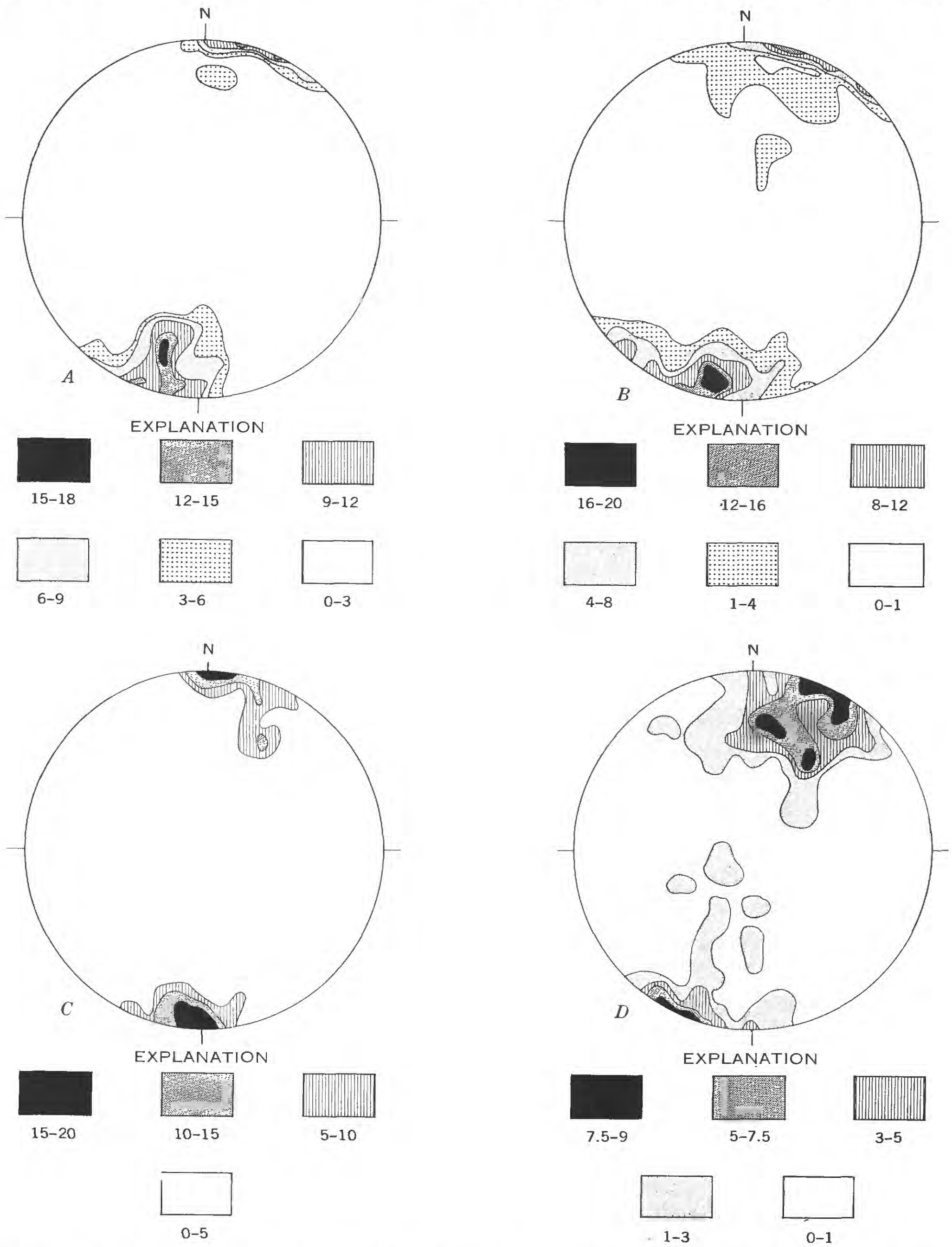

FIGURE 28.—Foliation diagrams, Marquette and Sands quadrangles. Poles of foliation surfaces plotted on lower hemisphere of Schmidt equal-area net. Shaded areas represent different percentage-concentrations of poles within 1-percent units of the total area of each diagram. *A*, 73 poles of foliation, Compeau Creek Gneiss north of Marquette synclinorium. *B*, 80 poles of foliation and foliation-layering in massive, schistose, and laminated greenstones of Mona Schist, north of Marquette synclinorium. *C*, 40 poles of foliation, Compeau Creek Gneiss, south of Marquette synclinorium, north third of Sands quadrangle. *D*, 109 poles of foliation, Compeau Creek Gneiss, south of Marquette synclinorium, southwestern part of Sands quadrangle.

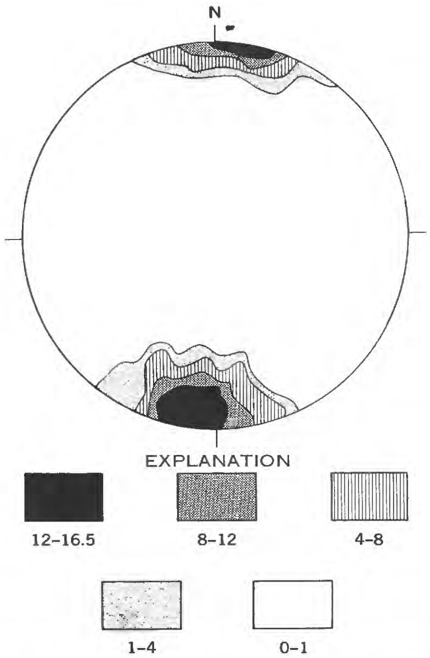

FIGURE 29.—Cleavage diagram; 219 poles of cleavage in rocks of Animikie age, eastern part of Marquette synclinorium. Poles of cleavage surfaces plotted on lower hemisphere of Schmidt equal-area net. Shaded areas represent different percentage-concentrations of poles within 1-percent units of the total area of the diagram.

points have been contoured and are shown in figures 30*A–C*. The existence of preferred trends is clearly established, and strong similarities in the patterns of the three diagrams indicates that most of the joints, if not all, are related to the same set or sets of deformation axes. Many joints in the gneiss, even though similarly oriented to joints of late Animikie age in the synclinorium, provided access for mafic dikes, now metadiabase, that were of earlier age than late Animikie. It seems certain, therefore, that joints of similar orientation are not necessarily of the same age. The difference in age of similarly oriented joints supports the conclusion, discussed in the sections on folds and foliation, that the Marquette synclinorium was folded at the close of Animikie time along an axis first established during pre-Animikie deformation.

The relationship of individual sets of joints to folds is largely conjectural because of (1) lack of knowledge of large-scale folds in the gneiss, (2) ignorance as to whether the major deformation resulted from simple compression, a force couple, or more complex movements, and (3) the impossibility of determining if all sets of joints formed during one or more deformations. Joint set B in figs. 30*A–C* is virtually parallel to foliation in the gneiss, to cleavage in the Animikie metasedimentary rocks, and to the axis of the synclinorium. Accordingly, set B is probably longitudinal joints related to the major deformation axis. The joints trending east-southeast in the outcrop of gneiss south of Wetmore Landing (pl. 3) probably belong to this set. The trend of joint set A is approximately perpendicular to that of set B in figs. 30*B* and *C* and about 75° from the trend of set B in fig. 30*A*. Set A is probably cross joints related to the major deformation axis. The joints that trend northeast in plate 3 may belong to this set. The other sets of joints are not so clearly related to the major axis, and lesser deformation axes to which they might be related are as yet unknown.

PATTERN OF JOINTS AND DIKES IN COMPEAU CREEK GNEISS NORTH OF MARQUETTE SYNCLINORIUM

Groupings of mafic dikes by trends and the total number of miles of dikes mapped in the gneiss north of the synclinorium are shown in figure 31.

Most of the greenstone (metadiabase) dikes in the gneiss trend between north and S. 75° E. (pl. 1, fig. 31). Locally, trends are more southeasterly, but only for short distances, especially in sec. 5 and in SW¼ sec. 4, T. 48 N., R. 25 W. Most of the dikes trending between N. 80° E. and S. 75° E. are considerably thicker than the dikes trending between north and N. 62° E. and may have been feeders for the thinner dikes. This relationship between the thicker and thinner dikes is also suggested by the dike pattern (pl. 1).

The diabase dikes range in trend from N. 55° E. to S. 45° E., but most trend between N. 70° E. and S. 85° E. Only a few trend more southeasterly than S. 70° E., and only for short distances. Hardly any diabase dikes trend southward, and few greenstone and diabase dikes trend more southeasterly than S. 70° E.

A comparison of these trends with dominant joint trends in the gneiss (fig. 30*A*) brings out clear associations; it also shows both conspicuous southeastward joint trends that are followed by very few dikes and conspicuous dike trends between N. 70° E. and East that parallel trends of poorly developed sets of joints. The dikes, as grouped by their trends in figure 31, are compared with the concentrations of joints that fall into corresponding ranges of trend in figure 30*A*. The grouping of dikes by trends, a, b, c, d, e, and f in figure 31 corresponds to the following percentage concentrations of joints in figure 30*A*:

Dike group (fig. 31)	*Corresponding grouping of joints and their percentage-concentrations (fig. 30A)*
a	N. 2° W. to N. 16° E. represented by 2–4 percent concentration.
b	N. 27° E. to N. 62° E. represented by 2–4.6 percent concentration.
c	No concentration of joint trends (0–1 percent).
d	S. 82° E. to S. 89° E. represented by 2–3 percent concentration.
e	S. 70° E. to S. 77° E. represented by 2–4 percent concentration.
f	S. 15° E. to S. 70° E. represented by 2–4.6 percent concentration.

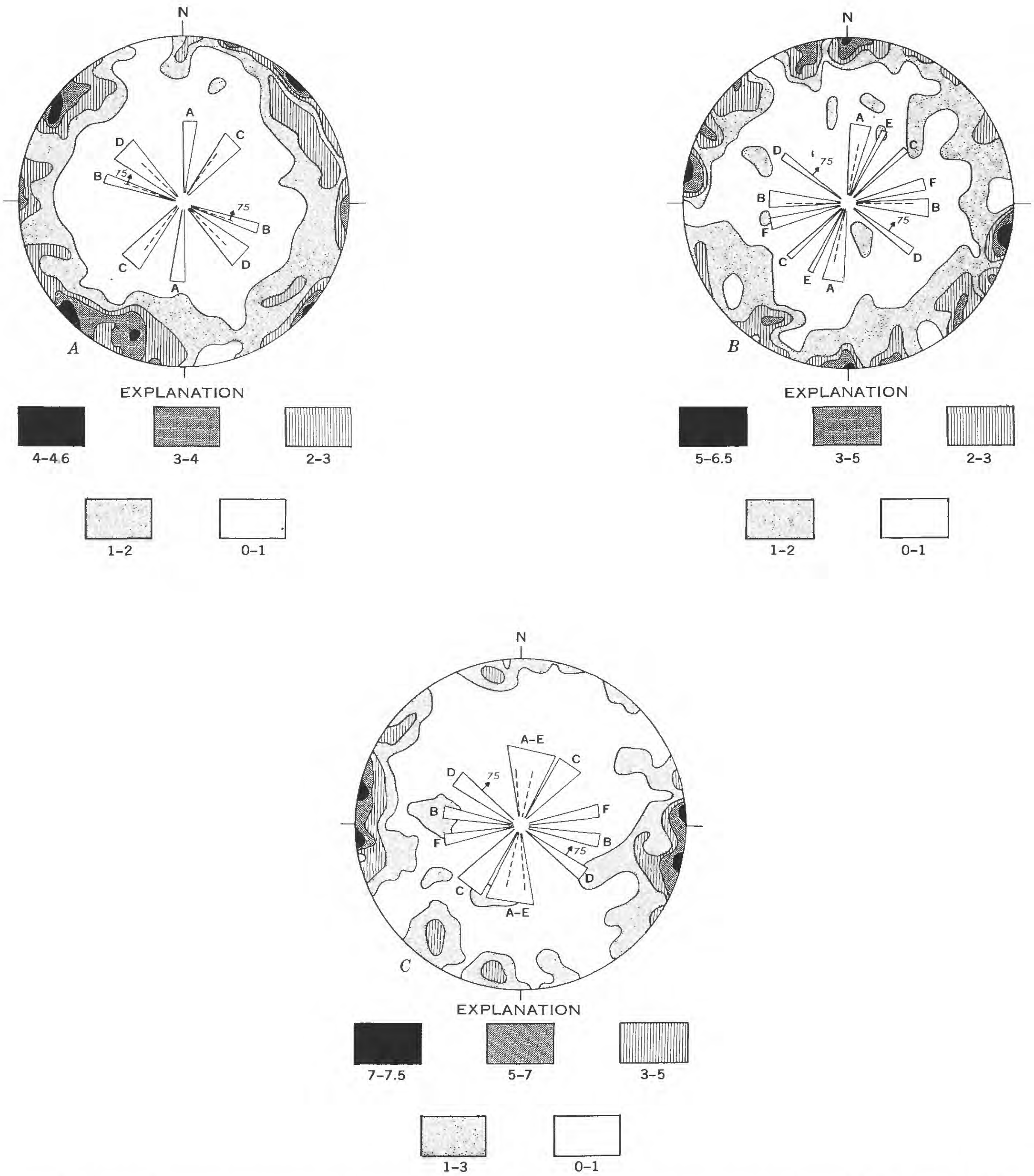

FIGURE 30.—Joint diagrams, Marquette and Sands quadrangles. Poles of joint surfaces plotted on lower hemisphere of Schmidt equal-area net. Shaded areas represent different percentage-concentrations of poles within 1-percent units of the total area of each diagram. Trends of major sets of joints indicated by superimposed rose diagram. Broken lines in some arms of the rose are trends of idealized joints derived from the approximate centers of maxima. Approximate dip of joint set or idealized joints shown only where dip departs more than 10° from vertical. *A*, 395 poles of joints, Compeau Creek Gneiss north of Marquette synclinorium. Rose diagram derived from areas of 3–4.6 percent. Broken lines in joint sets B, C, and D derived from centers of 4- to 4.6-percent maxima. *B*, 260 poles of joints, Compeau Creek Gneiss south of Marquette synclinorium. Rose diagram derived from areas of 3–6.5 percent. Broken lines in joint sets A and B derived from centers of 5- to 6.5-percent maxima. *C*, 105 poles of joints, rocks of Animikie age, eastern part of Marquette synclinorium. Rose diagram derived from areas of 3–7 percent. Broken lines in joint set A–E derived from 7- to 7.5-percent maxima.

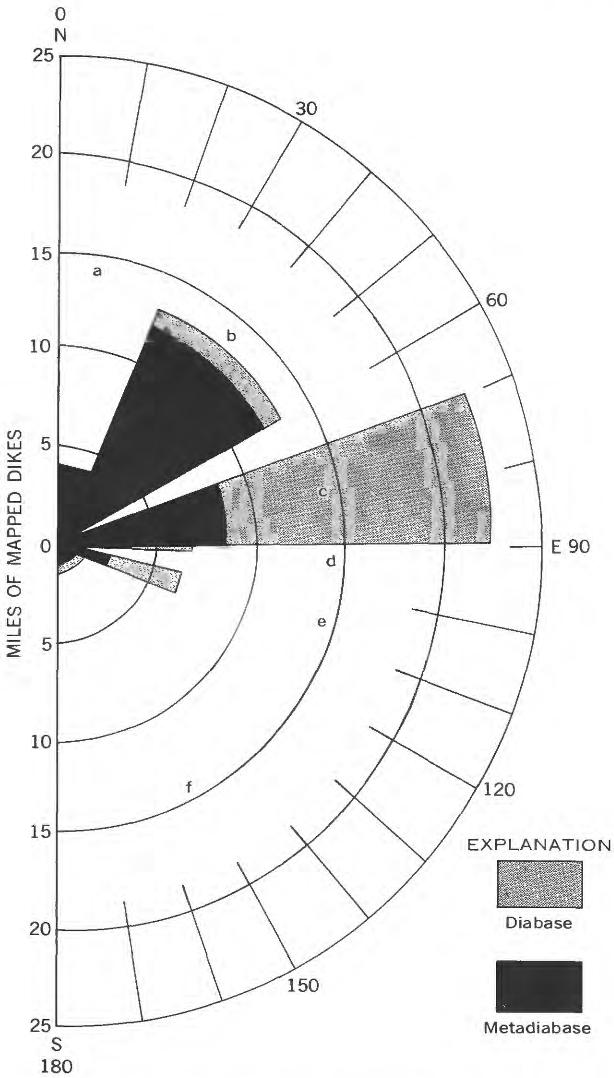

FIGURE 31.—Rose diagram showing grouping of mafic dikes by trends and total miles of dikes mapped, Compeau Creek Gneiss, north of Marquette synclinorium. (Dikes not distinguished on basis of width.)

The correspondence of dikes to some, but not all, conspicuous joint trends indicates that dikes tended to follow joints having certain orientations and to avoid those having other orientations. The deciding factor probably was the "openness" or accessibility of given joint sets at times of intrusion. Either different sets of joints were not contemporaneous, and some may have formed after intrusion of the dikes, or they formed contemporaneously but were under tension, that is, "open" at different times.

FAULTS

Faults in the area may be considered in two groups: those clearly related to the formation of the Marquette synclinorium and those in the basement rocks not clearly related to the synclinorium. No relationship is evident between faults of the two groups. Except for the inferred faults along the silicified cataclastic zone in the southwestern part of the Sands quadrangle (p. 23, 25), faults in the gneiss not in immediate proximity to the syclinorium are known only from north of the synclinorium. Only three small faults are known in the Mona Schist, two on Lighthouse Point (pl. 6) and one in the NW¼ sec. 28, T. 48 N., R. 25 W.

Faults in the gneiss have been determined by a combination of pronounced clefts cutting the gneiss and interrupted or offset mafic dikes. Additional evidence for the east-trending fault in the S½ of sec. 5, T. 48 N., R. 25 W., includes abundant white vein quartz p. 23).

Faults in the gneiss in secs. 31 and 32, T. 49 N., R. 25 W., and in secs. 3, 4, and 10, T. 48 N., R. 25 W., trend mainly west-northwest to north-northwest, roughly parallel to the shore of Lake Superior. One fault in the NW¼ sec. 32 and one in secs. 4 and 9 trend about north. The offsets of nearly vertical dikes along most of these faults indicate strike-slip movement, right lateral along the longer faults, but left lateral along two (or possibly three) of the shorter faults, such as the fault near the S¼ cor. sec. 32, the fault extending from near the center of sec. 4 to near the center of sec. 9, and possibly the north-south fault in the NW¼ sec. 32. The northwest-trending fault passing the edge of Wetmore Pond and along the cleft in the gneiss utilized by the Big Bay branch of the Lake Superior and Ishpeming Railroad appears to be along the same zone of weakness as the fault passing from sec. 4 across the SW¼ sec. 3 into sec. 10. The northwestern fault is crossed by Keweenawan diabase and is therefore pre-Keweenawan in age, whereas the southeastern fault offsets Keweenawan diabase in a right lateral sense. That the two faults are not actually one is indicated both by their differences in age and by the continuity of the greenstone dike that trends north-northwest between them in the SE¼SE¼ sec. 32. All but the fault passing from sec. 4 across sec. 3 into sec. 10 appear to be of pre-Keweenawan age. The northwest-trending joints along which some small right lateral offsets have occurred in the gneiss south of Wetmore Landing (pl. 3) may be related to the post-Keweenawan fault.

Most of the faults mapped in, or immediately bordering, the Marquette synclinorium trend generally westward, approximately parallel to the synclinorium (Gair, and others, 1961a). The major faults are in, or adjacent to, the core of the trough or are along the south margin of the synclinorium. Faults in, or adjacent to, the core of the synclinorium occur in secs. 5–8, T. 47 N., R. 25 W., and eastward from sec. 5 along the valley of the Carp River to Lake Superior. Most of the south margin of the synclinorium in this area is a fault contact between Animikie and pre-Animikie rocks. The faulted south margin can be seen in the Harvey quarry (pls. 4, 5), and effects produced very close to the fault can be seen in the SW¼ sec. 2 and at the sharp bend in the old highway in the south-central part of sec. 3, T. 47 N., R. 25 W. (pl. 4). There may be faults along the north margin of the synclinorium (p. 60–61), but their presence cannot be determined in the marquette quadrangle because of a lack of exposures.

The disposition of the faults in relation to the synclinorium as a unit, as well as in relation to individual folds, indicates that the faults are related genetically to the folds. Faults in the core probably arose out of the folding, but the faults along the south margin played a most significant part in the growth of the synclinorium itself, in producing the relative downward displacement beneath the synclinorium. Because of the evident relationship between folding and faulting, movement on the faults probably was virtually dip slip, roughly perpendicular to the major axes of folding. The apparent right lateral offsets of the generally west-dipping beds along the faults therefore indicate that the displacements on nearly all the west-trending faults, at least within the confines of the Marquette and Sands quadrangles, was north side downward.

The generally north-trending faults in sec. 4, T. 47 N., R. 25 W., and nearby parts of secs. 8 and 9, are probably related to the arching of the gneiss that folded the Animikie rocks in secs. 4, 8, and 9.

All the faults in and immediately bordering the synclinorium are of late Animikie age (or equivalent to the post-Animikie, pre-Keweenawan interval). The north-trending faults in secs. 4, 8, and 9, however, appear to be older than those that formed during folding in and adjacent to the core of the synclinorium in secs. 5–7.

AEROMAGNETIC SURVEY

The Marquette and Sands quadrangles are included in the aeromagnetic map of the Marquette iron range and adjoining areas to the north and south (Case and Gair, 1965). Buzas (1960) has previously interpreted aeromagnetic data obtained by the Jones and Laughlin Steel Corp. in parts of the Marquette-Sands area. Data for the aeromagnetic map (Case and Gair, 1965) were obtained with an AN/ASQ–3A airborne magnetometer installed in a DC–3 airplane operated by the U.S. Geological Survey. The pilot used topographic maps for guidance, and a gyrostabilized continuous-strip camera recorded the flight path of the aircraft. The locations of north- and south-trending flight traverses are indicated by tick marks on the aeromagnetic map. The distance from airplane to ground was 500 feet, measured by a continuously recording radio altimeter. J. Kirby and F. Petrafeso were in charge of the survey, and the compilation was supervised by J. Blanchett. Total-intensity profiles representing the parts of the flight traverses within the Marquette and Sands quadrangles are shown on plate 7.

The major features shown by the profiles are a northwest-trending positive anomaly (pls. 2, 7; anomaly k) in the southern part of the Sands quadrangle and numerous west-trending negative anomalies, mainly in the Marquette quadrangle. Probably all the negative anomalies represent diabase dikes of Keweenawan age. Some of the negative anomalies can be matched with diabase dikes mapped on plate 1, but close spacing of some of the diabase dikes, combined with probable small location errors in compiling the aeromagnetic data, make specific correlations of some dikes and anomalies impractical. The profiles are helpful in suggesting possible extensions of some of the mapped diabase dikes; such possible extensions are shown on plate 1. A negative anomaly a short distance south of the shore on the line between secs. 3 and 4, T. 48 N., R. 25 W. (pl. 1; anomaly n; pl. 7, profile 132, anomaly n) may represent an extension of one of the known dikes located to the west-northwest, the west-southwest, or the east-northeast, or it may represent an unmapped buried dike between these known dikes.

In the Sands quadrangle, a negative anomaly (i) extends westward across the central part of secs. 28–30, T. 47 N., R. 25 W., and sec. 25, T. 47 N., R. 26 W., and another (j) extends across the southern parts of secs. 1–6, T. 46 N., R. 25 W. Diabase is exposed along the westward extension of anomaly (j) in the Palmer quadrangle (Case and Gair, 1965), and in the Sands quadrangle these two anomalies probably represent buried diabase dikes.

Two diabase dikes are mapped in the N½ sec. 32, T. 48 N., R. 25 W., but in aeromagnetic profiles 126 and 128 a negative anomaly (g) occurs only above the southern dike. The adjacent profiles do not have this negative anomaly, so the dike evidently does not extend far east or west of its exposures. Above the northern dike the profiles flatten between negative anomaly (g) and positive anomaly (f), and the flattening may represent a negative anomaly over that dike, masked by the adjacent positive and negative anomalies. The slight dip in profile 132 at (m) is alined with the flattened parts of the two profiles to the west. The dip probably represents the diabase dike in the NE¼ sec. 33, which may well be the same as the northernmost of the two dikes in sec. 32.

The wide west-trending diabase dike crossing the south-central part of sec. 2 and the SE¼ sec. 3, T. 47 N., R. 25 W., is represented only by a small negative anomaly (l) in profile 136. In several places in the Marquette quadrangle, known diabase dikes are not indicated on profiles that cross them. This absence of a negative anomaly over known dikes of Keweenawan diabase suggests that diabase does not have negative remanent magnetism everywhere or that the negative remanent magnetism is low in places.

The conspicuous positive anomaly (k) in the southern part of the Sands quadrangle is probably related either to a large mafic intrusive body in the buried Compeau Creek Gneiss or to an infold of Animikie rocks into the gneiss (Case and Gair, 1965). The positive anomaly (d), trending west-northwest from near the NW. cor. sec. 24, T. 48 N., R. 25 W., is correlative with a swarm of dikes and irregular masses of metadiabase and metagabbro throughout most of its length. The correlation near the east end of the anomaly is uncertain owing to a lack of outcrops. The positive anomaly (f) trending westward across the southern part of secs. 28 and 29, T. 48 N., R. 25 W., is probably related to metadiabase intrusive into the Mona Schist. The positive anomaly (h) in profile 120 near the NW. cor. sec. 7, T. 47 N., R. 25 W., is probably caused by a magnetite-rich bed in the Siamo slate.

ECONOMIC GEOLOGY

Iron-bearing and copper-bearing rocks have been the principal incentive to exploration in the Marquette and Sands quadrangles, but until the present time the major mineral products have been sand and gravel, greenstone, Mesnard Quartzite, and Kona Dolomite for aggregate, breakwaters, and road rock; and the Jacobsville Sandstone for structural stone. Several test pits at the north end of Presque Isle also have been opened for nickel, silver, and lead, small amounts of which occur in a vein in the serpentinized periodotite (p. 56).

The Jacobsville was quarried in the NW¼ sec. 26, T. 48 N., R. 25 W., during the early years of Marquette, and many older buildings in the city are constructed of this rock. Abandoned greenstone quarries in intrusive greenstone and in the Mona Schist are located in the west-central part of sec. 10, the SE¼ SE¼ sec. 22, and the SW¼NW¼ sec. 29, T. 48 N., R. 25 W., and near the center of sec 1, T. 47 N., R. 25 W. Both quartzite and greenstone were taken from the Harvey quarry in the NW¼NW¼ sec. 6, T. 47 N., R. 24 W., and quartzite has been quarried along the railroad spur into the State prison in the central part of sec. 36, T. 48 N., R. 25 W. Dolomite has been quarried a short distance west of the center of sec. 8, and is now being taken from the north-central part of sec 17, T. 47 N., R. 25 W. Sources of sand and gravel are widespread. The largest sand and gravel pit is in secs. 8 and 17 in the northern part of the Sands quadrangle.

IRON

The principal occurrence of iron in the Marquette-Sands area was at the old Eureka mine, 3,850N-3,000E, SW. cor. sec. 21, T. 48 N., R. 25 W. Perhaps as much as several hundred tons of earthy goethitic iron ore was mined from a shear zone about 5 feet wide in chloritic slaty greenstone (Mona Schist). Secondary quartz and iron-bearing carbonate were deposited along the shear zone in places, prior to formation of the iron ore. The ore resulted from oxidation of carbonate and chlorite, the concentration of iron, and the removal of silica, MgO, CaO, and alumina by ground water circulating along the shear zone. Virtually all mining was done prior to 1880, and whatever ore was extracted was shipped to local charcoal-burning smelters. There were several unsuccessful attempts to renew mining in the 1880's, and three test holes were drilled at the site in the 1920's.

Test pits for iron are widespread along the belt of Animikie rocks and in the Mona Schist. They have been dug mainly in brecciated quartzose parts of the Kona Dolomite, in sheared oxidized ferruginous parts of the Mona Schist, and in some oxidized parts of the Wewe and Siamo Slates. A few test pits are also found in the Mesnard Quartzite. The iron in most of these places is a secondary concentration of hematite or goethite, or both, in cracks and shears, and cementing fragments of breccia. The likelihood of marketable concentrations of iron in these occurrences is slim indeed because of the small volume and erratic nature of controlling zones of breccia or shears.

The largest and greatest number of test pits for iron were dug in the NE¼NE¼ sec. 24, T. 47 N., R. 26 W., adjacent to the fault mapped between Wewe Slate and Kona Dolomite. Quartzitic parts of the Kona and some silicified rock and vein quartz are impregnated with secondary iron oxide, mainly specular hematite. The locations and geologic setting of test pits for iron ore are given in table 14.

COPPER

A considerable number of test pits and exploration shafts for copper have been dug in Kona Dolomite in the southeastern and northeastern parts of the Marquette and Sands quadrangles, respectively. The explorations have been made almost entirely in the north-central part of sec. 1, T. 47 N., R. 25 W., south of the crest of the principal west-trending ridge, and within 600 feet south and southeast of Buschell Lake (called Copper Lake on early maps) in sec. 2. One exploration shaft has been dug in the NE¼ sec. 32, T. 48 N., R. 25 W. These explorations are probably all more than 40

TABLE 14.—*Locations and geologic setting of test pits for iron, Marquette and Sands quadrangles*

[Measured in feet from southwest corner of section, except as indicated]

		Geologic setting
Marquette quadrangle		
T. 48 N., R. 25 W.:		
sec. 15	100N–850E	Oxidized slaty amphibole schist.
sec. 16	500N–4,550E	Oxidized shear zone in chloritic schist.
	750N–5,200E	Laminated magnetic slate phase of Mona Schist, adjacent to dike of Keweenawan diabase.
	1,000N–3,600E	Oxidized amphibole schist adjacent to dike of Keweenawan diabase.
sec. 21	2,700N–4,250E	Oxidized chloritic slate, Mona Schist.
	4,100N–3,300E	Oxidized schistose greenstone.
sec. 27	5,000N–3,100E	Oxidized chloritic slate in Mona Schist.
	5,200N–2,300E	Do.
sec. 31	150N–950E and vicinity.	Oxidized chloritic sericitic Siamo Slate.
	550N–550E	Oxidized chloritic Wewe Slate.
	550N–900E and vicinity.	Do.
	950N–550E	Do.
sec. 35	2,800N–500E	Zone of ferruginous breccia in Mesnard Quartzite.
T. 47 N., R. 25 W.:		
sec. 6	950S–4,550E of NW. cor.	Oxidized chloritic Siamo Slate.
Sands quadrangle		
T. 47 N., R. 25 W.:		
sec. 2	1,650N–200E	Brecciated bed of sericitic ferruginous quartzite in Kona Dolomite, containing secondary hematite and vein quartz.
sec. 6	2,500N–4,650E	Oxidized Siamo Slate.
	2,700N–4,050E	Do.
T. 47 N., R. 26 W.:		
sec. 12	950S–1,850W of NE. cor.	Do.
sec. 24	300S–1,000W of NE. cor. and vicinity	Secondary hematite and some vein quartz in and near beds of quartzite and ferruginous quartzite of Kona Dolomite. Beds locally brecciated.

years old and have not resulted in any known production. Since 1961 the general area in which these old test pits and shafts is located has been further explored by mining companies.

The locations of old test pits and shafts in secs. 1 and 2 are as follows:

sec. 1 (measured from NW. cor.):
- 1,100S–1,900E to 1,000S–1,750E (row of pits)
- 1,100S–2,500E to 1,000S–2,700E (row of pits)
- 1,200S–2,150E

sec. 2 (measured from SW. cor.):
- 2,200N–2,700E
- 2,250N–2,800E
- 2,250N–3,150E
- 2,400N–3,400E
- 2,600N–3,500E

The main copper minerals are chalcopyrite, chalcocite, native copper, and malachite(?). Reed (1965) also reported pyrite extensively replaced by chalcocite, alteration of chalcopyrite to bornite, and of some bornite and chalcocite to covellite or to covellite and specular hematite. The sulfides tend to oxidize to a brownish or bluish tarnish. Some azurite has also been seen. The copper minerals are disseminated in sericite slate, quartz-sericite slate, sericitic dolomite slate, quartz-sericite wacke, and sericitic quartzite in the lower part of the Kona Dolomite in secs. 1 and 2, and in a quartz vein that cuts the Kona in sec. 32. Tests for copper locally show as much as 7 percent but in most samples there is less than 2 percent copper (Reed, 1965, p. 46).

The mineralization in sec. 1 is close to, and may have been controlled by, a fault separating the Kona Dolomite from the Mesnard Quartzite and Enchantment Lake Formation (pl. 1, section *A–A′*; pl. 4). The source of copper is not known. The evident relationship between mineralization and the post-Animikie, pre-Keweenawan fault in sec. 1, and the association of copper mineralization with the probable late Animikie or post-Animikie quartz vein in sec. 32 indicate that the sulfide mineralization is probably late Animikie or younger in age. The quartz vein in sec. 32 probably followed north-trending cross joints formed in the Kona during late Animikie folding. Sulfide mineralization was probably either a result of the concentration of copper from small amounts in the lower Animikie sedimentary rock during the late Animikie deformation and metamorphism, or a result of emanations from mafic magma during Keweenawan time. Malachite (?), azurite, and possibly native copper are products of surface oxidation, by weathering.

The occurrences of copper sulfide mineralization in or near fault and breccia zones and in a quartz vein are not likely to contain large concentrations of copper, though locally the grade could be high. The nearby country rock, however, especially the slate of the lower part of the Kona Dolomite, may contain large total amounts of copper as low-grade disseminations.

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