GEOLOGY OF MINNESOTA: A Centennial Volume

P. K. Sims and G. B. Morey, *editors*

GEOLOGY OF MINNESOTA: A CENTENNIAL VOLUME

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In honor of George M. Schwartz

P. K. Sims and G. B. Morey, *editors*

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PREFACE

It seems appropriate on the 100th anniversary of the Minnesota Geological Survey and the Department of Geology and Geophysics of the University of Minnesota, Minneapolis, to summarize our current knowledge of the geology of the State of Minnesota. Summations of the geology of the Precambrian rocks, which have been the object of the greater part of geologic research in the state over the years, were made in 1951 by F. F. Grout, G. M. Schwartz, J. W. Gruner, and G. A. Thiel and later, in 1961, by S. S. Goldich and colleagues. Not since 1901, however, when the Final Report of the Geological and Natural History Survey of Minnesota was published by N. H. Winchell, has a review been undertaken of all aspects of the geology.

The long history of studies in the state has provided a step-wise growth of geologic knowledge. The early work by the Geological and Natural History Survey, prior to 1900, formed a solid foundation for later work. Reconnaissance geologic maps were prepared for each county and descriptions were made of nearly all the major rock series. One outcome of this work was the delineation of the important iron ores on the Mesabi range, which until recent years formed the backbone of this country's mineral industry. Subsequent to the termination of the "Winchell Survey," the geologic work of the Minnesota Geological Survey was carried out mainly by the faculty and graduate students of the Department of Geology. Most of these studies were topical, and were related to problems of specific interest to the faculty or to problems of economic significance. Notable among these studies were the following: F. F. Grout's research on the Duluth Complex, which we know today contains this country's largest combined copper and nickel resource; J. W. Gruner's mapping and research on the Mesabi range, which delineated the vast tonnages of taconite that currently are being mined; G. M. Schwartz's premier investigations of the environmental geology of the Minneapolis-St. Paul urban area; G. A. Thiel's investigations of the state's ground-water resources; and Goldich's radiometric age dating, which provided a time-stratigraphic framework for the Precambrian throughout the entire Lake Superior region. In addition, graduate theses prepared during this time—many of which were not published—contain a wealth of geologic data. Observations noted in these reports remain useful, and often have flavored the interpretations made by contributors to this volume.

Much progress has been made in our understanding of the geology of Minnesota during the last decade. A large part of the new knowledge has been gained in the course of systematic geologic mapping. During this decade, reconnaissance geologic maps (scale 1:250,000) were completed of all parts of the state in which outcrops are present, and mapping of the surficial deposits, at the same scale, was started. More detailed geologic maps were prepared in a few selected areas, especially in Precambrian terranes. Concurrently, the radiometric age dating and the geochemical studies that were started during the preceding decade were continued, also at an accelerated pace. These data have provided a time framework, critical to an understanding of time-stratigraphic relations and correlations, and have aided immensely in unraveling the geologic history.

The text of this report was written by persons actively engaged in the geologic and geochronologic studies in the state during the past decade. The specific individuals who had done the work were asked to write sections for which they had firsthand knowledge. The task of filling in the gaps was assumed by the editors. The text is organized in the format of a historical geology text. In this way, the geology can be described with respect to units that have comparable age and geologic history. Although an attempt is made to cover all significant aspects of the geology of the state, the coverage is unequal, as will be evident upon perusal of the text.

The progress that has been made in recent years in the knowledge of the geology of Minnesota largely was made possible by steadily increasing levels of funding by the Minnesota State Legislature. In the early 1950's, a substantial topographic mapping program, which was essential for later geologic mapping as well as for many other purposes, was initiated by funds appropriated by the legislature and contributed by the Minnesota Iron Range Resources and Rehabilitation Commission. In 1963, the topographic mapping program was accelerated, and a long-range geologic mapping program was initiated by funds provided from the Natural Resources Account by the Minnesota Resources Commission. Since that time, the geologic programs in the state have been sustained mainly by legislative appropriations and by supporting funds from the Departments of Iron Range Resources and Rehabilitation and Natural Resources. National Science Foundation grants have supported most of the geochronologic and geochemical studies pertinent to the geology of the state.

This volume could not have been completed without the able assistance of Ingeborg Westfall, who aided immeasurably in editing, and Richard N. Darling, who designed and drafted the illustrations. In addition, Lillian Krosch, aided by Jeanne Zaspel and Lynne Farmer, patiently and expertly struggled with the typing of a sometimes almost undecipherable manuscript.

P. K. Sims

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St. Paul, Minnesota

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AN APPRECIATION

Seldom has a geologist combined academic and practical careers as successfully as George Melvin Schwartz. As a teacher, Schwartz—as he is known to his former colleagues on the faculty of the Department of Geology and Geophysics—has had a significant impact on economic geology in this country, and at the same time he has been a dedicated public servant, highly respected by those with whom he has worked.

George Melvin Schwartz was born on a farm in Oakfield, Wisconsin, in September, 1892. He attended grade school in Byron Township, Wisconsin, and high school (1907-1911) at Fond du Lac, Wisconsin. His interest in the outdoors led him to enroll in geology at the University of Wisconsin, at Madison, where he received the B.A. degree in 1915 and the M.A. degree the following year.

Schwartz's teachers at the University of Wisconsin, and particularly C. K. Leith, had a profound influence on shaping his career in geology, and not surprisingly his first professional position was with the Wisconsin Geological Survey, during the summers of 1914 and 1915. Here, he got his "feet wet" in more ways than one, chasing down magnetic anomalies under the guidance of Ernie Bean, then State Geologist. Clearly, this experience did not dampen his spirits, for many years later he traversed similar swamps in the State of Minnesota while investigating magnetic anomalies that had been discovered by airborne magnetic surveys. For a brief time, from June, 1916 to the end of December, 1918, he was a geologist for the Copper Range Company, exploring for copper near Painesdale, Michigan. This position, which he fondly recalls still today, was terminated abruptly by a call to duty by the U.S. Army on January 1, 1918.

Schwartz's career in the U.S. Army spanned 16 months, and he was fortunate to survive front-line duty and the flu epidemic of 1918. As a Second Lieutenant in the Field Artillery, he was at the front from September to November, 1918, and was in the battles of St. Mihiel and the Argonne.

Schwartz entered graduate school at the University of Minnesota in 1919, to study under W. H. Emmons. He was awarded the Ph.D. degree four years later, in 1923, a remarkable accomplishment considering that he held an appointment as Instructor in Geology while doing graduate work. Probably it was at this time that he developed the efficient and productive working habits that he followed throughout his long professional career.

In 1920, he married Ruth Harriet Tucker. They have two sons, George Jr. and John, and one daughter, Ruth. Mrs. Schwartz survived early illnesses that could have crippled a person of less hardy stock, and has been an enormous asset to George in his personal life and his professional career. There are seven grandchildren.

Schwartz was appointed an assistant professor to the Minnesota faculty in 1923, an affiliation he retained until his retirement in 1961. While on sabbatical leave during the 1940-41 academic year he taught at Laval University, in Quebec. Some dozen students got their Ph.D. degrees under his direct supervision. His students were employed in teaching, by the U.S. Geological Survey, and by industry, and several have attained outstanding success in management as well as in geology.

In parallel with Schwartz's teaching career is a long and illustrious one with the Minnesota Geological Survey. Summer field work with the survey provided Schwartz and some of his students the opportunity to carry out research on a variety of problems. At the same time, his work with the survey provided an outlet for his desire to "bring geology to the people." Beginning in 1921 and continuing without interruption until 1942, Schwartz worked on a variety of problems ranging from a survey for sand and gravel, mainly for the Minnesota Highway Department, to a study of the anorthosites of the north shore of Lake Superior. He had a marked impact on the development of knowledge of the geology of Minnesota. Two of his most satisfying projects were investigations of the geology of the Minneapolis-St. Paul and Duluth urban areas. His report on the Minneapolis-St. Paul area, published in 1936, was a pioneering effort in environmental geology, and conservatively has been worth several hundreds of thousands of dollars to governmental units and the taxpayers of the area. The work that has been most satisfying to him, though, is *Minnesota's Rocks and Waters*, a popular book by Schwartz and his colleague George Thiel. Since publication in 1954, more than 18,000 copies have been sold. A third edition now is being prepared for publication.

During the latter part of World War II, Schwartz worked on the copper deposit at San Manuel, Arizona, for the U.S. Geological Survey. His accurate and comprehensive description of the deposit provided the basis for interpreting the structure and alteration in an adjacent area, which led to the discovery by private industry in 1965 of a faulted offset of San Manuel, the Kalamazoo ore body. Schwartz's last paper on copper deposits, written by invitation for inclusion in the volume on porphyry copper deposits, dedicated to Eldrich Wilson, was published in 1966, 45 years after his first paper on the subject.

From 1947 until his retirement in 1961, Schwartz was director of the Minnesota Geological Survey. During this period he completed a study of the geology of Cook County, which had been carried out mainly by F. F. Grout until his death, and supervised research on metamorphism of the Biwabik Iron-formation by the Duluth Complex. Among the many notable contributions during this period were his successful efforts to initiate a progressive topographic mapping program in Minnesota. With the strong support of Senator Elmer L. Andersen, later elected Governor of Minnesota, and the Minnesota Iron Range Resources and Rehabilitation Commission, an accelerated program of mapping was initiated. As a result of this and the impetus given later by the Minnesota Resources Commission, more than 85 percent of Minnesota is covered today by modern 7.5-minute quadrangles.

Following are some of the honors Schwartz has received and services he has performed.

Professional Societies

Geological Society of America
Councilor, 1955-57
Society of Economic Geologists
President, 1958
Société Géologique de Belgique
Honorary Fellow

Services to State of Minnesota

Chairman, Minnesota State Mapping Board, 1949-1961
Member, Upper Mississippi Drainage Basin Committee, 1934-1940

During Schwartz's long, active scientific career, which now spans nearly 55 years, he completed 143 papers of a technical nature or of broad general interest. The breadth of his interest and capabilities is indicated by the subjects he has covered, ranging from the geology and scenery of Minnesota state parks to complex mineral deposits and Precambrian stratigraphy.

Schwartz is a warm, generous person, with an almost too great modesty. It is doubtful that he has ever had an enemy. His former students speak fondly of him, a tribute to his many personable traits as well as his competence as a teacher and scientist. Those of us who know him as a friend or colleague are better because of it.

P. K. Sims

St. Paul, Minnesota
April 29, 1972

Chapter I

RÉSUMÉ OF GEOLOGY OF MINNESOTA

P. K. Sims and G. B. Morey

RÉSUMÉ OF GEOLOGY OF MINNESOTA

P. K. Sims and G. B. Morey

Minnesota lies on the southern margin of the Superior Province of the Canadian Shield, a great region of Precambrian rocks that comprises the exposed part of the North American craton (fig. I-1). Except on the northern side adjacent to Canada, the Precambrian rocks are overlain unconformably by strata of Paleozoic and Mesozoic age, which form a thin platform cover of relatively undisturbed rocks that thicken southward and westward. Inliers of Precambrian rocks protrude through the Phanerozoic cover in southwestern Minnesota. The present landscape is inherited largely from Pleistocene continental glaciation, which produced a variety of erosional and depositional landforms. The glacier ice scoured the bedrock in northeastern Minnesota, in much the same way as throughout most of Canada, and deposited materials of diverse lithology and provenance, as much as 500 feet thick, over most of the remainder of the state.

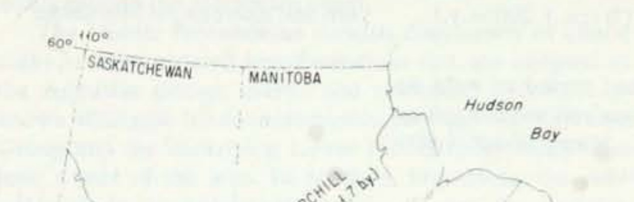

Figure I-1. Map of north-central North America showing (a) the Late Precambrian midcontinental rift system (horizontal ruling), (b) the boundary between contrasting magnetic anomalies in South Dakota (dashed line) and its possible connection with the boundary between the Superior and Churchill Provinces of the Canadian Shield, and (c) the outcrop area of the Upper Precambrian Sioux Quartzite (diagonal ruling), and the margin of the platform cover of Phanerozoic rocks. (Compiled by D. L. Southwick, 1970.)

PRECAMBRIAN ERA

The Precambrian rocks in Minnesota can be divided by means of classical stratigraphy into three sequences, which are separated from one another by major unconformities. Because each of the sequences has distinctive lithologic and structural characteristics, broad regional correlations can confidently be made for most of the major successions. Radiometric age dating of igneous and metamorphic rocks within the sequences has provided a time framework for the Precambrian (Goldich and others, 1961; Goldich, 1968), which has been widely accepted as a standard for the Lake Superior region. Although direct dating of time of deposition of the stratified rocks has been largely unsuccessful, certain limits can be placed on their times of deposition by determining the radiometric age of metamorphism and of the igneous rocks intrusive into them.

The time-stratigraphic framework of the Precambrian rocks in Minnesota, shown in Table I-1, is a slight modification of the stratigraphic succession and time classification of Goldich (1968; see also table II-3, this volume). In the table, the volcanic and sedimentary successions are described in terms of gross lithologies and environments of deposition rather than by formational names. The arbitrary time division placed between Early and Middle Precambrian is 2,600 m.y. and that placed between the Middle and Late Precambrian is 1,800 m.y., following earlier designations by Goldich (1968). However, there is some evidence that a mild deformation and accompanying low-grade metamorphism affected the Middle Precambrian rocks in and adjacent to the Mesabi range approximately 1,600-1,700 m.y. ago, and this event probably should be included in the Middle Precambrian. An arbitrary time division at 1,600 m.y. would be consistent with that between Precambrian X and Precambrian Y established by the U.S. Geological Survey (James, 1972).

The Early Precambrian includes rocks involved in or emplaced during the Algoman orogeny, and is equivalent to the Archean as used by the Geological Survey of Canada (Stockwell, 1964). The Algoman orogeny was a major tectonic-igneous event throughout northern Minnesota, adjacent areas in Ontario, and apparently also in southwestern Minnesota. In northern Minnesota, it comprised multiple folding and faulting, metamorphism, and emplacement of several types of granitic and syenitic rocks.

Table I-1. Time-stratigraphic framework of Precambrian rocks in Minnesota.

Era	Lithology	Event (Approximate age)	Intrusive Rocks
Paleozoic			
600 m.y.	------- Unconformity ------		
	Clastic rocks having dominantly fluvial attributes		
	------- disconformity ------		Duluth Complex and smaller mafic intrusions in northeastern Minnesota
Late Precambrian	Mafic lava flows, tuffs, and interbedded sedimentary rocks		
	------- Unconformity ------		
	Clastic rocks having shallow-water attributes	Deformation and metamorphism (?) (ca. 1,200 m.y.)	Diabasic gabbro dikes in northern and east-central Minnesota
		Unnamed period of mild deformation and metamorphism in NE. Minn. about 1,600-1,700 m.y.	
1,800 m.y.	------- Unconformity -------	Penokean orogeny (ca. 1,850 m.y.)	Granitic plutons in east-central Minnesota and small plutons of granitic and mafic rocks in Minnesota River Valley
	Clastic rocks and iron-formation having shallow-water to deep-water, or miogeosynclinal, attributes		
Middle Precambrian	------- disconformity ------		
	Clastic and carbonate sedimentary rocks having shelf-like attributes		Dikes of diabasic gabbro and diorite in northern Minnesota (stratigraphic position relative to the shelf-like strata unknown)
2,600 m.y.	------- Unconformity -------	Algoman orogeny (ca. 2,700 m.y.)	Small syenitic plutons in northern Minnesota; Granitic rocks of Vermilion, Giants Range, and Saganaga batholiths in northern Minnesota, McGrath gneiss in east-central Minnesota, and Sacred Heart and Ortonville plutons in southwestern Minnesota
	Lava flows, pyroclastic deposits, and graywacke-type sandstones having deep-water attributes		
Early Precambrian	------- Unconformity? -------		
	Gneisses of igneous and sedimentary derivation	Mortonian event (ca. 3,550 m.y.)	Morton Gneiss and Montevideo gneiss, at least in part, in southwestern Minnesota

The available radiometric ages from this region suggest that deposition of the volcanic and sedimentary rocks that comprise the several greenstone belts, deformation and metamorphism, and emplacement of the igneous rocks took place within an interval of about 50 m.y., from 2,700 to 2,750 m.y. ago. In southwestern Minnesota, metamorphism, deformation, and emplacement of granitic rocks about 2,650 m.y. ago largely obliterated an earlier record indicated by U-Pb data on zircons and by Rb-Sr data from whole-rock analyses. Two gneisses—the Morton Gneiss and the Montevideo gneiss—in this area, which represent segments of a probable widespread, interlayered sequence of gneisses, have been dated at about 3,550 m.y. (Goldich and others, 1970; Goldich, chap. II, this volume), and are the oldest rocks known on the North American continent. The two bodies that have been dated are interpreted by Goldich and others (1970) as syntectonic intrusions emplaced about 3,550 m.y. ago; they called this tectonic-igneous episode the Mortonian event.

The Middle Precambrian consists dominantly of clastic rocks and intercalated iron-formations that are assigned to the Animikie Group; clastic and carbonate rocks of unknown thickness lie disconformably between the Animikie Group and the underlying Lower Precambrian rocks in at least a part of the area. In northern Minnesota, the Animikie strata unconformably overlie Lower Precambrian rocks and are mildly deformed, whereas in the Cuyuna district of east-central Minnesota equivalent strata are strongly deformed. An isochron age of 1,850 m.y. obtained on metasedimentary rocks of the Cuyuna district is interpreted (Goldich, chap. II, this volume) as a minimum age for the time of folding of these rocks. This event was named the Penokean orogeny by Goldich and others (1961).

Some evidence exists that the rocks exposed on and adjacent to the Mesabi range were deformed more than once. The main structures are folds that are similar in orientation and style to those in east-central Minnesota, probably indicating that they were formed during the Penokean orogeny. Less conspicuous structures, mainly expressed by cataclasis, locally are superposed on the folds. Neither the extent nor the time of the younger deformation is known. However, a large number of K-Ar whole-rock and mineral ages (Goldich and others, 1961; Hanson, 1968; Hanson and Malhotra, 1971) and several whole-rock Rb-Sr isochron ages from these rocks cluster around 1,600-1,700 m.y.

An unconformity representing an unknown period of time separates the Middle and Late Precambrian. The first recognized event in the Late Precambrian was deposition of shallow-water shelf deposits, represented by the Sioux Quartzite in southwestern Minnesota. The Sioux Quartzite, which has a basal conglomerate that contains granule-bearing iron-formation of probable Middle Precambrian age, is inferred to unconformably overlie Lower and Middle Precambrian rocks. Rhyolite that apparently is interlayered with the Sioux Quartzite in northern Iowa has an apparent Rb-Sr age of $1,470 \pm 26$ m.y., which is interpreted (Lidiak, 1971) as a minimum age for the Sioux Quartzite.

All subsequent Late Precambrian events in Minnesota are assigned to the Keweenawan Period because of the lithologic and structural similarities of the rocks to those exposed on the Keweenaw Peninsula of Michigan. The Keweenawan was divided into three segments—Early, Middle, and Late. The Middle Keweenawan was defined as comprising a "short-lived" igneous event during which lava flows and interflow sediments (North Shore Volcanic Group in Minnesota) were formed and most gabbroic intrusions (Logan intrusions, Duluth Complex, Beaver Bay Complex, and other bodies) were emplaced. This igneous event was thought to have been preceded by and followed by deposition of dominantly clastic sediments. More detailed geologic studies over the past decade have shown that this classification, which is based largely on rock-stratigraphic units, is no longer tenable. However, the new data have not yet been integrated throughout the Lake Superior region, and a time-stratigraphic classification that is acceptable to all workers has not yet been established. The state of uncertainty is illustrated in this volume by the dissimilarities in the correlation charts of J. C. Green and Campbell Craddock. Although both writers are in general agreement, Craddock relies mainly on the classical methods of correlation for Keweenawan rocks, whereas Green utilizes paleomagnetic data as a major tool for correlation. The latter approach not only results in slightly different correlations of rock units than previously, but also leads to the assignment of substantial proportions of the lavas to the Lower Keweenawan. This implies that igneous activity occurred over a longer time span than previously believed. Radiometric age dating has not yet been successful in resolving the many correlation problems that plague Keweenawan stratigraphy, but it has shown that the major igneous events in Minnesota took place in the interval 1,200-900 m.y. ago (Goldich, this volume).

The end of the Precambrian and the beginning of the Phanerozoic has not been dated in Minnesota. However, Upper Cambrian sandstones unconformably overlie several types of Precambrian igneous, sedimentary, and metamorphic rocks on which a thick regolith had developed locally. Clearly, the Precambrian-Cambrian boundary in Minnesota is marked by a major unconformity that represents a time span of substantial duration.

Early Precambrian

Rocks of Early Precambrian age are exposed over wide areas in northern Minnesota and in a narrow window through the Phanerozoic cover along the Minnesota River Valley in southwestern Minnesota. The rocks in northern Minnesota are typical of the Archean greenstone-granite complexes in the Canadian Shield (McGlynn, 1970, p. 44-71). The older rocks in southwestern Minnesota are similar to the infracrustal rocks in many of the shield areas of the world, and are more intensely metamorphosed and apparently have had a more complex tectonic-igneous history than the rocks in northern Minnesota. The boundary between these contrasting terranes of different age is not exposed.

Northern Minnesota

The Lower Precambrian rocks in northern Minnesota consist dominantly of three rock assemblages—mafic to felsic volcanic rocks, graywacke-type sedimentary rocks, and granitic rocks—that have repeated patterns of distribution and structural habit. The volcanic and sedimentary rocks occur within curvilinear belts that trend northeastward and are enclosed by granitic rocks of batholithic dimensions (see pl. 1). Because of the characteristic greenschist-facies metamorphism, the volcanic-sedimentary complexes are referred to as greenstone belts.

The Vermilion district, in northeastern Minnesota, is the best exposed and most thoroughly studied of the greenstone-granite complexes in the state. It consists of a complicated metavolcanic-metasedimentary sequence, characterized by interfingering of lithologies and repetitions of rock types, which is bordered on the south by the Giants Range batholith, on the north by the Vermilion batholith, and on the east by the Saganaga batholith. The sequence trends northeastward, is steeply inclined, and is known from primary structures in the rocks, such as pillow tops and graded beds, to dominantly face northward. The oldest known strata are metabasalt and related rocks, formally named the Ely Greenstone. Pillowed or massive mafic lava flows and synvolcanic diabase dominate the formation, and andesitic lavas and pyroclastic deposits and lesser chert, banded iron-formation, and siliceous, carbonaceous tuff comprise the remainder. The Ely Greenstone is overlain stratigraphically by pyroclastic rocks of general dacitic composition and by volcanogenic sandstones, which together comprise the Knife Lake Group in the eastern part and the Lake Vermilion Formation in the western part of the district. The sedimentary rocks—dominantly interbedded graywacke and mudstone having compositions ranging from feldspathic and lithic graywacke to subgraywacke—are complexly interbedded with tuff and agglomerate; polymictic conglomerates occur locally. The proportion of graywacke to pyroclastic rocks increases as one proceeds away from apparent volcanic centers. Both the Knife Lake Group and the Lake Vermilion Formation contain scattered lenses and layers of mafic volcanic rocks, many of which are pillowed. Locally, a persistent iron-formation (Soudan) intervenes between the Ely Greenstone and overlying volcaniclastic rocks. In the central part of the district, in the vicinity of Ely, a second major mafic volcanic succession (Newton Lake Formation) conformably overlies at least a part of the Knife Lake Group. It consists mainly of andesitic and basaltic lavas and intrusive diabase, but also includes felsic-intermediate tuffs, agglomerates, breccias, and impure siliceous marble. Coeval hypabyssal porphyries of intermediate and felsic compositions as well as metadiabase locally intrude the sequence, especially the Ely Greenstone. Several thin, differentiated, sill-like bodies of mafic-ultramafic composition occur in the Newton Lake Formation. Typically, these consist from bottom to top of serpentinized peridotite, pyroxenite, hypersthene gabbro, and quartz-gabbro or granogabbro with granophyre.

The volcanic rocks and derivative graywacke-type sedimentary rocks constitute a complex volcanic pile accumulation that is grossly similar to the sequences in other greenstone belts in the Superior Province (Goodwin, 1968a). The sequence is interpreted from bedding characteristics and internal structures in the rocks and from analogies to more recent submarine volcanic rock sequences to have been deposited mainly in a subaqueous environment. The sedimentary rocks were derived largely from dacitic pyroclastic deposits (Ojakangas, 1972; this volume), and apparently were deposited by turbidity currents that moved the volcanic detritus down the slopes of the volcanic piles into adjacent basins. In those areas that have been studied, most of the graywacke represents reworked felsic pyroclastic rocks rather than epiclastic deposits. The presence of clasts of Saganaga tonalite and of greenstone in conglomerates interbedded with graywackes and tuffs in the Knife Lake Group (Gruner, 1941; McLimans, this volume) in the eastern part of the district, however, is indicative of at least local weathering and erosion of previously consolidated materials.

Granitic rocks having diverse compositions and structural characteristics intruded the metavolcanic-metasedimentary sequence during the Algoman orogeny. The dominant intrusive bodies—the Giants Range, Vermilion, and Saganaga batholiths—were emplaced virtually synchronously with regional deformation. The Giants Range batholith, composed of many separate plutons that range in composition from tonalite to granite, was intruded into and cuts out the lower part of the volcanic succession (Ely Greenstone), whereas the Vermilion batholith, herein renamed the Vermilion granite-migmatite massif (Southwick, this volume), was intruded into the upper part of the stratigraphic sequence, mainly by passive emplacement into folded sedimentary and volcanic rocks. The Vermilion massif is composed dominantly of gray and pink granite that is interleaved in varying proportions with biotite schist and amphibolite. The Saganaga batholith, at the eastern end of the district, is composed dominantly of tonalite (Goldich and others, in press) which was partly altered by late-stage fluids to more potassic rocks.

Small bodies of syenitic rocks and associated lamprophyres were emplaced into the supracrustal rocks near the end of the regional deformation, for they are generally discordant to the structures in the wall rocks. These bodies include, from east to west, the Kekekabic stock, the Snowbank stock, and several small bodies near Lost Lake, west of Tower, that appear to be apophyses from a shallow buried pluton. The bodies represent a family of syenitic rocks that was emplaced under a relatively shallow cover late in the Algoman igneous event, for the rocks contain strongly zoned plagioclase and augite and locally have miarolitic cavities.

The youngest plutonic rocks to be emplaced were the Linden pluton, west of Cook in St. Louis County, and the Icarus pluton, east of Saganaga Lake, in Ontario. These bodies are post-tectonic and have strong alkaline affinities.

The supracrustal rocks were deformed and metamorphosed approximately synchronously with emplacement of the batholithic rocks, during the Algoman orogeny. Details of the structural history of the belt as a whole are poorly known, but two generations of folding have been recognized in the western part of the district, and probably have a much broader areal extent. The older generation is well developed in the area between Embarrass and Lake Vermilion, and is represented by tight to close, northwest-trending folds. The folds have planar, steeply-inclined axial surfaces and, probably, gently-plunging axes, either to the northwest or southeast. Except locally, the younger generation of folding largely obscured structures of the older generation. In the vicinity of Tower, folds of this generation trend eastward, are mainly Z-shaped, and have planar, upright axial surfaces and steep plunges. The folds have an associated steep cleavage. High-angle faults of two trends, longitudinal and transverse, break the metavolcanic-metasedimentary sequence and the fringing batholithic rocks into several blocks or segments. Many of the longitudinal faults are significant crustal fractures, for they are long and continuous, have intense associated cataclasis—including mylonite—and apparently have large horizontal displacements. Some of them separate the supracrustal rocks from the batholithic rocks and associated infracrustal gneisses and schists; probably these faults formed virtually contemporaneously with emplacement of the batholithic rocks. The transverse faults trend mainly north-northeastward and have dominant left-lateral displacements of as much as 4 miles; probably, they formed after the initial movements on the longitudinal faults. Associated with these faults are systematic joints and kink bands. The causes and mechanisms of the deformation are not known, but it is probable that the deformation resulted from regional stresses and not from the diapiric emplacement of granitic plutons, as has been inferred for some Lower Precambrian terranes (Anhaeusser and others, 1969).

The supracrustal rocks within the district dominantly contain mineral assemblages characteristic of the greenschist and low-amphibolite metamorphic facies. Adjacent to the batholiths the rocks are metamorphosed to the middle-amphibolite facies. Primary textures and most structures remain in all but the most highly metamorphosed and deformed volcanic and sedimentary rocks.

The generally low grade of metamorphism, steep metamorphic gradients adjacent to the batholithic rocks, and internal textures and structures within the intrusive rocks that are indicative of emplacement at relatively shallow depths indicate that this part of the Canadian Shield has not been eroded to great depths. Clearly, as noted earlier by Anhaeusser and others (1969) for the Barberton Mountain Land of southern Africa, the Lower Precambrian rocks do not represent the "roots of mountains."

Judged from trace element data, both the volcanic and plutonic rocks were derived from mantle depths. Possible analogs with the Lower Precambrian rocks are recently evolved island arcs and circum-Pacific borderlands, as was suggested by Wilson and others (1965), but the actual environment in which these ancient rocks formed and the mechanisms by which the magmas were generated remain speculative.

Southwestern Minnesota

The ancient rocks exposed in the Minnesota River Valley comprise a grossly interlayered sequence of migmatitic granitic gneisses, amphibolitic gneisses, and intrusive granitic rocks. The gneisses have mineral assemblages characteristic of the upper amphibolite and granulite metamorphic facies (Himmelberg and Phinney, 1967). The early geologic history of these gneisses has been largely obliterated by metamorphism and emplacement of granitic rocks about 2,650 m.y. ago (Goldich, chap. II, this volume).

Interlayered gneisses crop out sporadically in the valley but are best exposed in the Granite Falls-Montevideo (Himmelberg, 1968) and Sacred Heart-Morton areas (Grant, this volume). In the Granite Falls-Montevideo area, they consist mainly of granitic gneiss (the Montevideo gneiss of Goldich and others, 1970), hornblende-pyroxene gneiss, and garnet-biotite gneiss, which constitute units that are 1,000 to 5,000 feet thick. Some of the hornblende-pyroxene gneiss contains garnet. The granitic gneiss is a pink, medium-grained, equigranular leucocratic rock having a compositional layering given by alternating, thin biotite-rich and quartzofeldspathic layers. Hornblende-pyroxene gneiss layers and lenses are common in the granitic gneiss, and pink leucocratic quartz monzonite locally intrudes it. Modes (see Grant, table III-44, this volume) and chemical analyses (Goldich and others, 1970, table 3) indicate that the granitic gneiss ranges in composition from granite to granodiorite.

In the Sacred Heart-Morton area, mappable units within the gneisses are less distinct than in the Granite Falls-Montevideo area, but four units constituting a succession a few thousand feet thick have been delineated (Grant, this volume). The lower three units are quartzofeldspathic gneisses that contain varying amounts of amphibolite, and the upper unit is a biotite-rich pelitic gneiss. Two varieties of the pelitic gneiss have been distinguished: (1) a gray, thinly layered rock containing the mineral association biotite-cordierite-garnet-anthophyllite in addition to quartz and plagioclase; and (2) a gray, layered gneiss containing quartz, plagioclase, and biotite with clots of sillimanite and K-feldspar and local garnet or cordierite. Distinguished from the gneisses is a generally coarse-grained quartz monzonite, which forms both concordant and discordant bodies of variable sizes in the gneisses. The quartz monzonite probably is related to the Sacred Heart pluton, which appears to have batholithic dimensions (Austin and others, 1970). The Morton Gneiss, named from exposures in quarries near Morton, is a variant of one of the quartzofeldspathic units that contain amphibolite (Grant, this volume). It is a migmatitic rock consisting of amphibolite, a gray phase having the approximate composition of tonalite, and a pink granitic or pegmatitic phase.

Throughout most of the valley, the rocks are folded on shallow-plunging, east-northeastward-trending axes; younger, small, asymmetric cross-folds related to steep shear planes occur commonly in both the Granite Falls and the Sacred Heart-Morton areas. Grant (this volume) interprets the east-northeastward-trending folds to have formed virtually contemporaneously throughout the valley, approximately synchronously with the main episode of metamorphism.

The geologic mapping shows that the gneisses constitute a grossly layered sequence, the origin of which is equivocal. In such a high-grade, plutonic environment, where all original structures and textures have been destroyed, introduction of magma, partial melting, and metasomatism can occur concurrently. Grant (this volume) suggests that the biotite-rich pelitic gneisses were graywacke and that the amphibolite and hornblende-pyroxene gneisses were either original mafic volcanic rocks or mafic sills. He considers the widespread quartzofeldspathic gneisses to represent either an original suite of igneous rocks ranging in composition from tonalite to granite or a more homogeneous igneous rock, originally of approximately quartz monzonite composition, that was fractionated during partial melting into more and less potassic members. Granitic material was added to these rocks to different extents during emplacement of the Sacred Heart pluton. Alternatively, the quartzofeldspathic gneiss could represent thoroughly reconstituted effusive rocks of dacitic-rhyodacitic composition. These interpretations are consistent with the gneisses being older, higher grade metamorphic analogs of the rock assemblages characteristic of greenstone-granite complexes. Goldich and others (1970, p. 3693) preferred a plutonic origin for at least a substantial part of the quartzofeldspathic gneisses. They stated that the chemical composition and texture and structure of the Morton Gneiss are best explained in terms of magmatic intrusions. In the same way, they interpreted the Montevideo gneiss as probably being a synkinematic intrusion.

Extensive geochronologic studies of the Precambrian rocks in the Minnesota River Valley in conjunction with geologic mapping have shown that at least substantial parts of the sequence of interlayered gneisses are older than 3,000 m.y. and possibly are as much as 3,600 m.y. old. This age is derived from U-Pb data on zircons and Rb-Sr data from whole-rock analyses of the Morton Gneiss and the Montevideo gneiss (Goldich and others, 1970; Goldich, chap. II, this volume), which are representative of the rocks comprising a large part of the exposed layered sequence. The metamorphism to upper amphibolite and granulite facies occurred at least 2,650 m.y. ago, when the Sacred Heart granite and related quartz monzonitic intrusions were emplaced. Both Himmelberg (1968) and Grant (this volume) believe that the rocks comprising the complexly layered sequence were deformed, metamorphosed, and intruded by the quartz monzonitic rocks approximately contemporaneously, although a culmination of metamorphism may have preceded igneous emplacement. Goldich and others (1970, p. 3693), on the other hand, interpreted the major structure to have developed, in part at least, about 3,550 m.y. ago, approximately contemporaneously with emplacement of the Morton Gneiss and the Montevideo gneiss, during the Mortonian event.

Middle Precambrian

Sedimentary rocks of Middle Precambrian age are inferred to cover a large part of northern and east-central Minnesota (pl. 1), but exposures are limited to a relatively few areas, including the Gunflint range in northern Cook County and the adjoining Thunder Bay district, Ontario, the Mesabi range, the Cuyuna district, and east-central Minnesota. In east-central Minnesota, these rocks were intruded by several types of Middle Precambrian igneous rocks, which were emplaced approximately contemporaneously with the Penokean orogeny. Diabasic dikes of Middle Precambrian age are exposed locally in the Lower Precambrian terrane in northern Minnesota, and similar dikes and small granite and gabbro-granite plutons occur in the dominantly Lower Precambrian terrane in the Minnesota River Valley.

Sedimentary Rocks

The Middle Precambrian rocks consist of at least two sequences of sedimentary strata, separated by a period of epeirogeny and erosion. The older sequence is not exposed, but is known from drilling in east-central Minnesota to consist of at least 100 feet of cherty dolomitic limestone (the Trout Lake formation of Marsden, this volume), which conformably overlies quartzite and slate that in turn are inferred to lie unconformably on Lower Precambrian rocks. Neither the geographic extent of these rocks nor the time represented by the disconformity is known, but the rocks appear to have a structural and stratigraphic position analogous to that of the Bad River Dolomite in Wisconsin. These rocks are overlain unconformably by the thick sequence of clastic strata and iron-formation that is assigned to the Animikie Group. The rocks of the Animikie Group comprise a northeastward-trending sediment wedge that thickens from less than 100 feet in the northern part of the state to at least 15,000 feet in areas to the south. They were deposited in an elongate basin, the configuration of which appears to have been controlled by a pre-existing northeastward-trending grain in the older rocks. Probably the Minnesota portion of the basin was not much larger than the present outcrop distribution.

The oldest rocks of the Animikie Group are shallow-water deposits consisting of several kinds of quartzite, argillite, argillaceous siltstone, and conglomerate. In general, the succession thickens southward. The Pokegama Quartzite, on the Mesabi range in northern Minnesota, probably was deposited by southward-flowing currents, and apparently was derived from an older Precambrian terrane to the north. The Mahnomen Formation, in east-central Minnesota, also appears to have been derived from a plutonic terrane, but its source is equivocal. Peterman (1966, p. 1039) concluded from a rather high initial Sr^{87}/Sr^{86} ratio that the sediments in the Mahnomen Formation were derived from a mixed-age terrane of rocks 2.5 to 3.5 b.y. old, such as is exposed in the Minnesota River Valley to the south. Thus, the shallow-water succession apparently represents an accumulation of sediments in a basin that was bordered by higher areas of older granitic rocks both to the north and south.

The next younger succession, consisting of the Gunflint Iron-formation of the Gunflint range, the Biwabik Iron-formation on the Mesabi range, and the Trommald Formation in the Cuyuna district—which are approximately equivalent—consists of chemical sediments that were deposited in water of varying depths. The Biwabik Iron-formation is as much as 750 feet thick in the central part of the Mesabi range (White, 1954), but thins to both the east and west. The Gunflint Iron-formation is less than 300 feet thick in Minnesota, but thickens to about 400 feet near Thunder Bay and then thins eastward. The Trommald Formation of the Cuyuna district ranges in thickness from 45 feet in the northern and western parts to more than 500 feet in the main part of the North range (Schmidt, 1963). The Biwabik Iron-formation and to a lesser extent the Trommald Formation are major sources of iron ores.

On the basis of textures, the iron-formations can be classified as apparently coarse-grained or "granular" types and fine-grained or "slaty" types. The textures of the cherty, or granular, rocks resemble those in some sandstones or clastic limestones, and possibly resulted from deposition of particulate detritus (Mengel, 1965); the slaty rocks have grain-size and bedding aspects similar to siltstone and argillite that were deposited under quiet conditions (LaBerge, 1967a). Accordingly, the thick-bedded cherty sediments are interpreted as having been deposited in a shallow-water, agitated environment, whereas the thin-bedded slaty sediments probably were deposited in a less energetic environment, which at places may have been some distance away from a strand line.

Strata having "cherty" or "slaty" textures are interlayered on all scales in each iron-formation. However, lithostratigraphic units composed of several beds having either dominantly cherty or dominantly slaty textures extend over wide areas, and it has been possible to subdivide each iron-formation into several facies that can be combined into members. In general, the Biwabik and Gunflint Iron-formations reflect two cycles of shallow-water deposition separated by intervals of deeper water deposition; the vertical succession of lithofacies clearly indicates that deposition started in shallow water and culminated in deep water. The depositional pattern of the Trommald Formation also is consistent with a general change from shallow-water to deep-water deposition.

The Rove Formation on the Gunflint range, Virginia Formation on the Mesabi range, Rabbit Lake Formation in the Cuyuna district, and Thomson Formation in east-central Minnesota, which are stratigraphically younger than the major iron-formations, comprise a succession of intercalated dark-gray mudstone, siltstone, and graywacke, and lesser amounts of quartzite, limestone, and iron-formation. The thickness of the succession is unknown but probably does not exceed a few thousand feet. Probably, the accumulation of these deep-water sediments resulted from tectonism at the close of deposition of iron-formation, together with some volcanism, that caused a rapid change in the geometry of the basin. The lower several hundred feet of the Rove, Virginia, and Rabbit Lake Formations are carbonaceous, suggesting deposition under reducing conditions. Later, much silt- and sand-size material was deposited, probably mainly by turbidity currents. Sedimentary structures, including cross-bedding and various kinds of sole marks, indicate that most of the currents that deposited the clastic sediments flowed southward, probably down a paleoslope perpendicular to a shoreline (Morey and Ojakangas, 1970). Sparse ripple marks on the tops of quartzite beds in the upper part of the Rove Formation indicate a westerly transport, and probably represent long-shore currents that reworked some of the previously deposited turbidite material. The major source of the graywacke detritus appears to have been a dominantly granitic terrane to the north.

Two types of iron-formation, one representing carbonate facies and the other representing sulfide facies, are intercalated with the graywacke and slate in the younger succession. Rocks of the carbonate facies occur mainly in the Virginia and Rabbit Lake Formations, and consist of argillite and intercalated beds of admixed siderite and chert; finely divided pyrite is a minor component. The sulfide-facies iron-formation occurs principally in the Rabbit Lake and Thomson Formations as massive or thinly laminated beds of graphitic slate containing abundant pyrite; the pyritiferous graphitic slate is intercalated with chert-rich beds consisting of graphitic material, pyrite, and an iron-rich carbonate.

Mafic dikes of Middle Precambrian age are widespread areally, and are potentially useful for dating the Middle Precambrian sedimentary strata. Altered diabase dikes within a northwestward-trending belt that traverses Lower Precambrian rocks on the north side of the range, and which apparently are older than the Middle Precambrian strata, provide a lower limit of about 2,200 m.y. for the time of Animikie deposition (Hanson and Malhotra, 1971). Other diabasic dikes in this belt, near Nashwauk, apparently cut the Biwabik Iron-formation, and have a minimum age of 1,395 m.y. Basaltic dikes that cut many of the plutonic rocks in east-central Minnesota have radiometric ages of about 1,570 m.y.

Penokean Orogeny

Structure and Metamorphism. The Middle Precambrian rocks were deformed during the Penokean orogeny. Deformation was most intense in the southern part of the depositional basin, within a westward-trending belt extending across east-central and central Minnesota, and was slight in northern and northeastern Minnesota.

In east-central Minnesota, the dominant structures are large asymmetric anticlines and synclines, the axial planes of which are steeply inclined and strike northeast to east. Within this region, the intensity of deformation decreases northward: the southern part of the Cuyuna district is characterized by isoclinal folds, the central part by doubly-plunging tight folds, and the northern part by open, eastward-plunging folds. Most of the folds plunge at angles less than 40°. Judged from the outcrop pattern of the Trommald Formation in the Cuyuna district, there was a second period of folding about northward-trending axes. A similar structural pattern may exist in the Emily district, for the outcrop pattern of the iron-formation suggests that the dominant northeastward-trending folds were deformed later about northward-trending fold axes.

In contrast, the Animikie rocks on the Mesabi and Gunflint ranges comprise a southeastward-dipping homoclinal succession. This structure generally has been interpreted as reflecting the north limb of the Lake Superior syncline, which formed in Late Precambrian time, but it probably reflects a more complicated structural history. For example, the northeastward-trending open folds in the Virginia horn area of the Mesabi range are subparallel to folds in the Cuyuna district, and probably are of the same (Penokean) age. Accordingly, a part at least of the apparent homoclinal dip may be a consequence of Middle Precambrian folding. Throughout most of the Mesabi range, however, joints and faults are the most conspicuous structures. The contrasting styles of deformation in these rocks can be related to differences in the lithology of the basement rocks. In general, the Animikie rocks that are underlain by Lower Precambrian metasedimentary and metavolcanic rocks yielded to stresses by folding, whereas those that are underlain by granitic rocks of the Giants Range batholith yielded by fracturing. Structures in the basement rocks also had a profound effect on the Middle Precambrian strata; for some faults in the basement rocks, which formed initially in Early Precambrian time, are contiguous with faults that transect the iron-formation. Recurrent movements along some of these faults apparently occurred during deposition of the iron-formation, affecting sedimentation patterns, whereas movements on others clearly took place after consolidation of the Animikie strata.

The Animikie strata in east-central Minnesota were metamorphosed about 1,850±25 m.y. ago, during the Penokean orogeny. The configuration of the metamorphic isograds corresponds generally to the spatial distribution of the Penokean igneous rocks. Adjacent to the igneous rocks, the metasedimentary rocks are characterized by a narrow zone of minerals assignable to Winkler's (1967) staurolite-almandine subfacies of the Barrovian-type facies series. Most of the metasedimentary rocks in east-central Minnesota, however, have metamorphic assemblages characteristic of the upper part (biotite zone) of the greenschist facies. The apparently steep metamorphic gradient cannot be quantitatively explained by conducted geothermal heat, but is readily explained as a product of regional contact metamorphism. Thus, deformation and metamorphism appear to be independent variables in the orogeny in this part of Minnesota.

Igneous Rocks. The Middle Precambrian igneous rocks in east-central Minnesota occupy an area of about 3,500 square miles and consist of several discrete intrusions of diverse lithologies that are in sharp contact with one another or are separated by thin septa of metamorphic rocks. The oldest igneous rocks are dikes, sills, and irregular bodies of pre-tectonic gabbro and diorite. These rocks have been deformed in accord with the regional structure, and are retrograded to amphibole- and chlorite-bearing assemblages. This period of intrusion was followed by the emplacement of several syntectonic plutons of intermediate composition, some of which are large, and include bodies of tonalite, monzonite, and granodiorite. These rocks have a strong to faint northeastward-trending foliation that is subparallel to the regional structure. Nearly all contain subconcordant inclusions of biotite or hornblende schist that probably are remnants of the Thomson Formation. The youngest igneous rocks, named the "Stearns magma series" by Woyski (1949), are large plutons of granite and associated rocks. Most of these rocks are massive, but they are strongly sheared and epidotized locally. Swarms of inclusions are common locally, and include large angular blocks of granodiorite, tonalite, biotite schist, and iron-formation. The wide variety of inclusions and the massive nature of the plutons suggest that these intrusive rocks are post-tectonic.

Post-Penokean Events. Many of the radiometric ages on rocks within and adjacent to the Mesabi and Gunflint ranges cluster at about 1,650±25 m.y., and are younger than the Penokean orogeny (ca. 1,850 m.y.). The interpretation of these ages is controversial, but they may be related to a period of mild deformation and metamorphism that occurred about 1,600-1,700 m.y. ago. That mild deformation was involved is indicated by the local presence of cataclasis in the pre-Animikie diabases that cut Lower Precambrian rocks north of the Mesabi range. Neither the extent of the cataclasis nor its exact age is known, but it seems to be intense near the Mesabi range and to decrease in intensity northward. Also, many textural features in the Biwabik Iron-formation suggest that it was subjected to a low-grade metamorphism that did not affect similar rocks on the Gunflint range. Possibly the northward-trending cross-folds in the Cuyuna district and the northward-trending folds in the Biwabik Iron-formation on the westernmost Mesabi range (pl. 1) are related to this same deformation (ca. 1,650±25 m.y.).

Late Precambrian

Two generally distinct sequences of rocks of Late Precambrian age occur in Minnesota: (1) an older clastic sequence, mainly represented by the Sioux Quartzite of southwestern Minnesota and adjacent areas; and (2) a younger volcanic-sedimentary sequence that is confined to a relatively narrow zone that extends from Lake Superior through eastern Minnesota, and in the subsurface, southwestward into Kansas. The Sioux Quartzite is a moderately thin deposit of wide areal extent that was formed in a shallow-water, near-shore tectonically stable environment. It has been deformed into broad, open, generally northward-trending folds (Austin, this volume). It is lithologically similar to the presumably equivalent Baraboo and Waterloo Quartzites of Wisconsin. These rocks probably are 1,400-1,500 m.y. old. The younger volcanic-sedimentary sequence appears to define a distinct petrologic and structural province, Keweenawan in age, which is characterized throughout much of its length by a central core of mafic volcanic rocks that is overlain and flanked by thin to thick wedges of clastic strata. The province is expressed geophysically by a gravity and magnetic high that is flanked by gravity and magnetic lows (pl. 2). Commonly, this geophysical anomaly is referred to as the Midcontinent Gravity High; the structure of the province is interpreted as an abortive continental rift.

Classification and Nomenclature

The traditional three-fold subdivision of the Keweenawan, based largely on rock-stratigraphic units, currently is being re-evaluated by the Minnesota Geological Survey. Pending adoption of a new time-stratigraphic classification, it seems appropriate to restrict the term "Keweenawan" to those events related in space and time to the development of the rocks and structures associated with the Midcontinent Gravity High. The currently available radiometric age data suggest that most of the Keweenawan igneous activity took place in the interval 1,200-900 m.y. ago (Goldich, this volume; Robertson and Fahrig, 1971). The significant Duluth Complex and North Shore Volcanic Group have been dated as 1,100 m.y. old (Silver and Green, 1963). Judged from radiometric analysis of one sample (Hanson and Malhotra, 1971), the Logan intrusions of northwestern Cook County may be distinctly older—about 1,300 m.y. Also, the Sibley Series, which has been assigned to the Early Keweenawan, is crosscut by dikes that probably are equivalent to the Logan intrusions, and it may be as much as 1,376 ± 33 m.y. old (Franklin and Kustra, 1970). The relationship of the Sibley Series to Keweenawan tectonism is unknown, but if the available radiometric age data are valid, it is younger than the Sioux Quartzite and therefore perhaps more closely related to the Keweenawan than to the Sioux. Weiblen and others (this volume) conclude that the Logan intrusions were emplaced during a period of arching that immediately preceded rifting. If this interpretation is correct, Keweenawan tectonism in the Lake Superior region started at least 1,300 m.y. ago and perhaps as early as 1,400 m.y. ago.

The definition of units within the Keweenawan is as difficult as definition of the Keweenawan itself as a time-stratigraphic unit. Paleomagnetic studies have shown that the Keweenawan volcanic rocks are characterized by a change in paleomagnetic polarity, and Books (1968) has suggested that the Lower-Middle Keweenawan boundary in Michigan be placed where rocks of reversed polarity are succeeded by those of normal polarity. According to this definition, the Lower Keweenawan contains a substantial thickness of igneous rocks that differ from Middle Keweenawan rocks in lithology and metamorphic grade as well as in magnetic properties (White and others, 1971). Green (this volume) has extended this definition to the Keweenawan rocks along the north shore of Lake Superior. Thus, the "Lower Keweenawan" in Minnesota, as used by Green, comprises the Puckwunge Formation, Logan intrusions, lower part of the North Shore Volcanic Group, and a part of the northern prong of the Duluth Complex, in Cook County. The remainder of the North Shore Volcanic Group and of the Duluth Complex as well as the lesser Beaver Bay Complex and the Pigeon River intrusions of Geul (1970) are assigned to the Middle Keweenawan. According to this usage, there is some evidence for a local structural discontinuity between Lower and Middle Keweenawan rocks in northeastern Minnesota. Green (this volume) suggests the existence of a disconformity at this position in the Pigeon River-Grand Portage area, and mapping in northwestern Cook County (Weiblen and others, this volume) indicates that the reversely polarized Logan intrusions and the layered series of Nathan in the Gunflint prong (Phinney, this volume) were deformed prior to emplacement of the anorthositic and troctolitic rocks that comprise the majority of the Duluth Complex. However, additional studies are needed before it will be known whether this structural discontinuity coincides everywhere with the paleomagnetic reversal.

Until recently, the Middle-Late Keweenawan boundary in Michigan has been placed at the top of the Lake Shore trap, and it has served to separate a dominantly volcanic succession from a younger, dominantly clastic succession. Because that contact is not definable everywhere, White (1972) has suggested that the boundary be placed at the base of the Nonesuch Formation. With this redefinition, a substantial thickness of sedimentary strata is now included in the Middle Keweenawan in Michigan. In Minnesota, lava flows that are overlain stratigraphically by clastic strata occur only in the subsurface in the southeastern part of the state. Because of inadequate knowledge of these subsurface rocks, the Minnesota Geological Survey continues to place the boundary between the Middle and Late Keweenawan at the contact between dominantly volcanic rocks and dominantly clastic strata.

Keweenawan Rocks in Northeastern Minnesota

In northeastern Minnesota, Keweenawan rocks comprise an arcuate mass of cogenetic intrusive and extrusive rocks of mafic composition that extends from Duluth northeastward to the vicinity of the Pigeon River, in extreme northeastern Cook County (pl. 1). Clastic strata occur locally, mainly as thin quartzite sandstones at the bases of some flows and thin lithic sandstones and shales intercalated with the flows. The lavas, which are assigned to the North Shore Volcanic Group, are intruded by several types of intrusive rocks, including the large Duluth Complex, the Beaver Bay Complex, several diabase sills at Duluth, the Hovland and Reservation River diabase complexes, and the Logan intrusions. In contrast to other areas in the Lake Superior region that are underlain by Upper Precambrian rocks, intrusive rocks far exceed volcanic rocks in areal extent.

North Shore Volcanic Group. The lavas that comprise the North Shore Volcanic Group are dominantly basaltic in composition but include substantial volumes of intermediate and felsic rocks (Green, this volume). They have the form of half a large, filled dish about 150 miles wide that is tilted gently southeastward toward Lake Superior. Judged from the outcrop pattern and geophysical data (Ikola, 1970), the lavas occupy two separate (?) basins within the larger Lake Superior basin, which are fringed by the Duluth Complex on the west and separated by rocks of the Beaver Bay Complex. In the southwestern basin, the flows are estimated to be about 23,000 feet thick; dips decrease from the base of the section at Duluth northeastward toward higher stratigraphic levels, suggesting that the axial region of the Lake Superior basin was sinking concurrently with lava extrusion. The northeastern basin contains about 21,500 feet of lavas; these lavas show no significant change in dip from the oldest flows at Grand Portage to the youngest flows at Tofte and Lutsen, implying that the tilting in that area took place after extrusion of the lavas.

The oldest flows—at Nopeming, west of Duluth, and in the Pigeon River-Grand Portage area in extreme northeastern Minnesota—overlie quartzite and have a reversed polarity. In the Grand Portage area, a structural discontinuity occupied by the Hovland diabase complex separates the lowermost flows from overlying (Middle Precambrian) flows having normal polarity. The lowermost flows at both localities appear to have been extruded in a subaqueous environment. The younger lavas are entirely subaerial; they have vesicular (now amygdaloidal) upper portions, massive interiors, and various types of jointing, surface features, and textures, depending on their specific compositions. Most flows are tabular, and individual flows or flow units can be traced along strike for distances of as much as 20 miles. Thicknesses vary with the compositions of the lavas; a few intermediate and felsic flows are more than 300 feet thick.

The North Shore Volcanic Group ranges in composition from olivine metabasalt to rhyolite. The most abundant rock types are olivine basalt and quartz tholeiite. The quartz tholeiite grades into more potassium-rich varieties. The intermediate varieties are nearly all porphyritic, and contain phenocrysts of plagioclase, augite, magnetite and, rarely, iron-rich olivine. They have the compositions of andesites, trachyandesites, and intermediate quartz latites. Felsic lavas (quartz latite) are common. Most are porphyritic and have quartz and feldspar phenocrysts. Although the lavas can be divided into several lithostratigraphic units having coherent petrographic character, there is no distinct compositional trend from the base to the top of the group.

The North Shore Volcanic Group resembles, both physically and chemically, plateau lava sequences of various geologic ages, and particularly the Tertiary plateau lavas of Iceland (Green, this volume).

Duluth Complex. The Duluth Complex, formerly called the Duluth Gabbro Complex (Taylor, 1964), is the largest and most complicated of the Keweenawan intrusive bodies in northeastern Minnesota. It is about 150 miles long and occupies an area of about 2,500 square miles. It is a composite body consisting of several anorthositic, gabbroic, and granitic intrusions that was emplaced along a major unconformity between Lower and Middle Precambrian metamorphic and igneous rocks and Keweenawan lava flows; locally, as at Duluth, Keweenawan lava flows directly underlie as well as overlie the complex. The base of the complex dips irregularly southeastward, toward Lake Superior, at angles ranging from less than 10° to 60°. Drilling associated with the exploration and development of copper and nickel sulfide deposits near the base of the complex has provided much data on the lithology and structure of the body.

The oldest rocks of the Duluth Complex are in the northern prong, in northern Cook County. They comprise a sequence of southward-dipping, relatively thin, sheet-like intrusions of general gabbroic composition. Several of the units are unique occurrences of oxide-rich or two-pyroxene gabbros. Some units show evidence of crystal settling. The rocks have a reverse magnetic polarity, and are transected to the west by rock types that are common to the remainder of the Duluth Complex and which have a normal magnetic polarity.

Anorthositic rocks containing from 80 to 99 percent calcic labradorite are the next younger rocks. They comprise more than 50 percent of the complex at the present level of erosion, and mainly occupy the central part of the body. A coarse gabbroic pegmatite occurs widely throughout the rocks. In the anorthositic rocks, plagioclase is the only cumulus phase; interstitial olivine, clinopyroxene, orthopyroxene, and opaque oxide predominate in different varieties. The rocks are interpreted (Phinney, this volume) as having been formed from multiple intrusions that were emplaced as crystal mushes; the complex foliation patterns given by aligned plagioclase crystals probably were controlled by currents in the melt, which had virtually the same density as the plagioclase crystals.

Intrusive into the anorthositic rocks are troctolitic rocks of several varieties, which locally contain layers and lenses of picrite, dunite, norite, olivine gabbro, and ferrogabbro. These rocks occur along the base of the complex, in a zone 2 to 4 miles wide, and further outward as basins, large dikes, elongate cones, and sheets. A steeply-dipping elongate cone—the Bald Eagle intrusion—in the northwestern part of the complex possibly is a major feeder zone for most of the troctolitic rocks in that area. Not uncommonly, anorthosite bodies of various dimensions occur as either "islands" or inclusions in the troctolitic rocks.

Inclusions of several types are common in the troctolite that comprises the basal zone of the Duluth Complex and occur locally at higher stratigraphic levels. Most of the inclusions in the basal zone are graywacke-slate and iron-formation, derived from the Middle Precambrian successions, and lava flows derived from the older Keweenawan volcanic successions; locally, they consist of granitic rocks derived from subjacent Lower Precambrian rocks. Probably many of the larger inclusions represent slabs of country rock that were pried off and broken up by stresses accompanying emplacement of the melts. Large units of granofels—probably metamorphosed lava flows, sills, or older gabbroic rocks—are common at or near the top of the complex, and represent either inclusions that foundered in the melt or remnants of the overlying metamorphosed country rock.

Felsic and intermediate rocks, including granophyric granite, occur discontinuously along the top of the complex —as small plutons and as relatively flat-lying sheets—from Duluth to the eastern margin. These rocks probably are late-stage differentiates of the various mafic intrusions, although some may result in part from partial melts of inclusions.

The close spatial relations of rocks of the Duluth Complex, the North Shore Volcanic Group, the Logan intrusions, and other mafic dike rocks suggest that they record a single, but complex, large-scale process of differentiation and intrusion. Much additional study is needed to determine definitely the relationships between specific rock units, but sufficient data are available to suggest the following sequence of events: (1) derivation of a parent magma at depths of at least 15-35 km (Green, this volume); (2) differentiation involving crystallization of olivine, plagioclase, and some clinopyroxene in the magma chamber at a relatively shallow depth (Phinney, this volume); and (3) periodic separation of melts of various compositions. Phinney (this volume) has shown that the sequence of lava flows of the North Shore Volcanic Group can be derived as differentiates of a magma of the composition of one of the early basalt flows. Similarly, the Logan intrusions have a bulk composition indicative of differentiates from the same parent magma (Weiblen and others, this volume). In this model, the several units of the Duluth Complex represent cumulus fractions, whereas the Logan intrusions and North Shore Volcanic Group represent fractions of the remaining melt.

Rifting provides a mechanism for developing magma chambers at intermediate and shallow depths. Later, as rifting proceeds, fracturing provides a mechanism whereby younger rocks derived from deep sources can be emplaced without significant differentiation.

Keweenawan Rocks in East-central Minnesota

Keweenawan lava flows and younger clastic sedimentary rocks crop out locally in east-central Minnesota and extend in the subsurface through southeastern Minnesota. The lava flows are physically separated from those on the north shore of Lake Superior, and, lacking evidence for correlation of the two lava successions, those in east-central Minnesota are informally referred to as the Chengwatana volcanic group.

The lava flows comprise part of a fault-bounded block, named the St. Croix horst (Craddock and others, 1963), which is flanked by and locally overlain by the younger clastic strata. The faults that bound the horst on the west and east are nearly vertical and have vertical displacements of at least 6,000 feet. Geophysical data suggest that the lavas are more than 20,000 feet thick. Approximately 6,000 feet of dominantly basaltic lavas are exposed along several rivers in Pine County. The exposed section consists of many flow units that range in thickness from less than 10 to more than 1,000 feet. In contrast to northeastern Minnesota, interflow sedimentary rocks are moderately abundant and intrusive bodies are rare.

In the Minneapolis-St. Paul area, the flows on top of the St. Croix horst define a broad basin which is filled with Keweenawan clastic strata. Because nearly all the clastic strata are red, they were assigned previously to an informal unit named the "Red clastic series." However, the Red clastic series can now be divided into three formations: (1) Hinckley Sandstone, a buff sandstone containing 95 percent or more quartz; (2) Fond du Lac Formation, consisting of intercalated moderate red shale and sandstone containing approximately 60 percent quartz, 30 percent orthoclase, microcline, and sodic plagioclase, and variable amounts of "granitic" and aphanitic rock fragments; and (3) Solor Church Formation, consisting of variable amounts of quartz, plagioclase of intermediate composition, and aphanitic mafic igneous rock fragments. The former two formations are contiguous with units exposed at the surface in east-central Minnesota, but so far as known the Solor Church Formation is confined to the subsurface.

In the flanking basin on the west side of the horst, the Fond du Lac Formation disconformably overlies Solor Church Formation and is overlain gradationally by Hinckley Sandstone. On top of the horst, the Solor Church Formation occupies an elliptical basin and is unconformably overlain by patches of generally thin Hinckley Sandstone; at places a thick regolith separates the two formations. Morey (this volume) suggests the following sequence of events in the evolution of these rocks: (1) the Solor Church Formation was derived from a basaltic terrane and deposited prior to formation of the St. Croix horst. (2) Subsequently, the Fond du Lac Formation was deposited in a subsiding basin that flanked the horst. Detritus was derived from two source areas, a distal granitic terrane and a proximal terrane consisting of basalt and older red beds, implying that the horst was a positive area during the time of Fond du Lac deposition. Sedimentation culminated with the Hinckley Sandstone, which is interpreted as a reworked phase of the Fond du Lac Formation. The thin and irregular nature of the Hinckley Sandstone on top of the horst as compared to the flanks indicates that movement and concurrent erosion occurred on the horst after the end of Hinckley deposition. Many structures in the overlying Paleozoic rocks occur over Keweenawan structures, and these can be accounted for by small, recurrent vertical movements on the major Keweenawan faults during Paleozoic time.

PALEOZOIC ERA

Sedimentary rocks of early Paleozoic age crop out in the southeastern part and occur in the subsurface in the extreme northwestern part of Minnesota. In southeastern Minnesota, strata ranging in age from Late Cambrian to Devonian occupy a shallow depositional lowland, called the Hollandale embayment, which is the northward extension of the Ancestral Forest City basin (Iowa basin) onto the cratonic shelf. The marine rocks that now remain within the embayment are bordered to the east by nearshore-facies Paleozoic rocks of the Wisconsin Arch, to the northeast by Precambrian rocks that constitute the Wisconsin Dome, and to the north and west by nearshore-facies paleozoic rocks that lie between the Hollandale embayment and the Transcontinental Arch. In northwestern Minnesota, the Paleozoic rocks comprise the easternmost margin of the Williston basin, and apparently are Middle and Late Ordovician in age. The margins of the Williston basin and the Hollandale embayment are about 300 miles apart, but inasmuch as the erosionally truncated edges of the sedimentary rocks in both basins are about 400 feet thick, the maximum transgressive shorelines of each basin were much closer, if not actually overlapping, in early Paleozoic time.

Southeastern Minnesota

The Hollandale embayment, which contains a maximum stratigraphic thickness of about 2,000 feet in southeastern Minnesota, overlies the Upper Precambrian rocks that produce the Midcontinent Gravity High, and there is a close spatial correlation of structures in the Paleozoic rocks with older ones in the Precambrian basement. The deepest part of the embayment approximately overlies the axis of the Midcontinent Gravity High, and the outline of the embayment coincides approximately with the distribution of the subjacent Keweenawan rocks. On a smaller scale, many structural features within the embayment coincide with structures in the underlying basement rocks. The Twin City basin in the Minneapolis-St. Paul area, for example, is underlain by a fault-controlled depositional basin of Keweenawan age. The eastern and southeastern margins of the Twin City basin are defined by the Hudson-Afton and Vermillion anticlines, respectively, and by several northeastward-trending, high-angle faults; each of these structures overlies the traces of major faults that were formed in Keweenawan time. The southern edge of the Twin City basin is defined by a structurally high belt of Upper Cambrian strata that extends from the general vicinity of Hastings westward to Belle Plaine, and the southwestern edge, near Belle Plaine, is defined by two northwestward-trending structures, the Waterville-Waseca anticline and the Belle

Plaine fault. Both these structures also have analogs in the Precambrian basement. On the western side of the basin, the Paleozoic strata overlap pre-Keweenawan rocks.

The Cambro-Ordovician rocks in the upper Mississippi valley consist of a series of alternating sandstones and carbonates with subordinate siltstones and shales. The alternation occurs in a broad and regular cyclical pattern that spans the entire succession. In the region as a whole, Ostrom (1964) has distinguished five regular sets of these alternating strata, each of which generally consists of four lithotopes indicating four depositional environments and is considered to represent a sedimentary cycle. The four recurrent lithotopes that have been recognized in the succession are (1) well sorted quartz arenite, (2) poorly sorted units of mixed lithologies, (3) shale or argillaceous sandstone, and (4) carbonate rock. The cycles resulted from repeated emergence and submergence, the former caused by rejuvenation of tectonically positive parts of the craton and the latter resulting from submergence of the Appalachian geosynclinal basin far to the southeast.

In Minnesota, nine cycles or recurrent lithotopes, some of which are incomplete, have been defined (Austin, this volume). These recurrent lithotopes resulted from cyclical variation in the amount, mineralogy, and grain size of the clastic influx into the basin, and reflect local depositional environments—produced by proximity to nearby positive areas and vertical movements in the underlying basement rocks—as well as the regional cyclic depositional environment. Because of the effect of the local features on sedimentation, the sedimentary succession within the Hollandale embayment differs in detail from that in adjacent parts of the upper Mississippi valley, both in lithology and in the position of unconformities.

The Paleozoic rocks in southeastern Minnesota show a gradual but detectable shift in sedimentation from predominant sandstone and subordinate carbonate in the lower cycles to predominant carbonate and subordinate sandstone in the upper cycles. The uppermost rocks, the Devonian Cedar Valley Formation, consist almost wholly of carbonate rocks.

The sedimentary rocks of Late Cambrian (St. Croixan) age in Minnesota were deposited during two transgressive-regressive cycles. During the first cycle, the Dresbachian sequence, including the Mt. Simon Sandstone, the Eau Claire Formation, and the Galesville Sandstone, was deposited and during the second, the Franconian-Trempealeauan sequence, including the Ironton Sandstone, the Franconia and St. Lawrence Formations, and the Jordan Sandstone, was deposited.

The Ordovician was initiated by accumulation of a succession of predominantly dolomitic rocks (Canadian Series), upon which the St. Peter Sandstone was deposited. The overlying strata of the Champlainian and Cincinnatian Series are dominantly carbonates. In the Hollandale embayment, sedimentation probably was continuous from the time of deposition of the St. Peter Sandstone, in early Middle Ordovician (Chazyan) time, to deposition of the Maquoketa Formation, in early Late Ordovician time.

After uplift and erosion of the Ordovician and older strata, the sea returned during Middle Devonian time and deposited the Cedar Valley Formation near the center of the Hollandale embayment. The Cedar Valley contains a local (?) marine siderite facies, which in Cretaceous or Tertiary time was oxidized, to provide the economically important limonite iron ores of the Fillmore County District (Bleifuss, this volume). Some time after deposition of the Cedar Valley Formation, renewed uplift of the Transcontinental Arch with subsequent erosion defined the limits of the present expression of the Hollandale embayment.

The Hollandale embayment and the smaller structures within it were not clearly defined until Early Ordovician time. The first indication of the embayment as a feature affecting sedimentation is given by the distribution of facies in the Cambrian Eau Claire Formation, but it was not until Early Ordovician time, when the area was covered by the shallow Prairie du Chien sea, that the subsiding center of the embayment was defined by thickening of the units.

The Twin City basin was outlined by Early Ordovician time. The basin probably formed initially by movements along structures in the basement rocks, for at least locally an unconformity exists between the Cambrian Jordan Sandstone and the Ordovician Oneota Dolomite in proximity to areas underlain by known basement structures. Also, isopach maps suggest that sedimentation in the basin was restricted during the Early Ordovician. Another major structural feature, the Red Wing-Rochester anticline, formed at about the same time, for the New Richmond Member of the Shakopee Formation is less than 10 feet thick west of the anticline whereas it is 65 feet thick east of the anticline, in extreme southeastern Minnesota.

Northwestern Minnesota

Sedimentation in northwestern Minnesota, at the eastern feather edge of the Williston basin, began in Champlainian time and continued into Cincinnatian time. The lower sandstone of the Winnipeg Formation marks the beginning of a transgressive sequence, and the upper sandstone marks a regressive phase of sedimentation. The Winnipeg Formation is overlain by the Red River Formation, a carbonate unit that has been partially eroded.

MESOZOIC ERA

Rocks younger than Devonian and older than Cretaceous are missing in Minnesota except for a small area of probable Jurassic strata in the extreme northwestern corner of the state. In northwestern Minnesota and in adjoining North Dakota, beds of red clay, red shale, and gypsum have been penetrated immediately beneath Pleistocene drift, and appear to comprise the southern segment of a southwestward-trending subcrop belt that extends from Manitoba into Kittson County, Minnesota. In Manitoba, where gypsum is being mined, the red beds are assigned to the Amaranth Formation of Jurassic age. The red beds in northwestern Minnesota are similar in appearance to those in Manitoba and North Dakota, and rocks of comparable age and lithology underlie and overlie them at both localities. The Amaranth Formation and its southern equivalent occupy a broad lowland cut onto a post-Silurian erosion surface that dips gently westward. The pre-Jurassic basement surface is irregular and is characterized by several topographic highs that separate the subcrop area into several small basins or embayments; post-Jurassic erosion locally has removed an unknown amount of Jurassic and older strata, particularly near the high areas. Therefore, the deposits generally are not continuous laterally, and where present are characterized by abrupt facies changes. Probably, the gypsum deposits and associated red beds were deposited under arid conditions in shallow lagoons and embayments that bordered more open seas to the west.

Rocks of Cretaceous age are nearly continuous beneath a thick cover of Pleistocene material throughout the western half and form numerous outliers in the southeastern part of the state. The rocks consist of a weathering-residuum and overlying clastic strata, and lie unconformably on a surface consisting of rocks ranging in age from Precambrian to Devonian. The basal residuum has been penetrated in the subsurface over a wide area in western Minnesota and adjacent areas in North and South Dakota and probably is equivalent to a weathered zone also known in Manitoba. It formed during a long interval of weathering that ended sometime in the Cenomanian. That the weathering was intense and persisted for a long period of time is indicated by the local presence of more than 200 feet of decomposed granite beneath the residuum. With transgression of the Cretaceous sea from the west, as much as 600 feet of shale, some sandstone, minor limestone, and several thin bentonite beds were deposited above a basal non-marine sandstone. These sedimentary rocks are correlative with the Dakota Formation, the Colorado Group, and perhaps the Pierre Shale of the western interior.

The basal residuum is best exposed along the Minnesota River Valley where it overlies Lower Precambrian granitic rocks and the Upper Precambrian Sioux Quartzite. Where it overlies felsic rocks, it consists primarily of kaolinite and quartz. Tubular halloysite is present in minor amounts, especially in the lower part of the weathering profile. Mafic rock types weather first to montmorillonite and under progressively more intense weathering to kaolinite. Probably, the residuum formed under humid tropical conditions during the encroachment of the warm epicontinental Cretaceous sea from the west. However, after the residuum was formed and prior to the advance of the Late Cretaceous seas into Minnesota a significant climatic change took place. The climate became more temperate and the water table rose, resulting in the formation of dominantly illite and montmorillonite.

With transgression of the seas from the west, thick marine shales and sandstone were deposited in western Minnesota. Sandstones of variable thickness containing thin lignite beds were deposited above the residuum. This sandstone unit is correlated with the Dakota Formation in South Dakota, where it is inferred to be a deltaic deposit; however, in Minnesota it is clearly continental in origin. As the seas continued to transgress eastward, the clastic detritus became finer grained and marine shales of the Colorado Group succeeded the Dakota Formation. Deposition of marine strata was nearly continuous in western Minnesota from Cenomanian to Santonian time (Sloan, 1964), and an equivalent of the Pierre Shale of Campanian age possibly is locally present at the top of the succession.

In southeastern Minnesota, the Cretaceous rocks are assigned to the Windrow Formation, a non-marine, clastic unit containing leaf imprints and carbonized wood. The unit is believed to be equivalent to the Dakota Formation of western Minnesota. As originally defined (Andrews, 1958), a residuum that lies beneath the clastic unit was considered to belong to the Windrow Formation, but the Minnesota Geological Survey now excludes it because it lacks kaolinite, which is diagnostic of the intensely weathered residuum in western Minnesota.

Predominantly marine rocks of the Coleraine Formation overlie the natural iron ores of the Middle Precambrian Biwabik and Trommald Formations of the Mesabi range and Cuyuna district, respectively, whereas elsewhere in northern Minnesota they overlie a kaolinite-bearing residuum. The stratigraphic position of these natural ores and their general conformity with the Cretaceous residuum suggest that the natural ores probably were formed during this interval of intense weathering. The overlying Coleraine Formation on the Mesabi range grades laterally eastward from dominantly marine to dominantly non-marine strata. Paleontologic evidence suggests that this formation most likely is equivalent to the Colorado Group in western Minnesota.

In summary, the dominant pattern of Cretaceous sedimentation in Minnesota is one of an eastward-transgressing sea characterized by a non-marine facies that migrated eastward with time. Thus, the marine Cretaceous rocks become younger in age upward and to the east, and those above the residuum in eastern Minnesota are younger than those above the residuum in western Minnesota.

CENOZOIC ERA

Glaciation during the glacial (Quaternary) period has dominated development of the landscape of Minnesota, leaving an almost continuous cover of glacial drift and related sediments such as glaciofluvial deposits, glacial-lake deposits, and upland loess. Although ice sheets covered the state several times, the landforms and surficial deposits for the most part record only the last (Wisconsin) glaciation. The glacial deposits obscure the record during the Tertiary Period, which apparently was a time of moderate weathering of a low-lying, rather flat geologic terrane.

Pre-Wisconsin Glaciation

In Minnesota, evidence for pre-Wisconsin glaciation is equivocal, and is based on extrapolation of older (Kansan) drift in adjacent parts of Iowa, on deep sections of drift exposed in the open-pit mines on the Mesabi range (Winter, 1971), and in the steep bluffs of the Minnesota River Valley (Matsch, this volume). The lower drifts in the deep sections have carbon dates >35,000 BP, but the successions lack weathering horizons or fossil beds by which an interglacial unit might be recognized, and Wright (this volume) tentatively considers them as being Wisconsin in age.

A small area bordering the Mississippi River valley in southeastern Minnesota commonly has been considered a part of the Driftless Area, which occupies primarily the southwestern quarter of Wisconsin. This area has a thick cover of Wisconsin-age loess, which seems everywhere to lie directly on Paleozoic bedrock. Glacial erratics can be found in the beds of streams that head in this area, however, so it is possible that glacial drift of pre-Wisconsin age occurs locally beneath the loess.

Wisconsin Glaciation

In contrast to the pre-Wisconsin drifts in Iowa and adjacent areas, which have essentially the same lithology, implying a common source, the Wisconsin drifts in Minnesota have a varied lithology and a complex stratigraphy, reflecting the configuration of the several ice lobes that protruded from the ice-sheet margin during various intervals of advance and retreat. This lobation of the ice margin was controlled by the major bedrock topography of the area over which the ice moved. In general, the ice lobes followed lowlands that were developed on the soft rocks that flank the exposed Precambrian rocks of the Canadian Shield. On the eastern side of the state, the major lowlands are the Lake Superior basin and the shallower Minneapolis lowland, which extends northeast through Minneapolis *en echelon* with the Superior lowland (Wright and Ruhe, 1965, fig. 4). The Superior lowland is localized by the relatively soft red sandstones and shales of Keweenawan age; the Minneapolis lowland is localized by relatively poorly cemented Cambrian sandstones and to a lesser extent by Keweenawan sandstones. On the western side of the state, the major lowland is the Red River Valley, which is underlain by soft Cretaceous shales that cover the Paleozoic rocks of the Great Plains. A southeastward extension of this valley is occupied by the Minnesota River Valley.

Four major Wisconsin ice lobes—Superior, Rainy, Wadena, and Des Moines—are recognized for the glaciation of Minnesota. On the east, the Superior lobe expanded out of the Lake Superior basin into the Minneapolis lowland, bringing generally red sandy till. Immediately to the northwest in northern Minnesota was the Rainy lobe, which deposited drift of variable color and composition, depending on the dominant rock types that it traversed. In western Minnesota, the Des Moines lobe followed the Red River Valley and moved across a low divide into Iowa; it sent sublobes eastward across northern Minnesota (St. Louis sublobe) and across southern Minnesota (Grantsburg sublobe), bringing gray calcareous silty till containing fragments of Cretaceous shale. Not so far west was the Wadena lobe, which brought gray to buff sandy calcareous till to the Red Lakes lowland, in northwestern Minnesota, from the carbonate terrane of Manitoba.

Because of the uncertainties in correlating ice advances from lobe to lobe in the Upper Midwest, for Minnesota the sequence of glaciation is recounted as a series of named glacial phases for each of the several interacting ice lobes (Wright and Ruhe, 1965; Wright, this volume). Where definite correlations can be made, the names of adjacent lobes are combined (*e.g.*, Nickerson-Alborn phase of the St. Louis and Superior lobes).

The oldest drift of probable Wisconsin age extensively exposed on the surface in Minnesota is in the Wadena drumlin field in the west-central part of the state, representing the Hewitt phase of the Wadena lobe. The south and western termination of the Wadena lobe in the Hewitt phase is equivocal. The Alexandria moraine complex, which rims the outer margin of the drumlin field, is considered to represent in its core the terminal deposit of the Wadena lobe when the drumlins were formed (Wright, this volume). Recent mapping in and adjacent to the Minnesota River Valley (Matsch, this volume), however, suggests that the Wadena lobe, with its shale-poor drift, extended across the valley into South Dakota. Matsch has named the till exposed in the valley the Granite Falls till. It overlies a still older till in the valley, the Hawk Creek till. Carbon dates on wood from beneath the Granite Falls till are 34,000 and >40,000 BP. This westward extension of the Wadena lobe may represent a phase of Wisconsin glaciation older than the Hewitt phase, or it may be an early maximum of the Hewitt phase, before retreat of the ice to the Alexandria moraine complex.

The other ice lobes in Minnesota probably were active at the same time as the Hewitt phase of the Wadena lobe but had not attained their maximum positions. In addition to the Rainy lobe, the Des Moines lobe must have occupied the Red River lowland at this time, and indeed the Wadena lobe may have been an offshoot of the Des Moines lobe but from a source farther north than its later counterpart (St. Louis sublobe) (Wright and Ruhe, 1965).

The second phase recognized by Wright (this volume) is the Itasca-St. Croix phase of the Wadena and Rainy/Superior lobes. At this time, the Wadena lobe retreated to the Itasca moraine, in western Cass County, and the Superior and Rainy lobes reached a common terminus at the St. Croix moraine, which forms a great loop from western Wisconsin on the St. Croix River southwestward to Minneapolis and thence to the northwest and north, where it becomes interlobate with the Itasca moraine. The St. Croix moraine has been traced for about 350 miles, and is one of the most sharply defined glacial features in the Great Lakes region. Wastage of the Superior and Rainy lobes of the St. Croix phase from the St. Croix moraine left a remarkable record of eskers and tunnel valleys (Wright, this volume). Concurrent wastage of the Wadena lobe from the Itasca moraine also resulted in the formation of tunnel valleys, now exposed in Itasca State Park and adjacent areas.

Following the general deglaciation of the St. Croix phase, the Superior and Rainy lobes readvanced to new positions, and this phase is called the Automba phase of the Superior lobe (Wright, this volume). The locations of the Wadena and Des Moines lobes at this time are unknown. The Superior lobe moved out of the Lake Superior basin westward across moderately high ground to the region of Mille Lacs Lake, in east-central Minnesota. Possibly stagnant ice in the vicinity of Sandstone, in Pine County, prevented movement of the ice southward into the topographically lower Minneapolis lowland. The (Highland) moraine formed at this time along the northwest margin of the Superior lobe, at the crest of the North Shore Highland, was continuous. At the end of the ice lobe, the very distinct Mille Lacs moraine was formed, at the west end of Mille Lacs Lake; this moraine fades out to the east. The Highland moraine truncates the conspicuous Toimi drumlins of the St. Croix phase Rainy lobe, and its outwash plains partially bury the drumlins. Near Isabella, in Lake County, the Highland moraine is joined from the west by the Vermilion moraine, which truncates the Toimi drumlin field on the north and extends west as a sharp morainic ridge across the eastern end of the Mesabi range, and thence west to bound Lake Vermilion and Nett Lake on the south.

The next major phase is marked by the concurrent movement of (1) a thin ice lobe out of Lake Superior and (2) the Grantsburg sublobe of the Des Moines northeastward from the Minnesota River Valley through the Minneapolis lowland to a maximum at Pine City. This is called the Split Rock-Pine City phase of the Superior and Grantsburg lobes (Wright, this volume). At this time, the Superior lobe deposited a thin, discontinuous blanket of red clayey till over the southwestern end of the Lake Superior basin. The Grantsburg sublobe overrode till left by the Superior lobe in the St. Croix phase, and in the process picked up red drift or drift-laden deadice to produce the intricate interlaminated red and buff structures that mark the contact between the two drifts (Cushing, 1963). Glacial Lake Grantsburg was formed north of the sublobe, from near St. Cloud eastward into Wisconsin, as a result of damming of the Mississippi River and other drainage from the north. As the Grantsburg sublobe withdrew from its maximum, Lake Grantsburg drained, and the meltwater formed a series of coalescing outwash plains, and, locally, small lake plains which together are known as the Anoka sandplain. As the Grantsburg sublobe wasted, the course of the outwash streams—both the Mississippi River and local streams—continued to shift generally southward, as lower ground was gradually opened by the wasting ice (Cooper, 1935). When the Mississippi River broke through the St. Croix moraine at Minneapolis it established its modern course, which ultimately became a gorge. Earlier gorges of similar type—now buried valleys—crisscross the Minneapolis-St. Paul area (Payne, 1965). Characteristically, the buried valleys underlie strings of lakes.

The Nickerson-Alborn phase of the St. Louis and Superior lobes (Wright, this volume) marks the maximum advance of the St. Louis sublobe of the Des Moines lobe across the Red Lakes lowland of northwestern Minnesota to southwest St. Louis County and the concurrent retreat of the Superior lobe to the Nickerson moraine, which is a striking feature of strong local relief in red clay till near the west end of the Superior lowland. The St. Louis sublobe was restricted in its southward expansion by the Itasca moraine and in its northern expansion by higher ground in Canada. The ice extended eastward beyond the interlobate junction of the Itasca and St. Croix moraines in Cass County, and then spread southward in a sub-sublobe as far as Mille Lacs Lake. Another sub-sublobe was diverted around the southwest end of the Giants Range, and it then flowed north to the range and east along the south flank (Winter, 1971). A third sub-sublobe overrode the west end of the Giants Range and extended as far northeast as Lake Vermilion, where it was blocked in part by the Vermilion moraine. As the St. Louis sublobe withdrew from its Alborn maximum, Glacial Lakes Upham II and Aitkin II formed at its front, west and northwest of Duluth. Later, with withdrawal of the Superior lobe into the Lake Superior basin, Glacial Lake Nemadji (altitude 1,060 feet) was formed as a proglacial lake at the southwestern end of the basin. As the ice withdrew further into the basin, a lower outlet was uncovered in Wisconsin, and the lake level lowered about 50 feet to 1,010 feet and stabilized as Glacial Lake Duluth. Further retreat of the Superior lobe uncovered still lower outlets to the east, to the Lake Michigan and Lake Huron basins. The strand lines that record these lower lake levels are all inclined gently to the southwest as a consequence of the southward tilting in response to removal of the load of glacier ice from the crust.

Meanwhile, as the Superior lobe was retreating into the Lake Superior basin for the final time, the Des Moines lobe withdrew into the Red River lowland, and Glacial Lake Agassiz formed at its front (ca. 12,000 years ago)—behind the Big Stone moraine, which essentially forms the divide between the Minnesota and Red River Valleys. As the ice withdrew farther north, its outlet stream dissected the Big Stone moraine, and as the outlet was eroded successive beaches were formed at altitudes ranging in the outlet regions from 1,060 feet (Herman) to 980 feet (Campbell). The outlet stream, Glacial River Warren, eventually became channeled into the Minnesota River Valley, which cut a deep outlet gorge through the glacial deposits to the Mississippi River at Minneapolis and beyond to Illinois. At the level of the Campbell strand line the outlet stream reached to granitic bedrock and downcutting ceased (ca. $9{,}200 \pm 600$ years ago). Outlets then were uncovered east to Lake Superior. The major remnants of Glacial Lake Agassiz in Minnesota are the Red Lakes in Beltrami County and Rainy Lake and Lake of the Woods on the International boundary.

Judged from available carbon dates, the Wisconsin glacial history in Minnesota correlates with the Cary (possibly Tazewell) to Port Huron advances of the Lake Michigan lobe (Wright and Ruhe, 1965).

In postglacial time, geologic changes in the Minnesota landscape have been minimal. Slopes soon became stabilized by vegetation, and most subsequent geologic action has been confined to the major river valleys and lakeshores.

Chapter II

HISTORY OF INVESTIGATIONS

HISTORY OF GEOLOGIC INVESTIGATIONS

P. K. Sims

Scientific investigations of the geology of the State of Minnesota span more than a century, and recorded observations of natural features date back to the late 1600's. Because investigations carried out since the establishment of the first organized geological survey reflect rather closely the structure and personnel of the state organizations, the history of investigations can be discussed with respect to five chronological periods: (1) the *pioneer period*, which includes the scattered observations made prior to the establishment of a formal geological survey; (2) the *period 1872-1900*, which coincides with the tenure of Newton Horace Winchell as State Geologist and Director, Geological and Natural History Survey; (3) the *period 1900-1911*, when there was no active State Geological Survey, but some excellent investigations were made by the United States Geological Survey; (4) the *period 1911-1961*, at which time the Minnesota Geological Survey was an integral part of the Department of Geology at the University of Minnesota; and (5) the *period since 1961*, when the Survey became a separate department with a permanent staff. Much of the history described here is summarized from Schwartz's earlier (1964) account.

PIONEER PERIOD

Early accounts by explorers and Jesuit missionaries provided considerable information on the geography and natural history of the state but little on the geology. An exception was Father Hennepin's 1680 description of St. Anthony Falls on the Mississippi River. His location of the falls was so accurate that N. H. Winchell later was able to use it in calculating the rate of migration since retreat of the glaciers from Minnesota.

N. H. Winchell (1884a, p. 34) credited Professor William Keating of the University of Pennsylvania with making "the first attempt to apply the accurate methods of modern science to the exploration of any portion of Minnesota." As a member of an expedition led by Major S. H. Long in 1817, Keating described the geology of the falls at St. Anthony, the stratigraphic section of the bluff at Fort Snelling, and some aspects of the geology of the Minnesota River Valley. He recognized the "primitive" character of some of the rocks in the Minnesota River Valley and in the northern boundary region.

In separate expeditions in 1835, G. W. Featherstonhaugh, titled U. S. Geologist, contributed to knowledge of the geology of the Mississippi and Minnesota Rivers and the Coteau des Prairies; and George Catlin, after whom catlinite was named, visited the pipestone quarry in what is now Pipestone County in southwestern Minnesota.

During the period 1836-1843, Joseph Nicolas Nicollet explored the upper Mississippi region as an employee of the Bureau of the Corps of Topographical Engineers. In addition to making the most complete and correct map of the region of that time, he recorded several observations on the geology.

The first systematic scientific report specifically based on geologic studies rather than on military reconnaissance was made by David Dale Owen during the period 1847-1850. Accompanied by able colleagues, notably J. G. Norwood, B. F. Shumard, and Charles Whittlesey, the Owen expedition established the paleontologic age of most of the sedimentary rocks of southern Minnesota and made the first serious attempts to correlate the strata with those in better known areas. Another major contribution of the Owen Survey was Whittlesey's description of the glacial deposits of northern Minnesota. N. H. Winchell (1884a, p. 99) credited Whittlesey with publishing (in 1866) the first correct theory of the origin of the glacial deposits of the state.

Between 1858, when Minnesota was admitted to the Union, and 1872, when a formal geological survey was established, several unsuccessful attempts were made to establish a state geological survey. During this interval, little geologic data of value were published. An exception was a series of papers by W. D. Hurlbut, published in the *Minnesota Teacher*, which described the geology in the headwaters of the tributaries of the Mississippi River. Because of the remoteness of this area, its rocks had not been described previously.

For the most part, geologic observations during this period concerned the sedimentary rocks in the southern part of the state, where they were found along the navigable rivers, the principal means of access. It was known that granitic and metamorphic rocks cropped out discontinuously from the northeastern to the southwestern part of the state, but little was known of the nature of these rocks. Selected areas had been prospected for gold, silver, and other metals, and iron ores had been discovered in what is today the Vermilion district, but the geology of these remote areas was largely unknown.

THE PERIOD 1872-1900

With the establishment of the Minnesota Geological and Natural History Survey in 1872, under the direction of Newton Horace Winchell, geologic knowledge advanced rapidly, and by the end of the century it compared favorably with that in other states. In addition to N. H. Winchell, the Survey roster during this period included such well-known men as U. S. Grant, A. C. Lawson, Charles Schuchert, E. O. Ulrich, and H. V. Winchell, as well as several younger men that later attained prominence, such as C. P. Berkey and J. E. Spurr.

A major objective of the Winchell Survey was to complete a geologic map of the state on a county basis. As counties were completed, the maps and accompanying texts were published in volumes of the Final Reports. Finally, in volume 6 of the Final Report, a geologic map of the state was published at a scale of about 42 miles to the inch. A lengthy explanation of the stratigraphic column to accompany the state map was published in the preface to volume 5 of the Final Report.

The mapping and petrographic studies carried out by the Winchell Survey provided a sound geologic basis for later investigations. As a result, the distribution and gross relationships of the major rock units were fairly well known, although many refinements remained to be made, especially for the crystalline rocks.

The stratigraphic succession of the Precambrian rocks, published in volume 5 of the Final Report (Winchell and Grant, 1900, p. 26) and reproduced below (table II-1), does not differ materially from that recognized today, although the age designations differ considerably. The Winchell Survey divided the older (Archean) rocks into two groups, the Lower Keewatin, consisting dominantly of mafic volcanic rocks, and the Upper Keewatin, consisting dominantly of clastic sedimentary rocks. It was inferred that an episode of folding, metamorphism, and igneous activity separated the two groups of rocks, and that the base of the Upper Keewatin generally was marked by a conglomerate.

Although local studies were made of the granitic rocks, no serious attempts were made to delineate separate bodies. The problem was partly a cartographic one — the difficulty of separating granitic material from schists and gneisses for mapping purposes — and partly consequent upon uncertainties in correlations. Both problems still plague us, at least in certain areas, and it is not surprising that the Winchell Survey was content to leave much of the granite problem to future generations.

The work of the Winchell Survey on the Animikie rocks of the Mesabi range was outstanding, and contributed greatly to rapid development of the natural iron ores. The geologic map of the Mesabi range (Winchell and others, 1899, opposite p. 616) was remarkably complete and accurate.

Substantial advances in knowledge of the stratigraphy and paleontology of the sedimentary rocks were made by the Winchell Survey. Volume 3 of the Final Report, published in two parts in 1895 and 1897, was an excellent geologic report, and remains a valuable paleontologic contribution. Through geologic mapping, the distribution and stratigraphic succession of the Paleozoic and Mesozoic rocks were defined quite accurately, and correlations of Paleozoic strata from area to area generally were reasonably correct. An excellent beginning in studies of the subsurface geology and the occurrence of ground water was made by collecting available water well cuttings and related data.

Geologic mapping added to knowledge of the distribution and lithology of the unconsolidated glacial deposits, but no serious attempts were made to integrate the data into a comprehensive statewide picture. The principal contributions were made by Warren Upham, notably on Glacial Lake Agassiz, and by N. H. Winchell.

Table II-1. Precambrian succession in Minnesota (after N. H. Winchell, 1900, p. 6).

	Upper Cambrian		The *St. Croix* and *Hinckley* sandstones of the Upper Mississippi valley. Seen at Fond du Lac. The lower portion interstratified with trap, and thus passing to the Keweenawan.
TACONIC	*Lower* (and probably *Middle*) *Cambrian*	Keweenawan	*Potsdam* (clastic) and *Manitou* (igneous) rocks. Puckwunge conglomerate at the base.
			NON-CONFORMITY
		Keweenawan	*Cabotian* (igneous). Gabbro and contemporary surface eruptives.
		Animikie	Slates, quartzytes; conglomerate at the bottom of the clastic rocks, but sometimes without a basal conglomerate. Numerous dikes and sills of the age of the Keweenawan.
			The *Mesabi* iron ores.
			NON-CONFORMITY
	Upper Keewatin		Quartz-porphyry Volcanic tuff Flint Quartzyte Sericite schist Jaspilyte Argillyte Graywacke Greenwacke Greenstone (clastic) Ogishke conglomerate
ARCHEAN			NON-CONFORMITY
	Lower Keewatin		Quartzyte Graywacke Flint Volcanic tuff Argillyte Greenstone (clastic) Jaspilyte Quartz-porphyry Greenstone conglomerate Greenstone (igneous), oldest known rock

THE PERIOD 1900-1911

Upon completion of Winchell's work in 1900, geologic investigations by the Minnesota Geological and Natural History Survey were discontinued. About this time, however, geologists of the United States Geological Survey started investigations in the state as part of a comprehensive study of the geology of the Lake Superior region. C. R. Van Hise was in charge of this program, and C. K. Leith and J. M. Clements were responsible for the field work. As a result of these investigations, U.S. Geological Survey

Monograph 43, *The Mesabi iron-bearing district of Minnesota*, by Leith, and Monograph 45, *The Vermilion iron-bearing district of Minnesota*, by Clements, were published in 1903. In addition, a report on water supplies, *Geology and underground waters of southern Minnesota*, done cooperatively with the Minnesota State Board of Health, was published as U.S. Geological Survey Water Supply Paper 256. Each report had a profound influence on later studies in the respective areas. Clements' (1903) report, in particular, on the older Precambrian rocks in northern Minnesota, established the general outline of these complex rocks, and together with the studies of Lawson in the Lake of the Woods and Rainy Lake areas, in adjacent Ontario, formed the basis for a standard geologic succession and nomenclature.

THE PERIOD 1911-1961

With the appointment of W. H. Emmons as Director in 1911, the Minnesota Geological Survey became a part of the Department of Geology, University of Minnesota. During the succeeding 50 years, the Survey staff consisted largely of faculty members and students employed during the summer months. This staffing arrangement provided continuity to many Survey projects and resulted in several outstanding scientific and economic contributions. However, broad investigations of all aspects of the geology were impossible because of the relatively small staff and financial support.

During this period, the U.S. Geological Survey was relatively inactive in the state. A notable exception was the mapping of the state's glacial deposits by Frank Leverett (U.S.G.S.) and F. W. Sardeson (M.G.S.). The map and report (Leverett and Sardeson, 1932) resulting from this cooperative project have been used widely to locate deposits of sand and gravel and to evaluate shallow groundwater supplies, and still are a major source of valuable information. Another cooperative investigation was that of E. C. Harder (U.S.G.S.) and A. W. Johnston (M.G.S.) in the Cuyuna range and adjacent areas in east-central Minnesota. Much later, a restudy of the Cuyuna North range was made by Schmidt (1963).

A substantial effort of the Minnesota Geological Survey during the early part of this period was the preparation of a new bedrock geologic map of the state. G. W. Stose of the U.S. Geological Survey edited the map, which was published in 1932 at a scale of 1:500,000.

Because of the economic importance of the iron ores, major emphasis was placed on studies of the Mesabi district. Prior to World War I, F. F. Grout and T. M. Broderick studied the eastern part of the range. Upon completion of this work, Broderick and J. W. Gruner began studies of the remainder of the Mesabi range (Gruner, 1924). Gruner continued work on the Mesabi during the ensuing years, which resulted twenty years later in the publication of his classic paper on the mineralogy and geology of the range (Gruner, 1946). Perhaps more than anyone else, Gruner was responsible for delineating the vast quantities of taconite eventually used for the making of iron pellets. Following World War II, research was continued on the Mesabi range by graduate students in the Department of Geology. A paper by White (1954) on the stratigraphy and structure of the range was a valuable addition to the earlier work of Gruner, and a paper on the detailed stratigraphy and mineralogy of the metamorphosed taconite at the east end of the range (Gundersen and Schwartz, 1962) provided data useful for interpreting problems related to concentration of the taconite.

Many years after the earlier study of the Vermilion district and adjacent areas by Clements (1903), Grout (1926) began reconnaissance investigations that extended from the district northward to the International boundary. Outcrop maps were made of each township and a generalized geologic map was compiled. This work clarified some aspects of the geology of the older rocks of the area, particularly in the Vermilion batholith (Grout, 1925b), but it led to conflict with A. C. Lawson and others regarding the Coutchiching problem, a problem that already had generated considerable debate but had brought forth few unequivocal facts. Later, S. S. Goldich and students studied the Rice Bay area in the Rainy Lake district of Ontario, the region in which the Coutchiching was named originally by Lawson (1888), and concluded (Goldich and others, 1961, p. 5) that the Coutchiching was a valid stratigraphic term, at least for the border region.

Another significant study during this period was an investigation of the Duluth Complex by F. F. Grout and his associates. Grout (1918a) proposed the term lopolith for the presumed intrusive form of the complex, a hypothesis now known to be incorrect. Although it is clear from observations recorded in his field notebooks that Grout recognized the composite nature of the Duluth Complex, he did not publish these data; and it was several years before R. B. Taylor (1964) pointed out the complex intrusive relations.

A major contribution to the structural geology of northeastern Minnesota was made by Gruner (1941) in the Knife Lake area. Through careful geologic mapping of the Knife Lake Group and associated strata, Gruner showed that these ancient rocks were faulted as well as folded. Gruner's map, which still is extremely valuable, was the first in Minnesota to show faults.

Several other Precambrian rock units in northern Minnesota also were investigated during this period. The Saganaga Tonalite, along the Canadian border, was studied by Grout and Schwartz (1933) and later by Grout and others (1959); rocks of the North Shore Volcanic Group were mapped by Sandberg (1938) and others; and the Giants Range Granite was studied by Allison (1925).

A summary paper by Grout and others (1951) on the Precambrian stratigraphy of Minnesota was the culmination of many years' work by Grout and his colleagues in northern Minnesota. It established a formal nomenclature for the rock units, and proposed a three-fold classification of the Precambrian (table II-2). The classification served as the basic framework for the geochronology studies of Goldich and coworkers on the Precambrian crystalline rocks.

The Minnesota River Valley, isolated from the classic Precambrian rock succession in northern Minnesota by many miles of covered area, was investigated by Lund (1956) in the early fifties. He delineated a sequence of metamorphic and igneous rocks that differs substantially from

Table II-2. Precambrian column in Minnesota; chronologic and stratigraphic sequence (after Grout and others, 1951).

	Eras and Rocks	Groups	Formations and Members		Other Names Used in Minnesota
Phanerozoic Eon	Cenozoic Mesozoic Paleozoic				
		unconformity			
Precambrian (or Cryptozoic Eon)	Later	Keweenawan Group	Upper: Sandstones and other sediments	Hinckley Fond du lac	Red Clastics*
			Middle: Intrusives, acidic and basic	Scattered granites* Duluth gabbro Beaver Bay Complex, and Logan intrusives	
			Middle: Flows, tuffs and sediments	Keweenaw Point Volcanics	
			Lower: Conglomerate and sandstones	Puckwunge formation	
		unconformity			
		? Sioux formation* ?			
		Animikie Group	Virginia slate (minor iron formation?) = Rove =		Upper Cuyuna slate*
			Biwabik iron formation series = Gunflint =		Deerwood*
			Pokegama quartzite		
		unconformity, great on the Mesabi and Gunflint ranges			
	Medial	Algoman batholithic intrusives, orogeny and erosion			
			Central and southwestern Minnesota granites*		
		Knife Lake Group (about 18 members)	slate	= Thomson*	(Carlton)*
			graywacke		Emily*
			iron-bearing beds		Agawa*
			conglomerate		Ogishke*
			tuffs, lavas and intrusives		
		unconformity, great west of Saganaga Lake			
	Earlier	Pre-Knife Lake batholithic intrusives, orogeny and erosion			
		Keewatin volcanics, and Soudan iron formation member. (No Couchiching recognized in Minnesota)			Ely greenstone

* Means uncertain as to place.

the Precambrian succession in northeastern Minnesota, and which served as a springboard for the subsequent intensive dating program initiated by Goldich.

Investigations of the Paleozoic sedimentary rocks in southeastern Minnesota and of the subsurface in the state were carried out intermittently. In the period immediately prior to World War II, G. A. Thiel, C. R. Stauffer, and I. S. Allison systematically collected subsurface data from various parts of the state and integrated them with outcrop data. Until recently, the reports resulting from these studies were the major sources of groundwater data in the state. An outstanding pioneering effort in environmental geology was a report, *The geology of the Minneapolis-St. Paul metropolitan area*, by G. M. Schwartz, published in 1936. A similar report on the Duluth area was prepared by Schwartz a few years later (1949). After World War II, W. C. Bell, R. E. Sloan, F. M. Swain, and their students, and R. L. Heller, restudied the Ordovician and Cambrian strata exposed in southeastern Minnesota. Much of the data obtained during this period was summarized in the Guidebook for the 1956 meeting of the Geological Society of America held in Minneapolis.

Many modifications and refinements of the earlier work on the glacial geology of the state were made during this period. Included among these studies are Cooper's (1935) careful analysis of the drainage changes between the Mississippi and St. Croix rivers during the Middle and Late Wisconsin, Wright's (1953) refinements of the glacial history in east-central Minnesota, and Zumberge's (1952) excellent synthesis and classification of the lakes in Minnesota.

A cooperative project with the U.S. Geological Survey to make an aeromagnetic map of the state was started in 1947 and continued intermittently into the 1960's. Aero-

Table II-3. Stratigraphic succession and time classification of the Precambrian rocks of Minnesota (after Goldich, 1968).

Era	Group	Formation	Event	Intrusive rocks
Paleozoic				
		— *Unconformity* —		
600 m.y.		Hinckley Sandstone		
		Fond du Lac Sandstone		
		— *Unconformity* —		
Late Precambrian	North Shore Volcanic Group	undivided, flows, tuffs, and sediments	Keweenawan igneous activity (1000-1200 m.y.)	Duluth Gabbro complex and smaller intrusions on North Shore
		Puckwunge Conglomerate(?)		
		— *Unconformity* —		
		Sioux Quartzite(?)		
		— *Unconformity* —		
1800 m.y. Middle Precambrian	Animikie Group	Rabbit Lake = Virginia = Rove = Thomson	Penokean orogeny (1600-1900 m.y.)	Granites in east-central Minnesota; small intrusions in Minnesota River Valley
		Trommald = Biwabik = Gunflint		
		Mahnomen = Pokegama = Kakabeka		
		— *Unconformity* —	Algoman orogeny (2400-2750 m.y.)	Vermilion and Giants Range granitic complexes; granite and gneiss (in part), Minnesota River Valley
2600 m.y.	Knife Lake Group	undivided, slate, graywacke, conglomerate, tuffs, lavas		
Early Precambrian		— *Unconformity* —	Laurentian orogeny (age ?)	Granite at Saganaga Lake
	Keewatin Group	Soudan Iron-formation		
		Ely Greenstone		
	Coutchiching(?)	Metasedimentary rocks(?)		
		Older rocks	(?) (3300-3550 m.y.)	Gneiss (in part), Minnesota River Valley

magnetic maps first were issued by the U.S. Geological Survey at a scale of 1:62,500; later, the data were republished with available geologic information at a scale of 1:250,000 on U.S.G.S. two-degree sheets. These maps disclosed some previously unknown magnetic anomalies that subsequently were investigated by private industry.

A program of radiometric dating was initiated in 1956 by S. S. Goldich and A. O. Nier, of the Department of Physics. This work clarified many of the problems in the Precambrian and resulted in several revisions in interpretation of the Precambrian history (Goldich and others, 1961). Notable contributions were a revision in the Precambrian time scale, and the recognition of an orogenic event, named the Penokean orogeny, which followed deposition of the Animikie Group. The time scale by Goldich and others (1961, table 2) retained the three-fold division introduced by Grout and others (1951). A modified version of the original in which results of more refined radiometric age measurements are incorporated is shown in Table II-3. The 1961 publication demonstrated the value of radiometric age determinations on a regional basis to delineate events in the Precambrian, and remains a useful work.

THE PERIOD SINCE 1961

Beginning in 1961, the Minnesota Geological Survey intensified geologic studies in the state. Modern standard topographic maps were then becoming available, as a result of the foresight of former director Schwartz, and a high priority was given to geologic mapping. Most of the existing geologic maps were 30 or more years old and many of these were simply outcrop maps. New geologic mapping was essential to a critical re-analysis of the geology and resources of the state. Major emphasis was given to preparing a new geologic map of the state at a scale of 1:250,000. To aid the geologic mapping, the aeromagnetic mapping started earlier in cooperation with the U.S. Geological Survey was completed as soon as possible, and gravity measurements were collected on a systematic basis throughout the state. An aeromagnetic map of the state, which was a compilation of all the separate airborne surveys (Zietz and Kirby, 1970), was published (scale 1:1,000,000) by the U.S. Geological Survey. A Bouguer gravity map of the state, which included all data compiled through 1966, was also published (Craddock and others, 1970), at the same scale. The availability of these geophysical maps provided the impetus to prepare a new state geologic map at the same scale (Sims, 1970), which will serve as an interim map until all the 1:250,000 map sheets are published. Through 1971, three (St. Paul, New Ulm, and Hibbing) of the 11 sheets for the state (bedrock) geologic map atlas had been published, and four others (Two Harbors, Duluth, Stillwater, and International Falls) were nearing completion. A complementary program to prepare a state surficial map at the same scale was not started until the late 1960's.

During this period, S. S. Goldich and colleagues continued investigations of the radiometric ages of rocks in Minnesota. In the Minnesota River Valley, it was shown that granitic gneisses near Morton and Montevideo have zircon ages of 3,550 m.y. (Goldich and others, 1970), and are among the oldest rocks known in North America. Less exciting but equally significant age data were obtained on rocks in northern Minnesota, the details of which are given elsewhere in this volume and in papers published in the Gruner volume, a Geological Society of America publication. In this work, Goldich, G. N. Hanson, and Zell Peterman, as well as others actively engaged in radiometric dating of rocks in the state, worked closely with the field geologists, a collaboration that has been most fruitful, as will be evident from many of the separate papers in this volume.

GEOCHRONOLOGY IN MINNESOTA

S. S. Goldich

In the 10 years since publication of Bulletin 41 of the Minnesota Geological Survey (Goldich and others, 1961) considerable progress has been made not only in the improvement of analytical techniques but also in the interpretation of the age measurements and in their geologic applications. Some of this progress has been reviewed in a summary paper (Goldich, 1968), and the present discussion is concerned primarily with the work accomplished during the past five or six years.

Studies have been continued in the Minnesota River Valley where the granitic gneisses in the vicinities of Morton, Granite Falls, and Montevideo are the oldest known rocks in North America. Like very ancient rocks in other parts of the world the gneisses have had a complicated history, and metamorphic changes have masked their original character and obscured their age. Conservatively the age may be given as 3300 to 3600 m.y. Goldich and others (1970) have attempted to probe the metamorphic history, and concluded that the gneisses date back to 3550 m.y. ago. Similarly old, or older, gneisses (3600 to 4000 m.y.) have been reported from the Godthaab district, West Greenland (Black and others, 1971). These ancient rocks are of special interest as they provide information on early Precambrian history, and detailed studies of the Greenland rocks are in progress. Geochronological and geochemical investigations are being continued in Minnesota.

Considerable progress has been made in northern and east-central Minnesota. New analytical techniques have been applied, and the new results have shown up earlier faulty conclusions. The new insights into Precambrian history are a significant achievement, but the complexity of Precambrian geology should not be underestimated, nor should the search for new and better tools be diminished.

DECAY CONSTANTS

The fundamental problems related to the uncertainties in the decay constants of K^{40} and Rb^{87} remain unresolved. Most laboratories use the constants given in Table II-4, and all ages in this discussion have been calculated using these constants.

The decay constants for K^{40} used in Bulletin 41 (Goldich and others, 1961, p. 9) were based on a value of 0.0118 atom percent for the relative abundance of K^{40} (Fuller, 1959, p. 69). In hindsight, it is apparent that introduction of the slightly different constants led to unnecessary confusion. The calculated ages differ by less than one percent from the values obtained with the constants (table II-4), which are based on a value of 0.0119 atom percent for the relative abundance of K^{40} (Nier, 1936).

In Bulletin 41, the decay constant of Rb^{87} determined by Flynn and Glendenin (1959), $\lambda = 1.47 \times 10^{-11} yr^{-1}$, was used. The decay constant of $1.39 \times 10^{-11} yr^{-1}$ for Rb^{87} (table II-4) was determined empirically by Aldrich and others (1956), and the recalculated ages are approximately 6 percent greater than the values originally reported.

Table II-4. Constants used for calculating isotopic ages.

PARENT	DAUGHTER	DECAY CONSTANT
U^{238}	Pb^{206}	1.54×10^{-10} yr^{-1}
U^{235}	Pb^{207}	9.72×10^{-10} yr^{-1}
Th^{232}	Pb^{208}	4.88×10^{-11} yr^{-1}
Rb^{87}	Sr^{87}	1.39×10^{-11} yr^{-1}
K^{40}	Ar^{40}	5.85×10^{-11} yr^{-1}
	Ca^{40}	4.72×10^{-10} yr^{-1}

Atom ratios:
$U^{238}/U^{235} = 137.8$
$Rb^{87}/Rb^{85} = 0.2785$
$K^{40}/K = 0.000119$

MINNESOTA RIVER VALLEY

Mineral Ages

On the basis of K-Ar and Rb-Sr mica ages, Goldich and others (1961) suggested two periods of orogeny accompanied by magmatic activity in southwestern Minnesota. A major orogenic event, approximately 2500 m.y. ago, was postulated for the Morton area in which granite near Sacred Heart was considered to be a late- or postkinematic intrusion. A younger event, approximately 1800 m.y. ago, was suggested in the Granite Falls area. This interpretation was soon found to be untenable when Rb-Sr determinations on K-feldspar samples from both areas gave similar ages of approximately 2500 m.y. (Goldich and Hedge, 1962).

The Rb-Sr ages on K-feldspar, like the mica ages, failed to give any indication that the gneisses actually were much older than 2600 m.y., and this was first revealed in zircon lead-alpha determinations by T. W. Stern in the U.S. Geological Survey laboratory, Washington, D.C. The startling ages of 2800 and 3000 m.y. led to the isotopic U-Pb analyses (Catanzaro, 1963). Catanzaro found that the U-Pb ages on two zircon concentrates from the Morton Gneiss and one from the gneiss (Montevideo granite gneiss of Lund, 1956) near Granite Falls are discordant, but the Pb^{207}/Pb^{206} ages range from 3100 to 3280 m.y. He also analyzed zircon from a granitic intrusion in the Montevideo gneiss and obtained discordant ages with a Pb^{207}/Pb^{206} age of 1825 m.y. Further work was then undertaken by Stern (1964), who reported preliminary U-Pb and Th-Pb ages on zircon from the gneisses and other rocks in the valley.

The first K-Ar hornblende measurements from the valley were by Thomas (1963) of the U.S. Geological Survey, who reported an age of 2630 m.y. for hornblende from the Morton Gneiss and 2760 m.y. for hornblende from the Sacred Heart granite of Lund (1956). Hanson and Himmelberg (1967) found an age of 2740 m.y. for hornblende from a hornblende-pyroxene gneiss near Granite Falls, and Hanson (1968) reported an age of 2590 m.y. for hornblende from the Morton Gneiss. Goldich and Gast (1966) studied the effects of weathering on the K-Ar and Rb-Sr ages of biotite from the Morton Gneiss, and in a companion paper, Stern and others (1966) showed that a large proportion of the lead was removed from zircon during the weathering.

Morton and Montevideo Gneisses

U-Pb Ages

The U-Pb zircon ages by Catanzaro (1963) demonstrated beyond any doubt that the mica ages and the Rb-Sr K-feldspar results reflect metamorphic events. A new attack on the problem was then initiated with the more refined Rb-Sr whole-rock technique and with additional isotopic analyses of zircons. The results are given in a paper by Goldich and others (1970) from which Figure II-1 is taken. In this diagram, mineral ages and Rb-Sr-isochron and U-Pb-concordia age interpretations reveal three major events. The oldest, the Mortonian event, at approximately 3550 m.y. ago, is subject to some interpretation because both the Rb-Sr and the U-Pb ages are discordant as a result of a high-grade metamorphic event that occurred 2650 m.y. ago and a lower-grade metamorphism at approximately 1850 m.y. ago.

The zircon analyses (Catanzaro, 1963, p. 2047) when plotted on a concordia diagram (Wetherill, 1956) defined a line intersecting the concordia curve at 3550 m.y. and 1850 m.y. The discordance in the U-Pb ages of the zircon can be attributed to loss of lead from zircon, 3550 m.y. old, as a result of a thermal event 1850 m.y. ago. Mineral ages in the range from 1700 to 1900 m.y. (fig. II-1) lend considerable support to this interpretation. Catanzaro (1963) also noted: (1) that in this episodic lead-loss interpretation the 2500-m.y. metamorphic event, although the stronger of the two events, apparently did not affect the zircons and (2) that the zircon from the granite pluton in the Montevideo gneiss also gave discordant ages, which to his mind, complicated the simple episodic lead-loss interpretation.

Figure II-2, from Goldich and others (1970, p. 3680), is a plot of zircons and an allanite from the gneisses and other rocks in the Minnesota River Valley. The chord, which represents a linear regression of the zircon samples from the gneisses, gives upper and lower intercept ages of 3560 and 1850 m.y., respectively, similar to those of Catanzaro (1963) and Stern and others (1966). Three zircon samples (Stern and others, 1966) from residual clay developed

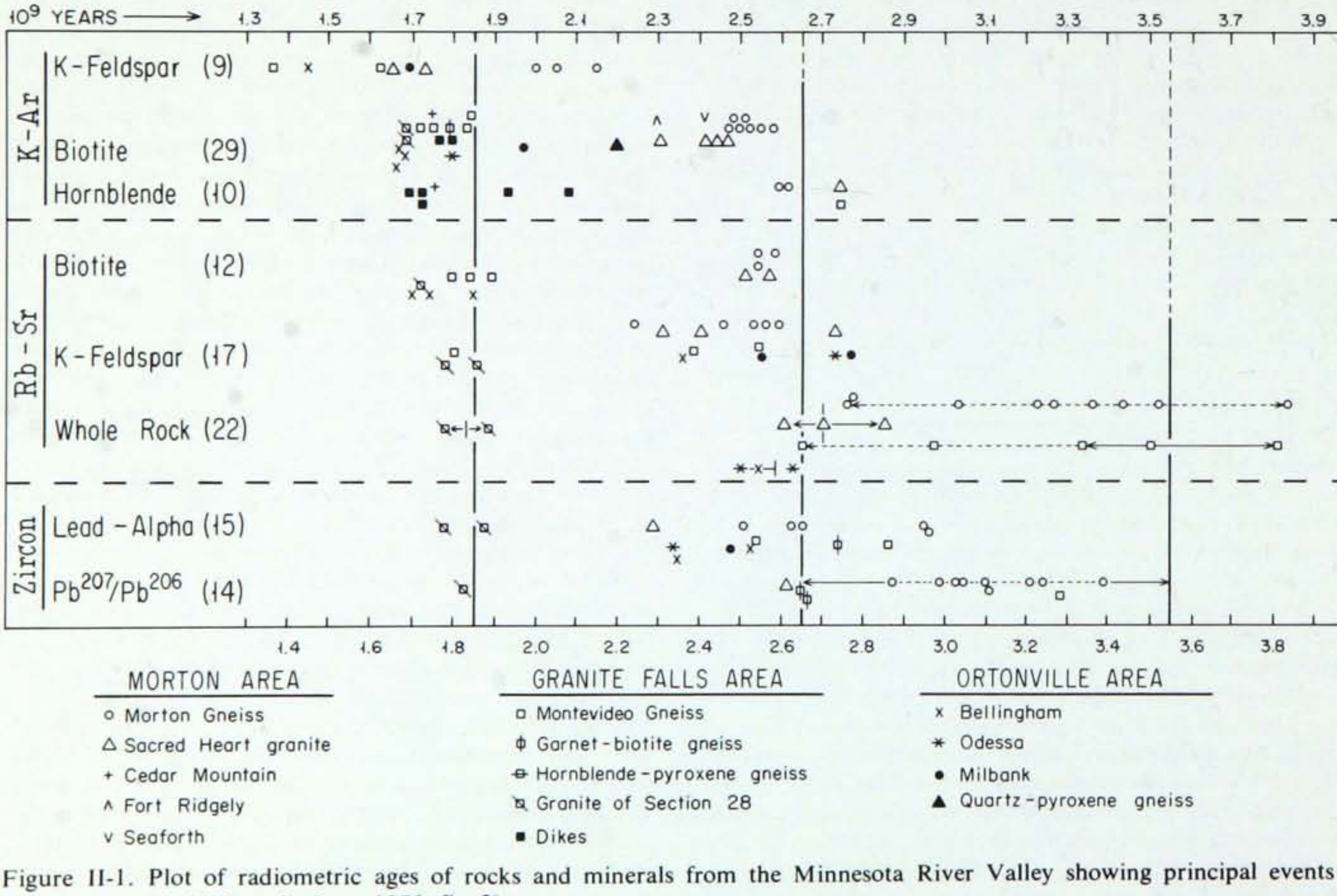

Figure II-1. Plot of radiometric ages of rocks and minerals from the Minnesota River Valley showing principal events (Goldich and others, 1970, fig. 2).

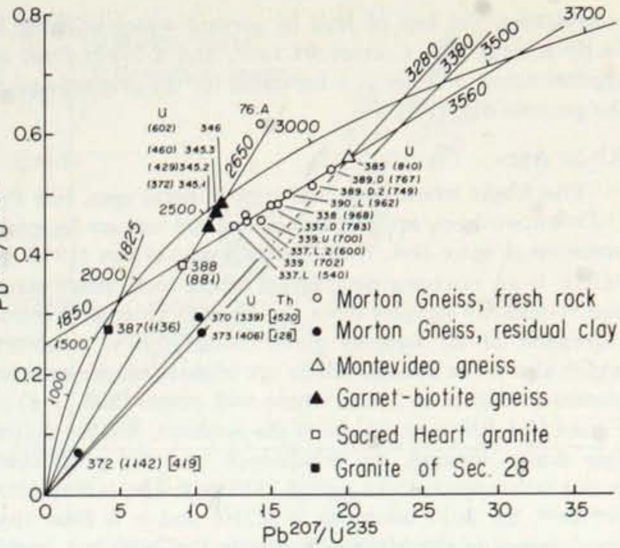

Figure II-2. Concordia diagram of zircon and allanite (76.A) samples. Contents of U and Th in ppm in brackets. The 1850-3650-m.y. chord is a linear regression of the zircon samples from the gneisses (Goldich and others, 1970, fig. 6).

on the Morton Gneiss also are included. Lines drawn from the origin through sample points intersect the concordia curve to give the Pb^{207}/Pb^{206} age which may be considered a minimum age. Sample No. 385 from the Montevideo gneiss is 3280 m.y. old, but if it is assumed that there has been no fractionation during the leaching of the lead from zircon sample 373 from the residual clay, a minimum age of 3380 m.y. is indicated for the Morton Gneiss.

In addition to the zircon samples from the gneisses, four samples from a garnet-biotite gneiss in Granite Falls and one sample from the Sacred Heart granite were analyzed, and these results have a significant bearing on the interpretation of the data from the gneisses. Sample 346, with nearly concordant ages, plots just below the concordia curve (fig. II-2), but three sized fractions of sample 345 are discordant and plot along a line which suggests recent lead loss. Sample 388 from the Sacred Heart granite, for which Rb-Sr whole-rock samples indicate an original age of approximately 2700 m.y., is more sharply discordant but also plots near the 2650-m.y. chord. These discordant ages, as well as those of sample 387, could be explained using Tilton's (1960) model which is based on loss of radiogenic lead by continuous diffusion. The discordance in the zircons from the gneiss, then, might be explained as the result of continuous diffusion combined with episodic lead loss. This possibility has been considered by Catanzaro (1963) and by Stern and others (1966).

The two-stage model (Goldich and others, 1970, p. 3681) to explain the age discordance of the zircons from the ancient gneisses combines episodic lead loss at 2650 m.y. ago as a result of a high-grade metamorphic event with a relatively recent loss of lead resulting from the dilatancy effects of uplift and erosion. Himmelberg and Phinney (1967) described the mineral assemblages of the metamorphic rocks near Granite Falls and Montevideo, and assigned them to the amphibolite facies and the hornblende-granulite and pyroxene-granulite subfacies. The assemblages are interlayered, and the presence or absence of hornblende is attributed to compositional variations (Himmelberg and Phinney, 1967, p. 331). Himmelberg (1968) concluded that the structure as well as metamorphic rock units, including the garnet-biotite gneiss (samples 345 and 346), resulted from a single period of dynamothermal metamorphism 2500 to 2700 m.y. ago. The newer radiometric data (Goldich and others, 1970) date this event at 2650 ± 50 m.y. ago.

The dilatancy model for discordant zircon ages was developed by Goldich and Mudrey (1969) to explain apparent episodic lead loss from zircon in areas where there is no known thermal event to account for the lower intercept. The lack of evidence for a younger thermal event in many areas in which discordant ages are found in part prompted Tilton (1960) to develop a model for lead loss by continuous diffusion. Goldich and Mudrey, however, rejected the continuous diffusion model, and pointed out that a principal cause of discordant zircon U-Pb ages is lead loss that occurs relatively late in the rock's history; hence, the Pb^{207}/Pb^{206} ages approach the true age, and the discordance closely resembles that produced by recent lead loss. As can be seen in Figure II-2, this is precisely the case in the zircons from the garnet-biotite gneiss (345, 346), the Sacred Heart granite (388), and the granite of section 28 (387).

The dilatancy model relates the loss of radiogenic lead to the loss of water from metamict zircon when quartz-bearing rocks are brought close to the surface through uplift and erosion. The radioactive decay of uranium and thorium produces radiation damage in the crystal structure of zircon with the development of crystallites and amorphous compounds. As has been well demonstrated by Silver (1962), the extent of the radiation damage depends on the original uranium and thorium contents and on the age of the zircon. It has long been known that water enters the microcapillary openings in metamict zircon, and Mumpton and Roy (1961) concluded that the water is largely molecular water that is strongly adsorbed within the metamict phases. Grünenfelder (1963) suggested that a positive correlation exists between the age discordance and the water content of zircon. Obviously a thermal event, even at low temperatures, would be effective in driving out water with loss of some radiogenic lead, producing age discordance. Goldich and Mudrey (1969), however, extended this concept to include the loss of water from metamict zircon as a result of uncovering by erosion. Crystalline rocks, especially those containing large amounts of quartz (R. H. Jahns, 1970, written comm.), undergo fracturing and rifting. Dilatancy with relief of pressure permits some of the water with dissolved radiogenic lead to escape from the metamict zircon.

The dilatancy model explains the age discordance of the zircon from the garnet-biotite gneiss, the Sacred Heart granite, and the granite of section 28. It should be noted that the zircon ages from the garnet-biotite gneiss and the Sacred Heart granite cannot be explained by a thermal event at 1850 m.y. If, however, the dilatancy model is used to explain the age discordance in these rocks, we should ex-

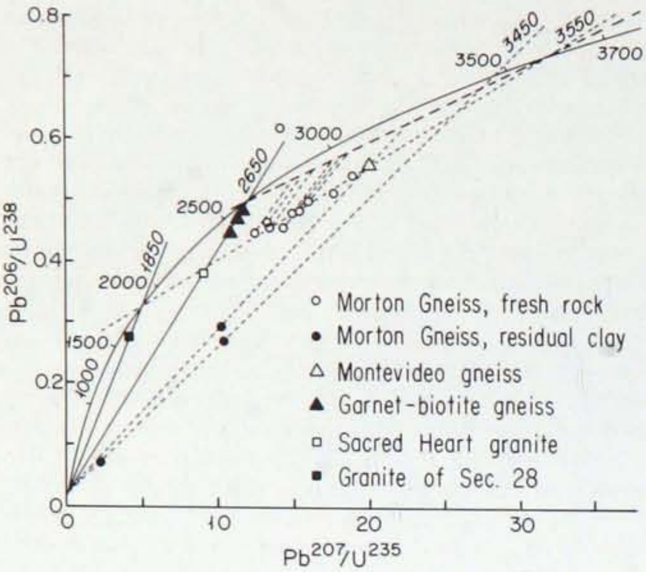

Figure II-3. Concordia diagram illustrating two-stage model. A primary age discordance (heavy chord) was produced in the zircon in the gneisses during the 2650-m.y. metamorphism. A secondary discordance in the gneisses and in the granites was developed through the dilatancy effects of uplift and erosion approximately 100 m.y. ago. Zircon from the residual clay shows additional loss of lead from ground water leaching. The dilatancy-model age for the gneisses is a minimum value of 3450 m.y.; for the Sacred Heart granite, it is 2650 m.y.; and for the granite of section 28, it is approximately 1850 m.y. (Goldich and others, 1970, fig. 7).

pect the zircons in the ancient gneisses to show a similar effect. The apparent linearity of the points (fig. II-2) is explained by Goldich and others (1970) as the result of the severe metamorphism at 2650 m.y. ago, and the intercept giving 1850 m.y. as the time of lead loss is a coincidence. A two-stage model to explain the age discordance of the zircon in the granitic gneiss involves an episodic event 2650 m.y. ago and a relatively recent loss of lead resulting from the dilatancy effects of uplift and erosion (fig. II-3).

Sedimentary rocks of Late Cretaceous age (Sloan, 1964) occur between New Ulm and Ortonville, and residual clay developed on the Morton Gneiss was incorporated in sediments near Morton (Goldich, 1938). Thus, during Cretaceous time the crystalline rocks in the Minnesota River Valley were brought close to the surface, and it is assumed that dilatancy occurred about 100 m.y. ago (fig. II-3). The heavy dashed chord in Figure II-3 is a reconstruction of the age discordance developed in the zircon of the granitic gneisses during the 2650-m.y. event which fixes the lower intercept. The upper intercept of 3550 m.y., giving the original age, cannot be rigidly demonstrated (Goldich and others, 1970), but there are limiting values of 3280 m.y. (sample 385) and 3380 m.y. (sample 373, fig. II-2). In the dilatancy model interpretation, the loss of lead by ground water leaching is ascribed largely to Cretaceous time, and a lower limit of approximately 3450 m.y. is indicated for the original age of the gneisses (fig. II-3).

Rb-Sr Ages

The Rb-Sr whole-rock and rock-mineral ages, like the U-Pb zircon ages, are discordant and are subject to interpretation. Figure II-4, from Goldich and others (1970, p. 3685), is an isochron diagram of whole-rock, plagioclase, and K-feldspar samples from the Morton Gneiss. A linear regression of all samples gives the 2550-m.y. isochron, which also is the average Rb-Sr age of three biotite samples plotted in Figure II-1. The whole-rock points (208, 338) in Figure II-4, however, lie above the isochron, and the dotted line drawn through the whole-rock and the plagioclase points corresponds to an age of 2650 m.y. The initial ratio, Sr^{87}/Sr^{86}, for both isochrons is 0.710, and it is clear that involvement of the Morton Gneiss in the 2650-m.y. metamorphic event caused a redistribution of radiogenic Sr^{87} among the mineral phases of the gneiss and, also, that the isotopic system in the minerals has not been closed since the 2650-m.y. event.

In Figure II-5, the 3800-m.y. isochron is a limiting line to which no special significance should be attached. An age of 3550 m.y. was computed for the Montevideo gneiss with an assumed initial Sr^{87}/Sr^{86} ratio of 0.700. This value is considered the best original age for the granitic gneisses. The subsequent history is complex and speculative, and the Rb-Sr data have been interpreted (Goldich and others, 1970) largely on the hypothesis of loss of Sr or gain of Rb during metamorphic events. In the Montevideo area, the gneiss consists of massive and foliated phases. Both rock

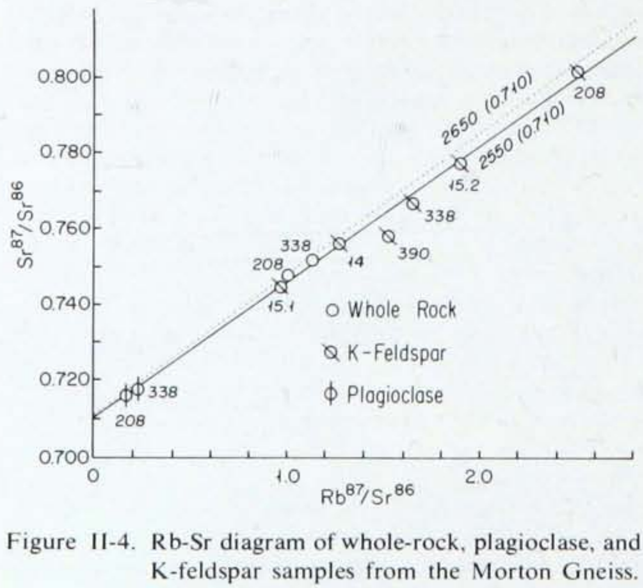

Figure II-4. Rb-Sr diagram of whole-rock, plagioclase, and K-feldspar samples from the Morton Gneiss. The 2550-m.y. isochron is a linear regression of all samples. The 2650-m.y. isochron was constructed for whole-rock and plagioclase samples (Goldich and others, 1970, fig. 8).

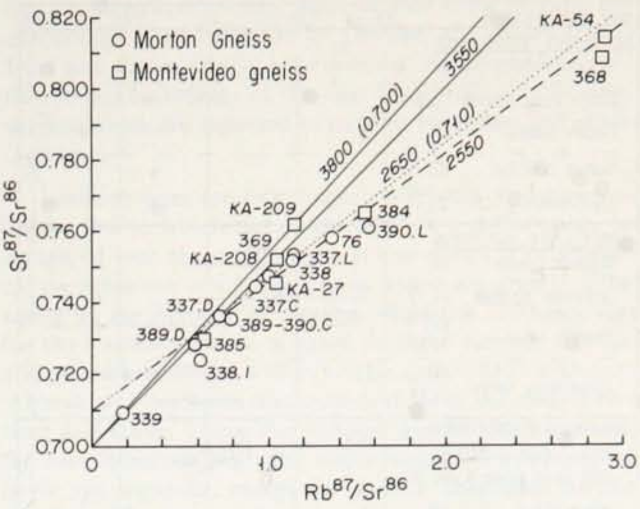

Figure II-5. Rb-Sr isochron diagram of whole-rock and feldspar samples from the Montevideo gneiss and whole-rock samples from the Morton Gneiss. Sample 337.C is a composite calculated from 337.D (dark-colored phase) and 337.L (light-colored), and 389-390.C is a composite of 389.D and 390.L. The 3550-m.y. isochron was constructed from the weighted average of the data for samples 385, 369, and KA-208. The 2550-m.y. and 2650-m.y. isochrons are from Figure II-4.

types were included in the sampling for the Rb-Sr study. Sample 368 (fig. II-5), representing the massive phase, gives a younger age than sample 369, the foliated biotite granite gneiss. Sample 384 from the same locality apparently is a mixture of both rock types. Re-examination of the outcrop indicates that the massive phase is younger and was intruded into the foliated phase.

2600-2700-m.y. Granites

Sacred Heart Granite

The main outcrop area of the Sacred Heart granite is in the Minnesota River Valley, 6 miles south of Sacred Heart. A small isolated outcrop, approximately 12 miles southwest, is the site of the Larsen quarry (Goldich and others, 1961, p. 130). The granite probably is of considerable extent. Age limits are set by the Pb^{207}/Pb^{206} age for zircon of 2605 m.y., with a model age of 2650 m.y. (fig. II-3) and by the K-Ar age of 2740 ± 140 m.y. on hornblende (fig. II-1) determined by Thomas (1963). The Rb-Sr isochron age (Goldich and others, 1970, p. 3687) is 2700 m.y., but the mineral ages indicate disturbance of both the Rb-Sr and K-Ar isotopic systems (fig. II-1).

Granites in the Ortonville Area

A reconnaissance survey (Goldich and others, 1970) of granites exposed in the upper part of the Minnesota River Valley in the vicinity of Odessa and Ortonville and in the Milbank area of South Dakota indicates that the granites probably are contemporaneous with the Sacred Heart granite. None of these rocks, however, has been studied in detail.

Granite Near Fort Ridgely

The Fort Ridgely granite of Lund (1956) is exposed in several small outcrops approximately 2.5 miles northwest of Fort Ridgely State Park. Preliminary radiometric age determinations (C. W. Keighin, 1972, written comm.) indicate that the porphyritic coarse-grained granite probably is 2600 to 2700 m.y. in age.

1800-m.y. Plutons

Several relatively small plutons in the Minnesota River Valley were emplaced approximately 1800 m.y. ago. The best dated of these is the granite of section 28, which cuts the Montevideo gneiss and a hornblende andesite dike northwest of Granite Falls. The Pb^{207}/Pb^{206} age for zircon is 1825 m.y. (Catanzaro, 1963), and the model age is 1850 m.y. (fig. II-3). The Rb-Sr isochron age is 1830 ± 160 m.y., but the mineral ages range from 1690 m.y. (K-Ar on biotite) to 1870 m.y. (Pb-α on zircon, fig. II-1); Goldich and others (1970, p. 3689) gave an age of 1850 ± 50 m.y. for the granitic pluton. A small gabbro-granophyre intrusive complex in the Morton Gneiss south of Franklin may be of similar age, as is suggested by K-Ar determinations on biotite (Goldich and others, 1961, p. 135) and on hornblende (Hanson, 1968, p. 5), both giving 1750 m.y.

Dike Rocks

Sill-like granitic masses in the Montevideo gneiss apparently are much younger than the gneiss and may be related to the 2650-m.y. high-grade metamorphic event. There are numerous aplitic to pegmatitic dikes in the Morton Gneiss which have not been dated, and these may have been formed around 2650 m.y. ago or possibly later. K-Ar age determinations (Hanson and Himmelberg, 1967; Hanson, 1968) on dike rocks range from 1690 to 2080 m.y. (fig. II-1). The range in ages may be interpreted as indicating different periods of intrusion or the resetting of the K-Ar isotopic system. If the K-Ar ages on hornblende are reliable in giving the time of intrusion, tholeiitic diabase dikes near Granite Falls were intruded more than 2000 m.y. ago, whereas the hornblende andesite dikes were intruded approximately 1800 m.y. ago. It should be recalled that the granitic pluton in section 28, northwest of Granite Falls, is dated at 1850 ± 50 m.y. and intrudes a hornblende andesite dike. The radiometric ages for the dike rocks probably should be considered minimum values.

Epeirogeny

The dilatancy model to explain discordant U-Pb ages of zircons is a relatively recent suggestion, but the possible effects of epeirogeny and cooling history on K-Ar mineral ages have been explored and discussed for some years. Harper (1967) has reviewed the contributions by a number of writers to the development of the concept that mica ages reflect the time of uncovering and of the stabilization of the K-Ar and Rb-Sr isotopic systems in micas. In this interpretation, mica ages do not date the time of original crystallization nor subsequent metamorphic events.

Goldich and others (1970, p. 3691) suggested that the biotite ages from the Minnesota River Valley are in a gross manner related to uplift and erosion of large blocks of the basement. The Rb-Sr and K-Ar systems were stabilized in biotite in the Morton Gneiss, the Sacred Heart granite, and in the granite near Fort Ridgely approximately 2500 m.y. ago, but final stabilization of the biotite in the Montevideo gneiss and in the granites of the Ortonville area was not achieved until approximately 1800 m.y. ago.

The apparent behavior of the Rb-Sr system in K-feldspar was found (Goldich and others, 1970, p. 3690) to be surprisingly similar to that of the K-Ar system in biotite (fig. II-1). Loss of radiogenic Sr^{87} from both biotite and K-feldspar occurred after the 2650-m.y. high-grade metamorphism. Some of the radiogenic Sr was trapped probably in secondary minerals such as epidote and sericite, but an appreciable part was lost; hence, the whole-rock samples do not represent a closed system and points on isochron diagrams have a considerable scatter.

In the Granite Falls area there is definite evidence that both the Rb-Sr and K-Ar systems in the Montevideo gneiss near the contact with the granite of section 28 were affected by the heat of the younger intrusion. The Rb-Sr age on K-feldspar from the gneiss near the contact is approximately 1800 m.y., and the K-Ar age on biotite from the same locality is 1840 m.y. Two K-feldspar samples (KA-27, KA-54) from the gneiss at other localities, however, do not show this effect and fit the 2550-m.y. isochron (fig. II-5). It might be concluded, then, that both the Morton and the Granite Falls areas were stabilized by uncovering approximately 100 m.y. after the 2650-m.y. metamorphic event, but the K-Ar and Rb-Sr biotite ages from the Montevideo gneiss and the granites in the Ortonville area are almost all in the range from 1650 to 1850 m.y. The interpretation that the mica ages are fundamentally uncovering or cooling ages is attractive because there is little evidence for widespread igneous activity at that time.

There are, however, some aspects of the mica ages that are not easily reconciled with the simple interpretation of cooling ages related to epeirogeny. The Rb-Sr age on biotite from the granite of section 28 is 1725 m.y. or roughly 100 m.y. less than the age of the granite, but the K-Ar age on the biotite is 1690 m.y., nearly 150 m.y. younger. Similarly, in the Sacred Heart granite, the Rb-Sr biotite ages are approximately 100 m.y. less than the age of the granite, and the K-Ar ages are approximately 200 m.y. less.

NORTHERN AND EAST-CENTRAL MINNESOTA

Early Precambrian

General Statement

The classical succession of Lower Precambrian rocks in northern Minnesota and adjacent parts of Ontario has been a main target of geochronological investigations. The principal results of recent studies are summarized in Figure II-6. The rock units (fig. II-6) in each area are arranged from youngest (top) to oldest (bottom) on the basis of field relationships. Two fundamental conclusions are derived:

(1) The events that can be interpreted from the geologic record, including the accumulation of a thick sequence

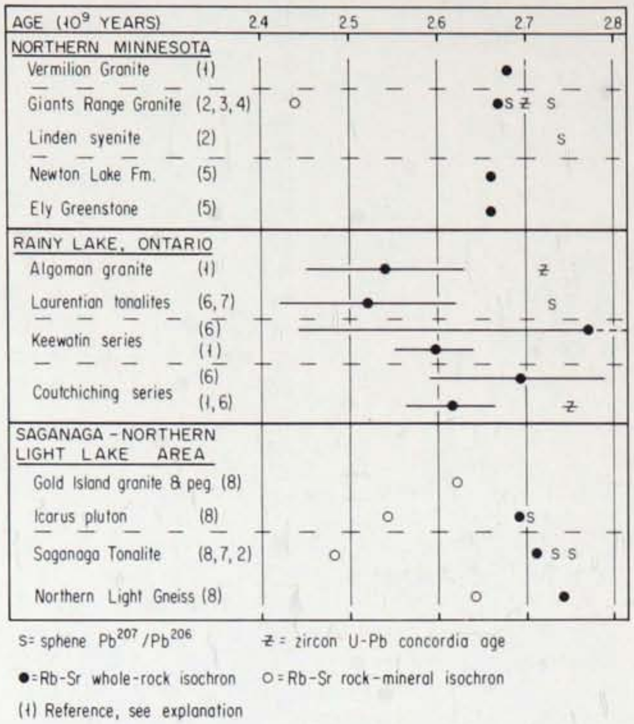

Figure II-6. U-Pb and Rb-Sr isochron ages for Lower Precambrian rocks from the Vermilion region, Giants Range, and Saganaga Lake area in Minnesota and adjacent parts of Ontario. (1) Peterman and others (in press), (2) Catanzaro and Hanson (1971), (3) Prince and Hanson (in press), (4) Hanson and others (1971a), (5) Jahn and Murthy (1971), (6) Hart and Davis (1969), (7) Tilton and Grünenfelder (1968), (8) Hanson and others (1971b).

of metasedimentary and metavolcanic rocks that was folded and intruded a number of times, all took place in a relatively short span of time from 2700 to 2750 m.y. ago.

(2) Whole-rock Rb-Sr isochrons, even with the use of the longer half-life, give lower ages than are found by U-Pb analyses of sphene or zircon.

Both of these conclusions have been more or less anticipated by others, particularly by Hart and Davis (1969) in their study of the Rainy Lake area, Ontario.

Vermilion Region

Samples collected along Echo Trail between Ely and Buyck define a whole-rock Rb-Sr isochron with an age of 2680 ± 95 m.y. and an initial Sr^{87}/Sr^{86} ratio of 0.7005 ± 0.0012 (Peterman and others, in press). The Vermilion Granite has been considered correlative with the Giants Range Granite, the type Algoman granite in Minnesota (Grout and others, 1951, p. 1039) and with the Algoman granites of Lawson (1913a) in the Rainy Lake district of Ontario adjoining on the north.

The classical Rainy Lake area, brought to world attention by Lawson (1913a) and well publicized through the

Coutchiching controversy, was included in the original geochronological investigation of Goldich and others (1961). Hart and Davis (1969) have made a significant contribution to our knowledge of the ages of the rocks, and current investigations are reported in part by Peterman and others (in press).

Isochron ages for four major rock units are shown in Figure II-6 in which the results of Hart and Davis can be compared with those of Peterman and others. The analytical uncertainties which have been added are greatly influenced by the number of samples. Thus the 2770-m.y. age for the Keewatin Series is based on three samples and has a large uncertainty of 330 m.y. The age of 2595±45 m.y. is based on 11 samples that include three of the analyses by Hart and Davis. Using the younger age for the Keewatin, the determinations place the major rock units in proper relative age sequence, except for a small aberration for the tonalites.

The isochron ages, however, cannot be the time of deposition of the Coutchiching, extravasation of the Keewatin volcanics, intrusion of the tonalites, or crystallization of the Algoman granites. A minimum age from U-Pb analyses for all the rocks is 2700 m.y. Tilton and Grünenfelder (1968) found nearly concordant ages on sphene with a Pb^{207}/Pb^{206} age of 2730 m.y. Hart and Davis (1969) reported a large number of U-Pb analyses of zircon from the Coutchiching, Keewatin, and Laurentian rocks which they treated as a single population using a modified time-dependent version (Wasserburg, 1963) of the continuous-diffusion model to arrive at an original age of 2750 m.y. The dilatancy model applied to the zircon from the Rainy Lake district gives an age of 2720 m.y. Four zircon concentrates from the Algoman granites, analyzed by T. W. Stern of the U.S. Geological Survey, fit the pattern of the zircon from the older units.

The isochron ages suggest a period of metamorphism at approximately 2550 m.y. ago. This is the isochron age of the Algoman granites, and it was first thought that the metamorphism was related to the magmatic activity. The Algoman granite isochron, however, although analytically precise, is fallacious. It dates neither the time of intrusion nor a thermal event of sufficient intensity to affect a redistribution of Sr or Rb in the rocks. Analyses of minerals reveal discrepancies that are difficult to explain, and the Rb-Sr ages on micas (Peterman and others, in press) are not concordant with the isochron age; biotite ages are younger (2150 and 2350 m.y.), and muscovite ages are older (2600, 2630, 2670 m.y.). Rb-Sr biotite ages in the Rainy Lake district range down to 2350 m.y. which, with two exceptions, is also the lower limit for K-Ar ages.

Peterman and others (in press) considered some mechanisms to explain the apparent conflict between U-Pb, Rb-Sr whole-rock, and K-Ar and Rb-Sr mineral ages. One of these involves low-grade metamorphism or hydrothermal alteration as first suggested by Brooks (1968). Brooks showed that rubidium and strontium migrated between K-feldspar and alteration products of plagioclase. A partially open system under these conditions will result in a low isochron age.

Giants Range

Prince and Hanson (in press) analyzed seven whole-rock samples of quartz monzonite from the central part of the Giants Range Granite and obtained an isochron age of 2670±65 m.y. with an initial Sr^{87}/Sr^{86} ratio of 0.7002±0.0019. This Rb-Sr age is similar to that found for the Vermilion Granite and supports the assignment of both to one period of magmatic activity, the Algoman. Two sphene samples from the Giants Range Granite were analyzed by Catanzaro and Hanson (1971). The Pb^{207}/Pb^{206} ages of 2680 and 2730 m.y. are plotted in Figure II-6. The first sample is concordant, but the second is sharply discordant.

Prince and Hanson (in press) also attempted to date the Linden Syenite, a small stock just north of the Giants Range Granite, approximately 30 miles north of Hibbing and 3 miles west of State Highway 73. The Linden stock is an aegirine-augite syenite, and the limited range in composition of samples that could be readily obtained did not permit a precise isochron age to be determined. The Rb-Sr data, however, are in approximate agreement with the U-Pb ages on sphene (Catanzaro and Hanson, 1971). The Pb^{207}/Pb^{206} age of 2740 m.y. is plotted in Figure II-6.

The mineral ages from the Giants Range Granite show the effects of low-temperature alteration since 2700 m.y. ago. The isochron age for a whole-rock sample and its component minerals is 2440 m.y. K-Ar ages on hornblende range from 2570 to 2660 m.y.; on biotite, from 2300 to 2650 m.y. Prince and Hanson (in press) suggested that the younger ages may reflect one or more low-temperature events, such as low-grade regional metamorphism at 1600 m.y. ago (Hanson and Malhotra, 1971).

Additional information on the age of the Giants Range Granite was presented in a paper by Hanson and others (1971a). They considered the effects of the intrusion of the 1100-m.y.-old Duluth Complex on the U-Pb ages for sphene and zircon and on the K-Ar ages for hornblende and biotite in the Giants Range Granite. The age of the granite, on the basis of the studies reviewed above, is taken as 2700 m.y. Both the sphene and the zircon samples gave discordant ages with the discordance in the zircons considerably greater than in the sphene. Four samples of sphene define a chord intersecting the concordia curve at 2700 m.y. and 1100 m.y., the latter being the time of the intrusion of the gabbro producing the lead loss and the discordance. Samples of zircon included in the concordia diagram fit the chord in its upper part, but fall below the chord with increasing discordance. These samples suggest a combination of episodic lead loss at 1100 m.y. ago followed by additional lead loss, a dilatancy effect, approximately 500 to 600 m.y. ago.

The results of two additional isochron studies by Jahn and Murthy (1971) are plotted in Figure II-6 and represent whole-rock samples from the Ely Greenstone and the Newton Lake Formation in the Vermilion district. The original ages of 2725 and 2720 m.y. for the Ely and the Newton Lake formations, respectively, have been refined by regression analyses to 2660 m.y. (Jahn and Murthy, 1972, written comm.). These ages are essentially the same as the Rb-Sr values for the Giants Range Granite and the Vermilion Granite (fig. II-6). Geological interpretations, however, require that the granites be younger and intrusive into the Ely Greenstone and the Newton Lake Formation.

Saganaga Lake Area

This area along the Minnesota-Ontario boundary has figured prominently in developing the succession and in deciphering the history of the Precambrian (Grout and others, 1951). For a summary of the geology see Hanson (this volume), Hanson and others (1971b), Goldich and others (in press), and Hanson and Goldich (in press).

Whole-rock Rb-Sr ages for the Saganaga Tonalite and the Icarus pluton are slightly less than the U-Pb ages on sphene (fig. II-6). The sphene ages by Catanzaro and Hanson (1971) are concordant, with a Pb^{207}/Pb^{206} age of 2750 m.y. An earlier analysis of sphene by Tilton and Grünenfelder (1968) gave somewhat discordant ages, with a Pb^{207}/Pb^{206} age of 2730 m.y. Three zircon samples from the tonalite were analyzed by Anderson (1965, unpub. Ph.D. thesis, Univ. Minn.) and one by Hart and Davis (1969, sample no. 43); all have sharply discordant ages. These data do not allow a unique model-age interpretation but indicate an original age of approximately 2700 m.y., and for this reason the zircon age is not shown in Figure II-6.

The rock-mineral isochron ages (fig. II-6) are considerably less than the corresponding ages for the whole-rock samples and range from 2480 to 2640 m.y. The lower ages cannot be explained as the result of a thermal event that redistributed the radiogenic Sr^{87} among the mineral phases. Hanson and others (1971b) concluded that low-temperature alteration accompanying or following shearing affected the isotope systems in the mineral phases. The Archean massif consisting of the Northern Light Gneiss and the Saganaga Tonalite is bounded on the north and south by faults along which it continued to rise after crystallization of the tonalite. Within a very short period the tonalite was uncovered by erosion and contributed boulders and cobbles to conglomerate beds in the Knife Lake Group in Cache Bay just west of the Saganaga batholith. K-Ar mineral ages range from 2760 m.y. on hornblende from the Saganaga Tonalite (Hanson, 1968) to 2430 m.y. on sericite from a pegmatite (Goldich and others, 1961). The range in mineral ages cannot be explained in terms of an event. In part, they may represent an epeirogenic effect, marking the time of stabilization of the K-Ar and Rb-Sr systems, but they may also reflect low-temperature hydrothermal alteration, local effects of faulting and shearing, or recent incipient weathering.

Algoman Orogeny

Two periods of folding accompanied by igneous and metamorphic activity were early recognized by geologists in Minnesota and were designated the Laurentian and Algoman orogenies, respectively. Erosion surfaces developed following each period of folding were used to separate major rock sequences (Grout and others, 1951). The geological basis for recognition and differentiation of the two orogenies is clear-cut both in the Saganaga Lake area and in the Rainy Lake district.

Goldich and others (1961, p. 154) were unable to differentiate the Laurentian and Algoman orogenies on the basis of K-Ar and Rb-Sr mica ages and concluded that the time interval was relatively short, favoring "the interpretation of the Laurentian folding as an early phase of the greater Algoman orogeny." On the basis of the mica ages, broad limits from 2400 to 2600 m.y. were placed on the Algoman orogeny, and the time division between Early and Middle Precambrian was set at 2500 m.y. This arbitrary time was selected with due regard to the consideration that metamorphic and igneous rocks related to the orogeny might provide samples amenable to radiometric dating techniques.

The recent studies summarized in Figure II-6 suggest that all Early Precambrian events probably took place during an interval of 50 m.y. from 2700 to 2750 m.y. ago. The data do not permit differentiating the Laurentian and Algoman orogenies or of setting close limits on the probable duration of the Algoman orogeny. The revised figure of 2600 m.y. for the time division between Early and Middle Precambrian (Goldich, 1968) remains useful as an arbitrary time division in Minnesota where the major orogenic event in the northern part of the state occurred 2700 m.y. ago but in the southwestern part appears to have been 2650 m.y. ago. There are similar variations throughout Precambrian time that argue strongly against using the time of an orogeny as a time division in a world-wide Precambrian time scale.

Middle Precambrian

General Statement

An important contribution of K-Ar dating in Minnesota was the recognition (Goldich and others, 1957) that the rocks of the Animikie Group are separated from the Keweenawan rocks by a major rather than a minor unconformity (Grout and others, 1951) and that there is a great difference in the age of these rocks. The Animikie Group in Minnesota and similar iron-bearing sequences in Wisconsin and Michigan were shown to have been folded and metamorphosed during an orogenic period which Goldich and others (1961) called the Penokean and dated at 1600 to 1800 m.y. ago.

Within recent years a considerable effort has been made to apply the Rb-Sr whole-rock technique to rocks of the Animikie Group in Minnesota (Peterman, 1966) and in the type locality of the Thunder Bay area, Ontario (Faure and Kovach, 1969). A Rb-Sr isochron age of 2155 ± 80 m.y. for the Nipissing Diabase (Van Schmus, 1965) placed a minimum age on the folding of the type Huronian rocks in the Blind River-Bruce mine area, Ontario. A similar age of 2162 ± 67 m.y. for the Nipissing Diabase and an age of 2288 ± 87 m.y. for the Gowganda Formation at Gowganda were reported by Fairbairn and others (1969). Volcanic and metasedimentary rocks from the Belcher Fold Belt in Hudson Bay, and the Labrador Trough and the Mistassini Lake area of Quebec have been dated by Rb-Sr whole-rock isochrons in the range from 1790 to 1870 m.y. by Fryer (1971).

Rb-Sr Whole-Rock Ages

The data from whole-rock studies of Middle Precambrian iron-bearing sequences that formerly were commonly referred to as Huronian iron-formations are summarized in Figure II-7. The plot suggests that the rocks might be di-

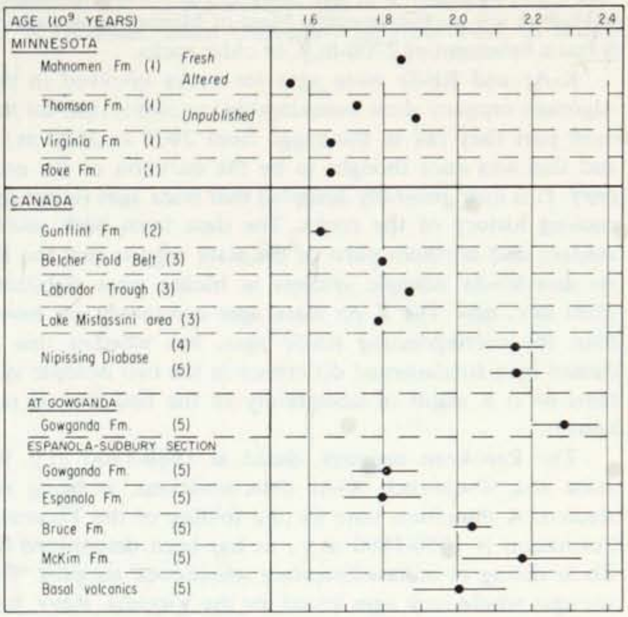

Figure II-7. Rb-Sr isochron ages for Middle Precambrian formations of Minnesota and Proterozoic rocks of Canada. (1) Peterman (1966), (2) Faure and Kovach (1969), (3) Fryer (1971), (4) Van Schmus (1965), (5) Fairbairn and others (1969). Bars represent errors quoted by the authors.

visible into two major groups, one older and one younger than 2000 m.y. This interpretation, however, cannot be accepted at present without serious reservations.

Peterman's (1966) isochron age of 1850 m.y. for the metasedimentary rocks of the Cuyuna district is supported by unpublished data for the Thomson Formation, and 1850 to 1890 m.y. (fig. II-7) may be taken as a minimum age for the time of folding of these rocks. The time of deposition is for the present unknown but may be much older than 1850 m.y. ago. The age measurements by Fairbairn and others (1969) are instructive in this matter. The Gowganda Formation in the type locality is dated at 2288 m.y., but in the Espanola-Sudbury area, at 1810 m.y. (fig. II-7). The latter age, 478 m.y. less, obviously must reflect younger events. Similarly, we can accept the older age of 2288 m.y. only as a minimum value for the Gowganda Formation; it may be older than the upper limit of 2374 m.y. given in the analytical error, not for analytical reasons, but because the Rb-Sr isotopic system in rocks, as demonstrated in the section on Lower Precambrian rocks, commonly has not been a closed system.

The geological correlation of the Rove, Virginia, and Thomson Formations in Minnesota is on good ground, and it is illogical to assign very different ages to these rocks. Hence, as Peterman (1966) concluded, the age of 1660 m.y. for the Virginia and Rove Formations is much too low, and the interpretation by Faure and Kovach (1969, p. 1725) that their isochron age of 1635 m.y. represents "approximately the time of deposition and diagenesis of the Animikie Series in the Thunder Bay district" is untenable. The low age may have resulted from low-grade metamorphism or may be related in some way to the Keweenawan igneous activity, 1100 m.y. ago (Hanson and others, 1971b, p. 1121).

Penokean Orogeny

The Penokean orogeny, as first defined (Goldich and others, 1961), applies to the time of folding and metamorphism of the Middle Precambrian rocks in Minnesota, Wisconsin, and Michigan. On the basis of the K-Ar and Rb-Sr ages, the limits for the event were given as 1600 to 1800 m.y. ago. Peterman's (1966) work indicates that the Penokean event is older than 1850 m.y., and work now in progress may further clarify the events of Middle Precambrian time (1800 to 2600 m.y.).

Late Precambrian

General Statement

The principal objective of radiometric dating in the Late Precambrian has been the volcanic and intrusive rocks formerly assigned to the Middle Keweenawan (Grout and others, 1951). These rocks include the North Shore Volcanic Group (Goldich and others, 1961, p. 81), the Duluth Complex (Taylor, 1956, p. 42; 1964), and numerous smaller intrusions such as the Logan sills (Lawson, 1893), the layered complex at Beaver Bay (Gehman, 1957, unpub. Ph.D. thesis, Univ. Minn.), the Endion, Northland and Lester River Sills at Duluth (Schwartz and Sandberg, 1940), and a variety of dikes. Goldich and others (1961, p. 161) dated the Duluth Complex and related intrusions along the north shore of Lake Superior at approximately 1.1 b.y. ago, the time of the Grenville orogeny in eastern North America. The name "Keweenawan igneous activity" was used by Goldich and others (1966, p. 5404) and tentatively assigned time limits from 1000 to 1200 m.y. ago.

Keweenawan Igneous Activity

A summary of radiometric ages on Keweenawan rocks is given in Figure II-8. The U-Pb zircon analyses (Silver and Green, 1963) represent a rhyolite from the North Shore Volcanic Group, granophyre from the Endion sill, and ferrogranodiorite and granophyre intruded into gabbro at Duluth, and granite near Mellen, Wisconsin. The linear pattern of the zircons led Silver and Green to conclude that all the rocks have an age of 1115 ± 15 m.y., indicating "a sharp pulse of igneous activity."

Seven samples from the Duluth Complex at Duluth and four from the Endion sill, analyzed by Faure and others (1969), gave isochron ages of 1115 ± 14 and 1092 ± 15 m.y., respectively. The quoted errors represent one standard de-

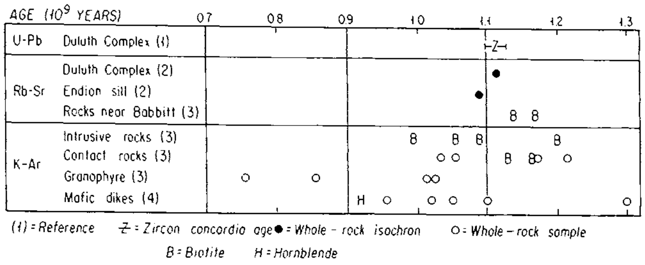

Figure II-8. Radiometric ages for rocks assigned to Keweenawan igneous activity. (1) Silver and Green (1963), (2) Faure and others (1969), (3) Goldich and others (1961), (4) Hanson and Malhotra (1971).

viation. Two Rb-Sr ages (Goldich and others, 1961) on biotite from gabbro and pegmatite (Gundersen and Schwartz, 1962, p. 71-75) near Babbitt are somewhat older (1140 and 1170 m.y.), and K-Ar ages on biotite range from 995 to 1200 m.y. (fig. II-8). K-Ar ages on biotite and on whole-rock samples of a variety of rock units near the gabbro range from 1035 to 1215 m.y. K-Ar age determinations on four whole-rock samples of granophyres from Duluth show a considerable spread from 755 to 1025 m.y. reflecting varying losses of radiogenic argon. The analytical data for the granophyre samples given by Goldich and others (1961, p. 54) are reported correctly, but the ages (900 to 1200) were improperly calculated and are erroneous.

Hanson and Malhotra (1971) reported K-Ar mineral and whole-rock ages for mafic dikes and one sill that intrude Lower and Middle Precambrian rocks in northeastern Minnesota and in adjacent areas in Ontario. The samples with K-Ar ages greater than 1300 m.y. were all found to be extensively altered by low-grade metamorphism (prehnite-pumpellyite facies), but the rocks with ages of 1300 m.y. or less do not show this type or extent of alteration. Hanson and Malhotra suggested low-grade regional metamorphism approximately 1600 m.y. ago. The 1300 m.y. age (fig. II-8) was found for a whole-rock sample of a sill in the Rove Formation near Suomi, Ontario. The K-Ar and Rb-Sr ages (fig. II-8) suggest a greater time span for the Keweenawan igneous activity than that suggested by Silver and Green on the basis of the U-Pb zircon data.

SUMMARY AND CONCLUDING REMARKS

Four major events are recorded in the Precambrian rocks of Minnesota. The granitic gneisses of Minnesota are among the oldest rocks known on earth and date back to 3550 m.y. ago. They have been considerably altered by subsequent metamorphic and magmatic activity, and the measurements of the disturbed U-Pb and Rb-Sr isotopic systems in the gneisses are subject to some interpretation. Additional studies should resolve some of the uncertainties, and hopefully shed some light on the geologic history between 3550 m.y. and 2650 m.y.

The interval from 2750 to 2650 m.y. was one of great metamorphic and igneous activity. The last major event was the intrusion of the granitic rocks of the Giants Range and Vermilion complexes, Algoman granites, 2700 m.y. ago. This was also the time of high-grade metamorphism in the Minnesota River Valley accompanied by emplacement of large amounts of granite. The McGrath Gneiss in east-central Minnesota, assigned to the Penokean orogeny on the basis of mica ages (Goldich and others, 1961) is now known to be Algoman, approximately 2700 m.y. old (unpublished age measurements). Most of Minnesota apparently has a basement of 2700-m.y. or older rocks.

K-Ar and Rb-Sr mica ages for rocks involved in the Algoman orogeny show considerable variability, but for the most part they fall in the range from 2400 to 2600 m.y., and this was once thought to be the duration of the orogeny. It is now generally accepted that mica ages reflect the cooling history of the rocks. The data from both southwestern and northern parts of the state suggest that the K-Ar and Rb-Sr isotopic systems in biotite were stabilized 2500 m.y. ago. The K-Ar mica ages commonly are lower than the corresponding Rb-Sr ages, but whether this is caused by a fundamental difference in the two isotopic systems or is a result of uncertainty in the half-lives is not known.

The Penokean orogeny, dated at 1680-1800 m.y. by mica and whole-rock K-Ar determinations, is being restudied. A minimum time for the folding of the Thomson Formation is 1850-1890 m.y., as has been determined by Rb-Sr dating of metasedimentary whole-rock samples. The younger whole-rock ages found for the Virginia, Rove, and Gunflint formations, however, indicate that these ages cannot be accepted at face value to represent either the time of deposition, early diagenesis, or time of original folding. The structure of the McGrath Gneiss was formed by cataclasis and recrystallization during the Penokean event. The time of shearing is now being investigated.

The gabbro at Duluth is well dated; however, K-Ar and Rb-Sr ages for other Keweenawan igneous rocks appear to be older than the U-Pb age of 1115 ± 15 m.y. found by Silver and Green (1963). If these older ages are shown to be reliable, further U-Pb investigations are needed. Most of the dated zircons come from localities that are relatively close to the gabbro intrusions at Duluth, and the U-Pb ages from some of the localities may have been reset by the gabbro.

The considerable progress that has been made in geochronology in Minnesota, in large part, resulted from the significant number of U-Pb analyses of zircon and sphene. The U-Pb method appears to be the most precise and useful of the various techniques in establishing the original age. The interpretation of discordant zircon ages has been shown by the work on the Morton Gneiss and related rocks to be complex. Similarly, complications in whole-rock Rb-Sr isochron studies have been revealed. The complications, both in the U-Pb and in the Rb-Sr isotopic systems, are the results of geological processes that are now being studied more carefully, such as the effects of low-temperature metamorphism, hydrothermal alteration, and weathering, as well as the effects of uplift and erosion. For this reason there is need for the maximum precision attainable with each analytical method. Only through the use of all available dating methods closely coupled with geological investigations are the maximum benefits to be realized.

ACKNOWLEDGMENTS

Contributors to the geochronology of Minnesota have been identified through citation of their publications, and I hope there are few omissions through oversight on my part. It is a pleasure to acknowledge the assistance received over the years from C. E. Hedge, Marcia Newell, Z. E. Peterman, and T. W. Stern of the U.S. Geological Survey; G. N. Hanson, State University of New York at Stony Brook; and G. B. Morey, P. K. Sims, and D. H. Yardley of the University of Minnesota. Many others have contributed in some way or other, and their assistance and cooperation are appreciated. Financial support through a number of grants from the National Science Foundation has been a major factor; current research is supported by National Science Foundation Grant No. 12316.

Chapter III

EARLY PRECAMBRIAN

NORTHERN MINNESOTA, GENERAL GEOLOGIC FEATURES

P. K. Sims

"The early pre-Cambrian, or 'basement complex' . . . is still a *terra incognita* and is the frontier of stratigraphic and historical geology. . . ."

Francis J. Pettijohn,
Early pre-Cambrian geology and correlational problems of the northern subprovince of the Lake Superior region: 1937, Geol. Soc. America Bull., v. 48, p. 153-202.

Rocks north of the Mesabi range in Minnesota are Early Precambrian, or in Canadian terminology, Archean in age. They comprise the Minnesota segment of the Superior Province of the Canadian Shield (see fig. I-1), and are remarkably similar to those rocks in better exposed parts of the shield. They consist mainly of belts of metavolcanic and metasedimentary rocks, which constitute heterogeneous volcanic complexes, and of enclosing granitic rocks. Granitic gneisses of diverse origin occur locally in intervening areas. Because of the characteristic greenschist-facies metamorphism, the volcanic complexes commonly are referred to as greenstone belts.

STRATIGRAPHIC FRAMEWORK

Repeated patterns of rock distribution and structural habit in the Superior Province provide a framework for interpreting events during Early Precambrian time. In general, the oldest exposed rocks at any particular locality are metabasalt and related synvolcanic intrusive bodies. These rocks attain thicknesses in some greenstone belts of as much as 25,000 or 30,000 feet. Many of the lavas are pillowed, indicating deposition in water. Lenses of banded iron-formation occur within the dominant volcanic successions, particularly in the upper parts. The mafic volcanic rocks in many areas give way rather abruptly upward to felsic volcanics, including abundant pyroclastic deposits. These rocks grade upward and laterally into graywacke-type metasedimentary rocks, and complex interfingering of volcanic and sedimentary rocks, particularly at the interface, is common. The felsic volcanic successions tend to be thinner than the mafic successions, commonly being in the range 5,000-10,000 feet, whereas the sedimentary successions attain thicknesses of as much as several thousand feet. As pointed out by Goodwin (1968a) and by others (see McGlynn, 1970), the volcanic complex in any particular area may consist of a single mafic to felsic sequence, or one or more of the assemblages may be repeated. The resulting assemblages can be quite complex.

The volcanic rocks in the greenstone belts of the Superior Province (Wilson and others, 1965; Goodwin, 1968b; Baragar and Goodwin, 1969; Clifford and McNutt, 1970; Green, 1970a; Irvine and Baragar, 1971) are chemically similar to the orogenic calc-alkaline trend and the tholeiitic trend, and differ markedly from other petrogenic series such as the spilite-keratophyre, the oceanic alkali basalt-trachyte, or the tholeiite-granophyre associations. They contain less K_2O and TiO_2 than most younger volcanic rocks; true rhyolite is rare.

The sources for specific accumulations of graywacke and associated clastic rocks in the volcanic complexes of the Canadian Shield are not generally known. Some successions of graywacke clearly are composed dominantly of volcanic minerals and rock fragments (Ojakangas, 1972; Ayres, 1969b; Weber, 1970), whereas others (Goodwin, 1968b, p. 4; Donaldson and Jackson, 1965; Walker and Pettijohn, 1971) contain substantial amounts of plutonic detritus. These observations of contrasting provenance for graywacke sandstones are in accord with younger orogenic regions, as, for example, recently evolved island arcs and circum-Pacific continental borderlands (Dickinson, 1970), which are possible analogs with the Lower Precambrian complexes.

COUTCHICHING PROBLEM

In two parts of the Superior Province, quartzofeldspathic metasedimentary successions of large areal extent have been interpreted to underlie major mafic volcanic successions (see Goodwin, 1968a, p. 5), and thus are inferred to be older "basement" rocks on which the volcanic rocks were deposited. These are the Pontiac Group of northwestern Quebec (see Kalliokoski, 1968, p. 1203) and the Coutchiching Series of northwestern Ontario. The Coutchiching is of particular interest to this volume because of the great controversy at and after the turn of the century between A. C. Lawson and F. F. Grout, and subsequent attempts to resolve the problem (Goldich and others, 1961; Hart and Davis, 1969; Ojakangas, this chapter).

Lawson (1888) gave the name Coutchiching to metasedimentary rocks in the Rainy Lake region, along the International boundary, which he interpreted to lie beneath mafic metavolcanic rocks. He correlated the metavolcanic rocks with similar rocks in the Lake of the Woods region, which he had mapped earlier and given the name Keewatin, for exposures in the Keewatin district, Ontario (table III-1). Later, a review of the geology of the region by a committee of Canadian and American geologists (Adams and others, 1905) discredited the designation of Coutchiching in a part of the mapped area—in the vicinity of Shoal Lake, Ontario and Rat Root Bay, Minnesota. It was shown in these areas that the metasedimentary rocks mapped by Lawson as Coutchiching are stratigraphically above a conglomerate of presumed wide areal extent that lies on the Bad Vermilion Granite (Shoal Lake area) and on schists equated with the Keewatin. Later, Lawson returned to the area to make a more detailed geologic map, which was published in 1913. He concluded that part of his earlier mapping was in error,

Table III-1. Sequence and classification of the Precambrian rocks of the Lake Superior region (after Lawson, 1913b).

Upper Cambrian
unconformity

Algonkian
- Keweenawan (Nipigon)
- unconformity
- Animikie

Eparchean Interval

Archean
- Algoman (granite-gneiss, batholithic in Huronian)
- Intrusive contact
- Huronian
  - Upper
  - unconformity
  - Lower
- unconformity
- Laurentian (granite-gneiss, batholithic in Ontarian)
- Intrusive contact
- Ontarian
  - Keewatin
  - Coutchiching

and he redefined the conglomerate and schist in the Shoal Lake area as the Seine series (Lawson, 1913a). He reaffirmed his earlier view, however, that a metasedimentary succession existed locally below the greenstone that he called Keewatin, and he retained the name Coutchiching for these strata. Subsequently, the geology of the Rainy Lake area was reviewed and studied by several geologists. Bruce (1925) and Tanton (1927) concluded that the Coutchiching was a valid formation underlying the Keewatin volcanic rocks, whereas Grout (1925a) and Merritt (1934) advanced contrary views. Grout agreed that a metasedimentary succession existed below greenstone in the Rice Bay area of the Rainy Lake district, but he argued that these rocks were insignificant quantitatively and did not deserve formational status. Grout and others (1951, p. 1020), therefore, did not recognize Coutchiching as a formal stratigraphic unit in Minnesota (see table II-2). Subsequent mapping in the Rice Bay area of the Rainy Lake district by S. S. Goldich and students (summarized in Goldich and others, 1961, p. 38-39), has confirmed the presence of a succession of metasedimentary rocks—actually several thousand feet thick—beneath the metavolcanic rocks that Lawson designated as Keewatin. Thus, the Coutchiching is a valid stratigraphic term in the type area. In their review of the Precambrian geology of Minnesota, Goldich and others (1961, p. 5) suggested the possible presence of Coutchiching rocks in Minnesota (see table II-3).

The Coutchiching controversy has stemmed from the assumption that all mafic volcanic rocks (greenstone) in Lower Precambrian (Archean) terranes have approximately the same geologic age, a common belief at the turn of the century. In a thoughtful paper, Pettijohn (1937) first questioned this philosophy, and pointed out that greenstone bodies in southern Ontario may indeed be of at least two ages. Recent careful mapping in several greenstone belts (Goodwin, 1962; Weber, 1970; Morey and others, 1970; Green, 1970a) has confirmed that mafic volcanism is repeated in the volcanic complexes of Early Precambrian age and that, not uncommonly, both mafic volcanic and sedimentary rocks are repeated in time. Thus, the Coutchiching problem can now be viewed from a different perspective. The fundamental question is whether the Coutchiching, in the Rainy Lake area, represents a "basement" of deformed and metamorphosed older rocks on which volcanic rocks were deposited, or represents metasedimentary rocks intercalated with metavolcanic rocks within a single volcanic-sedimentary cycle. In a recent study of zircon and whole-rock ages in the Rainy Lake area, Hart and Davis (1969) proposed a model (p. 610) whereby "The Coutchiching becomes a sedimentary sequence not at the base of a volcanic series but somewhere in a thick volcanic pile." The available isotopic age data from the area indicate that accumulation of the metavolcanic-metasedimentary assemblages and intrusion of the granites that cut them probably took place within a time span of no more than 100 million years. The interpretation of Hart and Davis is consistent with observations from other areas in the shield, including the Vermilion district, and should be tested further.

With one possible exception, a base for the major mafic metavolcanic successions has not been recognized in northern Minnesota. The lower parts of the volcanic complexes are intruded by and locally are cut out by granitic rocks, commonly of batholithic dimensions. The possible exception is northwest of Chisholm, in T. 59 N., Rs. 21 and 22 W. (Sims and others, 1970). In this area, a migmatitic biotite schist and amphibolite unit apparently underlies a large, northwest-trending body of mafic metavolcanic rocks. Judged from pillow tops near the base of the metavolcanic unit, which consistently face northeastward (S. Viswanathan, 1971, unpub. Ph.D. thesis, Univ. Minn.), the metavolcanic unit stratigraphically overlies the migmatite unit. It is not known definitely, however, whether the migmatitic unit is a distinctly older "basement" on which the volcanic rocks were deposited or whether it is a part of the same metavolcanic-metasedimentary succession that subsequently was migmatized during the development of the Giants Range batholith.

GRANITIC ROCKS

Granitic rocks, including gneisses, constitute two-thirds or more of the Lower Precambrian terrane in the southern part of the Superior Province in Canada (fig. III-1). Apparently, they are somewhat less abundant in northern Minnesota, although the actual volume of "granite" at the Precambrian bedrock surface may be greater than indicated on the geologic map (pl. 1, this volume). Reconnaissance sampling in the Canadian Shield (Eade and others, 1966; Fahrig and Eade, 1968; Shaw and others, 1967) indicates that the gneisses and granitic rocks have an average composition of granodiorite.

The granitic rocks in the southern part of the Superior Province have been considered from field relationships to

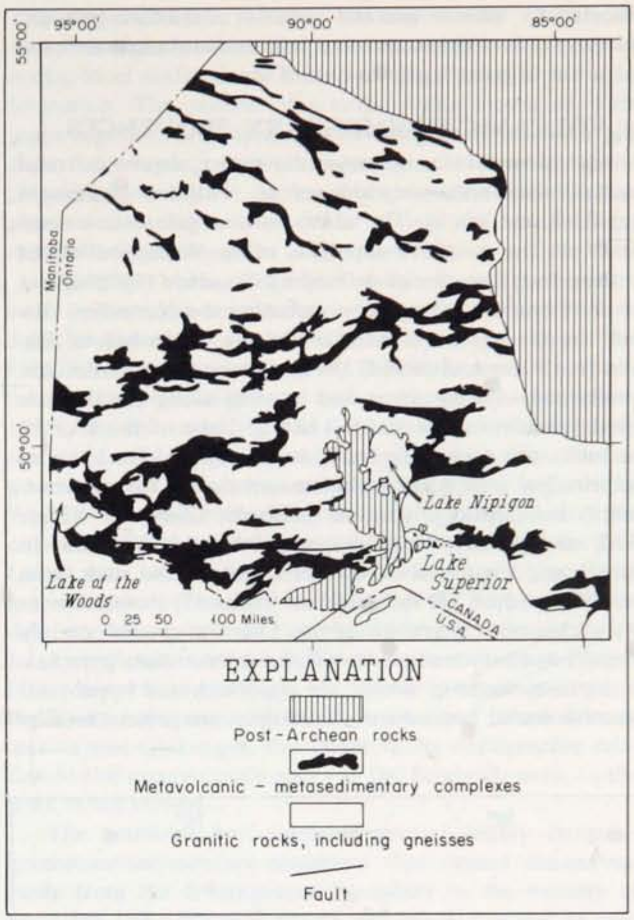

Figure III-1. Generalized geologic map of southern part of Superior Province, Canada. (Modified from Riley and others, 1971, fig. 2.)

be mainly of two geologic ages, Laurentian and Algoman (table III-1; see also table II-3). The term "Laurentian" was first applied by Lawson (1885, p. 100; 1888, p. 142) to granitic rocks that intrude Keewatin (mafic volcanic) rocks in the Lake of the Woods and Rainy Lake areas in northwestern Ontario. Subsequently, the term was widely used in Ontario and Minnesota for an older Archean "granite" (and orogeny) that was separated from a younger Archean "granite" (and orogeny), the Algoman, by the Epi-Laurentian unconformity. The Epi-Laurentian unconformity was considered to mark a widespread tectonic event that separated younger sedimentary successions, mainly referred to in Canada as the Temiskaming and in Minnesota as the Knife Lake Group, from older Keewatin volcanic rocks. As more detailed field data were gathered, it was shown that most of the granitic bodies previously referred to the Laurentian were actually younger than the sedimentary successions. Further, the existence of unconformities at the base of the clastic sedimentary successions in many areas was questioned. Ultimately, the criteria for older "granites" and a major unconformity were based solely on so-called granitic clasts in conglomerates at the base of the Temiskaming and the Knife Lake. In a regional study of the southern part of the Canadian Shield, Bass (1961) showed that many of the so-called granite clasts in Temiskaming-like conglomerates are hypabyssal porphyries or aphanitic volcanics rather than plutonic rocks. Also, he concluded that unconformities in the Archean sequences are local and temporally minor. However, he (Bass, 1961, p. 699) verified the earlier contention that "There is evidence in southwestern Ontario and adjacent parts of Minnesota for a major unconformity involving denudation of deep-seated plutonic rocks." In Minnesota, the major support for this conclusion is the occurrence of tonalite boulders in a conglomerate within the Knife Lake Group (Gruner, 1941, p. 1601; Grout and others, 1951, p. 1029) that overlies the Saganaga Tonalite at Cache Bay in the western part of Saganaga Lake. At this locality, the evidence for denudation of plutonic rocks and deposition of epiclastic sediments is unequivocal. However, the tonalite-bearing conglomerate at Cache Bay is neither at the base of the Knife Lake Group (Gruner, 1941), as inferred earlier, nor does it represent a major regional unconformity. For these reasons and the fact that radiometric dating does not support a major time break between emplacement of the Saganaga Tonalite and other batholithic rocks in the region, noted earlier by Goldich and others (1961, p. 73-74), the Minnesota Geological Survey has abandoned the term "Laurentian" in Minnesota.

Most of the granitic rocks in Minnesota have been referred to an Algoman age (table II-3) inasmuch as field relationships indicate that they intrude graywacke-type metasedimentary successions as well as mafic volcanic rocks (see Grout and others, 1951, p. 1029, 1038-1041). The name "Algoman" was first applied by Lawson (1913b) to rocks in the western Ontario district of Algoma. He stated (p. 5): "Having restricted Laurentian to the older we recognize the need of a name for similar intrusives that are post-Seine. . . ." In Minnesota, Grout (1925b, 1926) applied the name "Algoman" to both the Giants Range batholith and the Vermilion batholith, and later (Grout and others, 1951, p. 1041) used it to designate an orogeny (see table II-2). Grout and others (1951, p. 1041) stated: "The widespread igneous activity in Algoman time shows that it was an important period of orogeny. High mountains were formed with complexly folded structures, involving both the Keewatin and the thick Knife Lake Groups." Subsequently, the term "Algoman" has been used in Minnesota to designate both a tectonic and a plutonic event (Goldich and others, 1961, table 2; Goldich, 1968, table 1). There is no compelling evidence, though, that it represents a period of mountain building; instead it may have been a period of rock deformation or tectogenesis (Dickinson, 1971, p. 108) and granite emplacement.

The abundant geochronologic data obtained in recent years by S. S. Goldich, G. N. Hanson, Zell Peterman, and colleagues have clarified the temporal relations of the several granitic plutons in Minnesota and along the International boundary. All the major granitic plutons that have been dated—the Vermilion, Giants Range, and Saganaga in Minnesota, and the Bad Vermilion in Ontario—have approximately the same radiometric age (*ca.* 2,700 m.y.). The late, small, alkalic intrusive bodies—the Linden pluton in Minnesota and the Icarus pluton in Ontario—which postdate folding in the respective regions, also have radiometric ages (Hanson and others, 1971b) of approximately 2,700 m.y. Accordingly, within the limits of resolution of current radiometric techniques, the major plutons were emplaced at approximately the same time, and most likely within a time span of 100 m.y. or less (see Hanson and others, 1971b, p. 1120-1121). The radiometric age of the late Linden and Icarus plutons effectively fixes the date for cessation of the Algoman orogeny. Although some of the late faulting, described below, may postdate crystallization of these rocks, folding and prograde metamorphism had ceased prior to emplacement of the Linden and Icarus plutons.

The geologic and radiometric age data now available for the region are consistent with an interpretation that the major granitic plutons were emplaced into the volcanic-sedimentary sequences during the later stages of a volcanogenic cycle, which encompassed a time interval of approximately 50 to 100 m.y.; two of the plutons, the Saganaga and Bad Vermilion, were unroofed while the region was still unstable, and provided detritus to still-accumulating sediments. The other large plutons, the Giants Range and the Vermilion, were not unroofed at this time, perhaps either because of their being situated favorably with respect to topographic irregularities or because of their having been emplaced in a more stable environment. Evidently, emplacement of all the plutons was at relatively shallow depths, as indicated by narrow thermal aureoles and other geologic and geophysical characteristics, and erosion need not have had to cut to great depths to unroof them.

VOLCANIC-SEDIMENTARY SEQUENCES

Several metavolcanic-metasedimentary sequences trend generally east-northeastward across northern Minnesota (fig. III-2 and pl. 1). The northernmost greenstone complexes are the westward extension of the Wabigoon belt of northwestern Ontario, as defined by Goodwin (1970b), and the southernmost complexes, including the Vermilion district, are the westward extension of the Wawa belt of Ontario (see Riley and others, 1971). Except in the Vermilion district and adjacent areas and in areas along the International boundary from Rainy Lake to Lake of the Woods, the rocks are poorly exposed; accordingly, delineation of the principal rock types in the western part of the state necessarily is based largely on aeromagnetic (Zietz and Kirby, 1970) and gravity (Craddock and others, 1970) data. In Plate 1 and Figure III-2 only three generalized rock types are distinguished, (1) metavolcanic rocks, (2) metasedimentary rocks, and (3) granitic rocks. The first consists mainly of mafic and intermediate lava flows, intermediate pyroclastic deposits, some of which are reworked, and hypabyssal intrusive rocks; banded iron-formations are present locally.

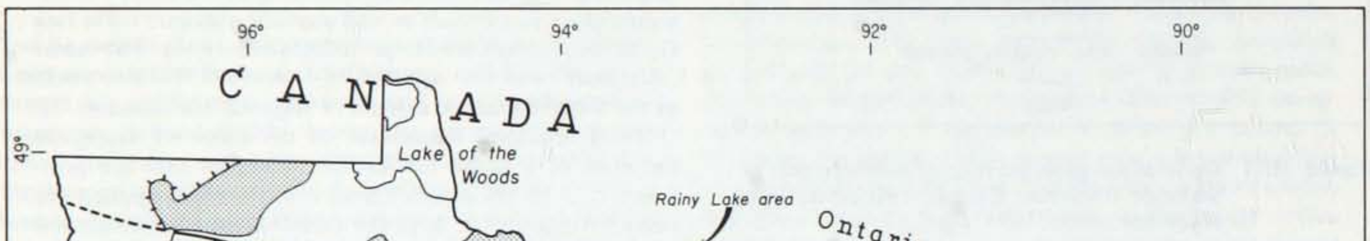

Figure III-2. Approximate outline of greenstone-granite complexes in northern Minnesota.

The metasedimentary rocks are dominantly graywacke-slate, but include substantial amounts of felsic volcaniclastic rocks, local mafic flows, and some conglomerate and iron-formation. The outline of granitic rocks shown on these maps represents the approximate margins of the main plutons, for mixed granite and country rocks cannot be distinguished readily on the basis of presently available geophysical data. Discussions of the gravity and aeromagnetic data in the state are given in Chapter VIII.

The major volcanic complex within the northern greenstone belt extends southwestward from the vicinity of Baudette, southeast of Lake of the Woods, nearly to North Dakota, where it is overlapped by Paleozoic rocks. It constitutes a conspicuous gravity feature (Craddock and others, 1970), and must contain several thousand feet of dominantly mafic volcanic rocks. At least one sulfide-facies iron-formation is present. The belt is transected along the Lake of the Woods County-Koochiching County line by granite, the northern extension of the large body of granite in the Red Lakes area, which forms the southeastern margin of the greenstone belt. The eastern segment of the belt, near Birchdale, contains substantial amounts of felsic volcanic rocks. In exposed parts of the belt, the rocks contain amphibolite-facies mineral assemblages. A narrow greenstone body in the Rainy Lake area (fig. III-2) is described in a following section (see Ojakangas, this chapter). Its stratigraphic relation to the metavolcanic rocks in the Birchdale area, to the west, is not known.

The southern belt contains several highly irregular greenstone-sedimentary sequences that extend discontinuously from the International boundary in the vicinity of Saganaga Lake across the state into North Dakota. In general, the belt is bordered on the north by the Vermilion granite-migmatite massif and, farther westward, by granitic rocks in the vicinity of Upper Red Lake; it is bordered on the south by the Giants Range batholith. Except for the Vermilion district, which is described later, and for parts of the area immediately to the west in northern Itasca, eastern Beltrami, and southern Koochiching Counties, the metavolcanic-metasedimentary complexes in this belt are covered by unconsolidated glacial deposits. Accordingly, the rocks are known only in very general terms. Judged from scattered outcrops and a few drill holes, the greenstone complex directly west of the Vermilion district consists mainly of mafic lavas and partly reworked pyroclastic deposits of intermediate composition. Scattered lenses of banded iron-formation occur in the dominant metavolcanic rocks. In the area between Deer Lake and Big Fork, several sill-like bodies of serpentinized peridotite and associated gabbroic rocks occur in reworked tuffs or sedimentary rocks. So far as known, most of these rocks are of low metamorphic rank except adjacent to the granitic rocks that border the belt on the south and intrude the center of it. Where observed, the strata are steeply inclined and folded on steeply-plunging axes. The horseshoe-shaped anomaly in Becker County (Zietz and Kirby, 1970), given by a folded banded iron-formation in amphibolite (metabasalt ?), was drilled in the 1950's.

The relative ages of the separate greenstone complexes are not known in detail. It seems probable from geologic mapping (Sims and others, 1970) that the metavolcanic rocks in Itasca County and adjacent areas are grossly equivalent in age to those in the Vermilion district, but precise correlations remain equivocal. Also, the age relation of the metabasaltic rocks on the north side of the Mesabi range with those in the Vermilion district remains questionable because of the intervening granite and the sparse exposures. It may be possible from geologic mapping in the Vermilion granite-migmatite massif (see Southwick, this chapter) to determine the approximate age relation of the greenstone at Rainy Lake with that in the Vermilion district. More likely, however, the relative ages of the separate bodies will not be determined until precise geochronologic dating methods are applied.

TECTONIC FRAMEWORK

The Lower Precambrian rocks of northern Minnesota have a pronounced northeastward-trending grain (fig. III-2), given by the long dimensions of the greenstone-granite complexes and by their internal structural elements. In detail, configurations of the major rock assemblages are more complex, and local segments may trend northwestward or northward. Faults, some of which are major crustal features, transect and disrupt the greenstone-granite complexes, and not uncommonly bring granite and high-grade schists into juxtaposition with greenschist-facies rocks. At any particular locality, folding of one or more generations was followed by faulting and related fracturing, some of which probably preceded emplacement of the major batholiths. Although the principal displacements on the faults were during Early Precambrian time, offsets of Middle Precambrian strata along the trace of some of them indicate subsequent renewed movement.

Although deformation was pervasive, primary structures remain in most of the low-grade metavolcanic-metasedimentary rocks. Primary features that commonly are preserved include graded bedding in the graywackes and pillow structures and variolitic textures in the mafic lavas. Such features are useful in determining the top directions of beds and local stratigraphic successions. At places, shearing has obscured bedding and other primary features, as in parts of the Rainy Lake area (see Ojakangas, this chapter). In the amphibolite-facies rocks, bedding may be retained, but other primary structures and textures commonly are obliterated by a pervasive foliation.

The folds characteristically are tight to close and have steeply-inclined axial surfaces. For the most part, they trend parallel to the long dimension of the rock assemblages, but locally they trend northwestward. The diverse trends in part reflect different generations of folding and in part are a consequence of inhomogeneous deformation or rotation of blocks by later faulting. Fold axes and related lineations mainly are steeply inclined, probably as a consequence of multiple folding.

Folding was developed to different degrees in the metavolcanic-metasedimentary sequences, apparently because of differences in the relative competence of the rocks. Judged from studies in the Vermilion district (Gruner, 1941; Hooper and Ojakangas, 1971), the well bedded rocks, such as graywacke-slate, tuff, and iron-formation, yielded primarily by folding, whereas the more massive lava flows and hypabyssal intrusive bodies yielded both by folding and by shearing. For example, nearly all strata in the type area of the Knife Lake Group (Gruner, 1941) and in the type area of the Lake Vermilion Formation (Morey and others, 1970) are folded on both a large and a small scale. Although much of the folding in these rocks is isoclinal, fold axes can be determined at many places from observed reversals in top direction and from changes in the angular relationships between bedding and cleavage. In contrast, the massive metavolcanic rocks, particularly the Ely Greenstone, tend to constitute homoclinal sequences several thousand feet thick; folding is relatively local in extent. Clearly, such thick lava successions yielded mainly by shearing, which is manifested by close-spaced fracturing that commonly can be observed on the flanks of bedrock hills. Minor structures associated with the folding in both classes of rocks include a weak to moderately developed cleavage and lineation in the low-grade rocks and a moderate or strongly developed foliation and lineation in the amphibolite-facies rocks.

The terrane is transected by abundant high-angle faults of both large and small scale. Many of the faults, or fault zones, are major structures that transect the greenstone complexes or are subparallel to them and effectively separate the supracrustal rocks from the granitic batholiths and associated high-grade metamorphic rocks. The largest of these are comparable in length to the remarkable strike-slip faults in the Yellowknife area, Northwest Territories, Canada (Joliffe, 1942) and to the Great Glen fault in Scotland (Kennedy, 1946). Others are less continuous and have smaller displacements, but nevertheless contribute to the structural complexity.

The major fault in northern Minnesota, so far as known, is the Vermilion fault (pl. 1). It is a long curvilinear fracture with many subsidiary strands that lies along the northern margin of the Vermilion district and is inferred from aeromagnetic data (Zietz and Kirby, 1970) to extend northwestward through the northwest corner of Minnesota, where it is overlapped by Paleozoic strata. Its length is inferred to be at least 250 miles. Probably, it intersects or merges with the major eastward-trending Quetico fault near the International boundary south of Lake of the Woods, as shown in Figure I-1. Other faults that are subparallel to it and have lesser displacements lie south of the Vermilion fault. Some of these are shown on Plate 1 and additional ones are shown on the Hibbing Sheet of the Geologic Map of Minnesota (Sims and others, 1970). In northeastern Minnesota (fig. III-3), the Vermilion fault forms the southern margin along most of the Vermilion granite-migmatite massif, and along much of its length brings granite and associated amphibolite-facies rocks into juxtaposition with greenschist-facies rocks. In eastern St. Louis County, it junctions with several faults that strike east-northeast, subparallel to the strike of the rock units. Faults that have the same trend and probably are part of the same system extend eastward through the eastern part of the Vermilion district (Gruner, 1941). The vertical component of movement on the Vermilion fault possibly is on the order of a mile, accounting for the marked difference in the metamorphic grade of rocks on opposite walls. Neither the direction nor the amount of horizontal displacement is known, although a displacement of several miles is possible. The longitudinal faults are expressed primarily as narrow, linear topographic depressions. Where exposed, they consist mainly of cataclastic rocks as much as 1,500 feet wide; at places these rocks are silicified and altered. A photomicrograph of a mylonite zone in Giants Range Granite along the North Kawishiwi fault is shown in Figure III-4.

Steeply-inclined faults that trend northeastward or northwestward, transverse to the greenstone-granite complexes, have been mapped in the Vermilion granite-migmatite massif and in the central part of the Vermilion district (fig. III-3). Those in the Vermilion massif have two dominant trends. In the northwestern part of St. Louis County, north of Cook, the faults trend northwestward and have both right-lateral and left-lateral displacements (D. L. Southwick, 1972, oral comm.); so far as known their horizontal displacements are small. Some nearly vertical movements on the faults are indicated by slickenside striae. In eastern St. Louis County (fig. III-3), the faults dominantly trend northeastward and have left-lateral displacements. They appear to have had dominant strike-slip movements. The Waasa fault, on the south side of the Vermilion fault (fig. III-3), has a left-lateral offset of more than 3 miles, and the adjacent Camp Rivard fault has a left-lateral offset of at least 2 miles (Griffin and Morey, 1969, p. 38). Most of the other known faults of this set have lesser displacements. In the same way as the longitudinal faults, the transverse faults form conspicuous lineaments, particularly in granitic terranes, and are marked by sheared and altered zones.

Although some of the faults may have formed initially before emplacement of the major granitic bodies, as suggested by Griffin and Morey (1969), most of the faulting probably was concurrent with or later than crystallization of the plutonic rocks. In those areas that have been studied in detail, faults and related joints and kink-bands generally postdate the folding and the prograde metamorphism. Also, faults appear to be as abundant in granitic rocks as in older metavolcanic-metasedimentary rocks. From these and other observations, it is inferred that the major episode of faulting was concurrent with and subsequent to emplacement of the granite batholiths. The faulting, therefore, is considered to be a late phase of the Algoman orogeny. That many of the faults subsequently were rejuvenated, however, is indicated by offsets of Middle Precambrian strata along the projections of faults in Lower Precambrian rocks. For example, the Dark River fault, which is subparallel to and south of the Vermilion fault and has an inferred right-lateral strike-slip movement (Sims and others, 1970), projects southeastward along its trace into a fault that displaces the Middle Precambrian Biwabik Iron-formation and associated strata in the same direction. Apparently as a result of renewed movements along the fault in Middle Precambrian or later time, the fracture was propagated upward into overlying strata, with resulting displacement of the younger rocks.

POST-ALGOMAN METAMORPHISM

Two lines of evidence indicate that the greenstone-granite complexes of northern Minnesota were subjected to low-grade metamorphism, at least locally, subsequent to the Al-

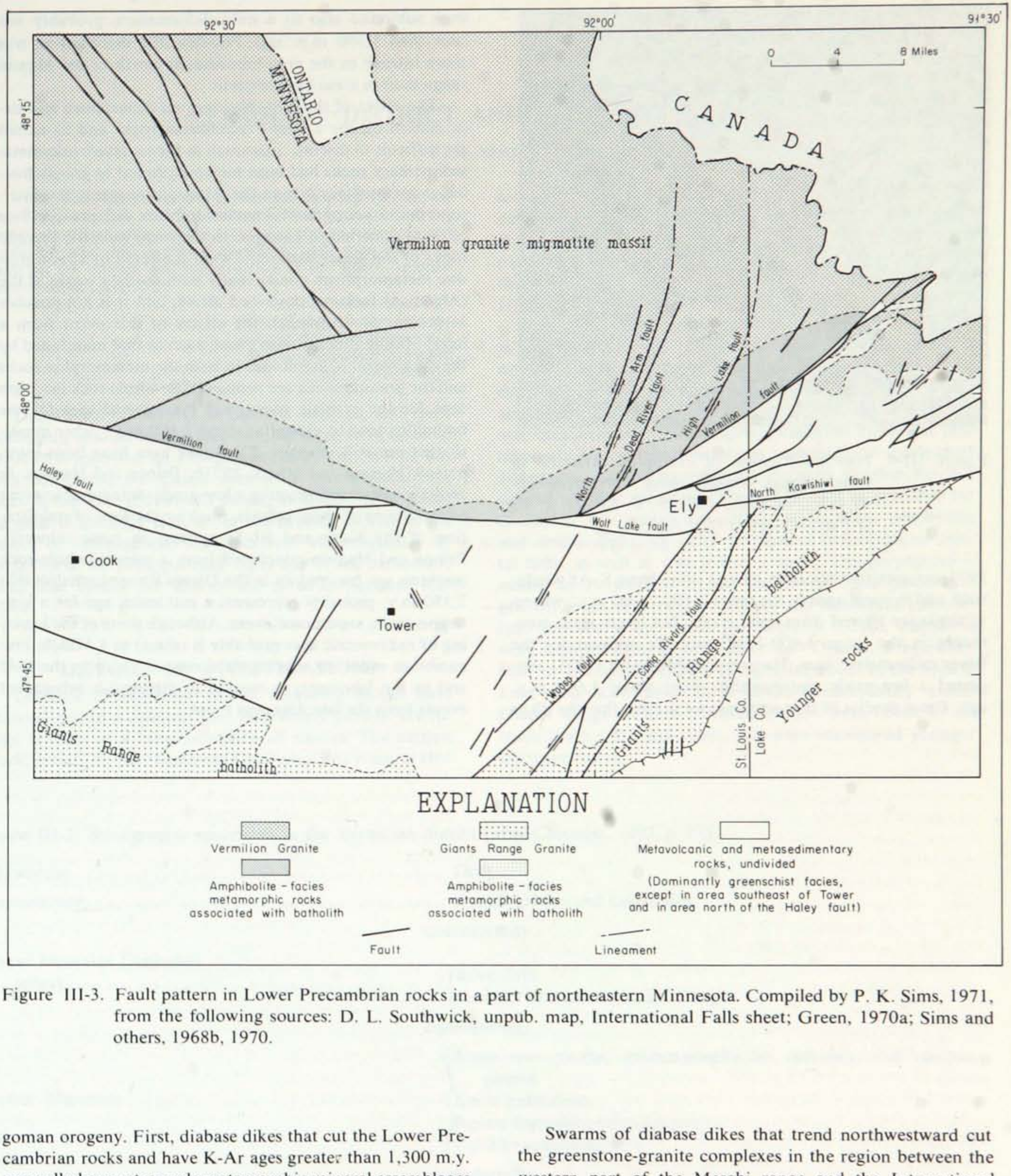

Figure III-3. Fault pattern in Lower Precambrian rocks in a part of northeastern Minnesota. Compiled by P. K. Sims, 1971, from the following sources: D. L. Southwick, unpub. map, International Falls sheet; Green, 1970a; Sims and others, 1968b, 1970.

goman orogeny. First, diabase dikes that cut the Lower Precambrian rocks and have K-Ar ages greater than 1,300 m.y. generally have retrograde metamorphic mineral assemblages and coincident lower apparent radiometric ages (Hanson and Malhotra, 1971), and second, radiometric age data from the major granitic batholiths suggest some retrogressive metamorphism subsequent to crystallization (Hanson and others, 1971b).

Swarms of diabase dikes that trend northwestward cut the greenstone-granite complexes in the region between the western part of the Mesabi range and the International boundary (see Sims and Mudrey, chapter on Middle Precambrian). The diabase is altered to different degrees to retrograde assemblages, and some of it in the area immediately north of the Mesabi range has cataclastic textures. Available radiometric age data (Hanson and Malhotra,

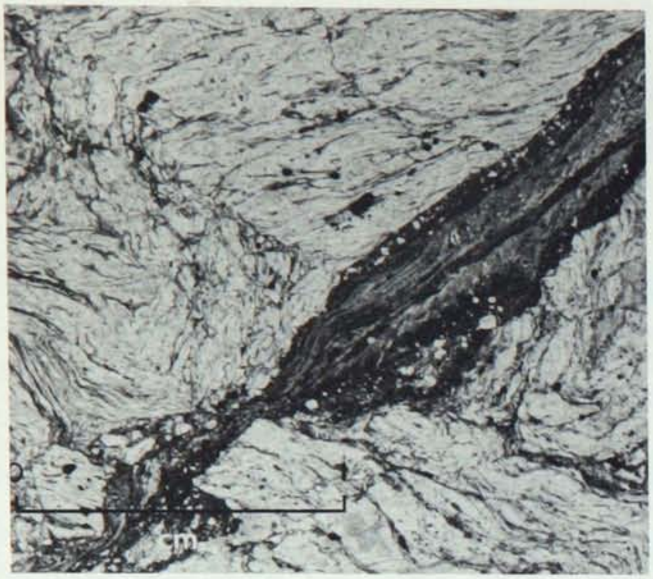

Figure III-4. Photomicrograph of crumpled mylonite in Giants Range Granite along North Kawishiwi fault. M-7594; sec. 28, T. 63 N., R. 10 W.; after Green, 1970a, fig. 35.

1971) suggest that the least altered dikes have K-Ar whole-rock and mineral ages in the range 2,100-2,200 m.y., whereas the more altered dikes have erratic but lower ages, commonly in the range 1,400-1,800 m.y. To account for the lower radiometric ages, Hanson and Malhotra (1971) suggested a low-grade metamorphic event about 1,600 m.y. ago. From studies of thin sections, we suggest that the dikes were subjected also to a mild deformation, probably not later than 1,500 m.y. ago. Possibly, the deformation was more intense in the area immediately north of the Mesabi range than in areas farther north.

The effect of the post-Algoman metamorphism and deformation on the Lower Precambrian rocks and its extent are difficult to discern. Inasmuch as the metavolcanic-metasedimentary rocks had been metamorphosed to greenschist-facies assemblages during the Algoman orogeny, the later, superposed retrogressive metamorphism did not produce obvious mineralogic changes. In the same way, the granitic rocks of the major batholiths were subjected to a retrogressive metamorphism concurrently with the late stages of the (Algoman) faulting, described above, and it is not possible at present to distinguish the effects of this event from a later, Middle Precambrian event, such as that manifested by the dike rocks. Mineral ages in both the metamorphic rocks and the granitic rocks are younger than whole-rock isochron ages for the granitic rocks, and the mineral ages for the batholiths tend to cluster at about 2,600 m.y.; other mineral ages are still younger. The lower ages have been interpreted (Hanson and others, 1971b; Prince and Hanson, in press) as either representing a low-grade metamorphic event or the effects of epeirogenesis, marking the time of stabilization of the K-Ar and Rb-Sr systems in some minerals. Prince and Hanson concluded from a mineral-whole-rock isochron age for epidote in the Giants Range batholith that 2,350 m.y. probably represents a maximum age for a low-temperature, superposed event. Although some of the lowering of radiometric ages probably is related to a Middle Precambrian event (or events), further study both in the field and in the laboratory is needed to distinguish subsequent events from the late Algoman event.

VERMILION DISTRICT AND ADJACENT AREAS

P. K. Sims

"... On first examining a new district, nothing can appear more hopeless than the chaos of rocks; but by recording the stratification and nature of the rocks . . ., always reasoning and predicting what will be found elsewhere, light soon begins to dawn on the district, and the structure of the whole becomes more or less intelligible."

Charles Darwin, 1839,
Journal of researches into the
geology and natural history of
the various countries visited
by H.M.S. Beagle.

The Vermilion district is the best exposed and most thoroughly studied of the greenstone-granite complexes in northern Minnesota. As a consequence, the region long has been considered the classic area of Lower Precambrian rocks in the state (Grout and others, 1951). In this section, the general geologic features of the district as well as the history of investigations are described. More detailed descriptions of the principal rock types and of the batholithic rocks that border the district are given in sections that follow.

HISTORY OF INVESTIGATIONS

Geologic investigations of the Vermilion district have spanned nearly a century, and accordingly present knowledge is built on a long succession of studies. The earliest work, mainly by the Minnesota Geological and Natural History Survey (1872-1900), was designed primarily to aid development of the iron ores. In 1897, the U.S. Geological Survey began a systematic investigation of the district, as part of a broader study of the entire Lake Superior region; this work established the general outline of the geology (Van Hise and Clements, 1901; Clements, 1903).

The sequence, as determined by Clements (1903, p. 33), is listed in Table III-2. Clements concluded that all like lithologies are correlative time-stratigraphic units, and he explained recurring lithologies as resulting from folding (see atlas accompanying Monograph 45). This interpretation was consistent with stratigraphic principles known at that time, which were based mainly on knowledge of stable-shelf and continental sedimentary environments. Further, he considered all the "acid rocks" in the district as granites. Included in this group were felsic-intermediate porphyries and similar-appearing rocks now known to be lithic or crystal tuffs, as well as true plutonic rocks. The porphyries—mainly in the western part of the district—and the coarse-grained rocks of Trout, Burntside, and Basswood Lakes—later referred to the Vermilion batholith (Grout, 1925b)—were interpreted to be older than the "Ogishke conglomerate," or Laurentian in age. The granitic rocks in the Moose Lake and Kawishiwi River areas—subsequently called Giants Range Granite (Allison, 1925)—were noted to intrude the Knife Lake slates, and were considered younger intrusive rocks.

Table III-2. Stratigraphic succession in the Vermilion district (after Clements, 1903, p. 33).

Pleistocene	Drift
Keweenawan	Duluth gabbro and Logan sills
	unconformity
Upper Huronian (Animikie series)	Rove slate Gunflint formation (iron-bearing)
	unconformity
Lower Huronian	Intrusives—granite, granite-porphyries, dolerites, and lamprophyres Knife Lake slates Agawa formation (iron-bearing) Ogishke conglomerate
	unconformity
Archean	Intrusive granites, granite-porphyries, and some greenstones Soudan formation (the iron-bearing formation) (Minor unconformity) Ely greenstone, and ellipsoidally parted basic igneous and largely volcanic rock

Later, Grout (1926) extended Clements' mapping to other areas in northern St. Louis County. He applied Clements' nomenclature, with some modification, to similar-appearing rocks throughout the region. Subsequently, Grout (1933a) substituted "Knife Lake series" for "Knife Lake slates," and later (Grout and others, 1951) called the succession "Knife Lake Group." As a part of studies of the batholithic rocks of northern Minnesota, Grout mapped the Vermilion batholith (Grout, 1925b) and the Saganaga batholith (Grout, 1929, 1936), at the eastern extremity of the district, and Allison (1925) mapped the Giants Range batholith.

In the late twenties and thirties, Gruner (1941) mapped in detail about 300 square miles in the eastern part of the Vermilion district, primarily to determine the structural pattern of the Knife Lake area. He modified the stratigraphic nomenclature of the earlier workers, and by careful mapping delineated many units within the "Knife Lake series." He showed that conglomerates occurred at several stratigraphic positions and not just at the base of the sedimentary succession, and concluded that the "Ogishke conglomerate" was invalid as a time-stratigraphic unit. His map was the first in the state to show faults, and remains very useful.

The geology of the Vermilion district was reviewed in a summary paper by Grout and others (1951), which provided the basis for the radiometric age studies of the Precambrian rocks by Goldich and others (1961), published 10 years later.

Recent geologic investigations in the western (Sims and others, 1968a and b, 1970) and central (Green and others, 1966; Green, 1970a) parts of the Vermilion district have added much new data. Some conclusions resulting from this work that differ substantially from earlier views are as follows:

(1) The metavolcanic-metasedimentary sequence in the district is a complex volcanic pile accumulation, deposited mainly in a subaqueous environment, which is analogous to that in other greenstone belts in Canada (Goodwin, 1968a).

(2) Mafic volcanism was repeated in time, and accordingly pillowed basaltic flows are not diagnostic of a unique time-stratigraphic unit, as inferred previously (*cf.* Clements, 1903; Grout and others, 1951).

(3) There is no evidence in the western and central parts of the district for a major unconformity between the mafic volcanics (Ely Greenstone) and younger sedimentary strata, as interpreted by Clements (1903) and many later workers. Specifically, the deposits of coarse-grained fragmental rocks at Stuntz Bay and vicinity, at the eastern end of Lake Vermilion, called "Ogishke conglomerate" by Clements (1903, p. 23), are agglomerate and tuff deposits which mark initial felsic volcanism that followed mafic volcanism and deposition of the Soudan Iron-formation.

(4) The widespread porphyries in the volcanic succession are shallow intrusive (hypabyssal) equivalents of the extrusive rocks of the district, and are not related to a pre-Knife Lake (Laurentian) orogenic event (*cf.* Clements, 1903).

(5) Metamorphism and deformation in the western part of the district followed deposition of the youngest clastic strata in that area and was virtually synchronous with emplacement of the Giants Range and Vermilion batholiths; there is no evidence in this area for tectonism prior to accumulation of clastic deposits.

GENERAL GEOLOGY

The Vermilion district, as defined originally by Clements (1903), constitutes a narrow belt of metavolcanic and metasedimentary rocks that extends from the vicinity of Tower, near Lake Vermilion, northeastward to the International boundary in the vicinity of Saganaga Lake (fig. III-5). The metavolcanic-metasedimentary sequence is bounded on the north by the Vermilion granite-migmatite massif, on the south by the Giants Range batholith, and on the east by the Saganaga batholith. Much of the eastern part of the district is truncated by the Duluth Complex of Keweenawan age.

The rocks within the district trend eastward and northeastward, generally dip steeply, and dominantly constitute a northward-facing stratigraphic sequence. Folding of at least two generations was broadly contemporaneous with emplacement of the fringing batholithic rocks, and was followed by a major period of faulting that was regional in scope.

Stratigraphic Succession

The metavolcanic and metasedimentary rocks of the district are assigned to five formations (Morey and others, 1970; fig. III-6, this report). The oldest formation, the Ely Greenstone, which consists mainly of mafic metavolcanic rocks, is overlain stratigraphically in the western part by the Lake Vermilion Formation and, locally, the Soudan Iron-formation, and in the central part by the Knife Lake Group (fig. III-7). Both the Lake Vermilion and the Knife Lake units consist primarily of metagraywacke-slate and felsic volcaniclastic rocks, but include some mafic and intermediate lavas. The Newton Lake Formation, a younger unit of mixed felsic and mafic metavolcanic rocks, overlies the Knife Lake Group in the central part of the district and may intertongue with it further eastward. The eastern part of the district is underlain mainly by the Knife Lake Group, as shown on the map by Gruner (1941); the stratigraphic position of the mafic metavolcanic rocks in the eastern area is equivocal, as discussed below.

Formerly (Clements, 1903; Grout, 1926, pl. 1), the mafic volcanic rocks now assigned to the Newton Lake Formation were called Ely Greenstone, or Keewatin Group (Goldich and others, 1961, pl. 1); also, the dominantly metasedimentary succession now assigned to the Lake Vermilion Formation was called Knife Lake Group (Goldich and others, 1961, pl. 1). The Lake Vermilion Formation was established formally (Morey and others, 1970, p. 20) because it is not continuous with the type section of the Knife Lake Group in the eastern part of the district and, as discussed in a later section, because it differs somewhat in lithology from the Knife Lake.

Rocks having lithologies similar to the formally designated units occur within the borders of the fringing batholiths, and probably are approximately equivalent in age; that

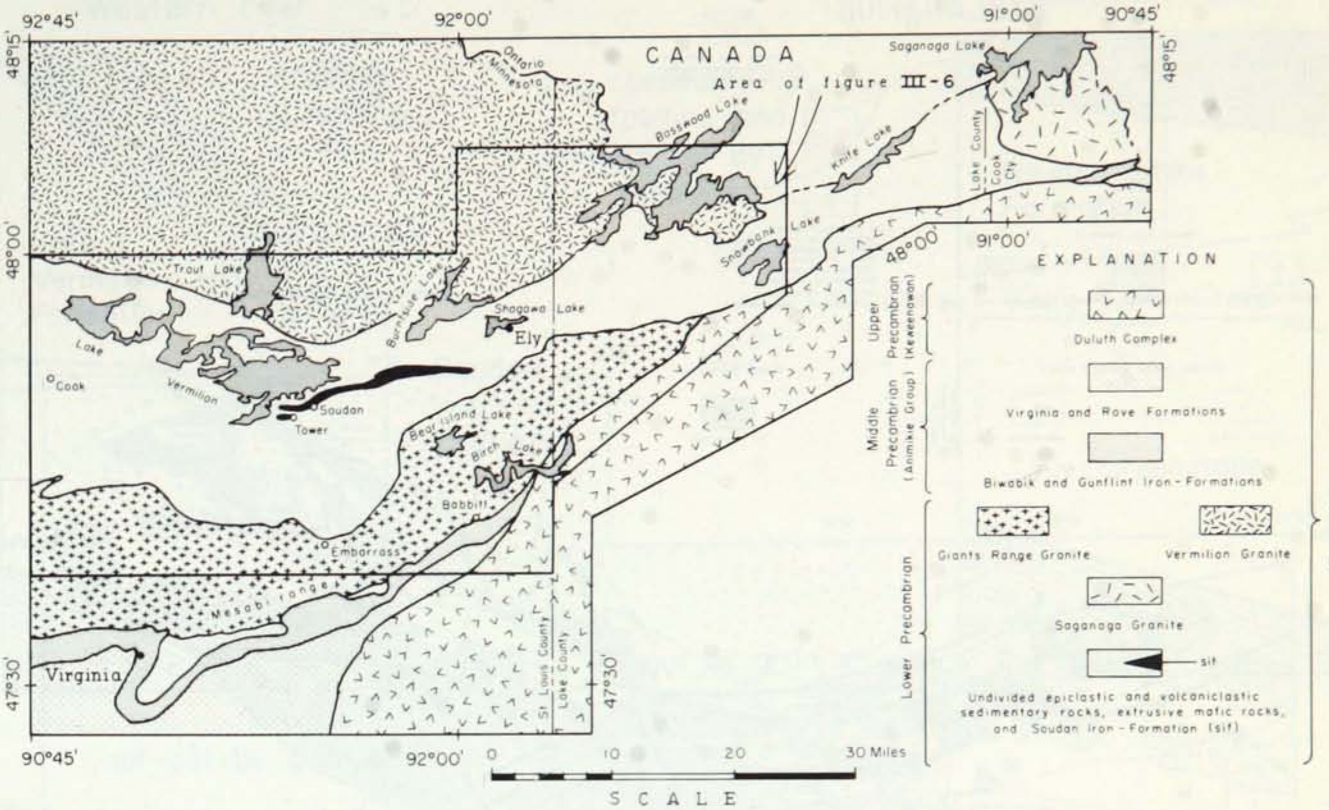

Figure III-5. Map showing geologic setting of Vermilion district.

is, they were deposited during the same cycle of volcanic-sedimentary accumulation as the formally designated units.

The revised stratigraphy (Morey and others, 1970) for the western part of the Vermilion district has implications regarding the nomenclature and stratigraphy of the eastern part. Previously, all mafic metavolcanic rocks in the eastern part of the district were equated with the Ely Greenstone (see Gruner, 1941, pl. 1), and all of the Knife Lake Group was interpreted as being younger than the Saganaga batholith at the eastern end of the district. A tentative reinterpretation of the geologic relationships is shown in Figure III-8. The long, continuous body of greenstone in Canada (unit ug on fig. III-8), along the northwestern side of the outcrop belt of the Knife Lake Group, apparently is continuous with the Newton Lake Formation, as mapped by Green (1970a, pl. 1) in Lake County, Minnesota, and tentatively is considered correlative with the Newton Lake. This interpretation is not inconsistent with Gruner's (1941) mapping, for although he inferred that the contact between the Knife Lake and the greenstone at most places was faulted, many graded beds in graywacke adjacent to the greenstone face northwestward, toward the greenstone. In mapping, he commonly placed faults along the contacts between these two units because he believed that the greenstone was stratigraphically older than the Knife Lake (J. W. Gruner, 1971, oral comm.). On the north shore of Saganaga Lake, pillow tops in the greenstone unit consistently face northward (Harris, 1968), and the lithology of the unit is similar to the Newton Lake Formation in the type area. In the Saganaga Lake area, the Saganaga Tonalite intrudes the greenstone, although the contact is mainly a fault that follows the north shore of Saganaga Lake (see map, Harris, 1968). If the correlation of this greenstone unit with the Newton Lake Formation is correct, the Saganaga Tonalite intrudes rocks correlative with the Newton Lake Formation and, inasmuch as the Newton Lake is younger than at least some of the Knife Lake Group, it must also intrude some rocks belonging to the Knife Lake Group.

On the basis of Gruner's (1941) mapping and petrographic studies by R. W. Ojakangas (this chapter), the Knife Lake Group can be interpreted as consisting of two rock successions of different ages, one younger than and one older than the Saganaga Tonalite. The younger succession, as tentatively interpreted on Figure III-8, contains locally abundant epiclastic deposits, including conglomerates having boulders and cobbles of Saganaga Tonalite as well as greenstone. The interbedded graywackes commonly contain some K-feldspar, indicative of a plutonic source (see Ojakangas, this chapter). At Cache Bay, at the western edge of Saganaga Lake, the epiclastic deposits dip steeply westward as a result of tight folding (Gruner, 1941) on northward- and northeastward-trending axes that postdate emplacement of the Saganaga batholith. This fold generation affected all the rocks in the younger succession. Judged from Gruner's geologic map, the rocks on the south side of the Saganaga batholith are clearly older than the Saganaga Tonalite. The

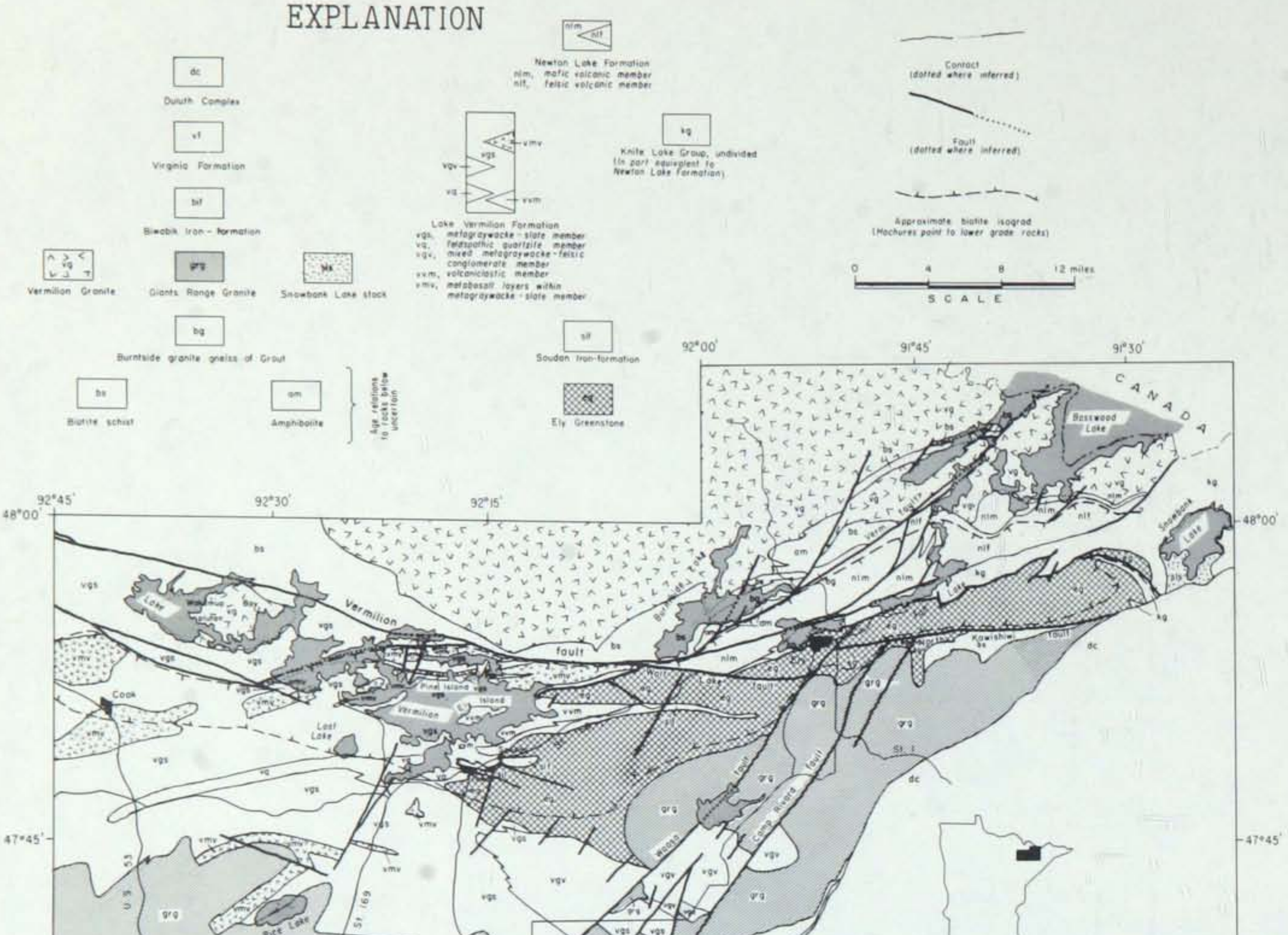

Figure III-6. Geologic map of western part of Vermilion district and adjacent areas.

mafic volcanic unit (Kl_{1a} on fig. III-8) adjacent to the batho-lith is intruded and metamorphosed by the Saganaga Tonalite, and the overlying rocks are conformable to it. This succession was folded on northwest-trending axes before or concurrently with emplacement of the batholith. Weiblen and others (1971) have suggested that this succession should not be included in the Knife Lake Group because it is not demonstrably continuous with typical Knife Lake rocks to the west, but it is tentatively included with it here (fig. III-8) because Gruner (1941, pl. 1) designated all the rocks in the succession except the greenstones as Knife Lake. The rocks in the Snowbank Lake area also are tentatively considered to belong to the older succession of the Knife Lake Group. In the same way as the succession on the south side of the Saganaga batholith, these rocks are folded on northwest-trending axes and, so far as known, lack detritus from the Saganaga Tonalite.

The interpretation of the geologic history of the eastern part of the Vermilion district given above is consistent with radiometric dating in the region. Radiometric dating has indicated that there is no appreciable difference in the age of the Saganaga Tonalite and the granitic rocks comprising the Vermilion and Giants Range batholiths. This fact led Goldich and others (1961, p. 73-74) earlier to relegate the Laurentian orogeny to a minor status. With the interpretation presented here, the Saganaga batholith was emplaced, probably at a shallow depth, and unroofed during the later stages of a volcanic-sedimentary cycle, to supply epiclastic debris locally to still-accumulating volcanogenic sediments. The epiclastic deposits shed from the batholith probably were deposited within a time span of a few million years, and there need not have been a long time span between emplacement of the Saganaga and the other batholiths.

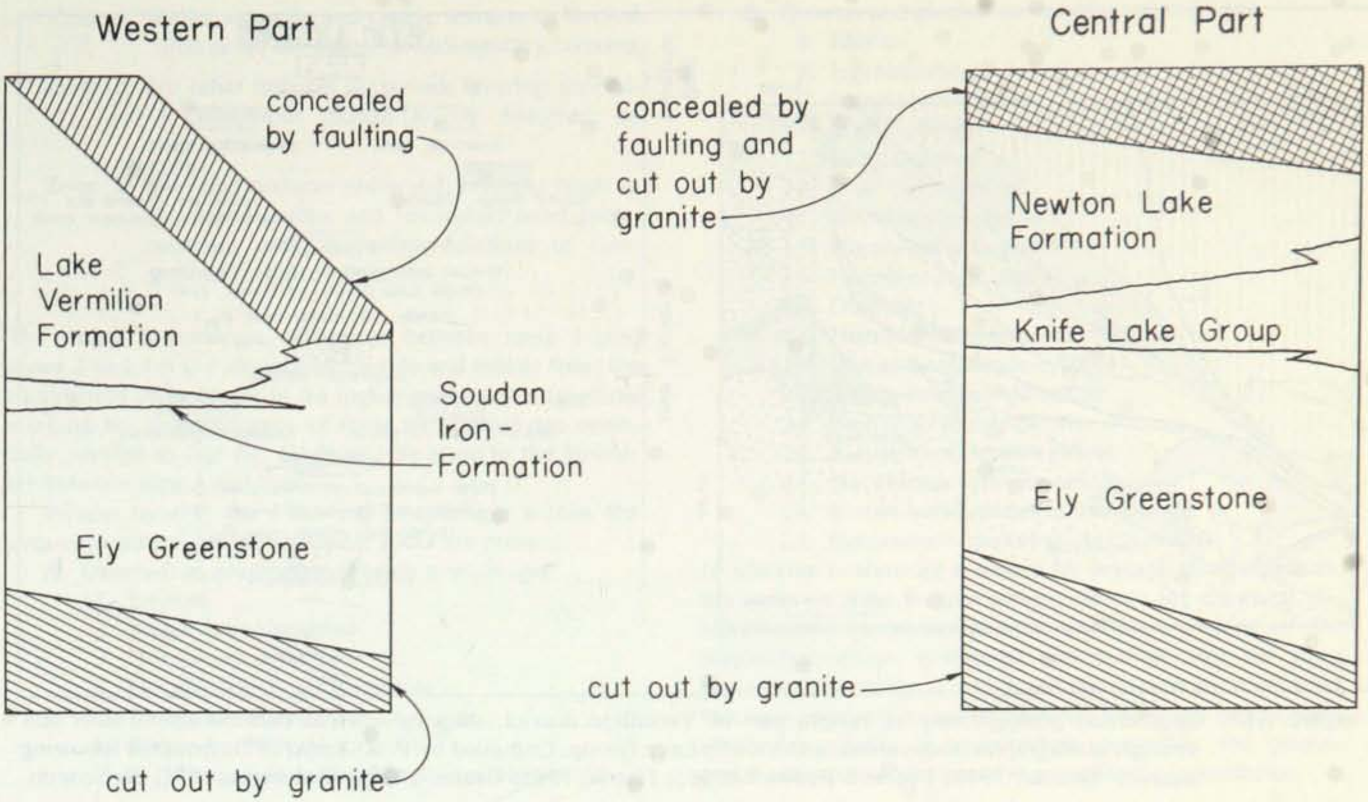

Figure III-7. Stratigraphic relationships in western part of Vermilion district and adjacent areas.

Intrusive rocks

The metavolcanic-metasedimentary sequence is intruded by a variety of Lower Precambrian hypabyssal and plutonic rocks, which were emplaced during four magmatic episodes. In approximate order of age, from oldest to youngest, the intrusive rocks are: (1) synvolcanic bodies, including hypabyssal porphyries and serpentinized peridotite; (2) granitic rocks of the Saganaga, Giants Range, and Vermilion batholiths; (3) small bodies of syenitic rocks and related lamprophyres; and (4) the alkalic Linden pluton. The plutonic rocks were emplaced during the Algoman orogeny, and are syntectonic or post-tectonic intrusive bodies. In addition, mafic dikes of both Middle Precambrian and Late Precambrian age cut the Lower Precambrian rocks.

The synvolcanic intrusive bodies are spatially and temporally associated with the metavolcanic rocks, and are altered and metamorphosed approximately to the same extent as the country rocks. They form small, irregular plutons, sill-like bodies, and dikes, and range in composition from gabbro to rhyodacite and include rather abundant andesite. Metadiabase is the most abundant and widespread rock type, and far exceeds the volume of porphyritic rocks. Serpentinized peridotite occurs as small, widely spaced lenses associated with gabbro within the Newton Lake Formation; it is interpreted as intrusive in origin. The largest known peridotite body is about 3 miles long and 200 feet wide.

The rocks comprising the major batholiths border the supracrustal rocks and partially engulf them, and accordingly have profoundly affected the volcanic-sedimentary sequence. The Giants Range batholith, on the south, transects the sequence and in areas southwest and east of Ely (fig. III-6) has cut out its lower part. The body consists of several distinct plutons that trend northeastward, subparallel to the regional structure. The eastern part is composed dominantly of hornblende adamellite-granodiorite; the western part is more diverse in composition and contains rocks ranging from tonalite to granite. The Vermilion batholith, on the northern side of the district, transects the upper part of the sequence. It consists dominantly of biotite granite but contains local, generally small, more mafic plutonic rocks, including trondhjemite. In contrast to the other batholiths, the granitic rocks contain abundant conformable layers and lenses of biotite schist and amphibolite, and accordingly Southwick (this chapter) has referred to the mass as a granite-migmatite massif. The Saganaga batholith at the eastern end of the district is composed largely of tonalite. It intrudes mafic volcanic rocks and the Northern Light Gneiss and is overlain by the younger succession of the Knife Lake Group (fig. III-8). All the batholiths were emplaced synchronously with regional deformation and metamorphism. At least the western part of the Saganaga batholith was deformed subsequent to cooling by folding, and all the batholiths were deformed locally by the late Algoman faulting.

Syenitic rocks that appear to be late syntectonic or post-tectonic, and thus possibly are younger than the rocks comprising the batholiths, intrude the Knife Lake Group at

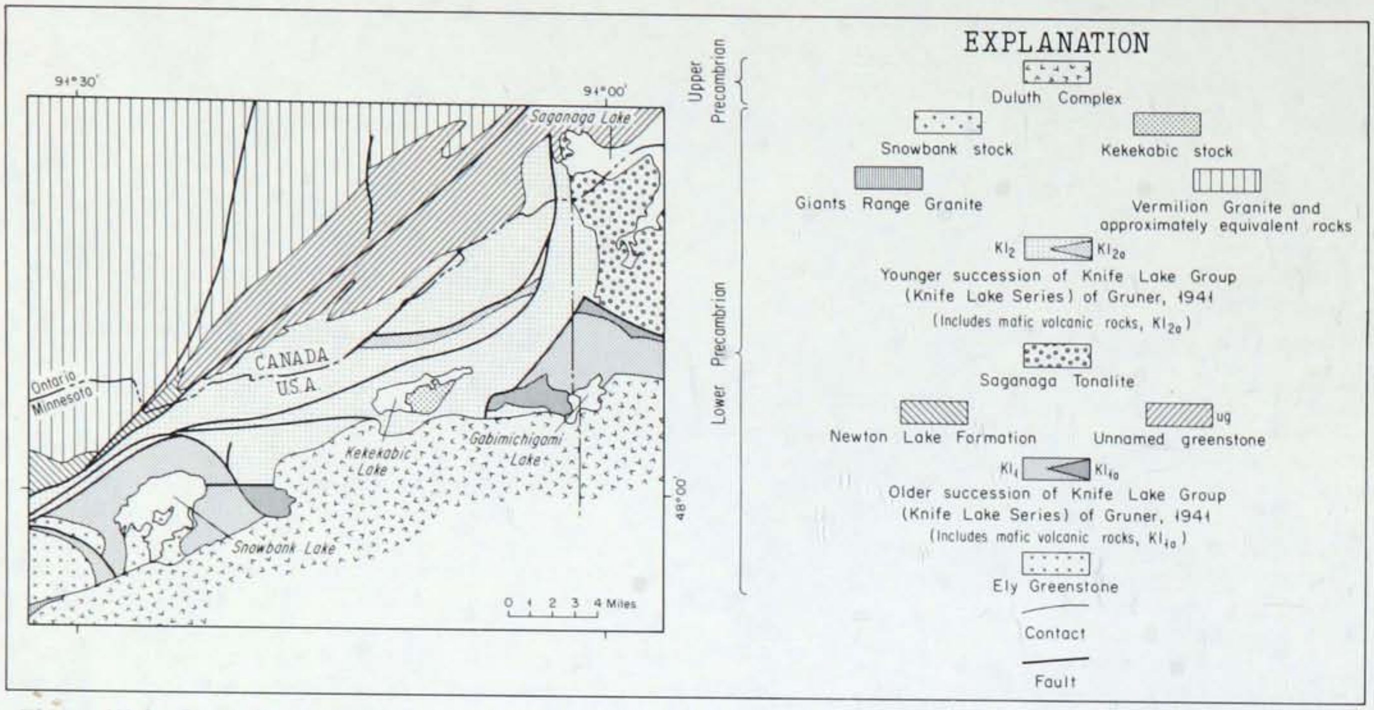

Figure III-8. Generalized geologic map of eastern part of Vermilion district, showing inferred distribution of older and younger stratigraphic successions in the Knife Lake Group. Compiled by P. K. Sims, 1971, from the following sources: Gruner, 1941; Pye and Fenwick, 1965; Harris, 1968; Green, 1970a; Ojakangas, 1971, oral comm.

Snowbank Lake and Kekekabic Lake in the eastern part of the district, and cut diverse metavolcanic and metasedimentary rocks in the western part. The bodies at Snowbank and Kekekabic Lakes are composite stocks of general syenodiorite composition. The bodies in the western part of the district are small and widely scattered, and many of them are interpreted to be apophyses and satellitic plutons related to a subjacent stock at Lost Lake (fig. III-6) that is cogenetic with the Snowbank and Kekekabic stocks. Lamprophyres are closely associated spatially with both the Snowbank stock and the subjacent stock, and most likely are related to this episode of magmatic activity.

The youngest of the Algoman intrusive rocks is the Linden pluton, west of Cook, a small, probably composite body of alkali-lime syenite. The pluton has a pronounced vertical flow-lineation and is interpreted as being post-tectonic. It is similar in composition and structure to the Icarus pluton in the Saganaga Lake-Northern Light Lake area in Ontario, which has been described by Goldich and others (in press).

Metamorphism

The supracrustal rocks within the district dominantly contain mineral assemblages characteristic of the greenschist facies (Turner, 1968). Adjacent to the Vermilion and Giants Range batholiths, and locally adjacent to some faults, the rocks are metamorphosed to the amphibolite facies. The configuration of the metamorphic gradients in the region is shown by an approximate biotite isograd in Figure III-6.

Most of the rocks are in the lower range (chlorite zone) of metamorphic grade in the greenschist facies. Quartzofeldspathic rocks contain chlorite, muscovite, albite, quartz, and epidote; and mafic rocks contain chlorite, calcite, tremolite or actinolite, epidote, albite, and quartz. Higher grade rocks in this facies contain a yellow-brown or red-brown biotite in addition to the phases listed above. Evidence of incomplete recrystallization is widespread, and is indicated by relict hornblende and zoned plagioclase in the metagraywacke and by relict augite and labradorite in the mafic lavas. In the metagraywacke, the coarser grained material—feldspar and quartz grains and volcanic rock fragments—typically shows little evidence of recrystallization, whereas the finer matrix generally has a mosaic texture indicating recrystallization, and only locally retains its primary texture.

Throughout the central part of the district (fig. III-6), amphibolite-facies rocks generally comprise narrow zones, commonly no more than 3,000 feet wide, adjacent to the Giants Range and Vermilion batholiths. In the western part, however, the aureole adjacent to the Giants Range batholith widens notably, and locally is as much as 8 miles wide. The marked increase in the outcrop width of amphibolite-facies rocks in the southwestern part of the area is interpreted to indicate that granite underlies this part of the area at relatively shallow depths (Griffin and Morey, 1969).

The metamorphic pattern adjacent to the Giants Range batholith in the southwestern part of the district has been studied in moderate detail (Griffin and Morey, 1969). In this region (fig. III-9) all the rocks are above the biotite isograd, and the metamorphic grade increases southeastward. Three metamorphic zones are distinguished on the basis of degree of recrystallization and deformation. The zones are defined as follows:

Zone 1: Relict volcanic and clastic textures preserved; little or no disruption of sedimentary layering.

Zone 2: No relict textures preserved; layering generally continuous though locally distorted; no migmatization.

Zone 3: No relict textures preserved; layering tends to be discontinuous and lenticular; most metamorphic rocks have intercalations of fine-grained leucotrondhjemite.

The major mineralogic difference between zone 1 and zones 2 and 3 is the absence of epidote and calcite from the equilibrium assemblages in the higher grade rocks. Isopleths marking the disappearance of these two phases are essentially parallel to (see fig. III-9) and lie close to the boundary between zone 1 and zone 2.

Within zone 1, the following assemblages within the metasedimentary and metavolcanic rocks are present:

A. Quartz-free, plagioclase-bearing assemblages
   1. Epidote
   2. Biotite-hornblende
   3. Hornblende-epidote
   4. Hornblende-diopside-epidote
   5. Biotite-hornblende-calcite
   6. Biotite-hornblende-actinolite
   7. K-feldspar-biotite-hornblende-epidote-calcite

B. Quartz- and plagioclase-bearing assemblages
   8. Biotite
   9. Hornblende
   10. Biotite-hornblende
   11. Biotite-epidote
   12. Biotite-garnet
   13. Biotite-muscovite
   14. Hornblende-epidote
   15. Hornblende-diopside
   16. Hornblende-garnet
   17. Diopside
   18. Hornblende-epidote-calcite
   19. Biotite-hornblende-epidote
   20. Biotite-hornblende-calcite
   21. Biotite-actinolite-calcite
   22. Biotite-hornblende-epidote
   23. Hornblende-chlorite-epidote
   24. Biotite-hornblende-calcite-epidote
   25. Hornblende-epidote-diopside-calcite

In addition to showing evidence for textural disequilibrium, the rocks in zone 1 show some evidence for chemical disequilibrium. In particular, the anorthite content of the plagioclase shows a bimodal distribution. For the most part, the bimodality is related to the preservation of relict volcanic phenocrysts in the volcanic rocks and large clastic plagioclase grains in graywackes; generally, the phenocrysts are more calcic than the granoblastic groundmass.

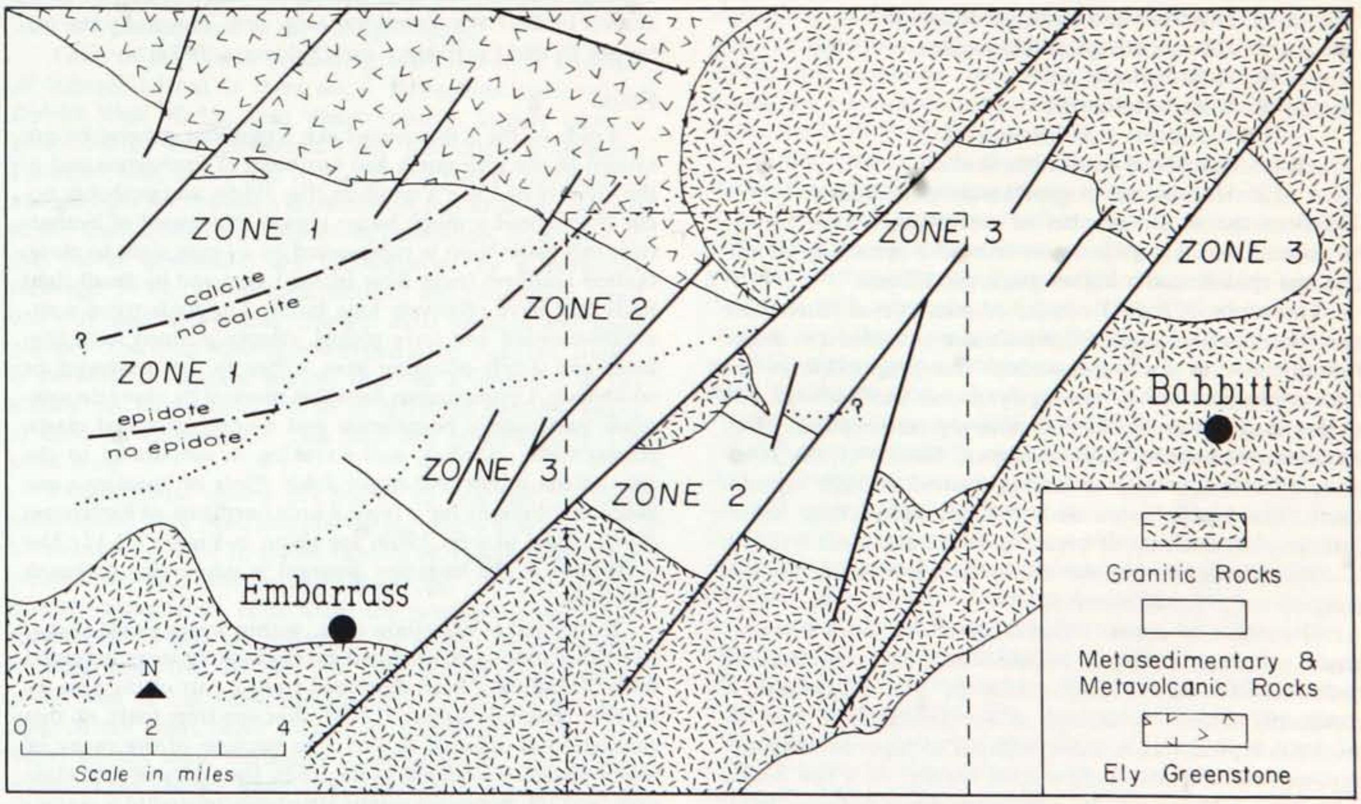

Figure III-9. Map showing boundaries of metamorphic zones and mineral isopleths in the Embarrass area.

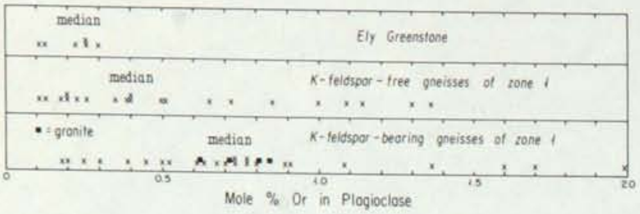

Figure III-10. Histograms of mole percent Or in analyzed plagioclases from zone 1, near Embarrass.

Histograms of the distribution of Or in analyzed plagioclases from zone 1 are shown in Figure III-10. Plagioclase coexisting with microcline has both a greater range and a higher median concentration (median = 0.71 percent Or) of Or component than plagioclase from rocks without microcline. The Or content of plagioclase from the Ely Greenstone is low (median = 0.24 percent Or), reflecting the low K_2O content of these rocks.

In zones 2 and 3, disequilibrium textures—zoned and antiperthitic plagioclase, myrmekite, and replacement of plagioclase by microcline—are widespread, and mineral assemblages cannot be rigorously determined. Common associations of non-retrograde minerals are as follows:

A. Quartz- and plagioclase-bearing assemblages
1. Biotite
2. Hornblende
3. Biotite-hornblende
4. Hornblende-plagioclase
5. Hornblende-diopside
6. Hornblende-plagioclase-diopside
7. Hornblende-plagioclase-quartz
8. K-feldspar
9. K-feldspar-biotite
10. K-feldspar-hornblende
11. K-feldspar-biotite-hornblende
12. Hornblende-plagioclase-diopside-quartz

The decrease in the number of assemblages from zone 1 to zones 2 and 3 partly results from the instability of calcite and epidote under higher grade conditions.

The rocks in zone 3 consist of interlayered leucocratic and amphibolitic gneisses, which are intruded by leucotrondhjemite to form migmatite. The plagioclase in the high-grade gneisses is complexly zoned and twinned and except in the mafic rocks is commonly antiperthitic. From textural and compositional evidence, Griffin (1969) interpreted the antiperthite as having formed through replacement. The biotite associated with the antiperthite shows petrographic evidence of breaking down, and Griffin (1969, p. 186) concluded that the minimum temperature for this reaction at P = 2,030 bars is 520° C.

The rocks of zones 1 and 2 show the series of retrograde reactions common to medium-grade metamorphic rocks (Griffin and Morey, 1969, p. 34). Plagioclase is commonly altered to sericite and/or fine-grained epidote, biotite is typically chloritized without changes in morphology, and hornblende is chloritized locally. In a few rocks, chloritized biotite contains pods of prehnite between folia. More generally, and especially in contorted biotite flakes, the pods consist entirely of low-Na (0.3-0.7 percent Ab) untwinned K-feldspar, rimmed in some cases by opaque granules. Wones and Eugster (1965) pointed out that this reaction might attend a rise in f_{O_2} at low temperatures.

The retrograde history of the gneisses of zone 3 is complicated by the effects of the metasomatism that produced the antiperthites. K_2O has been introduced into some amphibolites, altering hornblende to biotite and epidote along cleavage planes and shears. This metasomatic effect is most prevalent in the more schistose amphibolites. In some high-grade amphibolites, the hornblende is partially replaced by vermicular epidote; the epidote is separated from the hornblende by a narrow zone of cummingtonite. In addition to the reactions which may be linked to the metasomatic activity, the following generalized reactions have been observed: (1) chloritization of biotite; (2) breakdown of hornblende to chlorite and sphene; and (3) replacement of plagioclase by sericite and/or epidote.

Structure

In the western part of the district, two generations of folding and a younger generation of deformation including both faulting and kinking have been recognized (Hooper and Ojakangas, 1971). Similarly, in the eastern part of the district, at least two generations of folding can be inferred from interpretation of the geologic map of Gruner (1941), but their ages relative to the fold generations in the western part remain uncertain. The two generations of folding that have been identified in the western part are referred to herein as the Embarrass-Lake Vermilion and the Tower generations. Earlier, the two generations were described from studies in the Lake Vermilion area, near Tower, by the notations F_1 and F_2 (Hooper and Ojakangas, 1971).

Folds

Folds of the Embarrass-Lake Vermilion generation are known in the area north and northeast of Embarrass and in the vicinity of Lake Vermilion (fig. III-6), and probably occur throughout a much larger region. Northeast of Embarrass, this generation is represented by a large, tight to close, faulted antiform (near Bear Island Lake) and by small tight folds that have relatively long limbs. The folds trend west-northwestward and have planar, steeply-inclined axial surfaces and gently-plunging axes, either to the northwest or southeast. A conspicuous lineation marked by elongate minerals, particularly hornblende and biotite, stretched clasts, corrugations, boudins, and streaking is subparallel to the axes of the minor and major folds. Plots of lineations and poles to foliations for a typical area northeast of Embarrass (Griffin and Morey, 1969) are given in Figure III-11. The maximum in the lineation diagram is subparallel to major fold axes.

In the Lake Vermilion area, within lower grade rocks, this generation is represented by moderately large, refolded folds. These have been described from a part of the area by Hooper and Ojakangas (1971). Recognizing folds of this generation is difficult in this area because of the rarity of minor structures related to the folds, the rarity of fold closures, and the pervasive minor structures related to a younger deformation (Tower generation). The pattern of refolded

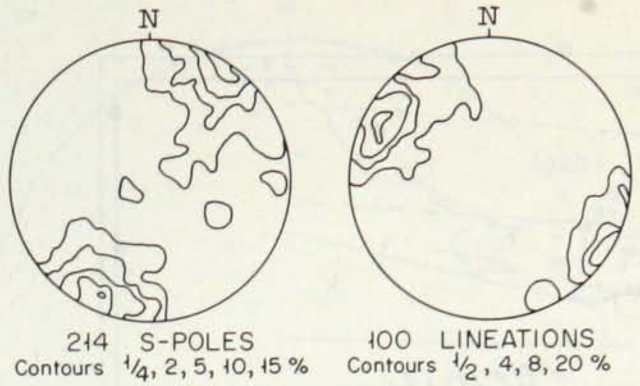

Figure III-11. Equal-area projections showing poles to foliation and lineation of Embarrass-Lake Vermilion structures in an area northeast of Embarrass.

folds of the Embarrass-Lake Vermilion generation has been defined rather accurately in the Echo Point-McKinley Park area, on the south shore of Lake Vermilion (fig. III-12). Here, several folds, spaced 1,000 to 1,500 feet apart, can be mapped on the basis of alternating top directions of graded beds in graywacke-slate, and these folds can be projected into areas containing other rock types. The folds trend either northwestward or northeastward, depending on the direction of closure of the younger, superposed folds. Judged from the map pattern, the major fold axes are nearly horizontal and the axial surfaces are steeply inclined. Some small-scale folds of this generation now have steeper axes as a consequence of the younger folding.

Folds of the Tower generation and small-scale structural features related to them occur throughout most of the district west of Ely, and where present, largely obscure structures of the Embarrass-Lake Vermilion generation. Probably, they also occur in the central part of the district, and account for the steep lineations in this area. In contrast to the Embarrass-Lake Vermilion generation, structures of the Tower generation are accompanied by a pervasive cleavage. In the Tower area, folds of this generation trend eastward and have planar, upright axial surfaces and steep plunges. The folds have an associated steep cleavage, roughly parallel to the axial planes, that has a tendency to fan around the larger folds and to be refracted between layers of different competence. The folds are of both large and small scale.

Larger folds, spaced one-fourth to one and one-half miles apart, have been mapped on the south shore of Lake Vermilion (fig. III-12), in the vicinity of Tower, and in adjacent areas (see Hooper and Ojakangas, 1971). They are strongly asymmetric, close folds; the dominant limbs trend northwestward and are separated by short northeastward-trending limbs (fig. III-13). The folds plunge steeply either to the west or east, depending on the attitude of the bedding. Minor folds occur locally on the limbs; Z-folds are found on the long, northwest-trending limbs, and S-folds occur on the shorter limbs. This pattern is consistent with the interpretation that the minor folds formed at the same time as the major folds.

Southwest of Tower, large folds of the type described above are absent, so far as known, but small asymmetric folds are abundant, and can be observed in almost all the larger outcrops. These are Z-folds that have long, straight northwest-trending limbs separated by highly crumpled north- or northeast-trending limbs (fig. III-14). Because this part of the area is on the southwestern limb of an older, tight anticline (Embarrass-Lake Vermilion generation), nearly all the Z-folds plunge moderately eastward and are downward-facing. Axial plane cleavage is neither as well developed nor as pervasive as in the area of larger scale folding. Lineations other than fold axes, on the other hand, are pervasive, and consist, in order of decreasing importance, of mineral alignment (biotite and hornblende), elongated particles, intersection of bedding-cleavage, mullions, and rodding, all of which are subparallel.

Contrary to an earlier interpretation (Hooper and Ojakangas, 1971, p. 434), folds of this generation are present in the middle amphibolite-facies rocks associated with the Vermilion batholith, on the north side of the Vermilion fault, as well as in the lower-grade rocks. In one area being mapped at present in the vicinity of Burntside Lake, a major antiform formed during this generation controls the distribution of the rock units. The axial surface of this fold trends about N. 70° E. and is inclined 80° N. Plunges along the axial surface trace are steeply inclined, either to the northeast or southwest, depending upon the attitude of the foliation (bedding). Small recumbent folds are found locally in the axial region. Minor, mainly S-folds, are common on the north limb. Associated lineations are mainly aligned minerals and elongated particles, but include rodding, corrugations, and boudins. Biotite granite, which intrudes and crosscuts the metavolcanic-metasedimentary rocks, has a foliation that is subparallel to the axial planes of the minor folds of this generation.

In the eastern part of the district, the metavolcanic-metasedimentary sequence trends either northwest or east to northeast, and apparently reflects at least two generations of folding, an older northwest-trending generation and a younger northeast-trending generation. These trends can be discussed with reference to Figure III-8. The rocks on the south side of the Saganaga batholith, which Gruner (1941, p. 1632) placed in his Gabimichigami Lake structural segment, are interpreted as having been folded during the older generation. They trend N. 60°-70° W., and apparently are deformed by tight to close folds on northwest-trending axes. Although exposures are poor, Gruner (1941, pl. 1) mapped several small folds that he regarded as comprising the Agamok synclinorium. The northwest-trending folded segment is truncated on the northwest by a major fault that separates this segment from rocks folded on northeast-trending axes. In the same way, the rocks in the Snowbank Lake area (fig. III-8) are folded on northwest-trending axes (see fig. III-58). The folds are tight to close and dominantly plunge southeastward. The largest of the folds is a syncline whose axis can be traced for a distance of at least 5 miles. Northwest of Snowbank Lake, just east of the northeast end of Moose Lake (see fig. III-58; Gruner, 1941, pl. 1), the rocks are deformed on northeast-trending axes, to produce a pattern resembling a basinal "eye." This pattern

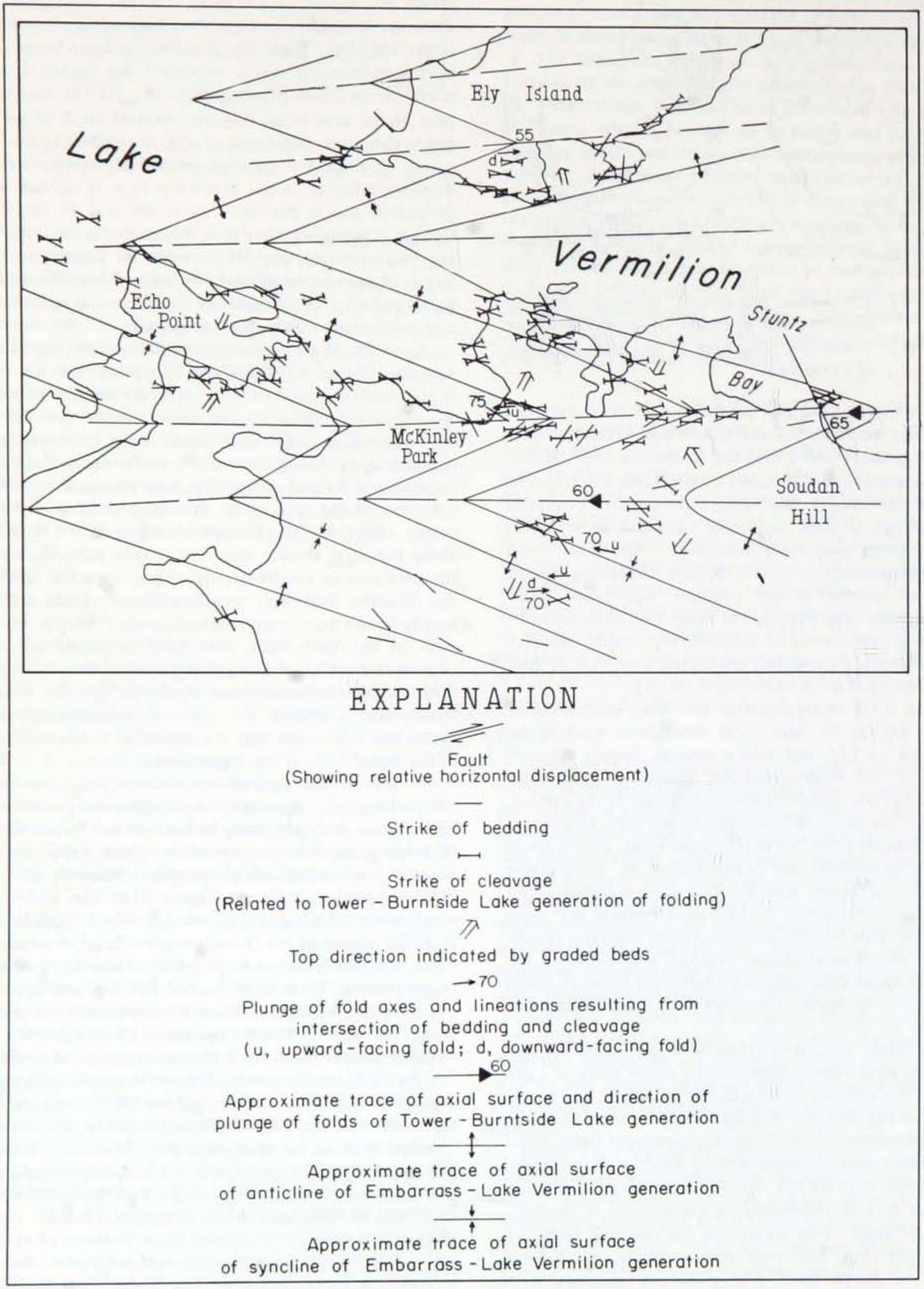

Figure III-12. Structure map of an area along south shore of Lake Vermilion.

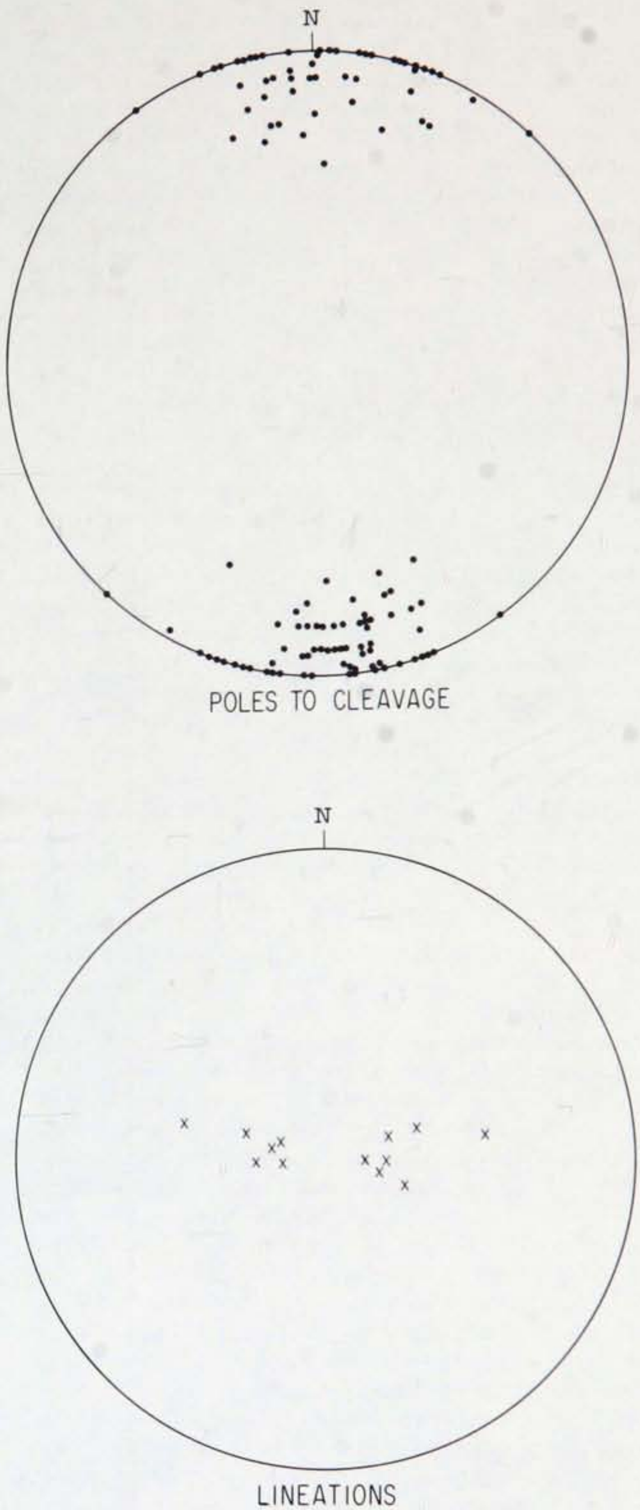

Figure III-13. Equal-area projections showing poles to cleavage (•) and lineation (x) of Tower generation structures in an area along south shore of Lake Vermilion.

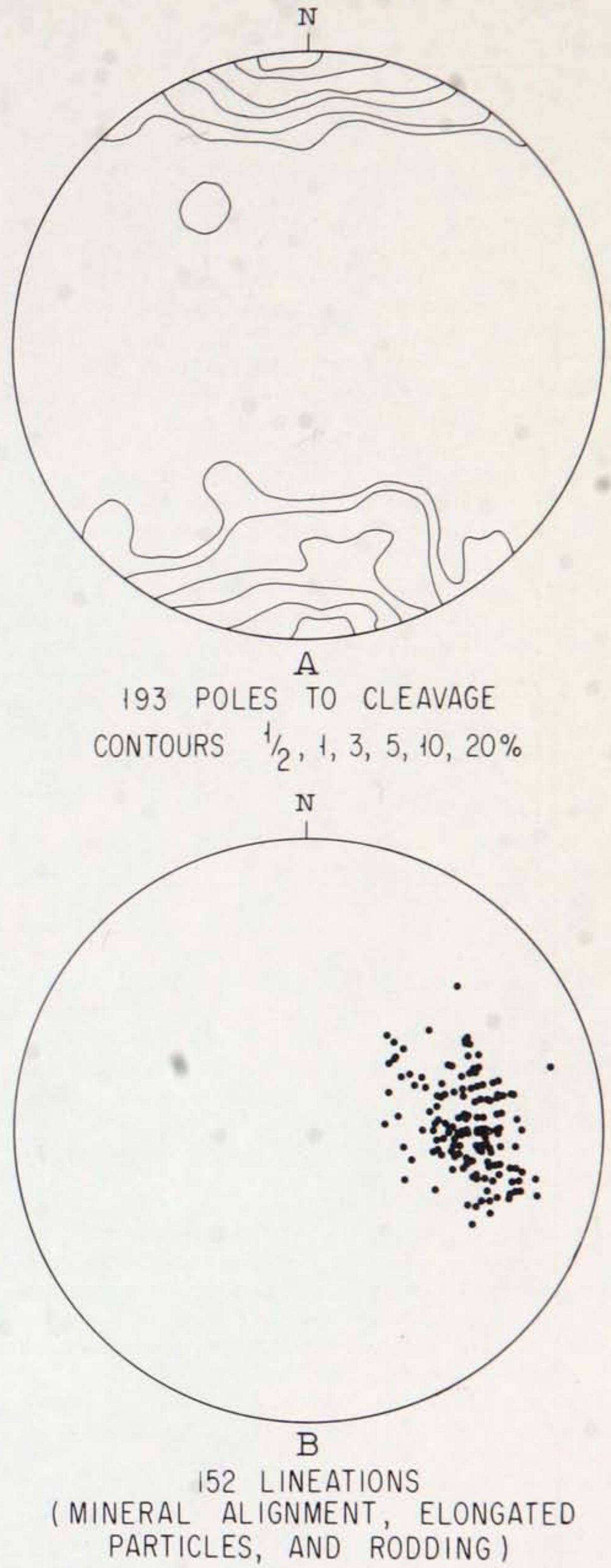

Figure III-14. Equal-area projections showing poles to cleavage and lineation of Tower generation structures in an area southwest of Tower.

is tentatively interpreted (M. G. Mudrey, Jr., 1971, oral comm.) as having resulted from two or more deformations having mutually perpendicular axial planes and fold axes. The older fold trends northwestward and has a vertical axial plane and a nearly horizontal fold axis; the younger fold trends northeastward, has a vertical axial plane, and plunges about 20° NE.

Evidence for multiple deformation of similar character is suggested by the map pattern (Gruner, 1941, pl. 1) on the west side of Kekekabic Lake. In this area, unit 11 of Gruner has the shape of a "mushroom" interference pattern, which can be accounted for by an older, northwest-trending, upright syncline having been refolded by a closely spaced east-northeast-trending anticline and syncline.

With this interpretation, remnants of the older, northwest-trending generation of folding remain in the Gabimichigami Lake and Snowbank Lake areas, and to a lesser extent elsewhere in the eastern part of the Vermilion district. The younger, northeast-trending folds dominate this region; judged from Gruner's field notebooks, these folds may not belong entirely to a single generation, for he described curved axial planes of northeast-trending folds and asymmetric minor structures that are contradictory to larger-scale structures. Remapping of this region using modern structural techniques should clarify the geologic history and contribute to a better understanding of the complex intrusive, stratigraphic, and structural relationships.

Faults and Related Fractures

Faulting had a profound effect on the rocks in the Vermilion district. The longitudinal faults slice the supracrustal sequence into several separate segments or blocks and at places cut out substantial amounts of both the upper and lower parts of the sequence; in addition, they greatly affected the distribution of the granitic rocks comprising the major batholiths and possibly partly controlled their emplacement. The transverse faults generally have lesser displacements, both horizontally and vertically, than the longitudinal faults, and mainly produced small or moderate offsets of rock contacts. Kink bands and joints that are related to the faults are widespread in the district, and are especially conspicuous in the granitic rocks.

Except in the Basswood Lake area and in the western part of the Lake Vermilion area (fig. III-6), the Vermilion fault effectively separates the Vermilion batholith and associated high-grade schists from the low-grade supracrustal rocks. Along this segment, the fault truncates the upper part of the metavolcanic-metasedimentary sequence; there is no granite on the south side of the fault. In the Basswood Lake area, on the southeast side of the fault, the Vermilion Granite intrudes and metamorphoses the Newton Lake Formation, but these rocks in turn are separated from low-grade rocks to the east by another longitudinal fault that is subparallel to the Vermilion fault. The same general pattern exists in the western part of the Lake Vermilion area (fig. III-6). Here, a satellitic pluton of Vermilion Granite, named the Wakemup Bay pluton, intrudes metagraywacke in a fault segment between the Vermilion fault and a parallel fault (Haley fault) to the south. The rocks south of the Haley fault are greenschist-facies rocks of the Lake Vermilion Formation. From the distribution of granite versus low-grade supracrustal rocks, it can be inferred that the north walls of the Vermilion fault and associated longitudinal faults have moved upward relative to the south walls, to bring granite into juxtaposition with greenschist-facies rocks. Probably, relative displacements on the order of a mile are required. It seems unlikely that horizontal displacements along the faults alone can account for the juxtaposition of contrasting metamorphic facies. An alternative explanation is that to some extent, the faults controlled emplacement of the granitic rocks, a suggestion made earlier by Griffin and Morey (1969, p. 38-39). With this hypothesis, the faults could have locally guided upward movement of magmas; the granite-wall rock interface would have been a likely site for later renewed movements along the fractures.

Another longitudinal fault that separates granite and associated high-grade schists from greenschist-facies rocks is the North Kawishiwi fault. It separates biotite schist and Giants Range Granite, on the south, from the Ely Greenstone, on the north. It truncates the lower part of the Ely Greenstone and cuts out an unknown amount. The probable displacement along the fault is puzzling. Assuming that the biotite schist on the south side of the fault (see fig. III-6) belongs to the Knife Lake Group, as tentatively inferred by Green (1970a, p. 75), the fault must have a stratigraphic throw of at least 12,000 feet, to bring Knife Lake rocks into contact with the lower exposed part of the Ely Greenstone. The displacement would have raised older rocks upward on the north against younger rocks. Yet, an opposite sense of movement is suggested by the presence of the higher grade rocks in the south wall of the fault; these rocks must have come from a deeper crustal level than those north of the fault. Possibly, there was some movement along the North Kawishiwi fault before granite intrusion.

Most other longitudinal faults in the western half of the district (see fig. III-6) as well as those in the eastern part (see fig. III-8) are subparallel to the strike of the rock units, and it is difficult to determine either vertical or horizontal displacements for them. That displacements generally were large is indicated, however, by the difficulty of correlating the rocks from one fault block to another. The problem is aptly discussed by Gruner (1941, p. 1622-1623): "The major longitudinal faults divide the district as a whole into long segments or belts, each one distinct in itself, but very difficult, if not impossible, to connect stratigraphically with any of the others. . . . When a unit of structure is completely delineated by faults it is left to one's imagination how it ever reached its present position, possibly by great displacements measurable in miles horizontally and thousands of feet vertically." A fault having a large displacement, probably largely strike-slip, is the Wolf Lake fault, southwest of Ely (fig. III-6). It truncates the Newton Lake Formation and the Knife Lake Group and brings them into fault contact with the upper (?) part of the Ely Greenstone. East of Ely, the contacts between the Ely Greenstone and the overlying Knife Lake Group and between the Knife Lake and the Newton Lake Formation are faulted at many places, and the Newton Lake Formation is sliced by several long, curvilinear faults that converge to the west. As shown in Figure III-6, several of the latter appear to splay from the Vermilion fault where the Vermilion fault changes strike from east to northeast.

The transverse faults that are known in the district (figs. III-3 and III-6) dominantly strike northeastward and have left-lateral strike-slip displacements. A few strike northwestward and have right-lateral strike-slip displacements. Most of the faults occur in a zone about 20 miles wide in eastern St. Louis County and adjacent areas in Lake County, on both sides of the Vermilion fault (see fig. III-3); others have been observed sporadically on the south side of the Vermilion fault as far west as Tower and vicinity. The Waasa fault has the greatest known displacement of any of the fractures; it has a left-lateral offset of 3¾ miles based on displacements of granite-gneiss contacts and an offset of 4½ miles based on displacements of internal contacts and fold axes in the gneisses (Griffin and Morey, 1969, p. 38). The nearby Camp Rivard fault has a left-lateral offset of at least 2 miles. Where the faults transect gneisses, they are marked mainly by topographic depressions; where they cut granite, however, they tend to form ridges as a result of intense silicification and cataclasis that accompanied the shearing. The zones of mylonite and silicification rarely exceed 100 feet in width. Several lines of evidence in addition to a greater offset of gneiss units than granite-gneiss contacts suggest some movement on the faults prior to intrusion of the Giants Range Granite (Griffin and Morey, 1969, p. 38): (1) locally, there are linear depressions parallel to the main faults that are marked by shearing and which do not extend into granite or offset the contact; and (2) several of the smaller faults and possibly also the Waasa fault apparently have localized small intrusions of Giants Range Granite; the granitic bodies are aligned along or parallel to observed faults; and (3) numerous granitic dikes are parallel to the strike of the faults, in both the gneisses and the granite itself, and have been sheared later along their strike.

The larger transverse faults on the north side of the Vermilion fault, as for example the Dead River and High Lake faults (fig. III-3), have left-lateral displacements on the order of half a mile. So far as known, granite-gneiss contacts are offset approximately the same amount as internal contacts in the metamorphic rocks. Many of the fault zones in this area have been gouged out by glacier ice, but where exposed they are seen to contain either broken, slightly altered rocks or cataclastic rocks.

Related to the faulting is a well-defined fracture pattern consisting mainly of joints and kink bands. All the fractures are steeply inclined, rarely deviating from the vertical by more than 20°. Data on the fracture patterns are available for two areas, the Tower area (Hooper and Ojakangas, 1971) and the Burntside Lake area, northwest of Ely (fig. III-15). In the Tower area (fig. III-15A), concentrations of fractures occur at orientations of approximately N. 5° E., N. 40° E., N. 80° W., and N. 30° W. In this area, the principal faults strike about N. 30° E.; a few faults, as at the Soudan mine (Klinger, 1956), strike about N. 30° W. Hooper and Ojakangas (1971, p. 425) noted that joints oriented N. 30° W. are common in the vicinity of the N. 30° E. faults. In the Burntside Lake area (fig. III-15B) north of the Vermilion fault, concentrations of fractures occur at orientations of approximately N. 5° W., N. 20° W., N. 40° W., and N. 60° W. The N. 40° and N. 60° W. joint sets are subparallel to steeply-inclined faults that occur between and in part at least transect the dominant N. 20°-30° E. fault set.

The longitudinal faults probably are shear fractures, for they are long and continuous and show evidence of intense shearing. The direction and cause of the shear stresses are not known. Possibly, the northeast-trending transverse faults are secondary fractures (Chinnery, 1966) that formed as a consequence of movement on the longitudinal faults. They appear to junction with the main longitudinal faults and to die out away from them; their pattern is consistent with their having formed in regions of overall tension.

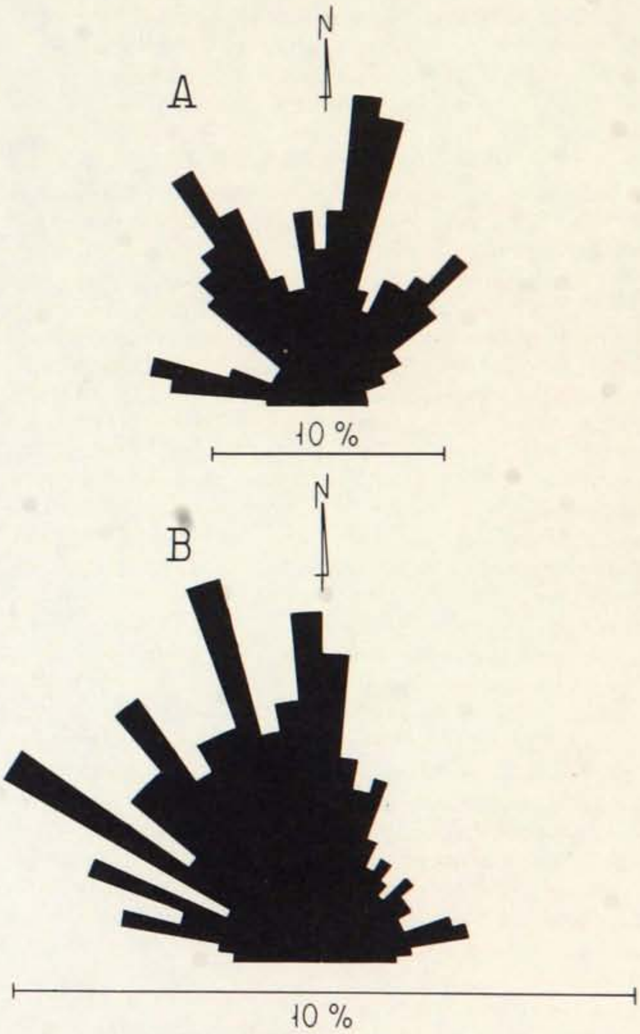

Figure III-15. Rose diagrams showing trends of steep minor faults, joints, and kink bands in Vermilion district. A, Tower area, after Hooper and Ojakangas, 1971, fig. 6; B, Burntside Lake area, north of Ely. 420 measurements.

The pattern of the longitudinal faults in the Vermilion district is remarkably similar to that in some of the much younger island arc-trench environments, as for example in the Puerto Rico region. There (Glover, 1971, fig. 34), the Greater Antilles fracture system consists of several west- or northwestward-trending faults that have dominant strike-slip movements of as much as 6 miles. In parts of the region, the major wrench faults are joined at acute angles by second-order shears having lesser displacements.

SUMMARY OF GEOLOGIC HISTORY

The early geologic history of the Vermilion district, insofar as known, can be summarized as follows:

(1) The first recognized event was mafic volcanism, accompanied by some pyroclastic discharge and intrusion of synvolcanic hypabyssal rocks. The volcanism was mostly under water, and magmas were mainly of basaltic composition. The nature of the basement rocks on which the volcanic rocks were deposited is not known. Small bodies of banded iron-formation formed locally in depressions on the surface of the volcanics, probably directly from volcanic emanations.

(2) The major Soudan Iron-formation and associated carbonaceous and tuffaceous sediments were deposited in the western part of the district during a quiescent period in the volcanism.

(3) Felsic pyroclastic discharge (volcaniclastic member, Lake Vermilion Formation) began abruptly in the west and mafic-intermediate volcanism was renewed in the west-central part of the district. Several thin, discontinuous banded iron-formations were deposited in areas of mafic volcanism, probably during quiescent periods. Contemporaneously, many hypabyssal intrusive bodies of felsic and intermediate composition intruded the volcanic pile.

(4) Clastic debris derived by erosion of the upper part of the volcanic edifice was deposited, in part contemporaneously with the felsic volcanism that produced the pyroclastic deposits of the Lake Vermilion Formation (in the west) and the Knife Lake Group (in the east).

(5) Mafic-intermediate volcanism was renewed in the central and eastern parts (Newton Lake Formation) of the district and sporadic mafic volcanism occurred during clastic sedimentation in the western part. Probably, some of the felsic volcanic deposits now designated as the Knife Lake Group were formed contemporaneously with the mafic volcanism.

(6) After an unknown interval of time, accumulated volcanic and sedimentary strata were folded and metamorphosed, virtually contemporaneously with early stages of emplacement of the batholithic rocks. Some faulting occurred approximately contemporaneously with emplacement of the batholiths.

(7) Near the end of major folding, small bodies of syenodiorite and related lamprophyres were emplaced. Probably, the Linden pluton, dated at about 2,700 m.y. (Prince and Hanson, in press), was emplaced slightly later; its age provides a lower limit for the time of folding, metamorphism, and magma generation associated with the Algoman orogeny.

(8) Faulting occurred on a regional scale, deforming local parts of all rock bodies and producing widespread cataclasis. Some of the faulting represented renewed movements on previously formed faults.

METAVOLCANIC AND ASSOCIATED SYNVOLCANIC ROCKS IN VERMILION DISTRICT

P. K. Sims

Metavolcanic rocks comprise most of the Ely Greenstone and the Newton Lake Formation and are common rock types in the Knife Lake Group and the Lake Vermilion Formation. In the following discussion, the metavolcanic rocks are referred to mainly in terms of their general chemical composition, inasmuch as only 28 chemical analyses are available. For those rocks that have been analyzed chemically, the names are based on the chemical classification of the common volcanic rocks by Irvine and Baragar (1971).

METAVOLCANIC SUCCESSIONS

The metavolcanic successions in the Ely Greenstone and the Newton Lake Formation are moderately well known as a result of recent mapping in the western half of the district. They consist dominantly of metabasaltic flows and intrusive metadiabase, and include pyroclastic rocks, some of which are reworked, porphyries of intermediate or felsic composition, and lesser derivative clastic rocks.

Ely Greenstone

Van Hise and Clements (1901, p. 402) named the Ely Greenstone from exposures at and near Ely, which include extrusive, intrusive, and fragmental rocks that are various shades of green because of contained secondary chlorite, amphibole, and epidote. Subsequently, Clements (1903) and most later workers applied the name to all major bodies of greenstone in the region, especially to pillowed metabasalts, regardless of stratigraphic position. As a result of recent mapping, however, the Ely Greenstone was redefined (Morey and others, 1970, p. 12) and restricted to that body of mafic metavolcanic and related rocks that is continuous with the exposures in the Ely area. As redefined, the Ely Greenstone extends from the vicinity of Tower northeastward nearly to Snowbank Lake, a distance of about 40 miles, and has an outcrop width of from 2 to 6 miles (fig. III-6). The formation is dominantly a steeply-dipping, northward-facing sequence. Folding is local, so far as known, and mainly confined to marginal portions in the Tower and Ely areas. The basal part of the sequence has been cut out by faulting and by the Giants Range Granite. The formation is overlain, apparently conformably, in the western part of the district by the Soudan Iron-formation and in the central part by the Knife Lake Group. The thickest homoclinal successions, so far as known, occur south-southwest of Ely, herein called the Twin Lakes area, and in the Gabbro Lake quadrangle (Green and others, 1966), just west of Snowbank Lake.

In the Twin Lakes area, an estimated 20,000 feet of strata are exposed. Neither the top nor the base of the formation is exposed; the base is cut out by granite and the top is truncated by the Wolf Lake fault (fig. III-6). The lowermost 9,000-10,000 feet of the succession is rather poorly exposed, but appears to consist mainly of metabasalt, much of which is pillowed. The succeeding 8,000 feet consists dominantly of metabasaltic flows with thin interbeds of chert and iron-formation. The iron-formation at the bottom of this part of the succession is the Soudan (Sims and others, 1968b); it contains interbeds of laminated tuff and agglomerate and is similar lithologically to iron-formation at the type area at Soudan. In this part of the district, at least 3,000 feet of pillowed metabasalt and metadiabase and lesser tuff were deposited above the Soudan Iron-formation. Included in these deposits is a distinctive black or dark-gray, carbonaceous, siliceous siltstone, probably a reworked tuff, which is as much as 600 feet thick; it is exposed discontinuously for a distance of 4 miles, and constitutes a good local marker bed (fig. III-16). It is overlain by and partly interbedded with a greenish-gray, sulfide-rich tuff, which is exposed in the vicinity of the Highway 169 overpass over the railway west of Ely. Metadiabase transects both the pillowed basalts and the bedded units.

An estimated 3,000 feet of strata belonging to the Ely Greenstone are exposed in the segment north of the Wolf Lake fault. Probably these strata are younger than the rocks on the south side of the fault, for they differ in lithology and are overlain by felsic tuffs of the Knife Lake Group. The lithologic units in this succession are shown in Figure III-16.

In the vicinity of Ely, the succession consists of pillowed metabasalt and lesser metadiabase, both of which are folded. These rocks pass southwestward along strike into pyroclastic rocks of intermediate composition and are overlain (on the northwest) by flows of andesitic or dacitic composition. This succession of lavas occurs directly below dacitic tuffs of the Knife Lake Group; the contact is in part faulted.

In the Gabbro Lake quadrangle (Green and others, 1966; Green, 1970a), a minimum of 12,000 feet of rocks belonging to the Ely Greenstone are exposed. They constitute a steeply-dipping, generally homoclinal succession that faces northward. The basal part is cut off by the North Kawishiwi fault; the formation is overlain by the Knife Lake Group, but the contact is faulted in most places. More than 90 percent of the rocks are metabasalt and metadiabase. The remainder are felsic volcanic rocks, chert and banded iron-formation, and metaclastic rocks. The succession across the Ely Greenstone in one part of the quadrangle, the Greenstone Lake area, is shown in Figure III-17. The lowermost 5,000 feet of the formation consists dominantly of metabasalt, which is intruded by small bodies of dacite and rhyodacite porphyries. Overlying this successive-

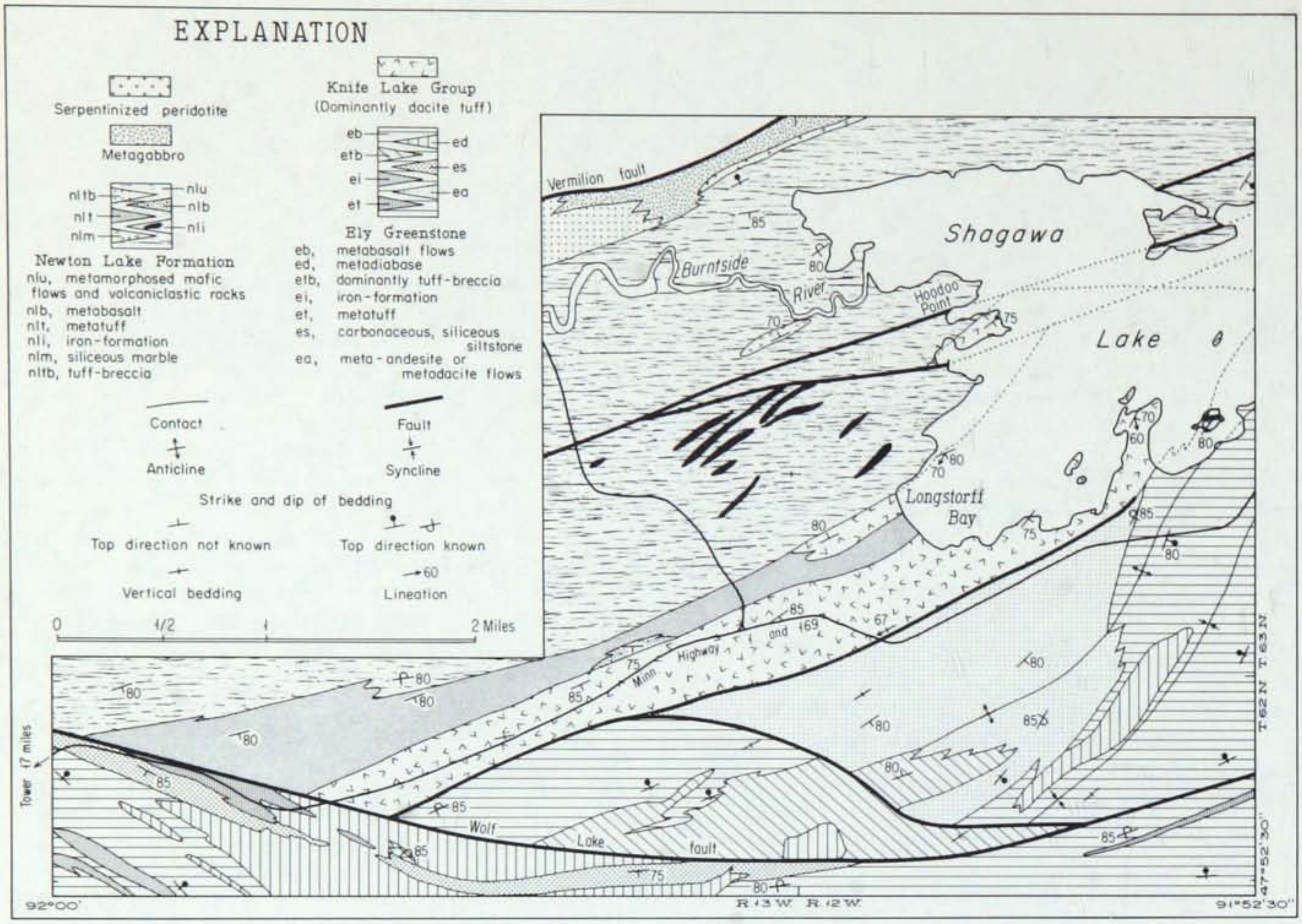

Figure III-16. Geologic map of southern part of Shagawa Lake quadrangle. Geology by P. K. Sims, 1967-70.

ly are a sill-like body of metadiabase (*ca.* 600 feet thick), metabasalt (3,000 feet thick), variolitic, pillowed metabasalt (200 feet thick), iron-formation (100 feet thick), and metadiabase (200 feet thick). Above these strata, the formation appears to consist mainly of metabasaltic lavas and metadiabase. To the east, near Jasper Lake (Green and others, 1966), an iron-formation that is underlain by spherulitic metabasalt and overlain by metadiabase probably represents the same succession described above, repeated by faulting. To the west, there are a few units of felsic lavas, and to the east a large, lenticular mass of dacite porphyry, which is interpreted as a lava dome; it is overlain by metabasalt and was a source for debris in interbedded clastic strata. Near the top of the Ely Greenstone there are a few thin, lenticular beds of conglomerate, breccia, and graywacke; and in the area near Snowbank Lake (fig. III-6) the upper part of the formation is characterized by interbedded pillowed basalts and volcanogenic conglomerates.

Newton Lake Formation

The Newton Lake Formation was named (Morey and others, 1970, p. 25) from exposures in the Newton Lake area (Green, 1970a) to include the body of dominantly mafic metavolcanic rocks that stratigraphically overlies the Knife Lake Group in the region between Ely and Moose Lake. As defined, the body extends from the vicinity of Wolf Lake, 8 miles west of Ely, northeastward to the vicinity of Basswood Lake, and is at least 30 miles long and 2 to 3 miles wide (fig. III-6). It consists of a mafic and a felsic member (Green, 1970a, p. 39-48). Formerly it was called Ely Greenstone, following the designation by Clements (1903), and was presumed to constitute the north limb of a syncline cored by the Knife Lake Group.

The Newton Lake Formation resembles the Ely Greenstone in gross aspects but differs from it in several respects: (1) the Newton Lake Formation contains a much greater proportion of felsic volcanic rocks than does the Ely Greenstone; (2) it contains several small lenses of impure siliceous marble and serpentinized peridotite, both of which are lacking in the Ely Greenstone; and (3) banded iron-formation and porphyries are uncommon in the Newton Lake, whereas they are widely distributed in the Ely Greenstone.

The formation is transected along its northwestern boundary by the Vermilion fault and, in the vicinity of Basswood Lake (fig. III-6), is irregularly intruded by granitic rocks of the Vermilion batholith (Green, 1970a, pl. 1). It is believed to be virtually conformable with the underlying strata of the Knife Lake Group east of Fall Lake and

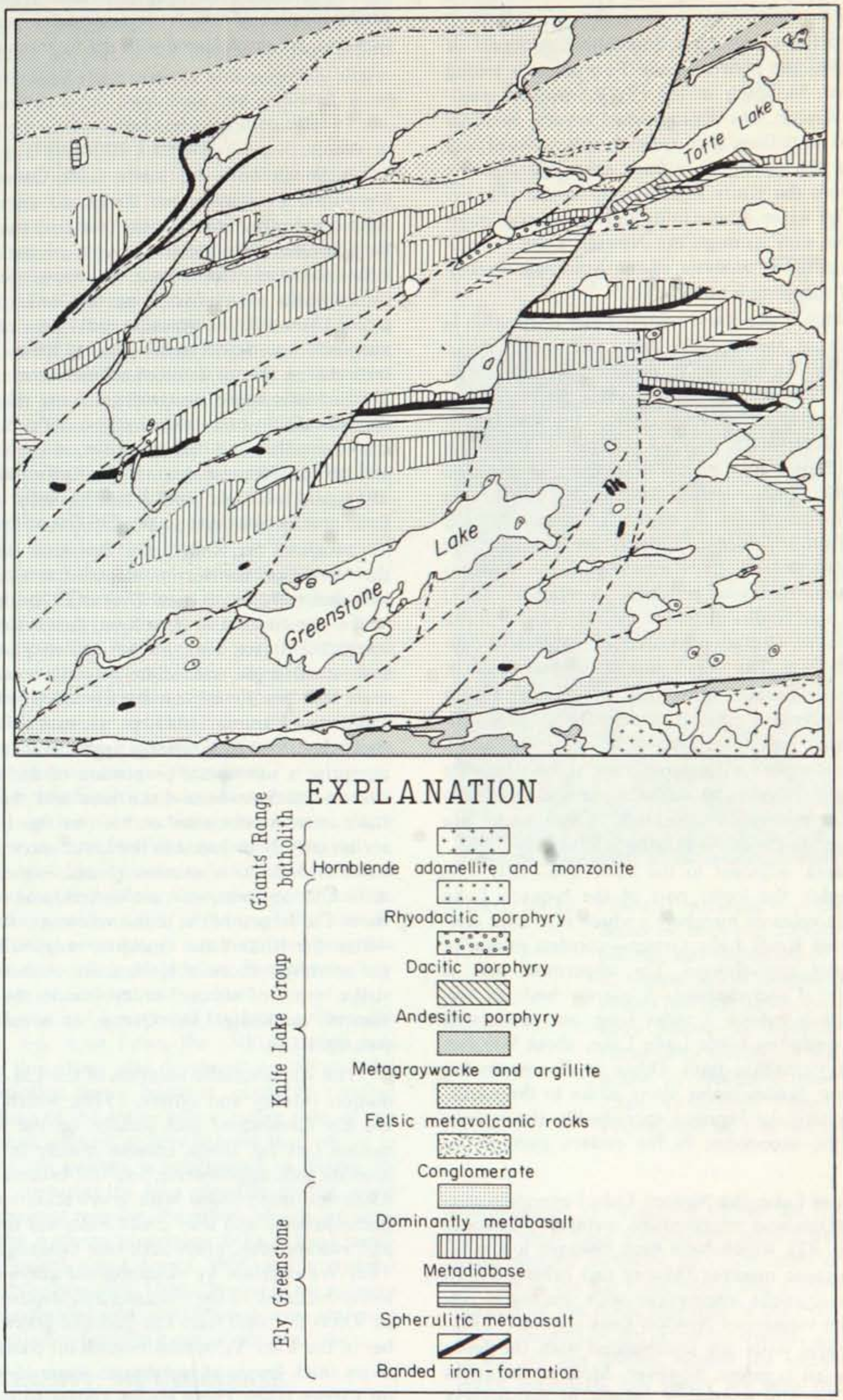

Figure III-17. Geologic map of Greenstone Lake area, Gabbro Lake quadrangle (after Green and others, 1966).

west of Shagawa Lake and to be separated from the Knife Lake by a fault in the intervening area (see Green, 1970a, p. 39).

In contrast to the dominantly homoclinal structure of the Ely Greenstone, the Newton Lake Formation is folded as well as faulted. Mapping in the Ely 7.5-minute quadrangle, by J. C. Green, has distinguished several isoclinal (?) synclines and anticlines that trend northeastward and are truncated to the south by a fault that separates the Newton Lake from the Knife Lake Group in the area between Ely and Fall River. These folds and the several longitudinal faults that slice through the Newton Lake Formation preclude a detailed analysis of the lithologic succession until further mapping is completed.

West of Newton Lake, the Newton Lake Formation is composed mainly of metabasaltic rocks and metadiabase, which characterize the informally designated mafic metavolcanic member (Morey and others, 1970, p. 25; Green, 1970a, pl. 1). Intrusive metadiabase is much more abundant than in the Ely Greenstone, and is particularly common in the upper stratigraphic part of the unit. West of Shagawa Lake (fig. III-16), the lowermost 5,000-6,000 feet of the Newton Lake Formation mainly consist of metabasalt and meta-andesite, both of which commonly are pillowed; the succession faces northward. At or near the base in this area is a nearly continuous crystal-tuff bed, as much as 1,000 feet thick. It is readily distinguished by abundant chlorite "eyes," which are an alteration product of amphibole crystals (see fig. III-19D). The upper part of the formation in this area, which is separated from the lower part by a fault of unknown displacement, consists principally of a greenish-gray quartz-bearing tuff, local beds of agglomerate, and a few thin lenses of impure siliceous marble; these strata are intruded by irregular bodies of metadiabase and small lensoid bodies of serpentinized peridotite. These rocks are rather well exposed north of the Burntside River.

In the Ely area, adjacent to the Echo Trail (Ely 7.5-minute quadrangle), the lower part of the Newton Lake Formation (mafic volcanic member)—which is in fault contact with the older Knife Lake Group—consists primarily of metabasalt and metadiabase. The uppermost part is composed largely of metadiabase. A narrow body of serpentinized peridotite, about 3 miles long, and associated gabbroic rocks, underlies Little Long Lake, about 900 feet southeast of the Vermilion fault. Other small, ultramafic, probably intrusive, bodies occur along strike to the northeast, and still others are exposed sporadically throughout lower parts of the succession in the eastern part of the quadrangle.

East of Newton Lake, the Newton Lake Formation consists mainly of felsic and intermediate metavolcanic rocks (Green, 1970a, p. 42), which have been assigned to the informal felsic volcanic member (Morey and others, 1970). The felsic volcanic rocks intertongue with the mafic volcanic rocks in the vicinity of Newton Lake (Green, 1970a, p. 40); a few mafic units are interbedded with the felsic rocks in the eastern segment, however. Most of the rocks are pyroclastic deposits consisting mainly of tuff-breccia and tuff, which have mineral and chemical compositions corresponding to andesite and dacite. Some of the breccias are scoriaceous. Massive lavas of dacitic and andesitic composition, some of which are amygdular, and pillowed dacitic lavas, are intercalated with the fragmental rocks.

Other Formations

Although both the Knife Lake Group and the Lake Vermilion Formation are composed dominantly of graywacke and other clastic rocks (see Ojakangas, this chapter), they contain substantial amounts of metavolcanic rocks. Felsic varieties predominate, and appear to be concentrated in, although not restricted to, the lower parts of the rock units. These definite volcanic rocks pass laterally as well as vertically into clastic debris derived from them, and are interpreted as the products of submarine eruption (see Fiske, 1969). Mafic lavas, many of which are pillowed, also occur sporadically in both units, as now defined. Generally these bodies are thin and discontinuous, as shown in Figure III-6 and on the geologic map of the Knife Lake area (Gruner, 1941, pl. 1). Gruner considered the mafic lavas in the Knife Lake area as belonging to the Ely Greenstone.

Much of the Knife Lake Group in the central part of the Vermilion district is composed of dacitic or andesitic tuff and tuff-breccia (see Green, 1970a, p. 32-36). These rocks constitute the dominant facies for the lowermost 1,000-2,000 feet, and, in the eastern part of the Gabbro Lake quadrangle, immediately overlie a polymict conglomerate that lies directly on the Ely Greenstone. In the Knife Lake area (Gruner, 1941, pl. 1), immediately to the east, rocks described by Gruner as agglomerate and conglomerate comprise a substantial proportion of the stratigraphic section in the Snowbank Lake area and the Gabimichigami Lake area. As discussed earlier (see fig. III-8), these rocks are tentatively included in the lower succession of the Knife Lake Group. At numerous places, higher in the section, tuffs and agglomerates are intercalated with graywacke-slate. The largest of the mafic volcanic units that are wholly within the Knife Lake Group as mapped by Gruner, is on the northwest shore of Knife Lake. Although thin, it has a strike length of about 9 miles. It is in the younger succession of the Knife Lake Group, as tentatively interpreted (see fig. III-8).

The volcaniclastic member of the Lake Vermilion Formation (Morey and others, 1970), which lies directly on the Ely Greenstone and, locally, on the Soudan Iron-formation (see fig. III-6), consists mainly of felsic and intermediate tuff, agglomerate, and tuff-breccia. At places these rocks are interbedded with graywacke, some of which is carbonaceous, and they grade westward into reworked tuff and volcanogenic graywacke (see Ojakangas, this chapter). They are overlain by volcanogenic graywacke. The maximum thickness of the volcaniclastic deposits is estimated to be 4,000 to 5,000 feet. The younger graywacke-slate member of the Lake Vermilion Formation contains several relatively thick lenses of metabasalt, some of which are shown on Figure III-6. These bodies appear to occur at different stratigraphic positions in the succession of graywacke-slate.

EXTRUSIVE MAFIC LAVAS

The mafic lavas are fine-grained green rocks, the weathered surfaces of which tend to bruise when struck with a hammer. At least 50 percent of the flows have pillow structures. The remainder are massive and generally slightly coarser grained (0.1-1.0 mm) than the pillowed flows (0.1-0.5 mm). Some of the metabasalts (as identified by texture, structure, and chemical analyses) are pale green, or even light tan, apparently as a result of bleaching by hydrothermal (?) solutions (Green, 1970a, p. 18).

Several different types of pillow lavas occur in the region. The most common type has rounded pillows that have average dimensions of $1 \times 1.5 - 2 \times 2\text{-}3$ feet, but are as much as 8-10 feet long and 3-4 feet thick. These pillows have chilled rims (rinds) that are generally 0.5-1 cm thick and typically paler than the interiors. Typical pillows are shown in Figure III-18. Interstitial material is locally muddy (tuffaceous?) material, chert, calcite, or pillow-rim breccia. Another common type of pillowed lava has dark chloritic rinds (altered palagonite ?) generally 1-2 cm across, and tends to have slightly flatter pillows than in flows of the first type. Typically, these pillows have a few vesicles in the outer zone. In the variolitic pillowed basalts, the variolites range in diameter from less than 1 to 3 cm, and generally increase in size toward the center of a pillow. Commonly the pillows also have a few vesicles. The pillows tend to be slightly larger than the types described above and to have thicker (1-3 cm) rims. The variolites are mainly composed of radially plumose actinolite, which in many rocks extends beyond the outline of the variolites until it meets and interferes with actinolite growing outward from other variolites. On weathered surfaces, the variolites tend to stand out in relief with respect to other parts of the pillows.

Typically, the metabasaltic rocks show intense retrograde alteration, making thin sections semi-opaque. Minerals recognizable in most sections are sodic plagioclase, quartz, actinolite, chlorite, epidote, calcite, leucoxene, and opaques. Relict augite and calcic plagioclase occur locally. An ophitic texture is recognizable in many of the least altered sections. Mineral assemblages resulting from prograde metamorphism include calcic plagioclase, hornblende, biotite, augite, opaque iron oxides, and epidote.

The pillow lavas are interpreted as having been extruded in water. For most flows, the chilling action was such that only a thin glassy skin developed on the globules of lava; the variolitic lavas, however, probably resulted from glassy chilling of the entire pillow rather than of just the rind. The small amount of lava breccia that occurs in the interstices of the pillows is interpreted as having resulted from the rupturing of lava globules during rapid chilling. In all essential respects, the pillow basalts are similar to those of the Archean greenstone belts in the Canadian Shield (see Henderson and Brown, 1966) and in younger terranes as well (Noe-Nygaard, 1940). Probably the massive lavas that are interbedded with the pillowed lavas represent basalt that was extruded beneath a cap of pillowed lavas.

INTRUSIVE METADIABASE

Concordant sill-like and irregular bodies of fine- to medium-grained metagabbro and metadiorite, herein called

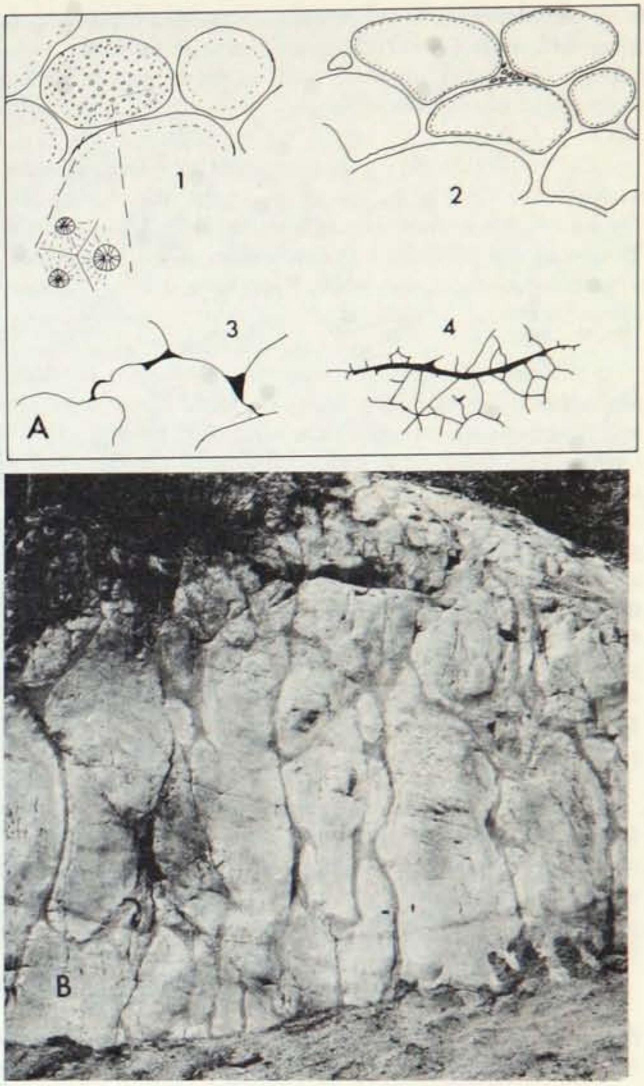

Figure III-18. Pillowed metabasalt in Ely Greenstone. A, sketches showing varieties of metabasalt pillows and variolites. 1, variolitic pillows; 2, typical pillow shapes; 3, irregular, thin-rinded pillows; and 4, contact between very large pillows or lobes in granular greenstone (after Green, 1970a, fig. 3). B, photograph of typical pillows, eastern end of Lake Vermilion. Tops are to the right. Photograph by J. L. McManus.

metadiabase, are common within the Ely Greenstone and the Newton Lake Formation. Judged from areas that have been mapped carefully, they constitute an estimated 15-20 percent of the Ely Greenstone and a somewhat greater proportion of the mafic member of the Newton Lake Formation. Contacts of the irregular bodies, and presumably also of the sill-like bodies, against the metabasaltic lavas generally are sharp, but rarely are seen because of poor exposures. The bodies are metamorphosed to the same extent as the mafic lavas and have similar compositions. They can be distinguished from the younger post-Algoman gabbroic dikes (see Sims and Mudrey, chapter IV) by being more pervasively altered, more irregular in shape, and in lacking distinct chilled margins.

The metadiabases are medium or dark green, massive, uniform, and tough. In the same way as the metabasalts, the weathered surfaces bruise when struck with a hammer. The rocks are difficult to distinguish in outcrop from massive metabasalts except where they have a coarse-grained ophitic texture.

The metadiabases differ from place to place in fabric, as a result of differences in primary compositions and textures and intensity of metamorphism. In the vicinity of Ely (Shagawa Lake and Ely 7.5-minute quadrangles), most of the metadiabase in the Ely Greenstone, which constitutes crosscutting bodies, is fine grained or at most medium grained and has distinct ophitic or subophitic texture. Typically, the rocks consist mainly of secondary minerals. Metadiabase from the center of sec. 5, T. 62 N., R. 12 W. (see fig. III-16), on the other hand, contains primary clinopyroxene, and although the plagioclase is completely saussuritized, the rock retains its primary subophitic texture. Pyroxene grains are strained and often broken, but not appreciably altered.

The metadiabases in the upper part of the Newton Lake Formation, in the Shagawa Lake quadrangle, are medium or coarse grained and commonly quartz-bearing. Although altered, they retain primary ophitic textures. Primary clinopyroxene is partly altered pseudomorphously to actinolitic hornblende.

In the Gabbro Lake quadrangle (Green, 1970a, p. 20), the most common texture of the metadiabases in the Ely Greenstone is relict poikilitic, in which relict augite oikocrysts (or actinolite pseudomorphs of augite) enclose small laths of saussuritized plagioclase. The actinolite grains and relict augite range in diameter from 3 to 8 mm, and commonly protrude slightly on weathered surfaces. Other bodies have a hypidiomorphic-granular texture with sub- to euhedral plagioclase laths and augite or hornblende prisms or actinolite pseudomorphs. Interstitial quartz and granophyre are present at places.

The close spatial association of the diabasic rocks with the basaltic lavas and their absence in felsic flows, the lack of distinct chilled borders, and the similarities in chemical composition and degree of metamorphism of the two mafic rock types are strong evidence for a coeval origin. Clearly, the diabases were intruded without disturbing the volcanic strata. Probably, the irregular bodies were emplaced virtually contemporaneously with the flows they intrude, and formed from basic magma that moved upward (?) through the partly crystallized flows but did not reach the surface. Such mafic intrusive bodies are reported from all Archean greenstone belts that have been studied in detail (Henderson and Brown, 1966; Weber, 1970). Although the bodies differ from place to place in the exact manner in which they intrude the volcanic strata, they have many common features that seem to confirm a common origin.

EXTRUSIVE FELSIC LAVAS

Metamorphosed lavas ranging in composition from sodic rhyolite to dacite occur locally in both the Ely Greenstone and the Newton Lake Formation. The felsic lavas probably constitute much less than five percent of the Ely Greenstone; because mapping has not been sufficiently detailed to differentiate felsic pyroclastic rocks from felsic flows in many parts of the Newton Lake, the relative abundance of felsic lavas in this formation is not known.

The extrusive felsites are greenish gray or light olive gray, dense, and tough. Weathered surfaces of the greenish-gray lavas are distinctly brownish; the chilled rims generally retain a greenish-gray color. Many felsites have small (1-2 mm) plagioclase and/or quartz phenocrysts which, because of alteration, are not readily visible megascopically. Nearly circular amygdules are not uncommon.

Some metadacite lavas are pillowed. The pillows are distinctly more bulbous than those in the metabasalts and generally are somewhat smaller. The chilled rinds are commonly 0.5-1.0 cm thick, but may be as much as 2 cm thick.

Massive felsite flows are common in the Newton Lake Formation (Green, 1970a, p. 42), and one has been identified in the Ely Greenstone. The latter is between Kempton and Sourdough Lakes, in the north-central part of the Gabbro Lake quadrangle (Green and others, 1966).

In the upper part of the Ely Greenstone, in the Jasper Lake-Moose Lake area (Gabbro Lake quadrangle), a large, irregular body of dacite porphyry is interpreted (Green, 1970a, p. 25) to be in part subaerially extrusive, although it is dominantly intrusive into metabasaltic rocks. Green suggested that the dacite lava dome (?) and flow at Jasper Lake may be essentially linked with the main part of the intrusive mass.

In addition to plagioclase, quartz, and hornblende phenocrysts, the felsites contain albite, chlorite, calcite, epidote, actinolite, sericite, sphene, opaque iron oxides, apatite, pyrite, and zircon. Textures range from felted to trachytic, depending on the habit of the plagioclase.

INTRUSIVE PORPHYRIES

Dikes and irregular small bodies of porphyritic rocks having compositions ranging from dacite to andesite are widely distributed in the upper part of the Ely Greenstone and the Soudan Iron-formation, and are found locally in the Knife Lake Group and the Lake Vermilion Formation. They are believed to be coeval with the volcanic rocks.

The porphyritic dacite is light or medium gray and rarely light olive gray, and weathers to light gray or white. Plagioclase is the most abundant phenocryst, and constitutes from about 10 to more than 40 percent of the rock

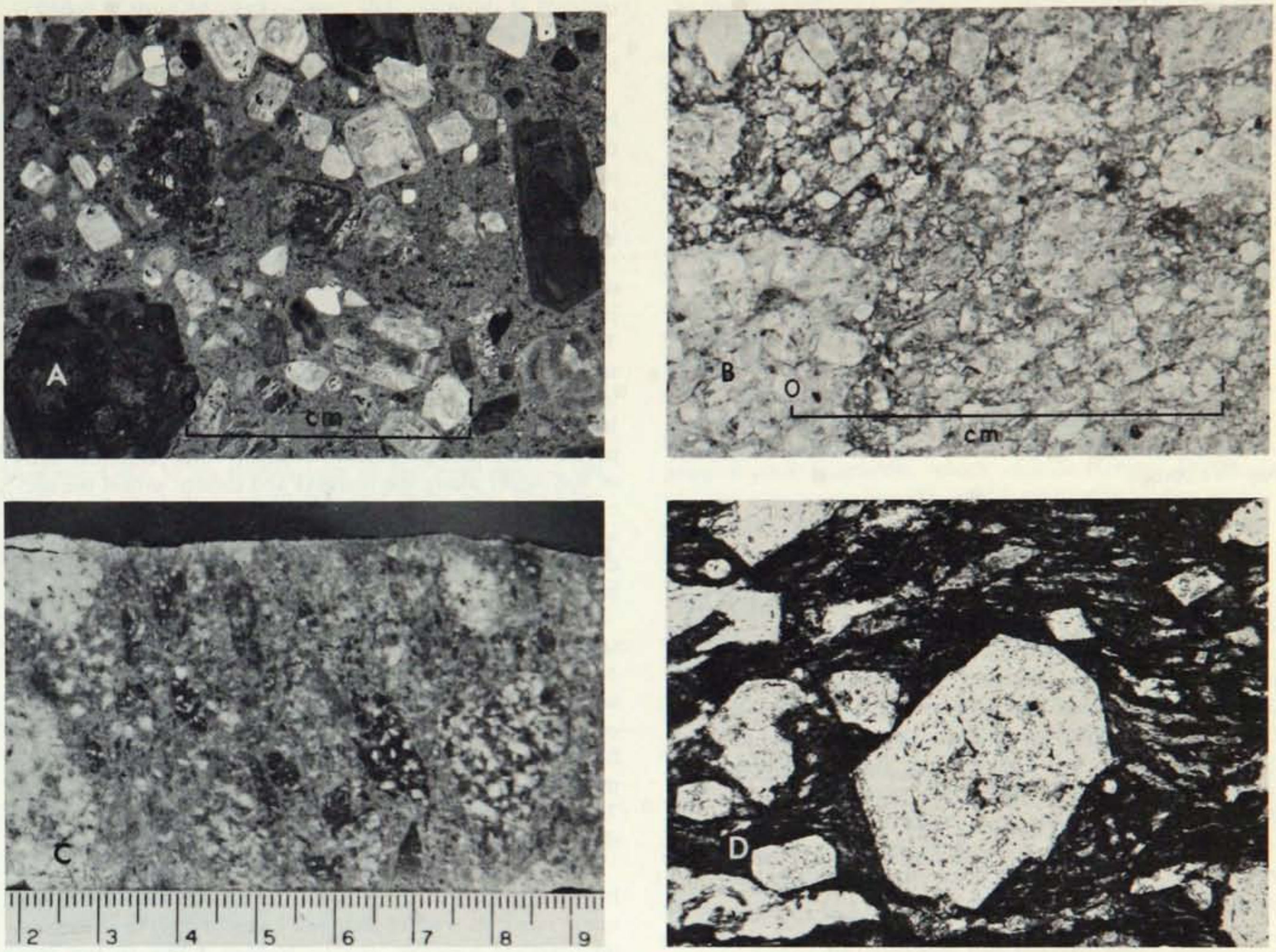

Figure III-19. Textures of typical volcanic and synvolcanic rocks. A, photomicrograph of typical dacite porphyry, M-7112; note large zoned, euhedral plagioclase grains and euhedral quartz grains; B, photomicrograph of felsic tuff in Knife Lake Group, M-7366; C, felsic-intermediate tuff-breccia in Knife Lake Group, M-7367; D, photomicrograph of crystal-tuff, Newton Lake Formation, west of Ely, Minnesota, ENW-379-D X20. (A through C after Green, 1970a; D, photomicrograph by R. B. Taylor).

(fig. III-19A). It occurs either as euhedral or subhedral, blocky albite crystals or as blocky zoned crystals with oligoclase cores and albite rims. The grain margins generally are somewhat serrated. The phenocrysts are slightly or strongly altered. Quartz phenocrysts are nearly ubiquitous, though less abundant than the plagioclase; generally they constitute 5 to 15 percent of the rock. The quartz occurs as nearly equant, subhedral to anhedral crystals from 1 to 5 mm across, some of which show resorption embayments. A few small bodies in the Gabbro Lake quadrangle (Green, 1970a, p. 49) contain potassic feldspar phenocrysts ranging in diameter from 3 to 10 mm. Hornblende phenocrysts, 0.2 to 1 mm across, which are largely retrograded to pseudomorphs of calcite, chlorite, and epidote, are common in the Gabbro Lake quadrangle but are rare in the Tower-Ely area. The groundmass of the porphyries is a microcrystalline, xenomorphic- or hypidiomorphic-granular aggregate of quartz and albite and, rarely, potassic feldspar. Generally, the groundmass appears less altered than the phenocrysts; alteration minerals include sericite, calcite, and lesser chlorite, epidote, sphene, opaque oxides, and sulfides. Most specimens from the Tower-Soudan area are cataclastically deformed, and locally this deformation has produced a moderately well developed foliation.

Judged from the occurrence of the porphyries, their emplacement overlapped in time the Ely Greenstone and older parts of the younger units. In the Gabbro Lake quadrangle (Green, 1970a, p. 48), porphyritic dacite intrudes the Ely Greenstone and the Knife Lake Group; and in the Jasper Lake-Tofte Lake area it apparently flowed onto the surface during late stages of Ely Greenstone time, and supplied detritus to interbedded conglomerate and graywacke.

A variety of porphyries that appear in hand specimen to be andesitic cut the Ely Greenstone and, locally, younger rocks. Nearly all the bodies are small, irregular plutons. Many of them have rather abundant plagioclase (albite-oligoclase) phenocrysts from 1 mm to 3 cm long, many of which are concentrically zoned. Typically, the plagioclase is more altered than in the dacitic porphyries. The tabular phenocrysts are roughly parallel, indicating a flow structure. Hornblende phenocrysts are next in abundance. Generally, they are 1-2 mm long, and some crystals are extremely ragged. Small quartz phenocrysts, about 1 mm in diameter, occur rarely. Both sphene and apatite form microphenocrysts. The groundmass is microcrystalline, and is composed of albite, quartz, epidote, opaque oxides, chlorite, zircon, calcite, and pyrite.

Clearly, the intrusive porphyries are related in space and time to the felsic-intermediate effusive rocks, and are the subjacent equivalents of some of the felsic volcanic rocks. The absence of chilled margins indicates that wall-rock temperatures were comparable to temperatures in the intrusive rocks, and supports intrusion as coeval dikes and plug-like bodies.

PYROCLASTIC ROCKS

Pyroclastic rocks occur locally in the upper part of the Ely Greenstone and are rather widespread in the Newton Lake and Lake Vermilion Formations and the Knife Lake Group. Judged from the available mapping, pyroclastic rocks constitute about five percent of the Ely Greenstone, 10-20 percent of the Newton Lake Formation, and at least 25 percent of the Knife Lake Group. Mapping of the Lake Vermilion Formation is inadequate to estimate the volume of pyroclastic rocks, but it is substantial in the area near Tower. The rocks are dominantly tuff-breccias, agglomerates, and tuffs.[1] Some of these rocks contain anomalous amounts of sulfide minerals.

Tuff-breccia and Agglomerate

Tuff-breccia and agglomerate form lenticular bodies from a few to several hundred feet thick and as much as several miles long. Most of them are composed of fragments having approximately the same composition as the enclosing lavas, and ranging in size from a few mm to a foot or more. The larger fragments are dominantly subangular but locally are elliptical. The smaller fragments are almost wholly angular. For the most part, deformation has not appreciably modified the dimensions of the blocks.

In the Ely Greenstone, moderately large deposits of tuff-breccia and agglomerate, with variable amounts of tuff, are known at three localities: (1) southwest of Ely, in the segment north of the Wolf Lake fault; (2) between Armstrong Lake and the east end of Lake Vermilion; and (3) west of Eagles Nest Lake. Only the body southwest of Ely has been mapped in detail.

The deposit southwest of Ely (fig. III-16) is a lenticular body as much as 2,500 feet wide and 2 miles long, and is in the upper part of the Ely Greenstone. Its stratigraphic thickness is not known. To the east, it interfingers with andesitic and dacitic lavas; to the west, it is cut off by the Wolf Lake fault. It is a heterogeneous pile of breccia, agglomerate, tuff-breccia, and tuff. The coarser rocks contain subangular or subrounded fragments from a few mm to as much as 10 inches in diameter in a greenish-gray, fine-grained matrix. The fragments are dominantly metadiabase but include a small proportion of intermediate effusive rocks. The matrix is approximately the same composition as the fragments. Intermediate lavas, which are mostly pillowed, occur at places within the pyroclastic deposit. Thin, lenticular beds of tuff occur along the margins and locally within the succession. These deposits contain substantial quartz. In the southwestern part of sec. 6, T. 62 N., R. 12 W. the pyroclastic rocks are intruded by a plug-like body of metadiabase.

A thin lens of greenish-gray tuff-breccia occurs on the south side of the Wolf Lake fault, in secs. 3 and 4, T. 62 N., R. 13 W. This rock overlies and locally is interbedded with a remarkably continuous bed of dark gray, carbonaceous, siliceous siltstone that probably is a reworked tuff. The fragments in the tuff-breccia are dominantly fine-grained metadiabase but include a quartz-bearing hypabyssal rock. The rocks contain a small percentage of disseminated pyrrhotite and traces of chalcopyrite and sphalerite.

Tuff-breccia and agglomerate are moderately abundant in two segments of the Newton Lake Formation: (1) in the mafic member, west of Shagawa Lake; and (2) in the felsic-intermediate member, east of Newton Lake. In the area west of Shagawa Lake, tuff-breccia and agglomerate occur in the upper part of the formation adjacent to the Vermilion fault and north of the Burntside River. The rocks are exposed along a strike length of nearly 4 miles and are at least 2,000 feet thick. Associated with the pyroclastic rocks are two lenses of impure marble and intrusive metadiabase and serpentinized peridotite. The pyroclastic rocks are greenish gray or grayish green, poorly sorted, and generally have a conspicuous cleavage along which the rock breaks. The fine-grained parts have a faint color banding and contain scattered quartz fragments. The agglomerates contain elliptical fragments of dacite porphyry as much as 12 inches in diameter in a matrix of intermediate tuff.

Tuff-breccias and related rocks also are common in that part of the Newton Lake Formation exposed on the west shore of Shagawa Lake, between Hoodoo Point and Longstorff Bay (fig. III-16). These are mainly intermediate in composition, and some are intensely cataclasized.

East of Newton Lake, the felsic member is made up dominantly of interbedded dacitic to andesitic tuff-breccias, tuffs, and some lava flows. It appears to be a few thousand feet thick at least. The clasts typically are plagioclase-bearing porphyritic rocks; quartz phenocrysts occur in a few units and chloritic clots, probably after hornblende phenocrysts, are common.

[1]The terminology used here is modified from Fisher (1961). *Tuff-breccia* refers to pyroclastic rocks containing coarse ($>2<64$ mm) fragments in a fine-grained (<2 mm) tuff matrix. The proportion of large pyroclastic blocks and fine-grained material varies considerably both along and across strike. The fragments are mainly angular or subangular. *Agglomerate* refers to rock aggregates composed mainly of large (>64 mm) pyroclastic fragments that are rounded, presumably by volcanic processes, whereas *tuff* refers to those rocks composed dominantly of small (<2 mm) pyroclastic fragments. All volcaniclastic material except that derived by weathering and erosion of older rocks is classed as pyroclastic.

Much of the Knife Lake Group in the central Vermilion district is composed of primary dacitic to andesitic tuff and tuff-breccia as well as reworked tuffs and tuffaceous sediments (see Green, 1970a, p. 32-36). Typical specimens are shown in Figures III-19B and C.

Tuff

Tuff of intermediate composition occurs locally in each of the major rock units. A distinctive tuff bed that forms a persistent stratigraphic marker occurs at the base of the Newton Lake Formation in the area west of Shagawa Lake (fig. III-16). It lies directly on the underlying dacitic tuff of the Knife Lake Group and, where observed, the contact appears to be conformable. Locally, pillowed lavas intervene between the two units. The bed is overlain by mafic lava, most of which is pillowed. The bed is exposed for a distance of 4 miles along strike, from Longstorff Bay in Shagawa Lake to its termination to the west against the Wolf Lake fault. It is about 600 feet thick near Shagawa Lake, and thickens westward, locally to more than 1,000 feet. The bed is a dark greenish-gray, massive crystal-tuff that contains a small percentage of coarse crystals of a zoned amphibole in a wavy, thin-bedded matrix composed of plagioclase, calcite, and unidentified semi-opaque materials (fig. III-19D). In most outcrops, the rock is schistose, and the amphibole is largely altered to dark-green chlorite, which produces a distinctly spotted appearance.

Much of the felsic-intermediate member of the Newton Lake Formation, at and east of Newton Lake, is composed of tuffs interbedded with coarser pyroclastic rocks and some lavas. Two or three thin beds of silicified felsic tuff, interbedded within more mafic lavas, continue southwestward from Newton Lake.

A large proportion of the Knife Lake Group is composed of tuff, generally of dacitic to andesitic composition. Some of the tuff may be reworked, and it appears to grade into volcanogenic graywacke (Green, 1970a, p. 32-36).

Probably the tuffs formed on the sea bottom during intervals between outpouring of the lava flows. Their greater abundance in the upper part of the Ely Greenstone and in the younger formations and their general absence in more mafic parts of the Ely reflect the tendency for the volcanic activity to become more explosive when the parent magmas attained intermediate and salic compositions.

EPICLASTIC ROCKS

Immature metaclastic rocks that are epiclastic[2] in origin are locally interbedded with the metavolcanic rocks of the Ely Greenstone and the Newton Lake Formation, and are common in the upper succession of the Knife Lake Group (see fig. III-8). These rocks can be distinguished by the presence of plutonic rock fragments in the strata. Except for the upper part of the Knife Lake Group, epiclastic rocks probably constitute a small percentage of the mechanically deposited sediments in the volcanic units. Instead, most of the metasedimentary rocks in the units appear to be reworked tuffs.

[2] *Epiclastic* is used here for mechanically deposited sediments that consist of products derived by weathering and erosion of older lithified or solidified rocks.

A conglomerate in the Ely Greenstone that contains a small percentage of granitic clasts in a dominant assemblage of volcanic material was found by Green (1970a, p. 25) in SE¼ sec. 8, T. 63 N., R. 10 W., Gabbro Lake quadrangle (Green and others, 1966). The granitic clasts, which could be distinguished only in thin section, consist of both granophyric and coarser intergrowths and independent crystals of quartz, potassic feldspar, and plagioclase. The matrix of the conglomerate consists of sand- and clay-size particles of the same dominantly metabasaltic materials as the clasts. Several small lenses of conglomerate in the upper part of the Ely Greenstone contain clasts of metabasalt, chert, and dacite porphyry; also, rare beds of feldspathic to chloritic metagraywacke occur in the upper part of the Ely (Green, 1970a, p. 26).

A coarse conglomerate interbedded with pillow metabasalt in the Newton Lake Formation contains a variety of granitic rock fragments, which include (Green, 1970a, p. 46) (1) a massive, quartz-poor hornblende-biotite granodiorite and (2) a foliated, coarser grained, quartz-rich tonalite. The source of the plutonic clasts is not known. Also, the Newton Lake Formation contains local lenses of metagraywacke as well as a thick (>1,000 ft.) folded succession of metagraywacke and conglomerate near Camp Lake (sec. 28, T. 64 N., R. 11 W.).

CHEMICAL COMPOSITIONS

Twenty-eight chemical analyses of metavolcanic and synvolcanic rocks in the Vermilion district are available. Of these, 20 are from the Ely Greenstone (table III-3), six from the Newton Lake Formation, and one each from the Lake Vermilion Formation and the Knife Lake Group (table III-4). All but four samples represent flows or shallow intrusive bodies. Twenty-one of the samples are from the Gabbro Lake 15-minute quadrangle (Green, 1970a, p. 87-90); the others are random samples collected earlier by Winchell (1895) and by Grout (1926). Although the samples are not adequate for determining systematic regional or stratigraphic chemical characteristics and trends, they are thought to be representative of the more common lava types, especially in the Ely Greenstone. The analyzed samples are wholly inadequate for determining the chemical characteristics of the volcaniclastic rocks.

The rocks are classified on the basis of their chemical composition according to the scheme proposed by Irvine and Baragar (1971) for common volcanic rocks. All the rocks are subalkaline, as distinguished by a plot of (Na_2O+K_2O) against SiO_2 (fig. III-20A). They show characteristics of both the tholeiitic basalt series and the calc-alkaline series (fig. III-20B). Most of the analyzed samples are basalt (fig. III-20C); and of these, the majority are quartz-normative and all are hypersthene-normative. One sample (M-7250) is a picritic basalt; a few are olivine basalt or olivine tholeiite. With some exceptions (fig. III-20D), the analyzed samples contain less potassium than "average" rocks, as determined by Irvine and Baragar (1971, fig. 8). The felsic rocks are sodic dacites and sodic rhyolites.

Judged from the available analyses (table III-3), the lavas in the Ely Greenstone can be characterized as repre-

Table III-3. Chemical analyses, in weight percent, and norms of metavolcanic and hypabyssal intrusive rocks in Ely Greenstone (compiled from Green, 1970, except for nos. 1, 2, 3, 4, 6, and 7; rock names based on chemical classification of Irvine and Baragar, 1971).

	M-7258	M-7286	EG-16	M-7360a	EG-26	6	M-7228	7	EG-17	M-7251	M-7201	EG-27	4	3	M-7288a	M-7509	M-7202	M-7112	1	2
SiO_2	39.71	48.15	49.30	49.55	49.65	49.72	50.14	50.47	50.90	51.06	51.20	51.30	51.73	52.94	69.11	69.55	60.95	66.75	62.01	60.61
TiO_2	0.87	0.84	0.79	0.98	0.78	0.89	1.59	n.d.	1.33	1.44	1.47	0.76	0.78	n.d.	0.61	0.37	0.51	0.28	0.45	n.d.
Al_2O_3	19.70	13.80	14.68	16.05	15.44	16.76	13.54	18.48	16.05	13.85	14.91	14.29	15.28	14.70	16.24	17.05	15.14	15.56	16.72	16.61
Fe_2O_3	3.62	2.34	4.42	2.45	3.22	1.92	2.77	2.13	2.33	1.79	2.18	4.46	3.41	2.52	1.10	0.62	4.34	1.42	1.53	1.97
FeO	8.46	10.20	8.04	8.32	7.96	7.33	11.81	7.74	8.66	10.86	9.16	6.76	7.30	7.80	1.37	0.45	1.16	1.20	1.99	5.09
MnO	0.16	0.22	0.22	0.22	0.18	0.16	0.30	n.d.	0.18	0.22	0.25	0.19	0.15	n.d.	0.02	0.01	0.07	0.03	0.05	n.d.
MgO	9.93	8.59	6.78	4.66	7.42	7.62	5.07	6.90	6.15	5.31	5.45	7.53	6.72	4.49	0.98	0.68	4.38	0.92	3.70	3.10
CaO	9.67	8.97	10.42	11.94	9.56	9.35	8.25	6.61	9.30	10.66	7.60	9.54	9.40	6.56	2.63	1.77	3.32	3.18	5.93	4.46
Na_2O	1.40	2.70	1.90	2.04	1.64	3.14	2.95	2.58	2.06	2.49	2.78	1.69	3.83	3.09	3.36	5.84	5.48	5.60	5.20	3.11
K_2O	0.04	0.10	0.12	0.62	0.32	0.71	0.25	0.30	0.76	0.25	0.00	0.16	0.76	0.04	2.09	1.72	2.12	1.73	0.69	0.25
H_2O+	6.10	3.81	2.80	2.34	3.02	1.57	2.95	2.34	1.74	1.83	3.93	2.83	2.86	2.04	2.19	1.81	1.83	1.42	1.08	2.45
H_2O	0.20	n.d.	n.d.	0.12	n.d.	0.06	0.31		n.d.	0.06	n.d.	n.d.	0		0.30	0.18	n.d.	n.d.	0.06	
CO_2	0.80	0.00	0.17	1.39	0.05	0.10	0.06		0.00	0.33	0.74	0	n.d.	4.86	0.30	0.58	0.43	1.62	0.07	1.57
P_2O_5	0.07	0.16	0.08	0.09	0.13	0.09	0.24		0.20	0.13	0.14	0.12	n.d.		0.17	0.09	0.13	0.07	0.17	n.d.
S	n.d.	n.d.	n.d.	n.d.	n.d.	0.04						n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	0.28	
BaO	n.d.	n.d.	n.d.	n.d.	n.d.	0.06														
						99.52													99.93	
Less O						.01													.10	
Total	100.73	99.88	99.72	100.77	99.37	99.51	100.23	97.52	99.66	100.28	99.81	99.63	102.22	92.04	100.47	100.72	99.86	99.78	99.83	99.22
Q			5.4	3.1	5.6		3.8	4.4	4.8	3.2	7.3	9.6		10.9	35.5	24.8	9.2	20.7	12.5	25.1
or		0.6	0.6	3.9	2.2	4.3	1.7	1.8	4.4	1.7		1.1	4.5	0.2	12.8	10.6	12.8	10.6	4.1	1.6
ab	12.6	23.6	16.8	17.8	14.1	27.2	25.7	22.9	17.8	21.5	24.6	14.7	32.6	28.3	28.8	50.3	47.2	48.7	44.7	27.7
an	50.6	26.4	32.2	33.9	35.0	30.2	23.6	34.4	33.1	26.4	29.8	32.0	22.4	28.4	12.2	8.3	10.8	12.8	20.6	23.2
c								2.0							4.2	2.6				3.3
di		15.5	16.7	22.1	10.9	13.6	14.2		10.3	22.2	7.1	13.0	19.9	4.8			4.1	2.4	6.6	
hyp	4.4	20.6	19.8	13.4	25.6	9.6	23.8	31.1	22.9	19.2	24.5	21.1	6.9	23.3	3.2	1.7	9.2	1.9	8.0	16.2
ol	24.6	7.7				10.2							7.2							
il	1.8	1.7	1.5	2.0	1.5	1.7	3.2		2.6	2.7	2.9	1.5	1.5		1.2	0.8	1.1	0.6	0.9	
mt	5.6	3.5	6.7	3.7	4.9	2.8	4.2	3.2	3.5	2.6	3.2	6.7	5.0	4.0	1.6	0.5	2.6	2.1	2.2	3.0
hm																0.3	2.7			
ap	0.2	0.4	0.2	0.2	0.3	0.2	0.6		0.5	0.3	0.4	0.3			0.4	0.2	0.3	0.2	0.3	
Analyst	H.A.	K.R.	K.R.	H.A.	K.R.	E.O.	H.A.	C.S.	K.R.	H.A.	K.R.	K.R.	S.D.	C.S.	H.A.	H.A.	K.R.	K.R.	D.T.	C.S.

Analysts: H. A.: Hiroshi Asari, Japan Anal. Chem. Res. Inst., Tokyo, 1967; K. R.: K. Ramlal, Univ. of Manitoba, 1968; C. S.: C. F. Sidener; S. D.: S. Darling; E. O.: Eileen H. Oslund; D. T.: Doris Thaemlitz

M-7258: Pale gray-green, fine-gr., pillowed greenstone (high alumina picritic basalt); Fernberg rd. old jct. with eastern Moose L. rd., S½ sec. 6, 63N/9W

M-7286: Gray-green, massive, fine-gr. greenstone with relict ophitic augite lumps (K-poor olivine tholeiite); Fernberg rd., ¼ mi. W of BM 1419 in NW¼ sec. 9, 63N/10W

EG-16: Med. to pale green, fine-gr., pillowed greenstone with few small amygdules (quartz tholeiite); E shore Triangle L. in SW¼ sec. 13, 63N/10W

M-7360a: Dark green, fine-gr., pillowed greenstone with relict labradorite laths (K-rich quartz tholeiite); logging rd. S from Fernberg rd. in SW¼ sec. 8, 63N/9W

EG-26: Gray-green, massive, fine- to medium-gr. greenstone with relict ophitic texture (quartz tholeiite); E shore Triangle L. in SW¼ sec. 13, 63N/10W

6: Pillowed metabasalt (olivine basalt), NW¼ SE sec. 9, 61N/14W; collected by J. W. Dalrymple (M4011); previously unpub.

M-7228: Dark green, dense, variolitic, pillowed greenstone (K-poor quartz tholeiite); N shore Garden L. in NE¼ sec. 20, 63N/11W

7: Greenstone (quartz basalt) NW¼ sec. 4, 63N/9W; Ref.: N. H. Winchell, 1895, p. 123

EG-17: Dark green, massive, fine-gr. subdiabasic basalt, conformable beneath pillows with fresh labradorite and augite and pseudomorphs after olivine (K-rich quartz tholeiite); E shore Triangle L. in SW¼ sec. 13, 63N/10W

M-7251: Dark green, fine-gr., pillowed greenstone with few small labradorite phenocrysts and devitrification texture (quartz tholeiite); Fernberg rd. nr E edge of Gabbro L. quad., SW¼ sec. 8, 63N/9W

M-7201: Pale gray-green, dense greenstone (K-poor quartz tholeiite); SW of Wood L. in SW¼ sec. 34, 64N/10W

EG-27: Gray-green, massive, fine-gr. greenstone with few small amygdules (quartz tholeiite); E shore Triangle L. in SW¼ sec. 13, 63N/10W

4: Greenstone, average of three phases at Ely, Minn. (K-poor olivine basalt); Ref.: Grout, 1926, p. 12, table 1, no. 1

3: Greenish felsite (K-poor quartz basalt), country rock at Ely, Minn.; Ref.: N. H. Winchell, 1895, p. 204 (no. 221); CO_2 eliminated in calculating norm

M-7288a: Light gray, dense, massive felsite flow (?), few amygdules (dacite); powerline nr Fernberg rd., in SE¼ sec. 7, 63N/10W; no K-spar detected

M-7509: White, dense, trachytic felsite flow (?) with plagioclase phenocrysts, few possible amygdules (sodic rhyolite); ¼ mi NE of BM 1425 in SW¼ sec. 14, 63N/11W; no K-spar detected

M-7202: Dark green, intrusive andesitic porphyry with abundant plagioclase, fewer hornblende phenocrysts (sodic dacite); SW shore Tofte L., NE¼ sec. 10, 63N/10W; oxidation ratio of Fe in analysis is questionable

M-7112: Med. gray, slightly sheared, dense porphyry with abundant plagioclase, fewer quartz, and altered mafic phenocrysts (sodic dacite); Moose L. rd., SW¼/SE¼ sec. 6, 63N/9W

1: Hornblende-feldspar porphyry (sodic andesite), NENWSE sec. 9, 61N/14W; collected by J. W. Dalrymple (M4025); previously unpub.

2: Greenish felsite from interior of rounded mass in agglomerate (K-poor andesite); railway cut in Ely, Minn.; Ref.: N. H. Winchell, 1895, p. 204 (no. 222)

Table III-4. Chemical analyses, in weight percent, and norms of metavolcanic rocks in Newton Lake Formation, Lake Vermilion Formation, and Knife Lake Group (compiled from Green, 1970, except for no. 5; rock names based on chemical classification of Irvine and Baragar, 1971).

	Newton Lake Formation						Lake Vermilion Formation	Knife Lake Group
	M-7527	M-7152	M-7194	M-7441	M-7548	M-7560	5	M-7499
SiO_2	51.45	57.83	58.79	63.61	60.55	54.72	51.95	70.52
TiO_2	0.98	0.79	0.67	0.61	0.53	0.76	1.03	0.58
Al_2O_3	11.60	13.39	15.81	13.91	15.99	14.77	12.58	15.50
Fe_2O_3	3.24	2.44	3.70	1.64	2.44	2.37	0.90	0.60
FeO	8.24	5.07	4.09	3.90	3.13	4.92	8.77	1.26
MnO	0.19	0.20	0.10	0.09	0.08	0.11	0.15	0.03
MgO	7.73	2.69	4.44	4.00	3.50	3.77	8.90	0.52
CaO	11.09	7.16	2.86	4.27	6.95	7.94	7.00	1.90
Na_2O	2.62	3.84	5.59	5.23	4.14	3.29	2.79	5.59
K_2O	0.08	0.23	0.20	0.43	0.86	0.68	1.38	1.38
H_2O+	2.44	3.44	3.33	2.27	2.17	3.16	2.67	1.86
H_2O-	n.d.	0.14	0.16	0.11	0.06	0.23	0.14	0.13
CO_2	0.00	2.87	0.53	0.53	0.10	3.06	1.02	0.42
P_2O_5	0.22	0.24	0.14	0.14	0.11	0.17	n.d.	0.14
Cr_2O_3	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	0.17	n.d.
S	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	0.11	n.d.
Total	99.88	100.33	100.41	100.77	100.61	99.95	99.22	100.47
Q	3.7	21.4	11.9	16.4	14.8	11.9		28.4
or	0.6	1.7	1.1	2.8	5.0	4.4	8.5	8.3
ab	22.5	33.5	49.2	45.0	35.6	29.9	24.7	48.2
an	20.3	16.4	14.7	13.6	23.1	25.0	18.6	8.9
c		1.0	1.5					1.6
di	28.2			5.8	8.9	12.9	14.5	
hy	17.3	13.5	15.0	12.3	7.4	10.2	28.9	3.2
ol							1.2	
mt	4.9	3.7	5.6	2.6	3.7	3.7	1.4	1.6
hm								
il	2.0	1.5	1.4	1.2	1.1	1.5	2.0	1.2
ap	0.5	0.6	0.3	0.3	0.3	0.4		0.4
cc		6.7						
cm							0.3	
Analyst	K. R.	H. A.	H. A.	H. A.	H. A.	H. A.	S. D.	H. A.

Analysts: H. A.: Hiroshi Asari, Japan Anal. Chem. Res. Inst., 1967; K. R.: K. Ramlal, Univ. of Manitoba, 1968; S. D.: S. Darling

M-7527: Gray-green, fine-gr., recrystallized, variolitic, pillowed greenstone with relict augite microphenocrysts (K-poor quartz tholeiite); NW of Fall L., NW¼ sec. 9, 63N/11W

M-7152: Dark green, subtrachytoid greenstone (flow) with amygdules of calcite and chlorite (K-poor dacite); NW shore Wood L. in SW¼ sec. 27, 64N/10W; norm computed with calcite; if CO_2 is eliminated, there is no c in norm

M-7194: Dark green, massive greenstone (flow), rare amygdules (sodic andesite); E end Wood L. in NW¼ sec. 25, 64N/10W

M-7441: Med. gray-green, bulbous-pillowed metavolcanic flow with albite phenocrysts, quartz, feldspar, amphibole, epidote, and chlorite groundmass (sodic andesite); S shore Witness L. at Tp. line, secs. 25/30, 64N/9-10W

M-7548: Light gray-green, lineated tuff-breccia, unwelded (K-poor andesite); W side Newton L. in SE¼ sec. 27, 64N/11W

M-7560: Light greenish-buff, lineated tuff-breccia (quartz basalt); NE of Mud L. in SW¼ sec. 30, 64N/10W; contains about 7% dissem. calcite

5: Greenstone (olivine tholeiite) from Pine Island in L. Vermilion; Ref.: Grout, 1926, p. 12, no. 2

M-7499: Light gray, dense, schistose felsite with plagioclase phenocrysts (sodic rhyolite); Knoll WNW of Sourdough L. in NE¼ sec. 6, 63N/10W; no K-spar detected

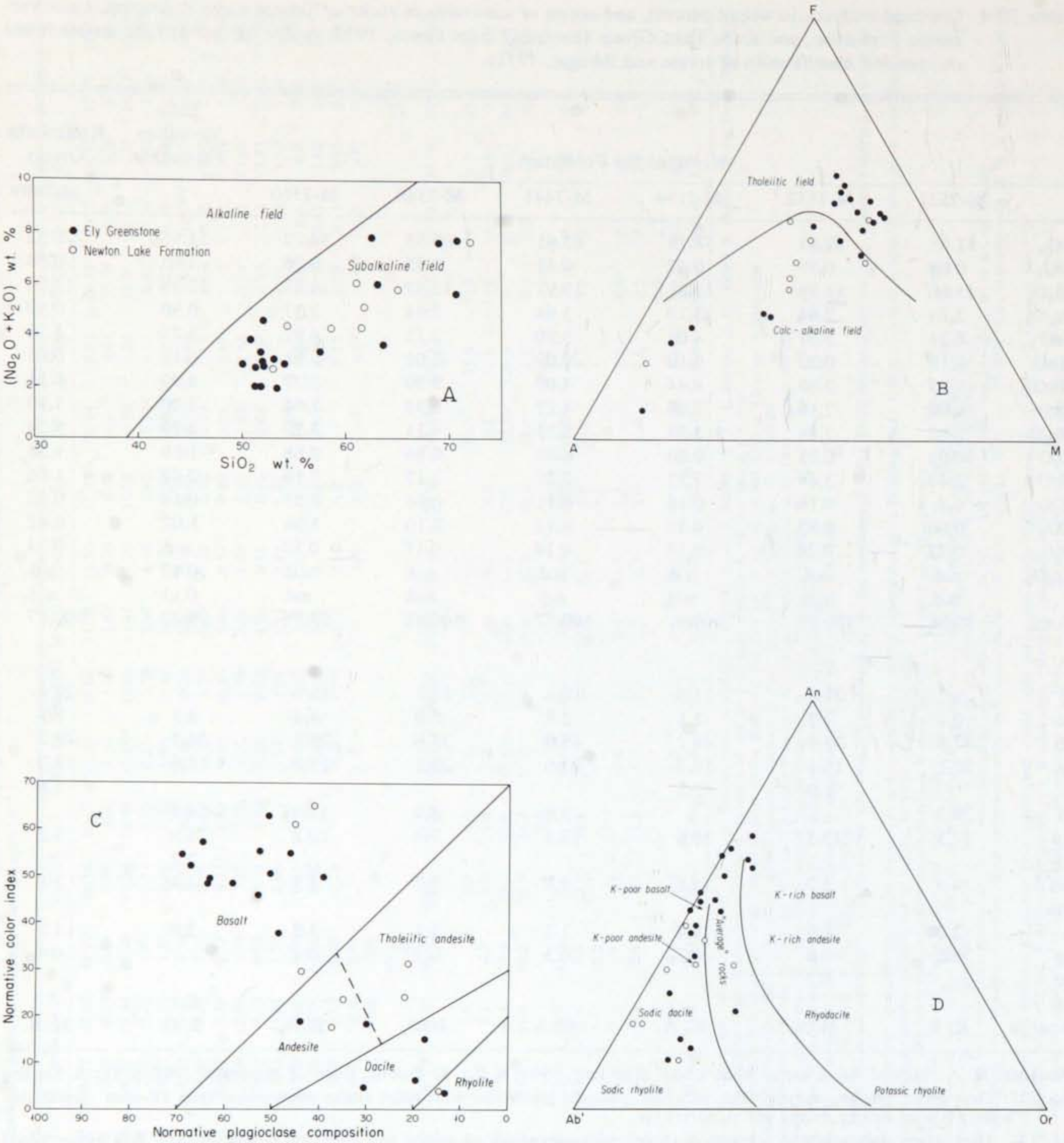

Figure III-20. Diagrams showing chemical characteristics of metavolcanic rocks in Vermilion district. A, alkali-silica plot, in weight percent; the solid curve is after Irvine and Baragar (1971, fig. 3); B, AFM plot; $A = Na_2O + K_2O$; $F = FeO + 0.8998\ Fe_2O_3$; $M = MgO$, all in weight percent. The line serves to separate tholeiitic and calc-alkaline compositions, after Irvine and Baragar (1971, fig. 2); C, plot of normative color index against normative plagioclase composition showing dividing lines for distinguishing different rock types; plot in percent cation equivalents, after Irvine and Baragar (1971, fig. 7); D, An-Ab'-Or projection, showing Irvine and Baragar's (1971, fig. 8) boundaries for distinguishing K-poor, "average," and K-rich variants. Plot in percent cation equivalents.

senting either a moderately well advanced low-potash tholeiite or a basic calc-alkaline rock series (fig. III-21). The more acidic rocks follow the calc-alkaline trend (see figs. III-20B and C).

The analyses from the Newton Lake Formation clearly indicate a calc-alkaline trend, suggesting that the Newton Lake may not be related to the same magma source as the Ely Greenstone.

The metavolcanic rocks in the Ely Greenstone and the Newton Lake Formation are chemically similar to others from the Superior Province, as can be seen by comparing the diagrams in Figure III-20 with those compiled by Irvine and Baragar (1971). From the chemical data now available for the Canadian Shield (see, for example, Wilson and others, 1965; Baragar and Goodwin, 1969), it can be concluded that the compositions of the Lower Precambrian volcanic sequences are remarkably uniform from belt to belt, and that ". . . whatever the processes responsible for the volcanism and for the fractionation of its products they acted with uncanny uniformity throughout the Archean of the Canadian Shield" (Baragar and Goodwin, 1969, p. 140).

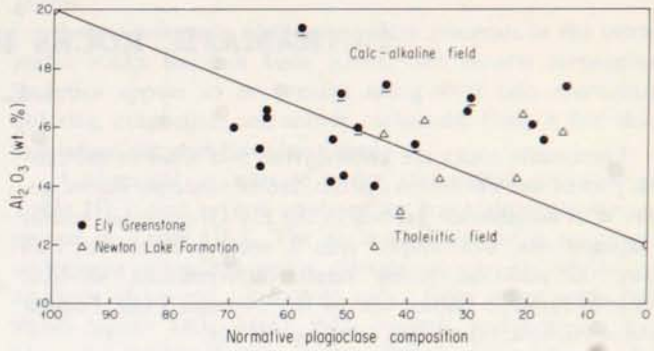

Figure III-21. Plot of Al_2O_3, in weight percent, against normative plagioclase composition.

ULTRAMAFIC ROCKS IN VERMILION DISTRICT

J. C. Green

Ultramafic rocks are known from two areas in the central part of the Vermilion district. Hornblende peridotite occurs with hornblende gabbro in the Ely Greenstone, within a square-mile, oval-shaped area 2 miles southeast of Fall Lake. In addition, many small, differentiated, sill-like, mafic-ultramafic bodies are in the Newton Lake Formation northwest, north, and northeast of Ely.

HORNBLENDE PERIDOTITE IN ELY GREENSTONE

The hornblende peridotite in the Ely Greenstone occurs in an area of poor exposures, and its structural relations to the low-grade greenstone that surrounds it have not been established. The peridotite is cut by small red syenitic and granitic intrusions that crop out in the same general area. Adjacent to the granitic dikes, the ultramafic rocks are retrograded, probably as a result of hydrothermal alteration; augite is converted to actinolite and plagioclase is saussuritized.

The peridotite is generally a black, medium-grained, fresh rock that is massive or weakly foliated. It is composed primarily of hornblende and augite, and contains minor amounts of plagioclase, biotite, magnetite, and apatite. It appears to grade, with increasing proportions of plagioclase, into hornblende-augite gabbro, but exposures are too sparse to establish this relationship clearly; certainly, both rock types are present in the area labeled "ea" on the geologic map of the Gabbro Lake quadrangle (Green and others, 1966). Estimated modes of ultramafic and mafic rocks of this complex are given in Table III-5.

From textural evidence, these rocks appear to have formed by accumulation of crystals from a hydrous mafic magma. Augite, magnetite, hornblende, biotite, and possibly apatite occur in subhedral to euhedral crystals that suggest cumulus origin; interstices contain overgrowths on hornblende, biotite, and magnetite, and in some specimens small amounts of interstitial plagioclase and biotite. In the more gabbroic types, laths of plagioclase have a normal subhedral habit. The plagioclase is rather strongly altered, and some of the biotite is altered to chlorite and epidote. The biotite and hornblende are strongly pleochroic. The cumulus augite crystals generally range in diameter from 0.3-1.5 mm. The hornblende crystals are somewhat larger; cores are mostly 1-2 mm across, but some large poikilitic crystals are as much as 6 mm across. The magnetite and apatite are smaller than the augite crystals, and the biotite books are about the same size as the augite.

Table III-5. Estimated modes of hornblende peridotite and gabbro, Ely Greenstone.

	M-7517c	M-7536	M-7656
Plagioclase	46		2
Augite	12	45	28
Hornblende	30	45	59.5
Biotite	3	5.5	0.5
Magnetite	4	3	8
Apatite	2	1.5	2
Chlorite	2	Tr	
Epidote	1	Tr	
Calcite		Tr	

DIFFERENTIATED MAFIC-ULTRAMAFIC BODIES IN NEWTON LAKE FORMATION

Reconnaissance mapping in the Newton Lake Formation has disclosed several generally small, differentiated, sill-like bodies of mafic to ultramafic composition. The mapping has not yet fully outlined all the bodies, but one immediately southeast of the Vermilion fault, near Little Long Lake, north of Ely, appears to be about 3 miles long and 600 feet thick. The others are considerably smaller. Most bodies intrude the metavolcanic rocks, but some of the smaller lenses may be lava flows, for they contain about 20 percent of an altered groundmass that is tentatively interpreted as having been glass. Contacts are rarely seen, no doubt because the sheared serpentine is readily eroded, but the map pattern shows them to be concordant. According to Klaus J. Schulz (1972, written comm.), these bodies resemble the gravity-stratified sills and ultramafic lenses that Naldrett (in press) has recognized in many other Archean greenstone belts.

All the rocks are altered to some extent. The plagioclase in the gabbroic rocks is strongly saussuritized in most samples, the olivine is serpentinized to varying degrees, and the orthopyroxene is largely altered to mixtures of talc, chlorite, and serpentine. The clinopyroxene in many samples is fresh. Other minerals still recognizable are magnetite, chromite, phlogopite, apatite and, in some rocks, brown hornblende.

Despite the alteration, original igneous textures generally are well preserved. The gabbros are medium grained and have hypidiomorphic-granular textures; some of the later differentiates contain long, skeletal mafic crystals. The ultramafic rocks also are medium grained and have excellent cumulate textures; in these rocks, olivine, chromite, and/or pyroxene euhedra occur in a poikilitic mesostasis that generally is pyroxene but in the inferred lava flows appears to have been glass. The gabbroic and pyroxenitic rocks are grayish green, but the serpentinized peridotites are black and have a thin, pale-weathering rind in which the outlines of the rusty olivine grains are evident. Cleavage reflections from the enclosing poikilitic pyroxenes can generally be seen on fresh fracture surfaces.

The largest body, and most of the smaller ones as well, appear to be gravity-differentiated sills containing peridotite, pyroxenite, and gabbro layers. Modes are listed in Table III-6. Judged from pillow shapes in adjacent greenstones, the peridotite (harzburgite or lherzolite) occurs at the base of the sills, followed by pyroxenite, then hypersthene gabbro, and finally quartz gabbro or granogabbro containing granophyre. More than one magmatic cycle may be present. This order is consistent with textures indicating gravity settling of mafic crystals in place, perhaps alternating with renewed injection of magma.

Olivine and chromite are found only in the harzburgites and lherzolites. The olivine occurs as subhedral to euhedral grains 0.2-4 mm across, and constitutes from 30 to 70 percent of the rock. One sample was determined by microprobe analysis to be Fo_{74}. The chromite is euhedral, and constitutes as much as five percent of the samples studied. Reconnaissance microprobe analysis of chromite grains indicates that the Cr_2O_3 content varies but lies in the range 37 to 54 percent; the Cr/Fe ratio decreases outward from the core of each crystal.

The orthopyroxene is nearly all altered, but a few relict crystals and parts of crystals remain in some specimens. Orthopyroxene occurs as poikilitic oikocrysts as much as 7 mm across in some of the harzburgites but as euhedral to subhedral, blocky crystals in the pyroxenites and hypersthene gabbros. Relict exsolution lamellae appear to be present. Augite also occurs as large oikocrysts in the peridotites and as euhedral cumulus grains, 1.0-2.5 mm across, in the pyroxenites. Some of the peridotites contain augite as well as olivine as cumulus crystals. In the gabbros, the pyroxenes are also mostly subhedral and not poikilitic, and some have reaction rims of green hornblende.

Pale-brown hornblende occurs in some of the peridotites, including those which had a chilled glass (?) groundmass. It occurs as both euhedral and interstitial-poikilitic grains.

A detailed study of the secondary minerals in the ultramafic rocks has not been made, but several serpentine varieties appear to be present along with talc, tremolite, chlorite, magnetite, and minor carbonate. Only a few thin veinlets of asbestos have been seen.

A chemical analysis of major elements is reported in Table III-7, and several analyses for trace elements are reported in Table III-8. The rock analysis—of a specimen containing abundant relict clinopyroxene—is similar to analyses of several ultramafic rocks from other areas but shows higher TiO_2, MnO, P_2O_5 content, and a higher Fe/Mg ratio (reflected also in the olivine composition), and lower alkalis than typical mantle-derived ultramafic rocks (analyses 5-8, table III-7). Except for Ni, which has values as much as 50 to 100 times lower than expected, trace-element abundances are comparable to those in other similar ultramafic bodies. Sulfides were not observed in or near the peridotite bodies.

SIMILARITIES TO ULTRAMAFIC ROCKS IN NORTHWESTERN ONTARIO

Watkinson and Irvine (1964) have described ultramafic rocks from greenstone belts in northwestern Ontario that are similar to those in the Vermilion district. Near Quetico, they found several bodies of hornblende peridotite that are similar both in mineralogy and occurrence to the rocks in the Ely Greenstone; they differ mainly from the Ely rocks in containing olivine. Watkinson and Irvine concluded that the rocks crystallized from a hydrous tholeiitic olivine basaltic magma at a depth of at least 5 miles, sufficient to retain the necessary partial pressure of water to form the primary hydrous silicates. They also reported serpentinized peridotites from the vicinity of Shebandowan

Table III-6. Estimated primary modes of serpentinized peridotite in Newton Lake Formation.

	M-7549**	E-1	E-44	E-46	E-48	E-56	E-62	E-85	E-89	E-95	E-98	E-100
Olivine	47	X*	40-50	50†	X	X	50-60	30	70	35	60	70
Augite	18	X	X	X	20	X	X	7	10		5	6
Orthopyroxene	32	X	X	X	X	X	X	40	10	15	30	15
Hornblende			X	X			X	4		15		
Opaque	2	X	X	X*	X*	X	X	4	3	5	4	9*
Biotite	1	Tr	Tr	X	Tr					Tr		
Apatite			Tr		Tr		Tr		Tr	Tr	Tr	Tr
Unidentified								15	7	30		
Size of olivines, mm	0.5-2	0.3-1	1	0.2-1.5	0.4-2	1-2	1-2	1-3	2-4	0.2-1	2-4	1-5

* Includes identified chromite
† Composition Fo_{71}
x = Present
** Analyzed: see table III-7

Table III-7. Chemical analyses of selected peridotites.

	1	2	3	4	5	6	7	8
SiO_2	38.50	37.26	42.70	40.87	43.56	44.77	44.57	44.93
TiO_2	0.26	0.22	0.29	0.24	0.04	0.19	0.12	0.08
Al_2O_3	3.69	3.86	4.09	4.23	2.36	4.16	4.10	3.21
Cr_2O_3	n.d.*	0.52	n.d.	0.57	0.40	0.40	0.46	0.45
Fe_2O_3	6.65	6.72	7.37	7.37	1.00		1.17	0.09
FeO	7.76	8.10	8.61	8.89	7.77	8.21	6.85	7.58
MnO	0.23	0.19	0.26	0.21	0.10	0.11	0.13	0.14
NiO	n.d.†	n.d.	n.d.	n.d.	0.34	0.24	0.25	0.26
MgO	30.10	31.72	33.38	34.80	41.53	39.22	39.07	40.03
CaO	2.69	2.29	2.98	2.51	2.51	2.42	2.87	2.99
Na_2O	0.13	Tr	0.14	Tr	0.32	0.22	0.32	0.18
K_2O	0.02	Tr	0.02	Tr	Tr	0.05	0.07	0.02
H_2O+	9.39	7.31						
CO_2	0.00				n.d.			
P_2O_5	0.08	0.05	0.09	0.05	0.07	0.01	0.02	0.04
Total	99.50	98.41	99.93	99.74	100.00	100.00	100.00	100.00

1: M-7549, serpentinized peridotite, Newton Lake Fm., SW¼ sec. 33 64N/11W; Anal.: K. Ramlal, 1968
2: Ave. of two serpentinized peridotites, Shebandowan area, Ont. (Watkinson and Irvine, 1964, p. 70)
3: M-7549 as above, recalculated water-free
4: Anal. 2 as above (Shebandowan, Ont.), recalculated water-free
5: Ave. "Type C" serpentinized peridotite, Mayaguez, Puerto Rico, water-free and reduced (Hess, 1964, p. 172)
6: Ave. Lizard peridotite, Cornwall, water-free and reduced (D. H. Green, 1964, p. 184)
7: St. Paul's Rocks peridotite, North Atlantic Ridge (Hess, 1964, p. 172)
8: Tinaquillo peridotite, Venezuela (D. H. Green, 1963, p. 1398)

* Spectrographic analysis by U.S. Geol. Survey gives 0.3% Cr or 0.5% Cr_2O_3
† Atomic absorption analysis by U.S. Geol. Survey gives 0.0034% Ni or 0.0043% NiO

Table III-8. Trace element content of serpentinized peridotite (analyses by U.S. Geol. Survey; amounts in parts per million).

	Cr	Pt	Pd	Rh	Cu	Cu	Ni	Co	Zn	Ag
E-13	>5000	<0.010	<0.004	<0.005	100	88	900	200	56	0.60
E-44	3000	<0.010	0.011	<0.005	30	68	68	150	60	0.40
E-46a	5000	<0.010	<0.004	<0.005	50	n.d.	840	150	40	0.60
E-48	>5000	<0.010	<0.004	<0.005	20	n.d.	n.d.	150	n.d.	0.20
E-56	5000	<0.010	0.009	<0.005	20	330	60	150	25	0.20
E-62	>5000	<0.010	0.008	<0.005	20	105	24	150	50	0.20
M-7549	3000	n.d.	n.d.	n.d.	20	180	34	150	n.d.	0.40
Method:	spec	spec	spec	spec	spec	a.a.	a.a.	spec	a.a.	a.a.

that have similarities to both alpine-type bodies and differentiates of stratiform mafic intrusions, and which are closely analogous to the lenses in the Newton Lake Formation. The bodies differ, however, in having a much higher and apparently more normal Ni content (1,000-3,500 ppm). The authors concluded that the rocks formed from tholeiitic magma by early crystal concentration of olivine within the crust; subsequently, the peridotites were intruded as a crystal mush lubricated by a small amount of gabbroic liquid that may eventually have been squeezed out during final consolidation. They also concluded that the subsequent serpentinization has not materially affected the bulk composition of the peridotite, except for the addition of H_2O, O_2, minor CO_2, and local redistribution of CaO. Until further studies are made, these conclusions can be tentatively applied to the rocks near Ely as well.

BANDED IRON-FORMATIONS IN VERMILION DISTRICT

P. K. Sims

Iron-formation of the type commonly called Algoma (Gross, 1970) or simply Archean volcanogenic iron-formation (Goodwin, 1970a) is widely distributed in the Vermilion district and in other volcanic-sedimentary sequences farther west in the state. It is most abundant in the volcanic successions, but is present also in the sedimentary successions. The best known and most thoroughly studied deposit in Minnesota is the Soudan Iron-formation, which was mined until recently for high-grade hematite ores that occurred locally at Soudan (see section on mineral deposits, this chapter).

The iron-formations consist of several intergradational types of fine-grained ferruginous chert. A common variety is jaspilite, a thin-bedded or laminated rock consisting of alternating layers of chert or jasper and magnetite or hematite (fig. III-22). Other common varieties are more or less massive chert or jasper containing disseminated magnetite or hematite, or both. Chert-siderite beds, iron-silicate mineral facies, and iron-sulfide mineral facies are lesser, more local lithologic varieties. All varieties may occur together. Magnetite is the major oxide except near the high-grade hematite ore bodies or other local concentrations of massive hematite. The iron-formations yield strong magnetic anomalies, which are discussed in a later chapter (Sims, this volume).

Figure III-22. Jaspilite in Soudan Iron-formation, Soudan hill. Dark bands contain substantial hematite; white bands are mainly chert. Two generations of folds are present; an older set (in upper part of photograph) is modified by a younger fold set parallel to hammer head.

The iron-formations are closely associated with basaltic and andesitic lavas, tuffs and other pyroclastic rocks and, locally, volcanogenic graywacke. Fine-grained rocks—dominantly tuffs or reworked tuffs—commonly are interbedded with the chert-oxide beds. At the type locality, the Soudan Iron-formation, which is the most continuous deposit known in the region, occurs at the transition from mafic flows and pyroclastics to younger dacitic tuff and agglomerate. The occurrence of iron-formations at major volcanic contacts such as this is common throughout the Canadian Shield (Goodwin, 1970a). Except for the Soudan, the iron-formations are thin, discontinuous lenses, generally a maximum of a few tens of feet thick and a few thousand feet long.

SOUDAN IRON-FORMATION

The Soudan Iron-formation was named by Van Hise and Clements (1901, p. 1402) from exposures on Soudan hill, near the town of Soudan. Subsequently, it became common practice to designate all iron-formations, except those in the Knife Lake Group, which were called Agawa formation (see Gruner, 1941, p. 1616), as Soudan Iron-formation. It was generally presumed that the distribution of banded iron-formations could be accounted for by complex folding of a single, major, continuous unit (Grout and others, 1951, p. 1027). More detailed regional mapping (Sims and others, 1968b; Green and others, 1966) has shown, however, that iron-formations occur at several stratigraphic positions in the greenstone belt and that they cannot be accounted for by any combination of multiple folding of a single iron-formation. Accordingly, the term "Soudan Iron-formation" is now restricted (Morey and others, 1970) to that body that is stratigraphically continuous with exposures at the type locality. The Soudan, as redefined, can be traced from the type locality eastward to the vicinity of Twin Lakes, a distance of 16 miles (fig. III-6). It appears to lens out near Twin Lakes, southwest of Ely, and is not continuous with the iron-formation mined in the Ely trough.

Near Soudan (fig. III-6), the Soudan Iron-formation overlies mafic volcanic rocks of the Ely Greenstone and is overlain by agglomerates, tuffs, and other volcaniclastic rocks assigned to the Lake Vermilion Formation (Morey and others, 1970). However, the eastern part of the formation is overlain by at least 7,000 feet of mafic metavolcanic rocks, accompanied by local lenses of iron-formation, assigned to the Ely Greenstone.

The outcrop width of the Soudan Iron-formation is as much as 3,000 feet, but the true thickness is much less. At

Soudan hill, the formation is repeated by folding of at least two generations, and at most is a few hundred feet thick. Probably, similar repetitions of the iron-formation occur in other areas.

At the type locality, the Soudan consists of several types of fine-grained ferruginous cherts that are interbedded with fine-grained tuffaceous rocks and, rarely, mafic flows. Mafic dikes, now largely altered, and dacite porphyry dikes transect the bedded rocks and locally are moderately abundant. Klinger (1956; 1960, unpub. Ph.D. dissert., Univ. Wisc.) considered all the rocks except the ferruginous cherts to belong to the Ely Greenstone. He interpreted the Soudan, therefore, as lenses or zones of iron-formation in the upper part of the Ely Greenstone. In the same way, Schwartz and Reid (1955) concluded that ". . . these jasper or jaspilite beds do not comprise a formation separate from the Ely Formation. . . ." The Minnesota Geological Survey now (Morey and others, 1970, p. 15) considers all the bedded rocks that are intercalated with the ferruginous cherts as part of the Soudan Iron-formation. This interpretation is consistent with observed lithologies in other areas of less complexly folded Soudan Iron-formation. The fine-grained rocks intercalated with the ferruginous cherts are laminated or thin bedded and, rarely, massive. They contain sericite, chlorite, and quartz as the dominant minerals; quartz grains as much as 2 mm in diameter and angular volcanic rock fragments as much as an inch in diameter are common (Klinger, 1960, *op. cit.*). Some beds, commonly called graphitic slate or schist, contain rather abundant carbonaceous material, some of which has microstructures that may represent primitive life (Cloud and others, 1965). Interbedded with these rocks are several types of chloritic schists that represent altered lava flows and, probably, also diabasic intrusive rocks.

Klinger (1956; 1960, *op. cit.*) distinguished three types of ferruginous chert in the Soudan mine area. In decreasing order of abundance these are: (1) greenish-white chert, composed principally of quartz but containing minor amounts of pyrite and chlorite; (2) lean jasper, a rock composed of quartz, hematite, and martite or magnetite, which has an iron content of less than 20 percent; and (3) jaspilite, a banded rock composed of quartz, hematite, and martite or magnetite, which contains somewhat more than 30 percent iron. Outside of the mine area, magnetite is the dominant iron oxide, and the color of the siliceous component generally is gray or dark gray, depending upon the amount and proportion of magnetite in the chert. Local beds contain siderite, iron-silicate minerals, and pyrite. The pyrite appears to be paragenetically later than the associated magnetite.

Partial analyses of representative samples of different types of iron-formation in the Soudan mine area are listed in Table III-9.

Judged from available analyses (table III-10), the tuffaceous rocks intercalated with the ferruginous cherts dominantly have compositions in the rhyolite-dacite range. Interpretation of the analyses is equivocal, however, because of possible additions or subtractions of various constituents during rock alteration related to development of the high-grade hematite ores at the Soudan mine. Schwartz and Reid (1955, p. 299) suggested that the sericite schist represented by analysis 1 in Table III-10 might be either a metamorphosed felsic rock or a hydrothermally altered basic greenstone. Subsequent studies in the mine area by Klinger (1960, *op. cit.*, p. 78) showed that the sericitic rocks are not

Table III-9. Partial chemical analyses, in weight percent, of different types of iron-formation in Soudan mine area. (Analyses by Oliver Iron Mining Division, U. S. Steel Corp.; source: Klinger, 1956.)

Type	No. of Samples	Total Fe	P	SiO_2	Al_2O_3	Mn
Greenish-white chert	4	1.8	0.02	93.2	0.3	0.04
Lean jasper	25	15.3	.04	76.1	0.3	.06
Siderite iron-formation	5	16.4	.025	64.9	.15	.45
Jaspilite	11	34.1	.07	50.8	n.d.	.10

Table III-10. Chemical analyses, in weight percent, of bedded rocks intercalated with ferruginous chert in the Soudan Iron-formation, Soudan mine.

	1	2	3
SiO_2	59.71	73.58	64.38
Al_2O_3	16.05	10.61	15.52
Fe_2O_3	0.73	3.80	2.79
FeO	1.81		
MgO	2.78	1.95	1.76
CaO	4.55	0.23	0.36
Na_2O	0.51		
K_2O	4.18		
H_2O+	2.07		
H_2O-	0.29		
TiO_2	0.28	0.32	0.36
P_2O_5	0.16		
CO_2	6.40		
MnO	0.15		
Loss on ignition		2.65	2.66
Total	99.67		

1: Sericite schist from hole 708 (175-210 ft.), 12th level; Anal.: Eileen K. Oslund; source: Schwartz and Reid, 1955, p. 299
2: Chlorite-quartz-sericite schist, 19th level; analysis by Oliver Iron Mining Division, U. S. Steel Corp.; iron reported as total Fe
3: Chlorite-sericite schist, occurring within unit represented by sample 2. Analysis by Oliver Iron Mining Division, U. S. Steel Corp.; iron reported as total Fe

altered mafic lavas. Further, he showed that there is no direct association of sericitic schists and hematite ore bodies. More recent mapping (Sims and others, 1968b) has indicated that the sericitic bedded rocks occur throughout the exposed length of the Soudan Iron-formation, and probably contain regional metamorphic mineral assemblages. It is possible, though, that the alteration related to the hematite ores did modify these rocks to some extent adjacent to hematite ore bodies.

IRON-FORMATION IN ELY TROUGH

The iron-formation in the Ely trough (Reid, 1956), which is the host rock for the hematite ore deposits formerly mined in the Ely area, is a folded (synclinal) lens about 1¾ miles long and a maximum of one-fourth mile wide (see fig. III-77). It has been called Soudan Iron-formation, but is neither continuous with it nor does it occur at the same stratigraphic position. The iron-formation is enclosed within metabasalt of the Ely Greenstone, much of which is pillowed.

There are no accessible exposures of the iron-formation because of subsidence of the surface over the mined-out areas in the syncline. Judged from earlier observations by company geologists (Reid, 1956, p. 138), however, a substantial part of the iron-formation is similar to the banded jaspilite in the Soudan mine area. Machamer (1968, p. 14) reported that some of the iron-formation also is a jasper consisting of fine-grained quartz and hematite. An argillaceous facies of iron-formation is inferred from the local presence of paint rock in one of the mines (Pioneer). Also, lenses of clastic strata are inferred from the presence of small bodies of carbonaceous and sericitic schists in the Zenith mine (Reid, 1956, p. 139; Machamer, 1968, p. 14).

OTHER IRON-FORMATIONS

Other iron-formations having little economic significance occur in the Ely Greenstone and to a lesser extent in the Knife Lake Group, the Newton Lake Formation, and the Lake Vermilion Formation. Most of these iron-formations are thin lenses or irregular beds of magnetite-bearing chert, generally associated with pillowed greenstone. More extensive and thicker beds include the lenses of iron-formation in the Ely Greenstone that extend southeastward from Tower (Sims and others, 1968b), the Section 30 deposit north of White Iron Lake, which has local hematite bodies, and the deposits between White Iron Lake and Garden Lake. Sparse grunerite and carbonate are known to occur in the latter group (Green, 1970a).

Gruner (1941, p. 1616-1617) has demonstrated that some of the beds of so-called Agawa formation, as defined by Clements (1903, p. 324-325), are not primary deposits but instead are replacement bodies along faults. He stated: "A striking example . . . on the east shore of Ogishkemuncie Lake . . . is a lens of supposed Agawa formation more than 1500 feet long and several hundred feet wide in a shear zone of a big fault. The rock is made up of iron carbonate, chlorite, sericite, and quartz and is weathered conspicuously to a chocolate brown. . . . This zone actually cuts across the beds . . . in many places and replaces them by iron carbonate."

ORIGIN

The close association of iron-formation with volcanic rocks suggests a volcanic source for its chemical components. The known iron-formations occur mainly in mafic-intermediate volcanic rocks, and are nearly completely lacking in the graywacke-type metasedimentary rocks that are some distance from the volcanic centers. Within the Vermilion district, most of the deposits are in the upper part of the Ely Greenstone, at or near the contact with overlying felsic volcanic rocks. Probably, most deposits accumulated in shallow basins during periods of relative quiescence in the volcanism. The fine-grained rocks commonly interbedded with the iron-formations may have been deposited directly into the basins as ash falls or transported into them by water that reworked the pyroclastic deposits. The environment of deposition was favorable for the development of the oxide facies of iron-formation. Although carbonates and sulfides occur sparsely with the oxide-facies rocks, and seem to be relatively more abundant in the central part of the district, they are not dominant in any known iron-formations in the district.

By analogy with the Michipicoten iron-formations, which have been studied intensively by Goodwin (1962), the Vermilion deposits can be inferred to be direct volcanic products, possibly formed by volcanic exhalative processes.