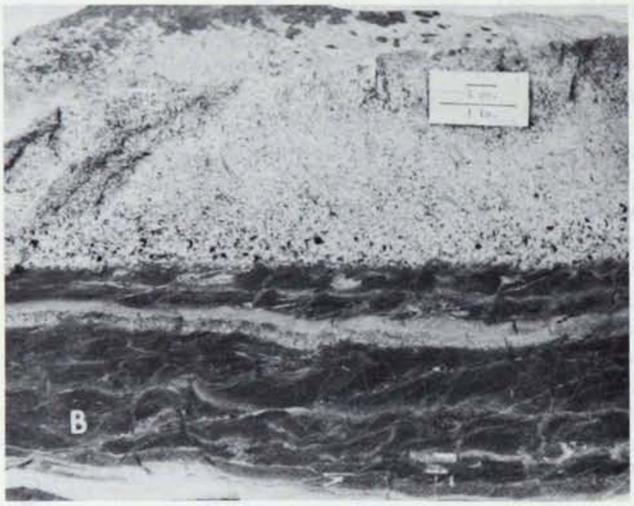
GRAYWACKES AND RELATED ROCKS OF KNIFE LAKE GROUP AND LAKE VERMILION FORMATION, VERMILION DISTRICT

Richard W. Ojakangas

Graywacke is the dominant rock type in both the Knife Lake Group, in the eastern part of the Vermilion district, and the Lake Vermilion Formation, in the western part (fig. III-6). It is also a minor component of the Ely Greenstone and the Newton Lake Formation in the central part of the district (Green, 1970a). Numerous other rock types also occur in the Knife Lake Group and the Lake Vermilion Formation, including slate, conglomerate, dacitic tuff and agglomerate, mafic to felsic flows, dikes, and sills. All are metamorphosed to varying degrees, but the prefix "meta" generally is omitted in this paper.



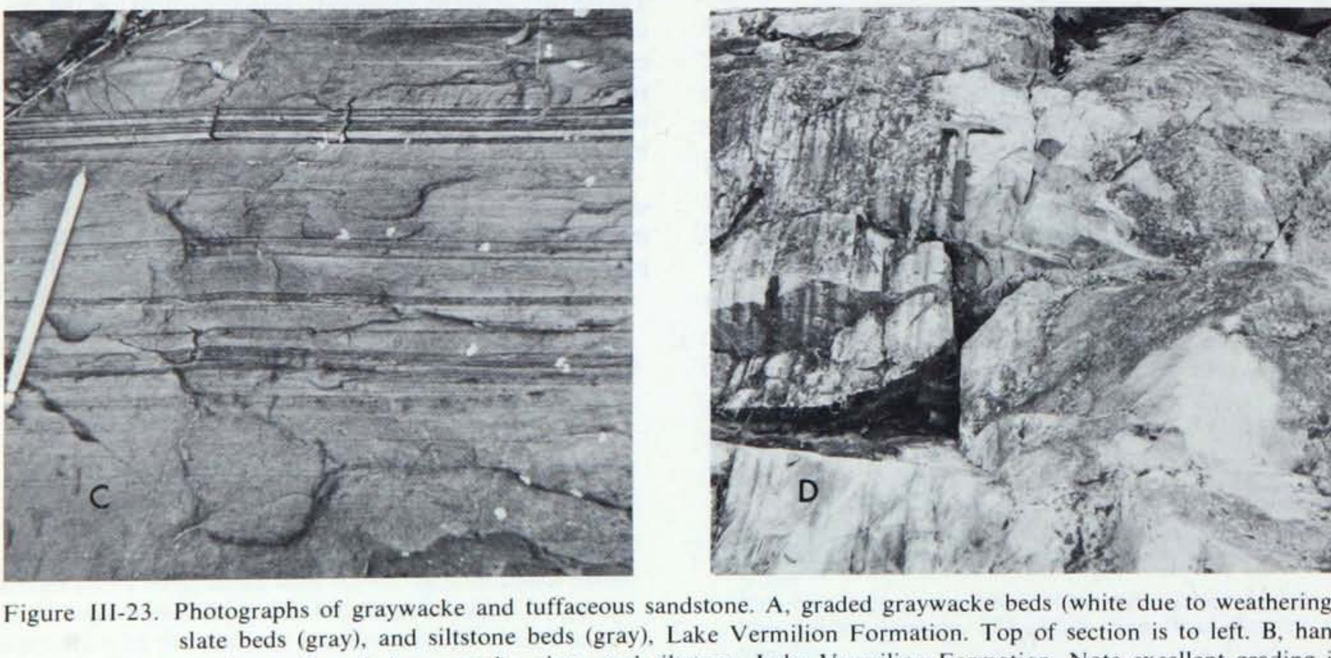


Figure III-23. Photographs of graywacke and tuffaceous sandstone. A, graded graywacke beds (white due to weathering), slate beds (gray), and siltstone beds (gray), Lake Vermilion Formation. Top of section is to left. B, hand specimen of graded graywacke, slate, and siltstone, Lake Vermilion Formation. Note excellent grading in graywacke. C, thin tuffaceous beds in graywacke unit of Knife Lake Group, Ensign Lake. D, graywacke-slate beds in left of photo; massive tuffaceous sandstone on right; Knife Lake Group, Knife Lake.

The graywackes resemble those of typical turbidite sequences described in the literature (for example, see McBride, 1962; Ojakangas, 1968), and have several characteristic sedimentary structures, especially graded bedding (figs. III-23A and 23B). A detailed study of two good exposures showed that nearly two-thirds of the graywacke beds and some interbedded siltstone beds are visibly graded (Ojakangas, 1972). Original sedimentary features such as load casts and flame structures are present on the bottoms of some beds (Ojakangas, 1972); however, soles are not exposed in three dimensions. Green (1970a) noted a few crossbeds in graywackes of the Knife Lake Group, and some also were noted in the Knife Lake during my investigation, but a paleocurrent analysis does not seem warranted. Most other, more subtle sedimentary features are obscured by a combination of deformation, low-grade metamorphism, weathering, and lichen on the outcrop surfaces.

Most of the graywacke is fine to medium grained (0.1 to 0.5 mm average grain size), but some is coarse grained (0.5 to 2.0 mm average grain size). Siltstone and slate interbeds are thin and vary significantly in amount from locality to locality; generally, they are composed of very thin laminae and are intimately intercalated with one another and with thin sandy laminae (fig. III-23B). Most beds are continuous across exposures, but exposures are too small to resolve the lateral extent of individual strata (fig. III-23A). The beds generally are less than one meter thick, but some are as much as four meters thick.

The term "graywacke" is used here and elsewhere in discussions in this chapter as a somewhat imprecise field term for well bedded, medium-gray to dark-gray sandstones. Petrographic studies show that not all these sandstones are true graywackes; many are actually subgraywacke and some are arkose. As defined by Pettijohn (1957), graywacke contains more than 15 percent detrital clayey matrix, less than 75 percent quartz, and abundant feldspar and/or rock fragments.

In the field, the graywackes of the two map units are very similar. In fact, all were mapped previously as Knife Lake slates (A. Winchell *in* Winchell, N. H., 1888; Van Hise and Clements, 1901; Clements, 1903), Knife Lake Series (Grout, 1933a; Gruner, 1941), or Knife Lake Group (Grout and others, 1951). However, the Knife Lake strata of the eastern part of the district cannot be traced westward into the Lake Vermilion Formation; they terminate against the Wolf Lake fault about 5 miles west of Ely (see fig. III-6). Because the correlation of the rocks in the eastern and western parts of the district was based on tenuous lithologic similarity, the Lake Vermilion Formation was established formally (Morey and others, 1970) for the body in the western part of the district.

Also, there are petrographic similarities between the graywackes of the Knife Lake Group and the Lake Vermilion Formation. In both units, the graywackes are comprised largely of volcanic detritus which is dominantly dacitic in composition (Ojakangas, 1972). They are immature rocks, both texturally and compositionally; their quartz contents are low and the grains are subangular to angular and poorly sorted.

KNIFE LAKE GROUP

The rocks of the Knife Lake Group, in the eastern part of the Vermilion district, were first studied by A. Winchell (*in* Winchell, N. H., 1888) and Clements (1903). Later, Gruner (1941) made an excellent map of the type area at Knife Lake and of adjacent areas. Several other workers have described certain geologic aspects of the district. Recently, Green (1970a; Green and others, 1966) mapped the Knife Lake rocks within the Gabbro Lake quadrangle in the central part of the district in detail. Only Green (1970a) and I (Ojakangas, 1972) have done detailed petrography on the graywackes.

The Knife Lake area is the most suitable part of the Vermilion district to study original textures and structures. Although the rocks in this region are deformed and the beds are generally steeply dipping, most of the rocks have undergone little recrystallization. Locally intense shearing parallel to bedding, as at Ensign Lake, however, has obliterated original textures and has produced well foliated phyllitic and phyllonitic rocks.

Gruner (1941) divided the Knife Lake Group into 21 lithologic members, eight of which consist dominantly of graywacke, slate, and tuff. Major longitudinal faults extend through the area and divide the folded Knife Lake rocks into seven structural segments; correlation of members from segment to segment is not possible with present data. Gruner estimated the total thickness of the Knife Lake Group to be between 11,500 and 21,000 feet (Grout and others, 1951, p. 1033).

The Knife Lake Group consists of a great variety of rock types. Igneous rocks, which are a minor but integral part of the group, include basalt flows and intrusions and porphyritic andesites, which are either flows or intrusive rocks. Most of the rocks in the group are epiclastic or volcaniclastic; all show evidence of a volcanogenic origin. These rocks, in approximate order of decreasing abundance, are graywacke, slate, agglomerate, conglomerate, and tuffaceous sandstone. Gruner (1941) gave field descriptions of all these lithologies. In this report, only the graywacke-slate units and the interbedded tuffaceous sandstones (called arkosites by Gruner) are discussed in more detail.

The graywacke-slate units are composed of thin intercalated beds of graywacke, slate, and siltstone. The individual beds are commonly less than 10 cm thick and some on Ensign Lake are still thinner (fig. III-23C); however, beds as much as 30 cm thick are not uncommon, and some are much thicker. Massive, coarse tuffaceous sandstone is interbedded with graywacke and slate at several localities, and individual beds are as much as 5 m thick (fig. III-23D). Bedding is difficult to discern in some of the more slaty units; some of these, as on the south shore of Knife Lake, are siliceous, weather white, and break with a conchoidal fracture.

Graywackes in the Knife Lake Group are various shades of green and most have relatively unaltered original textures and compositions (fig. III-24A). In those samples that have been point-counted (tables III-11 and III-12), volcanic rock fragments generally constitute from 25 to 50 percent of the rock, plagioclase from 10 to 36 percent, volcanic quartz from a trace to 11.5 percent, and detrital hornblende from zero to 22 percent. Photomicrographs of two samples are shown in the report by Green (1970a, p. 33). Comparisons of framework grains of plagioclase and hornblende with the phenocrysts of associated volcanic rock fragments and with Knife Lake volcanic rocks indicate that these

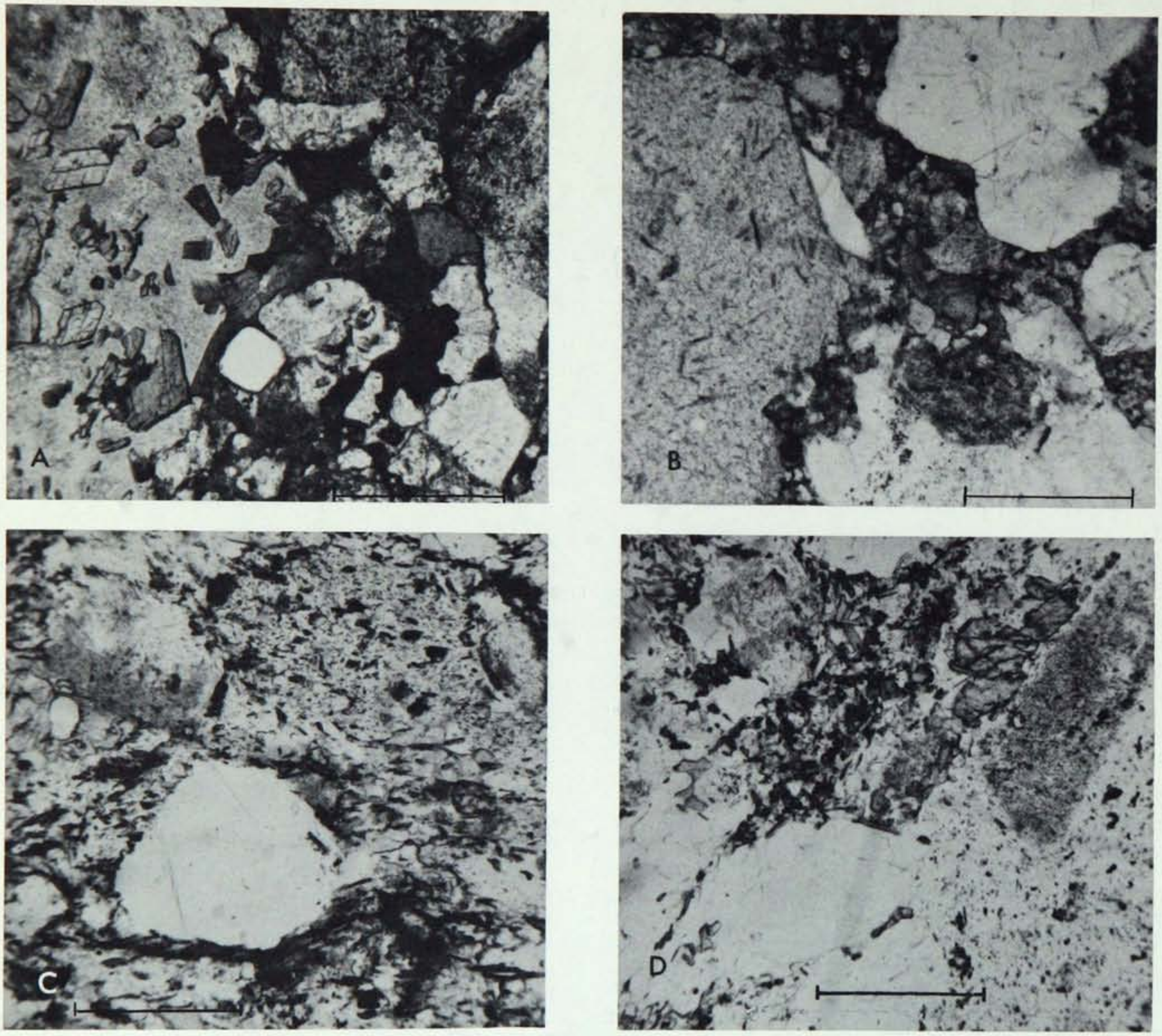


Figure III-24. Photomicrographs of graywacke and tuffaceous sandstone. A, graywacke (KL-204, table III-11) from Hanson Lake, Knife Lake Group. Volcanic quartz grain and plagioclase grains near center; volcanic rock fragment at left contains hornblende phenocrysts; volcanic rock fragment at right contains quartz and plagioclase phenocrysts (one nicol). B, graywacke (KL-82, table III-11) from Ogishkemuncie Lake, Knife Lake Group. Volcanic rock fragment at left; coarse-grained, plutonic (?) quartz-feldspar rock fragment at lower right; composite quartz grain (plutonic) at upper right (one nicol). C, graywacke (T-1087, table III-11) from west of Tower, Lake Vermilion Formation. Probable volcanic quartz grain left center; plagioclase grain upper left; volcanic rock fragment upper center; micaceous matrix (one nicol). D, tuffaceous sandstone (T-230, table III-11) from west of Tower, Lake Vermilion Formation. Volcanic rock fragment with plagioclase phenocryst at right; plagioclase grain upper left center; recrystallized composite quartz grains bottom center; fine-grained quartz-plagioclase masses at left probably represent recrystallized volcanic rock fragments. Minor metamorphic mica, amphibole, and carbonate (one nicol). Bar scale is 0.5 mm long.

Table III-11. Modal analyses, in volume percent, of graywackes and tuffaceous sandstones.*

	LAKE VERMILION FORMATION									KNIFE LAKE GROUP										
	West of Vermilion district		Western Vermilion district							Central Vermilion district				Eastern Vermilion district						
Field Name	Graywacke†	Tuffaceous Sandstone	Graywacke†					Tuffaceous Sandstone		Graywacke†				Graywacke†					Tuffaceous Sandstone	
Sample Number	ASW-3	LG-3	T-Echo	T-411	T-944	T-1087	T-1040	T-24	T-230	7156	7160	7200	7101	KL-170C	KL-99	KL-121	KL-204	KL-82	KL-144	KL-167
Volcanic Quartz	2.2	0.8	1.8					2.3	0.2	9.0	0.7	1.8	0.3	11.5	8.0	1.2	1.5	1.1	1.0	4.0
Composite Quartz	6.3	20.7	4.3	6.5	2.8	5.7	1.7	6.0	11.2	1.5	1.8	0.8	2.0		0.5	1.3		37.5	13.5	8.1
Plagioclase	9.2	14.3	42.8	49.8	15.0	20.0	10.2	29.1	21.3	35.7	24.2	29.3	10.0	22.1	19.8	21.7	28.1	17.0	19.5	35.0
K-spar[a]												1.0			2.0	4.0				
Volcanic rock fragments:																				
Felsic	27.7	2.7	22.3	22.8	14.0	24.8	19.2	16.7	28.3	26.3	8.7	36.7	9.0	36.1	24.3	35.1	11.7	13.0[d]	26.8	34.1
Felsic with hornblende											3.2	9.7	0.2	17.8	16.3	2.8	16.5			
Intermediate-mafic										1.3	12.5	2.5	37.5		9.1	6.1	3.8	1.3	1.8	
Recrystallized quartz-plagioclase, fine-grained[b]	30.3	55.8			42.7	32.3	46.0	28.3	29.2											
Quartz-feldspar rock fragments, coarse			0.3		1.0	1.0		0.3				0.3	0.5				0.5	10.7	8.3	
Hornblende	1.8						3.2[c]				11.8	2.2	12.2	1.1	7.1	7.3	22.3		0.8	
Epidote	0.2			0.2			7.3	4.7	1.8	0.2	0.5	11.0		7.2		3.3	0.3		2.7	7.8
Carbonate			10.8	2.8	7.3	5.0	0.2	2.5	1.0	6.3	0.8	0.8	1.5	0.2	0.3	0.2	1.7	8.8	0.3	0.7
"Matrix," micaceous	22.0	5.7	17.0	17.1	15.8	10.5	11.7	9.8	6.7	19.0	30.2	3.5	18.0	3.7	12.3	16.3	11.8	10.3	20.2	10.2
Opaques	0.5		0.3	0.5	1.2	0.7	0.7	0.2		0.7	0.8	0.3	1.0	0.2	0.3	0.5		0.2	0.3	
Miscellaneous			0.2	0.2					0.5		4.9[e]		7.8[f]		0.5	0.3	2.0	1.8	1.2	1.7

Sample Number	Location
ASW-3	Railroad cut just N of Cook, nr SE cor. sec. 12, 62N/19W
LG-3	About 5 mi. E of Sherman's Corner, NE cor. sec. 16, 61N/19W
T-Echo	Echo Point, L. Vermilion, nr cen. sec. 19, 62N/15W
T-411	Island 1 mi. WSW of Ely Island, sec. 24, 62N/16W
T-944	About 3 mi. S of L. Vermilion, SW¼NW¼ sec. 24, 61N/16W
T-1087	About 1 mi. S of L. Vermilion, NW¼NW¼ sec. 12, 61N/16W
T-1040	About 1 mi. S of L. Vermilion, NW¼NW¼ sec. 11, 61N/16W
T-24	Roadcut Hwy 169, about 2 mi. W of Tower, SW¼NE¼ sec. 1, 61N/16W
T-230	About 1 mi. E of Pike Bay, L. Vermilion, NE¼-NW¼ sec. 34, 62N/16W
7156	Wood L., SW¼, sec. 27, 64N/10W (collected by J. C. Green)
7160	Wood L., sec. 34, 64N/10W (collected by J. C. Green)
7200	Rookie L., NW¼NW¼ sec. 10, 63N/10W (collected by J. C. Green)
7101	Moose L., SW¼NE¼ sec. 31, 64N/9W (collected by J. C. Green)
KL-170C	Little Knife L., SE¼SW¼ sec. 6, 65N/6W
KL-99	Knife L., NE¼SE¼ sec. 24, 65N/7W
KL-121	Pickle L., SE¼NE¼ sec. 34, 65N/7W
KL-204	Hanson L., cen. NE¼ sec. 10, 65N/6W
KL-82	Ogishkemuncie L., island SE¼NE¼SE¼ sec. 23, 65N/6W
KL-144	Island in southern baylet of westernmost bay of Cache Bay, Saganaga L., Ont., about 1½ mi. N of International boundary
KL-167	Little Knife L., SW¼SW¼ sec. 32, 66N/6W

* 600 points counted per thin section along traverses normal to bedding

† Imprecise field names

(a) K-spar percentage estimated by staining, and plagioclase content decreased accordingly

(b) Recrystallized quartz-plagioclase masses represent original volcanic rock fragments, plagioclase, quartz, and matrix

(c) Metamorphic origin

(d) May include minor chert

(e) Includes argillite, chert, augite, and metadiabase

(f) All chert

Table III-12. Description of components in graywackes and related rocks.

Volcanic quartz:	Clear, unit grains; sharp extinction; no inclusions; some crystal faces; some well rounded (magmatic resorption); commonly large; some recognizable in spite of cataclasis
Composite quartz:	Grains with more than one extinction unit; commonly undulose; some inclusions and bubble trains; irregular shapes and sizes. Some grains probably cataclasized volcanic quartz. Much probably plutonic in origin
Plagioclase:	Altered and dusky; commonly poorly twinned or untwinned (verified by staining techniques); some zoned crystals; generally albite or albite-oligoclase
K-feldspar:	Altered; difficult to distinguish orthoclase from untwinned plagioclase, and therefore staining of slabs necessary; some perthite
Volcanic rock fragments:	In eastern and central Vermilion district, mostly fine-gr. equigranular but some coarser; of dacite composition, commonly with phenocrysts of quartz, plagioclase and/or hornblende; minor rhyolites and rhyodacite fragments identified by staining for K-feldspar; minor intermediate to mafic fragments identified by dark color and/or felted groundmass of plagioclase laths. May include minor chert. In western Vermilion district and west of Vermilion district, all fine-gr. equigranular dacitic fragments, some with phenocrysts of quartz and/or plagioclase
Recrystallized quartz-plagioclase (fine-grained):	Present in western area where recrystallization was more intense; masses of quartz and plagioclase between grains with distinguishable boundaries; originally dacitic volcanic rock fragments, plagioclase, quartz and matrix; includes minor micas, epidote, amphiboles, etc.
Quartz-feldspar rock fragments (coarse):	Grains with coarse quartz and feldspar grains; probably hypabyssal or plutonic rather than extrusive
Hornblende:	Detrital crystals and fragments in eastern Vermilion district, identical to phenocrysts in associated volcanic rock fragments; pleochroic in yellows and greens. Actinolitic (blue-green) and metamorphic in western Vermilion district and west of Vermilion district; some metamorphic in central part of Vermilion district
Epidote:	Generally fine-gr. dark-colored masses
Carbonates:	Fine- to medium-gr.; late
Matrix:	Fine-gr. interstitial material generally less than 0.03 mm in diameter. Includes chlorite, epidote, sericite and minor fine-gr. quartz and feldspar in eastern and central Vermilion district; chlorite, sericite, biotite and minor fine amphiboles in western Vermilion district and west of Vermilion district
Opaques:	Small sulfide masses and grains, probably pyrite
Miscellaneous:	Includes sphene, tourmaline, zircon, augite, greenstone fragments, metadiabase, quartz veinlets, argillite fragments, chert

grains also have a volcanic origin. A matrix of chlorite, sericite, and epidote comprises from 3 to 30 percent. The mode of a tuffaceous sandstone (table III-11, sample KL-167) is similar.

The volcanic rock fragments in the graywackes differ in composition. Most are felsic, generally dacitic, and many have phenocrysts of quartz, plagioclase, and/or hornblende. The intensity of a sodium cobaltinitrate stain on some slabs ranges from deep to light yellow, indicating that some of this detritus is rhyolitic and rhyodacitic. As much as 37 percent of the grains in some samples are intermediate to mafic in composition and are identified readily by their darker color and/or felted groundmass of plagioclase laths; some have been altered to greenstone.

A few clastic rocks within the Knife Lake Group have plutonic components mixed with the volcanic detritus (fig. III-24B). The best example is the Ogishke granite pebble conglomerate of Gruner (1941). This member was studied by Clements (1903) and Stark and Sleight (1939), and was mapped by Gruner (1941). McLimans (this chapter) has restudied it and other granite pebble conglomerates of the Knife Lake Group. He found that virtually all the granite pebbles probably were derived from the distinctive Saganaga Tonalite, which intruded the lower part of the Knife Lake Group and, after rapid unroofing, shed detritus into the upper part of the group. A related granite pebble conglomerate rests unconformably upon the Saganaga Tonalite at Cache Bay, on Saganaga Lake, in Ontario, just north of the International boundary, and grades upward into arkose. A modal analysis of this arkose (table III-11, sample KL-144) shows that it contains about 13 percent composite (probably plutonic) quartz and about 8 percent coarse quartz-feldspar (plutonic) rock fragments; probably some of the plagioclase is plutonic. However, volcanic components are abundant in the arkose. A graywacke sample collected from just beneath the conglomerate on Ogishkemuncie Lake (table III-11, sample KL-82) contains about 37 percent composite (plutonic) rock fragments mixed with volcanic components (fig. III-24B). The plutonic rock fragments contain quartz, plagioclase, and K-feldspar. Staining of more than 100 slabs of rocks from the Knife Lake Group shows that detrital grains of K-feldspar are present in minor amounts in nearly all the clastic units. Although Green (1970a) has noted K-feldspar phenocrysts in some volcanic rocks of the Knife Lake Group, it is likely that most of these detrital grains were derived from the Saganaga Tonalite, which intruded the Knife Lake volcanic-sedimentary pile during volcanism and sedimentation (Ojakangas, 1972). However, the plutonic detritus is volumetrically important in only a few units.

A few thin beds within the graywacke-slate units show definite evidence of having been deposited as tuffs. Some contain devitrified shards (Green, 1970a, p. 32), and others are made up largely of plagioclase crystals. Some of the very thin beds (fig. III-23C) apparently are thin fine-grained tuffs, as suggested by Gruner (1941, p. 1606).

LAKE VERMILION FORMATION

The Lake Vermilion Formation consists of four informal members (Morey and others, 1970). A metagraywacke-slate member is most extensive, and has a minimum thickness of 3,000 feet. The total thickness may be several times greater, however, for the rocks are tightly folded, making accurate estimates of thickness very difficult. As defined, this member is composed of interbedded metagraywacke and slate and subordinate amounts of conglomerate, dacite tuff, mafic volcanic and volcaniclastic rocks, iron-formation, and metadiabase.

Two varieties of graywacke-slate have been mapped. Greenish-gray, chloritic graywacke-slate is restricted to the vicinity of Lake Vermilion; gray to black, biotitic graywacke-slate has a wide geographic distribution south and west of Lake Vermilion, and extends (W. L. Griffin, 1967, unpub. Ph.D. thesis, Univ. Minn.) as far as 8 miles southeast of Tower. The biotitic variety is somewhat more recrystallized than the chloritic variety, and is interpreted as representing the biotite zone of the greenschist metamorphic facies. The mineralogic differences between the two varieties, however, may be in part the result of original differences in bulk chemical composition.

Clements (1903) described the rocks in the vicinity of Lake Vermilion, but did little petrography. The following is a summary of my (Ojakangas, 1972) studies. The graywackes are recrystallized to varying degrees, but many of the original textures and compositions can be determined. They are composed wholly of volcanically derived components. Equigranular, felsic volcanic rock fragments, volcanic quartz grains, and large sodic plagioclase grains make up nearly all the detrital grains (fig. III-24C). Probably, masses of fine-grained, recrystallized quartz and plagioclase represent, in large part, recrystallized volcanic rock fragments and some recrystallized quartz and plagioclase grains, and most of the composite quartz is volcanic quartz that later was deformed. There is virtually no plutonic debris in the graywackes, but metamorphism has made unequivocal determination of components difficult.

Although largely recrystallized, biotite-quartz-plagioclase schist in the westernmost part of the district and in areas still further west has relict structures suggesting that it is a higher-grade equivalent of the volcanogenic graywacke in the Lake Vermilion Formation (fig. III-25). Original clastic textures given by grains of quartz, plagioclase, and fine volcanic rock fragments are visible in some samples as far as 25 miles west of Tower (fig. III-26). In areas further west and north of the Vermilion fault, which transects the northern part of the Lake Vermilion Formation, granitic plutons (see Hibbing Sheet, by Sims and others, 1970) have caused thorough recrystallization of the graywacke and slate to biotite- and amphibole-bearing schist. Epidote is common and staurolite, garnet, sillimanite, and muscovite are less commonly present. Where present, chlorite apparently is a retrograde product of biotite.

Beneath the metagraywacke-slate member is a light grayish-green, coarse-grained tuffaceous sandstone that is at least 1,500 feet thick. It is well exposed near Tower and Lake Vermilion, and crops out intermittently as a long narrow band that extends 25 miles west from Tower (see Hibbing Sheet, by Sims and others, 1970). This unit was designated feldspathic quartzite in the field during initial mapping of the area (Morey and others, 1970). Subsequent petrographic study has indicated, however, that it has a tuffaceous origin.

Figure III-25. Outcrop of biotite-plagioclase-quartz schist west of Vermilion district, at same locality as sample LG-3. Note excellent preserved bedding.

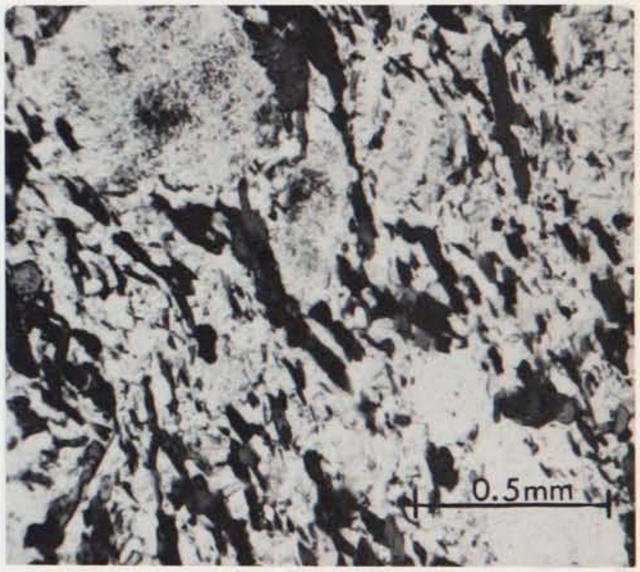


Figure III-26. Photomicrograph of biotite schist from 12 miles west of Tower. Plagioclase grain at upper left; quartz grain lower center; fine-grained quartz-plagioclase masses comprise remainder of light areas and probably represent recrystallized volcanic rock fragments. Dark grains are metamorphic biotite and amphibole (one nicol). Note resemblance to graywacke in Figure III-24C.

The tuffaceous sandstone strongly resembles more massive dacitic tuff and dacitic agglomerate on Lake Vermilion, but is better bedded and sorted. It contains the same major components as the dacitic tuff, including large volcanic quartz grains, large sodic plagioclase grains, and equigranular, felsic volcanic rock fragments (table III-11, samples T-24 and T-230; see also fig. III-24D). The quartz grains are conspicuous because of their bluish color and large size (as much as 4 mm in diameter).

Inasmuch as the Lake Vermilion graywackes, the associated tuffaceous sandstone, and the dacitic tuff and agglomerate contain the same major components, a common origin is indicated for these rocks. Apparently, explosive felsic volcanism provided the detritus. Dacite flows occur with the tuffs, and dacite porphyry dikes cut the Ely Greenstone and the Soudan Iron-formation, which lie stratigraphically beneath the rocks described here. These probably represent "feeder dikes" related to the volcanic centers.

CONCLUSIONS

Volcanic rocks in the volcanic-sedimentary belts of the Canadian Shield have compositions comparable to the basalt-andesite-rhyolite association, which is typical of continental and island arc regions (Wilson and others, 1965; Goodwin, 1968b). New analyses reported by Green (1970a) indicate that the volcanic rocks of the Vermilion district have similar compositions.

Table III-13 lists available chemical analyses of graywacke and a biotite schist from the Vermilion district, as well as analyses of other rocks. The data permit some generalized comparisons. The graywackes of the Vermilion district differ somewhat from other Precambrian graywackes and "the average graywacke" in being lower in SiO_2, Fe_2O_3, and Na_2O and generally higher in other major components. They have compositions similar to dacites in general, and especially to Archean dacites.

A dominant volcanogenic origin for the graywackes of the Vermilion district is indicated by their mineralogy and scant chemical data. The close field relationships and petrographic similarity of the graywackes and tuffaceous sandstones to the dacitic tuffs, dacitic agglomerates, and dacitic porphyries indicate their proximity to the felsic-intermediate volcanic centers from which the volcanic rocks were derived. Apparently, at least two such centers existed during Early Precambrian time, one in the western part of the district, in the vicinity of Lake Vermilion, and the other in the eastern part of the district, in the vicinity of Knife Lake. There may well have been more than one eruptive center, however, in both the eastern and western parts of the district.

The sedimentary structures, especially the abundant graded beds, yield some additional insight into the origin of the graywackes. Although graded beds can result from pyroclastic debris settling directly into a standing body of water, the presence of sedimentary structures characteristic of turbidite sequences (convolutions, flame structures, mud chip zones, internal bedding features) and the poor sorting and abundant matrix suggest that turbidity currents at least moved the volcanic materials from their points of origin (Ojakangas, 1972). The accumulations of pyroclastic and effusive materials at the volcanic centers would have provided the slopes favorable for the development of turbidity currents.

Possibly, much of the graywacke had a pyroclastic origin or was only slightly reworked. Distinguishing a true volcanic tuff from a volcanically derived, immature graywacke, however, is difficult. Any glass that was present would have been devitrified, and metamorphism would have readily altered glassy material and probably obliterated most original shards. The uniformity in the compositions of the Lake Vermilion graywackes and tuffaceous sandstones may be an indication of a rather direct pyroclastic or slightly reworked pyroclastic origin, similar to that for the dacitic, lithic-crystal tuffs in the Lake Vermilion Formation. Also, the abundance of free grains of volcanic quartz and plagioclase in the Lake Vermilion Formation as well as in the Knife Lake Group, and the detrital hornblende in the latter, suggests tuffaceous origins; it would have been difficult to free these minerals as phenocrysts from porphyritic volcanic rocks. The mixed volcanic detritus in the graywackes and tuffaceous sandstones of the Knife Lake Group indicates derivation from a source area containing diverse volcanic rocks. Flows, hypabyssal rocks, and tuffs may have been included as sources.

The composite quartz, coarse quartz-feldspar rock fragments, and the sparse K-feldspar grains in the graywackes and related rocks of the Knife Lake Group, as well as most or all of the granitic pebbles in the conglomerates (McLimans, this chapter), probably were derived from the Saganaga Tonalite at the eastern end of the district. This body has been dated at about 2,700 m.y. (Hanson and others, 1971b), and is approximately contemporaneous with the Vermilion Granite and the Giants Range Granite. All these granites were intruded during the Algoman orogenic event, but the Saganaga appears to have been the first to be unroofed. Apparently it was intruded into older greenstones, and provided sediments to younger sedimentary rocks that lie unconformably on the granite. There is little evidence for a granitic basement for the Vermilion belt (Ojakangas, 1972), although Green (1970a, p. 11) noted a few granitic pebbles dissimilar to the Saganaga Tonalite in conglomerate lenses within the Ely Greenstone, the Newton Lake Formation, and the Knife Lake Group. See McLimans (this chapter) for a further discussion of this problem.

Dickinson's (1970) review of the relationships of Cenozoic-Mesozoic andesites, granites, and derivative sandstones of the circum-Pacific region, and Hamilton and Myers' (1967) work showing that batholiths are commonly emplaced under a thin cover of their own volcanic ejecta, have direct applications to Lower Precambrian rocks. In the Vermilion district, the dominant volcanic detritus and the minor plutonic detritus in the sedimentary rocks appear also to indicate approximately contemporaneous volcanic and plutonic activity.

The graywacke sequences in the Knife Lake Group and the Lake Vermilion Formation are thick, although structural complexities complicate estimates of actual thickness. More than 100 square miles are underlain by graywacke and slate of the Knife Lake Group and about 150 square miles are underlain by graywacke and slate of the Lake

Table III-13. Chemical analyses, in weight percent, of graywackes and related rocks.

Rock type	Graywacke, Vermilion district	Biotite schist, Vermilion district	Graywacke, average	Graywacke, Pre-cambrian	Dacite, Archean	Dacite, average	Rhyodacite, average	Andesite, average	Grano-diorite, average	Dacite porphyry, Ely	Andesite porphyry, Ely	Andesite, pillowed, Ely	Tuff-breccia, Ely
Number of samples	3	1	61	12	272	50	115	49	137	1	1	1	1
Source	Grout, 1933a	Grout, 1933a	Pettijohn, 1963	Pettijohn, 1963	Goodwin, 1968b	Nockolds, 1954	Nockolds, 1954	Nockolds, 1954	Nockolds, 1954	Green, 1970	Green, 1970	Green, 1970	Green, 1970
SiO_2	62.40	63.04	66.75	64.67	61.5	63.58	66.27	54.20	66.88	66.75	60.95	63.61	60.55
Al_2O_3	15.20	16.45	13.54	13.41	15.7	16.67	15.39	17.17	15.66	15.56	15.14	13.91	15.99
Fe_2O_3	0.57	1.32	1.60	1.24	1.83	2.24	2.14	3.48	1.33	1.42	4.34	1.64	2.44
FeO	4.61	4.89	3.54	4.53	4.49	3.00	2.23	5.49	2.59	1.20	1.16	3.90	3.13
MgO	3.52	5.04	2.15	3.23	2.38	2.12	1.57	4.36	1.57	0.92	4.38	4.00	3.50
CaO	4.59	3.17	2.54	3.04	4.41	5.53	3.68	7.92	3.56	3.18	3.32	4.27	6.95
Na_2O	2.68	2.62	2.93	2.99	3.15	3.98	4.13	3.67	3.84	5.60	5.48	5.23	4.14
K_2O	2.57	2.14	1.99	2.02	1.16	1.40	3.01	1.11	3.07	1.73	2.12	0.43	0.86
TiO_2	0.50	0.54	0.63	0.57	0.63	0.64	0.66	1.31	0.57	0.28	0.51	0.61	0.53
P_2O_5			0.16	0.14	0.12	0.17	0.17	0.28	0.21	0.07	0.13	0.14	0.11
CO_2	0.87		1.24	2.15	2.18					1.62	0.43	0.53	0.10
MnO		0.22	0.12	0.13	0.16	0.11	0.07	0.15	0.07	0.03	0.07	0.09	0.08
H_2O	1.63	0.69	2.97	2.14	2.27	0.56	0.68	0.86	0.65	1.42	1.83	2.38	2.23

Vermilion Formation. In addition, the biotite-quartz-plagioclase schist to the west covers more than 400 square miles. As most of these rocks now dip vertically, they probably represent a much wider original areal extent. Apparently, tremendous quantities of volcanic material, largely dacitic in composition, were deposited originally in the region, after extrusion of the basaltic rocks of the Ely Greenstone.

The thick volcanic-sedimentary sequence in the Vermilion district is consistent with Goodwin's (1968b) model for greenstone belts, in which mafic volcanic rocks give way upward to felsic volcanic rocks. In the Lake Vermilion Formation of the western part of the district there is no evidence for rocks more felsic than dacite. In the Knife Lake Group in the eastern part of the district, however, rhyolitic and rhyodacitic detritus is present in the sediments and a few rhyolitic flows are known.

GRANITE-BEARING CONGLOMERATES IN THE KNIFE LAKE GROUP, VERMILION DISTRICT

Roger K. McLimans

Several granite-bearing conglomerates in the Knife Lake Group were mapped by Gruner (1941) in the type area in the eastern part of the Vermilion district. In this report, they are discussed with respect to four separate units that have different geographic occurrences within the Knife Lake area (fig. III-27). Unit 1 is exposed at Cache Bay on Saganaga Lake, Ontario; unit 2 is exposed in the vicinity of Ogishkemuncie Lake, in the southeastern part of the map area; unit 3 is exposed at Nawakwa Lake, between Cache Bay and Ogishkemuncie Lake; and unit 4 crops out in the area of Ensign and Explorer Lakes, in the western part of the map area. In addition, thinner, less extensive conglomerate beds in the Knife Lake Group in the central part of the Vermilion district, mapped recently by Green and others (1966), are discussed briefly.

The only plutonic rock type recognized in the conglomerates is the Saganaga Tonalite, which is exposed at the eastern terminus of the Vermilion district in the vicinity of Saganaga Lake (fig. III-27). Typically, it is a distinctive rock that contains large ovoid quartz aggregates that resemble eyes and which have slight undulose extinction and, locally, crenulated borders. Formerly called the Saganaga Granite (A. Winchell *in* N. H. Winchell, 1888), the rock has been renamed (Goldich and others, in press) the Saganaga Tonalite because tonalite is the dominant lithology in the batholith. As noted in Table III-14, however, the rock is granodioritic on the western margin of the batholith. Modes of samples from typical outcrops of the Saganaga Tonalite and from clasts of the same rocks in the conglomerates are given in Table III-14.

DESCRIPTION OF CONGLOMERATE UNITS

The conglomerate at Cache Bay (unit 1) lies directly on Saganaga Tonalite, and consists entirely of cobbles and boulders of the Saganaga in a matrix of finer granitic detritus. Megascopic modal analyses (table III-15) on outcrops of the conglomerate indicate proportions of about 74 percent granite clasts and 26 percent arkosic matrix. The unit grades upward into massive arkose.

The conglomerates of unit 2, at Ogishkemuncie Lake, comprise the type area of the Ogishke conglomerate, as originally defined (A. Winchell *in* N. H. Winchell, 1887); they contain clasts of all rock types in the area that are older than the conglomerates, including Saganaga Tonalite, greenstone, mafic and intermediate volcanics (mostly andesitic), felsic volcanics (rhyodacite and dacite porphyry), aplite, chert, jasper, slate, tuff, and other sedimentary rocks. Jasper is distinguished separately in Table III-15 from other cherts because of its restricted occurrence and stratigraphic importance. Stark and Sleight (1939) divided the conglomerate at the type locality into three distinct lithotopes based on the presence of certain clasts: (1) a basal granite facies; (2) a jasper facies; and (3) an upper porphyry facies. Gruner (1941) discredited this subdivision, stating that the facies were local and therefore not mappable. However, my work shows that the conglomerates in the Ogishkemuncie Lake area can be divided into a non-jasper-bearing lower facies (unit 2A) and a jasper-bearing upper facies (unit 2B).

Unit 2A, as defined, consists of pebble- or boulder-conglomerate and interbedded graywacke. The conglomerates contain all the types of clasts named above except jasper (table III-15), and the matrix consists of detritus of the same rock types. The conglomerate units range in thickness from a foot or less to about 40 feet (figs. III-28A and B). The interbedded graywackes range from thinly laminated beds only a few inches thick to massive beds about 12 feet thick. Many of the graywacke beds are graded as are some of the conglomerate beds.

Unit 2B includes jasper-bearing conglomerate exposures at Ogishkemuncie, Alpine, Kingfisher, Spice, and Jenny Lakes. These conglomerates also contain all the rock types given above (table III-15). The lowermost beds are cobble-size conglomerate, which are interstratified locally with graywacke beds. The cobble-size beds grade upward as well as laterally into finer grained beds. The scarcity of interbedded graywackes and of regular bedding is notable.

The unit (3) exposed in the vicinity of Nawakwa Lake is a boulder conglomerate that overlies a graywacke-slate unit that lies above the conglomerate at Ogishkemuncie Lake (unit 2). Unlike the conglomerate at Ogishkemuncie Lake and Cache Bay, it lacks visible grading and is coarse throughout. Except for jasper, the clasts consist of all rock types previously named. Saganaga Tonalite clasts are particularly noteworthy because of their size and abundance (fig. III-28C). Two megascopic modal analyses on outcrops along the western shore of the lake (table III-15) show that granitic clasts account for about 50 percent of the conglomerate; across the lake, however, greenstone is the dominant clast, and the matrix is composed largely of greenstone debris.

The granite-bearing conglomerate of unit 4, at Ensign Lake, is very similar to that of unit 2A at Ogishkemuncie Lake, for it contains interbedded, graded graywacke and has the same types of clasts (table III-15); the clasts are more highly deformed (fig. III-23D), however. Unit 4 may be contemporaneous with unit 2A (Ogishkemuncie Lake), and possibly originally was continuous with it.

In the central part of the Vermilion district, conglomerate constitutes a widespread basal unit of the Knife Lake Group (Green, 1970a). Metavolcanic clasts are dominant,

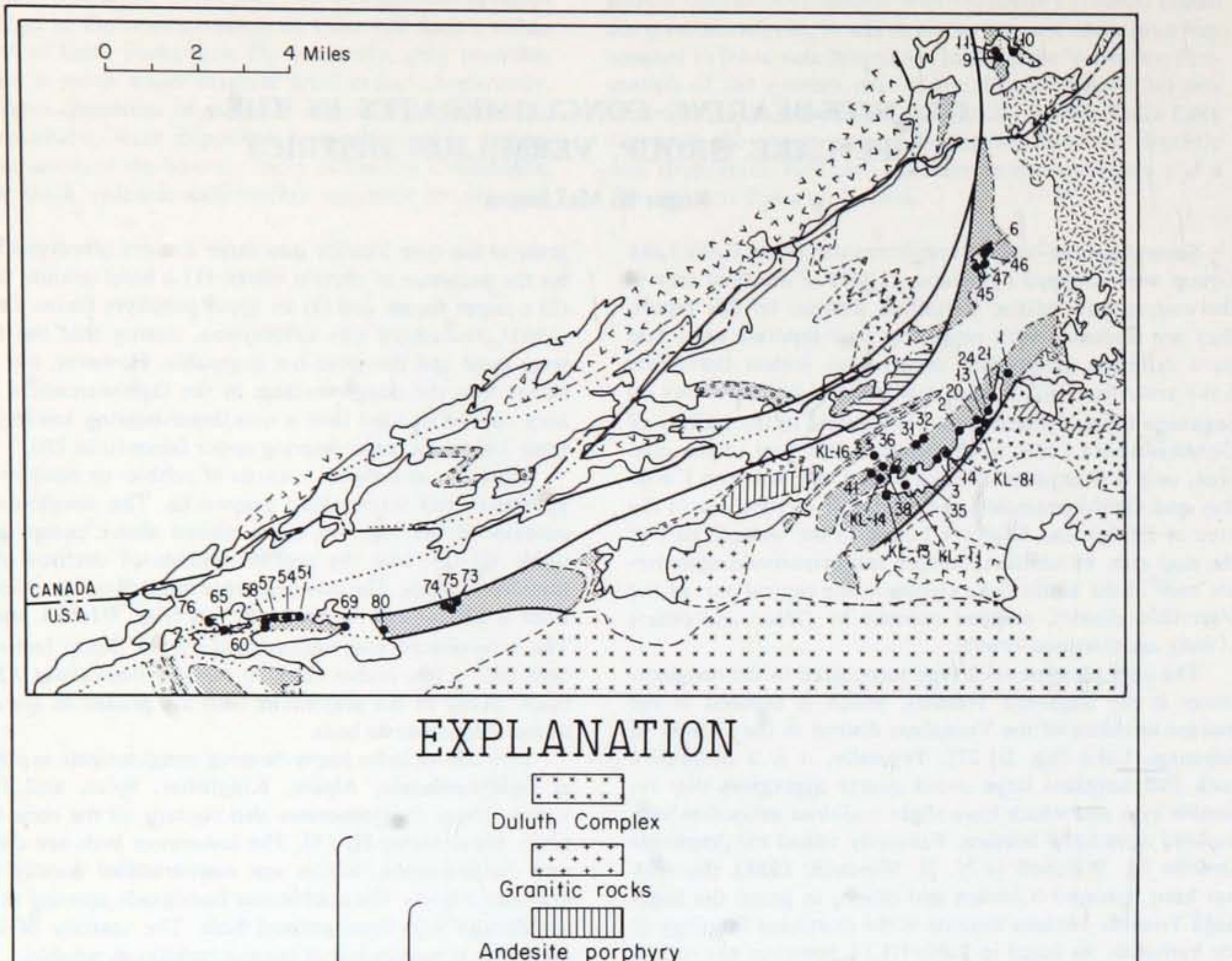


Figure III-27. Generalized geologic map of the Knife Lake area, showing locations of samples discussed in text (modified from Gruner, 1941, pl. 1).

Table III-14. Modes, in volume percent, of Saganaga Tonalite and of granitic clasts in conglomerates.

Unit		Outcrop or sample number	Plagioclase	K-spar	Quartz	Hornblende	Chlorite	Sphene	Miscellaneous*
Saganaga Tonalite		9	43.6	15.7	30.0	8.3	2.0	.2	.2
		KL-137	49.2	15.5	21.3	11.0	2.0	.5	.5
		KL-146	65.4	6.3	18.8	5.7	3.0	.3	.5
		KL-216	49.1	25.0	13.2	7.3	3.7	.5	1.2
		KL-217	60.2	5.3	20.2	6.3	5.7		2.3
Average:			53.3	13.6	20.7	7.7	3.3	.3	.9
Granitic clasts in conglomerates	1	11	43.2	15.0	32.2	7.7	1.7	.2	
	2A	2-3	36.0	18.0	38.6	4.0	2.3	.6	.5
		3	49.4	13.3	24.0	6.7	5.3	.3	1.0
	2B	31	54.9	17.7	18.8	1.3	4.8	.3	2.2
		6	56.3	18.2	16.8	6.2	1.8	.7	
	3	45-5	50.3	12.0	19.7	14.0	4.0		
		45	51.9	10.0	19.5	11.5	4.7	.2	2.2
	4	57-1	47.8	9.2	30.8	6.5	3.2	.5	2.0
		76	56.0	19.8	16.2	1.6	1.8	.4	4.2
Average:			49.5	14.8	24.1	6.6	3.3	.4	1.3

* Miscellaneous includes accessory minerals and chlorite, carbonate, and epidote which occur as veinlets. Locations of samples are given in Figure III-27.

but granitic clasts have been reported from near Wood and Moose Lakes. Except for lacking distinctive quartz eyes, the granitic clasts resemble the Saganaga Tonalite. In contrast, granitic clasts (Green, 1970a) from the North Kawishiwi River area do not resemble the Saganaga Tonalite; instead, they are foliated and some are layered.

CONGLOMERATE MATRICES AND ASSOCIATED GRAYWACKES

Modal analyses of the conglomerate matrices, of intercalated graywacke beds, and of graywackes from units that are stratigraphically above and below the conglomerate units, are similar (table III-16). Volcanic rocks constitute the dominant clasts, as they do in the coarse fraction of the conglomerates. The compositions are similar also to the graywackes described by Ojakangas (1972 and this chapter). In general, the graywackes and the conglomerate matrices are poorly sorted, consist of angular to subangular grains, and are texturally and compositionally immature.

PROVENANCE

The conglomerates show a mixed provenance, with volcanic rocks and the Saganaga Tonalite being the dominant source materials. As shown in Table III-16, volcanic rock fragments far exceed plutonic rock fragments in both the matrices of the conglomerates and in the graywackes. Interestingly, the conglomerate matrices and the graywackes contain an average of about 20 percent quartz, whereas the graywackes that underlie and overlie the conglomerate units contain about 12 percent quartz (Ojakangas, this chapter). Probably, the higher quartz content of the finer grained part of the conglomerate units reflects the greater amount of detritus from the Saganaga Tonalite in these beds, for the Saganaga contains an average of about 21 percent quartz.

The granitic clasts in all four conglomerate units delineated in the eastern part of the district are entirely Saganaga Tonalite. The granitic clasts in conglomerates near Moose and Wood Lakes, in the central part of the Vermilion district, may also have been derived from the Saganaga batholith, but those from the North Kawishiwi River area may have been derived from other granitic rocks that were exposed during Knife Lake time, as suggested previously by Green (1970a).

SEDIMENTATION

Sedimentation proceeded throughout Knife Lake time with little interruption, for unconformities between members of the Knife Lake Group have not been recognized (Gruner, 1941). The majority of the sequence is composed of graywacke and associated mudstone (now slate) that are poorly sorted and immature in composition, implying rapid erosion and deposition from source areas of substantial relief. The compositions of the graywackes indicate that they were derived from a basalt-andesite-rhyolite volcanic suite (Ojakangas, 1972 and this chapter). The graywackes contain sedimentary structures typical of turbidite sequences.

Unroofing of the Saganaga Tonalite led to deposition of the granite-bearing conglomerates. These conglomerates, some of which are graded, are interbedded with graded graywackes. The sharp contacts between conglomerate beds and the lack of cross-stratification probably are indicative of deposition by turbidity currents or submarine slumping.

The slope at the margin of the basin must have been relatively steep, for clasts of Saganaga Tonalite as much as 15 cm in diameter were transported as far as Ensign Lake, some 25 miles from the source area. Probably, temporary accumulations of sediments built up on the slope until they became unstable and then slumped, with some of them generating turbidity currents. The slumping may have been triggered by volcanic eruptions, earthquakes, or perhaps even by storm waves. Similarly graded conglomerate turbidite sequences have been described from California by Fisher and Mattison (1968). Recently, Walker and Pettijohn (1971) attributed deposition of the granite-bearing conglomerates in the Minnitaki basin, Ontario, to turbidity currents.

The boulder conglomerate at Cache Bay (unit 1) is overlain conformably by the same graywacke-slate sequence that overlies the conglomerate at Ogishkemuncie Lake (unit 2), and accordingly these conglomerates may be approximately contemporaneous, as suggested by Gruner (1941). The conglomerate at Cache Bay probably represents granitic debris that accumulated on the shelf near the exposed granitic land mass. Also, the conglomerate at Ensign Lake (unit 4) probably is approximately contemporaneous with the conglomerate at Ogishkemuncie Lake. Following deposition of these granite-bearing conglomerates (units 1, 2, and 4), volcanism within the basin again became a major process, and several thousand feet of volcanogenic graywacke and mudstone were deposited. At the time of deposition of conglomerate unit 3 (Nawakwa Lake), which is younger than

Table III-15. Megascopic modes of clasts in conglomerate units of the Knife Lake Group.

Unit	Outcrop number	Saganaga Tonalite	Coarse green-stone	Fine-Med. green-stone	Mafic to inter. volcanics	Felsic to inter. volcanics	Aplite	Chert	Jasper	Slate	Tuff and sedimentary clasts	Miscel-laneous*	Matrix
1	11	73.7											26.3
	24	6.8	11.3	20.5	8.2	9.7	.5	.2		1.1	.5	.2	41.0
	13A	21.1	5.4	7.6	9.5	11.8	7.3				5.7	.3	31.3
	13B	3.5	1.8	27.0	6.7	6.8		.3			1.2	1.0	51.7
	20	9.1	.9	6.3	8.3	9.1	.2	6.1			3.2	1.9	54.9
2A	14A	5.7	11.9	1.3	13.1	15.2	.2	.5		1.0	4.0	.5	46.6
	14B	4.0	15.7	2.4	11.1	24.7	2.9			1.5	3.1		34.6
	3	7.9	6.6	8.7	4.8	10.5		.5			4.6		56.4
	35	4.2	5.0	5.6	6.0	6.4		.9		.7	2.2		69.0
	AVE	7.8	7.0	10.0	8.0	12.0	1.0	1.0		.5	3.0	.5	49.1
	1	10.5	2.1	2.9	3.8	35.2			3.9		.4	2.0	39.2
2B	36	6.3	3.8	12.5	10.0	7.2	4.4	2.8	3.2		.4	3.4	46.0
	AVE	8.5	3.0	7.7	6.9	21.2	2.2	1.4	3.6		.4	2.7	42.4
	46	6.0	7.0	9.2	.2	4.3	.2	.2		.1	.2	.2	72.4
	47	50.7	9.6	5.0	.7	1.0	.5					4.0	32.1
3	45	53.4	1.6	7.3	9.5						1.1		29.1
	AVE	36.7	6.7	7.2	3.5	1.8	.2	.1		Tr	.4	1.4	44.0
	74	13.1	6.2	8.4	15.0	1.1		.5		.2	4.7	.9	49.9
	54	7.8	.1	19.9	21.9	4.3	.8	1.0		1.8	7.2	.4	34.8
4	57	6.2	6.1	8.4	12.0	1.0	.7	1.5		.9	1.4		61.8
	AVE	9.0	4.1	12.2	16.3	2.1	.5	1.0		1.0	4.4	.4	50.0

* Miscellaneous includes vein quartz and some unknowns. Outcrop numbers are arranged from east to west within each unit. Outcrops are plotted in Figure III-27. (See R. K. McLimans, 1972, unpub. M.S. thesis, Univ. Minnesota, Duluth, for detailed descriptions of clast lithologies.)

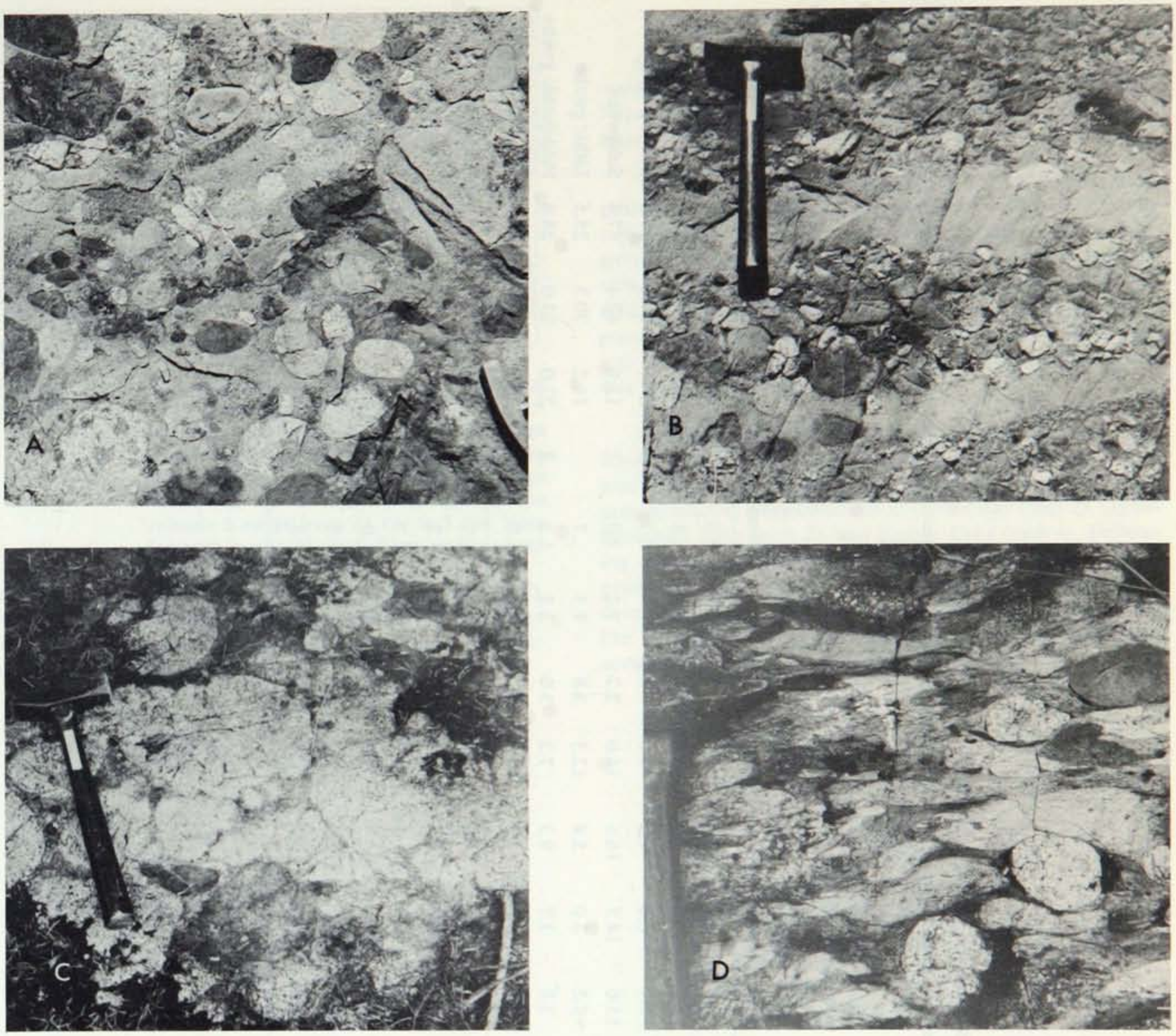


Figure III-28. Conglomerate and associated rocks in Knife Lake Group. A, massive conglomerate at Ogishkemuncie Lake (unit 2A). Most of the light gray clasts are Saganaga Tonalite. Modes from this outcrop are given in table III-15, numbers 14A and 14B). B, interbedded conglomerate and graywacke at Ogishkemuncie Lake (unit 2A). Modes from this outcrop are given in table III-15, numbers 14A and 14B. C, conglomerate along the western shore of Nawakwa Lake (unit 3). Dominantly Saganaga Tonalite clasts. Mode from this outcrop is given in table III-15, number 45). D, deformed conglomerate on south shore of bay in sec. 9, Ensign Lake (unit 4). Note stretching of the clasts. The matrix is chloritic schist. Modes from this outcrop are given in table III-15, numbers 54 and 57.

the other units, the topography in the source area had changed so that the Saganaga batholith again was a major source of debris. Boulders of granite and other material were deposited near the shore at the base of a probably high source area. Farther offshore, volcanic sediments still were accumulating.

Determination of the history and paleogeography of the basin is impaired by the lack of paleocurrent indicators. Only one graywacke bed with cross-stratification was observed during my study, and it indicated transport from northeast to southwest. However, a study of the change in the maximum and average size of Saganaga Tonalite clasts shows a decrease in size westward away from the Saganaga batholith (fig. III-29). It appears that transport by turbidity currents and slumping was along the present tectonic strike, with the granitic clasts being supplied at the eastern end of

Table III-16. Modes, in volume percent, of graywacke and conglomerate matrices.

	Unit	Outcrop number	Plagioclase	K-spar	Quartz	Felsic volcanic rock fragments	Felsic-intermediate volcanic rock fragments	Mafic-intermediate volcanic rock fragments	Mafic volcanic rock fragments	Plutonic rock fragments	Hornblende	Chert	Miscellaneous*	Matrix**	Total rock fragments	Total feldspar	Name
Matrices	3	46	15.1	.8	14.2	9.0	4.5	6.8	4.5	.8	11.2	3.0	2.7	27.4	25.8	15.9	Lithic gwcke
	3	45	21.0	1.2	24.0	5.2	5.5	5.3	7.0	1.7	3.8	.5	1.0	23.8	24.7	22.2	Lithic gwcke
	2B	17	19.0	1.0	19.7	5.7	6.0	7.5	7.0	.8		6.7	2.5	24.1	27.0	19.0	Lithic gwcke
	2B	32	5.7	1.7	5.0	15.0	14.8	16.6	16.0	2.2	7.2	3.0	.2	12.6	62.4	7.4	Subgwcke
	2A	2-1	23.7	1.8	26.0	5.2	3.0	5.8	12.5	3.8	1.3	.7		16.2	30.3	25.5	Lithic gwcke
	2A	2-3	23.6	5.0	17.8	3.8	3.7	9.3	7.2	3.0	3.2	1.2	.2	22.0	27.0	28.6	Feldspathic gwcke
	2A	3	19.7	3.6	22.0	5.0	5.2	5.3	4.7	3.2	3.2	1.0	1.5	25.6	23.4	23.3	Lithic gwcke
Graywacke***	2A-i	2-2	34.2	2.3	20.7	3.0	3.5	4.8	1.0	.8	4.2	1.7	1.5	22.3	13.1	36.5	Feldspathic gwcke
	2A-i	KL-5	31.4	3.3	21.2	5.7	4.5	3.5	2.5	1.0	.5	1.2	1.2	24.0	17.2	34.7	Feldspathic gwcke
	2	KL-14	32.0	3.5	4.8	17.0	7.3	1.5	1.0	1.7	4.5	2.7	.3	23.7	28.5	35.5	Feldspathic gwcke
	2	KL-16	21.5	3.0	10.0	10.0	8.5	11.5	4.5	2.0	2.5	1.0	.5	25.0	36.5	24.5	Lithic gwcke
	4	KL-249	20.0	1.0	13.0	5.0	6.0	3.5	1.0	1.0	9.5	1.0	2.0	37.0	16.5	21.0	Feldspathic gwcke
	4	KL-243A	22.0	4.0	20.5	6.2	6.0	3.3	2.8	.3	2.0	.5	3.8	28.6	19.3	26.0	Feldspathic gwcke

* Includes accessory minerals and opaques
** Includes epidote and chlorite veinlets
*** i = Interbedded graywacke

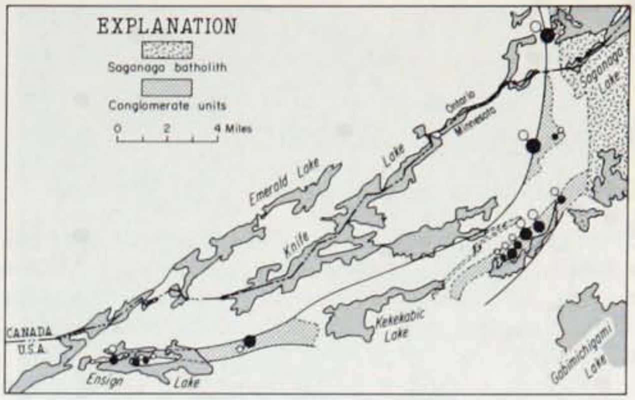


Figure III-29. Map showing the apparent maximum diameter (solid circles) and the apparent average diameter (open circles) of Saganaga Tonalite clasts in selected exposures of conglomerates (solid circles indicate a range from >50 cm to 15 cm and open circles indicate a range from 21 cm to 5 cm. Note decrease in size of clasts from east to west).

an elongate basin or trough. Walker and Pettijohn (1971) similarly concluded that conglomerate transport in the Minnitaki basin, Ontario, took place along the present tectonic strike.

CONCLUSIONS

Source areas for the granite-bearing conglomerates were mainly mafic to felsic volcanic rocks and the Saganaga Tonalite, which was emplaced during the time of volcanic activity in the district.

The conglomerates and associated graywackes were deposited in an island-arc environment, characterized by rapid erosion and little weathering. Transport of the conglomerates apparently was parallel to the present regional strike, and was accomplished by turbidity currents generated by slumping of debris on a steep initial slope. Slumping possibly was triggered by eruptions and earthquakes. The size distribution of granitic clasts indicates that transport was from east to west.

Physiographic conditions and topography were constantly and rapidly changing during Knife Lake time. Initial exposure of the Saganaga Tonalite gave rise to the granite-bearing conglomerates at Cache Bay (unit 2), Ogishkemuncie Lake (unit 2), and Ensign Lake (unit 4). Following deposition of these conglomerates, the sediments were primarily volcanically derived until conditions again changed, causing renewed erosion of the Saganaga Tonalite. This gave rise to the deposition of the granite-bearing conglomerate at Nawakwa Lake (unit 3).

BURNTSIDE GRANITE GNEISS, VERMILION DISTRICT

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

The Burntside Granite Gneiss, named by Grout (1926, p. 29) from exposures on Burntside Lake, is of interest because of previous uncertainties concerning its relationship to adjacent rocks. Grout (1929, p. 796) mapped the gneiss and determined it to be an early phase of the igneous activity accompanying the Algoman orogeny. Goldich and others (1961, p. 58-59) suggested, however, that the gneiss possibly is older than the Ely Greenstone. Detailed mapping by us in the Shagawa Lake quadrangle and consideration of the regional geology confirm Grout's views. The rock is a foliated leucocratic tonalite (trondhjemite) that intrudes the adjacent country rocks and in turn is cut by apophyses of the main pink granite of the Vermilion granite-migmatite massif.

The unit is exposed on the islands and shores of the main part of Burntside Lake, and underlies much of the lake (fig. III-30). It extends eastward into the Ely 7.5-minute quadrangle, and terminates against the Vermilion

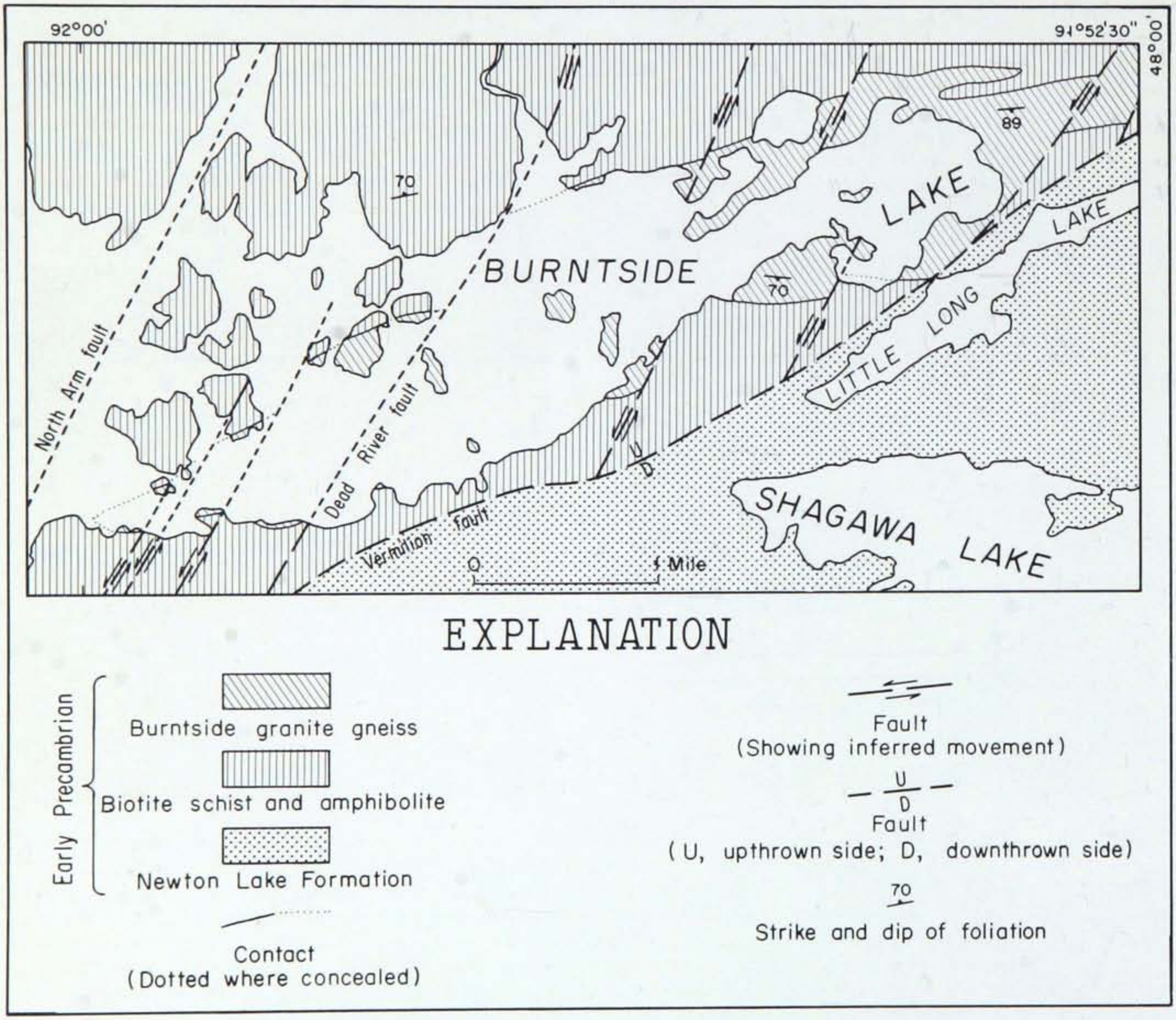


Figure III-30. Geologic map of part of Burntside Lake area; geology by P. K. Sims, 1967.

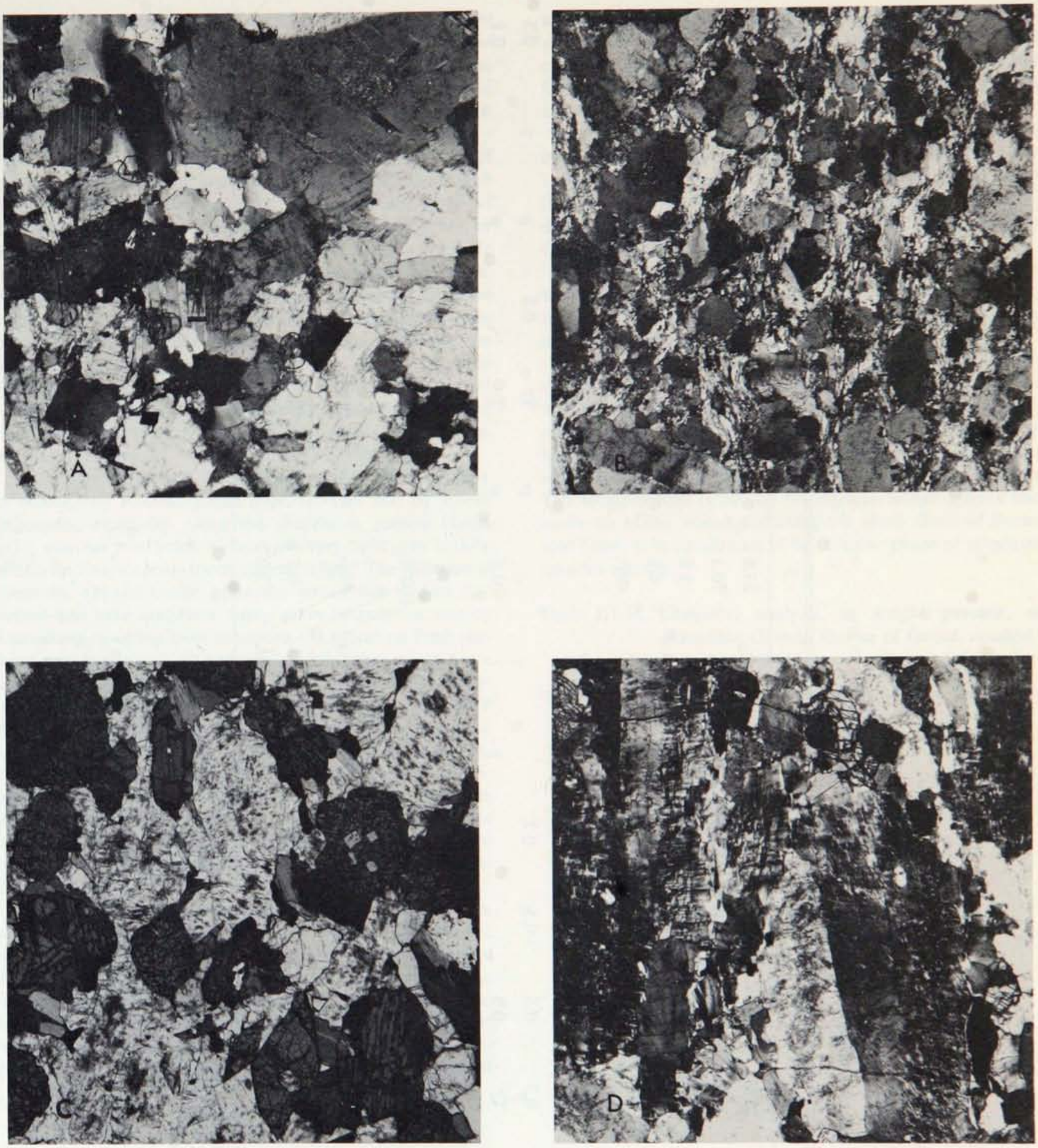


Figure III-31. Photomicrographs of Burntside Granite Gneiss of Grout and of Linden pluton. Photomicrographs by R. B. Taylor. A, least deformed facies of Burntside Granite Gneiss, from east shore of Burntside Lake (ENW-48A). Crossed nicols. X20. B, cataclastic facies of Burntside Granite Gneiss, from east shore of Burntside Lake. (Note that larger grains have strain shadows and generally are separated by finely ground material.) ENW-796. Crossed nicols. X20. C, syenite from Linden pluton. Consists of microcline microperthite (blotchy, light gray), subhedral green clinopyroxene, and lesser sphene, apatite, and biotite. GNW-7-B. Plane light. X20. D, same as C. Large twinned grains of microcline microperthite have shadowy extinction, indicative of strain. Small blebs of plagioclase occur along grain boundaries and probably are a result of exsolution. Crossed nicols. X20.

Table III-17. Modes, in volume percent, of Burntside Granite Gneiss of Grout.[1]

	1	2	3	4	5	6	7	8	9	10	11	12	Average
Quartz	21.9	12.1	34.4	41.4	p	p	23.6	p	25.6	23.9	p	p	26.1
Plagioclase	71.2	74.2	58.8	50.5	p	p	70.1	p	69.7	66.8	p	p	65.9
Microcline	2.0	6.6	3.3	7.1	p	p	1.8		1.9	5.6	p	p	4.0
Biotite	3.5	5.3	2.3	Tr	p	p	4.2		1.7	Tr			3.2
Muscovite							Tr		0.1				Tr
Hornblende	0.7	0.8										p	0.2
Chlorite		0.4	0.6	0.8	p			p		2.9	p	p	0.1
Magnetite	0.7	0.3	0.5	0.2				p	Tr	0.5	p	p	0.3
Sphene	Tr	0.3			p		0.3	p	Tr		p		0.1
Other accessories	Tr	Tr	0.1	Tr	p	p	Tr	p	1.0	0.3	p	p	0.1
Total	100.0	100.0	100.0	100.0			100.0		100.0	100.0			100.0
Composition of plagioclase	An_{23}	An_{30}	An_{23}	An_{20}	An_{18}	An_{14}	An_{21}	An_{24}	An_{19}	An_{19}	An_{22}	n.d.	An_{22}
Grain size, in mm, groundmass	0.1	0.2	0.4	0.2	0.1	n.d.	0.5	0.1	0.3	n.d.	0.1	n.d.	n.d.
Crystals	1.5	1.1	1.5	1.4	1.5	n.d.	3.0	2.0	2.1	n.d.	0.8	n.d.	n.d.

[1] Sample localities shown on Figure III-30
(p, present; Tr, trace; n.d., not determined)

fault. The body trends east-northeast, subparallel to the regional trend, and has an exposed length of about 9 miles and a maximum width of 1.1 miles. It is bounded on the north by biotite schist and on the south by biotite schist and amphibolite, and is intrusive into each of these rock types. At places, small inclusions of both types of country rock occur in the trondhjemite. The intrusive relations of the trondhjemite to the country rocks are particularly well shown along the south contact, on the south shore of Burntside Lake. From a point on the shore due south of Miller island to east of the township line, the trondhjemite intertongues repeatedly with biotite schist and amphibolite, and on a broader scale crosscuts the structure of these rocks. The crosscutting relationship is dramatically illustrated by the difference in rock types exposed along the shore and on the upland surface in this area. Along the shore only trondhjemite is exposed, whereas on the upland surface only country rock is exposed, yet the two types of rocks and the foliations within them trend (eastward) toward one another.

The trondhjemite is a light-gray, nearly white, weakly to moderately foliated gneiss (figs. III-31A and B). Fresh exposures, especially along the shorelines, appear bluish gray, whereas weathered surfaces are very light gray (nearly white) and have conspicuous quartz "eyes." The foliation is given by aligned mafic minerals, which are widely dispersed and have indistinct, fuzzy grain boundaries, and by a streaking resulting from cataclasis. At places on fresh surfaces a faint mineralogic layering is visible that may lie at a small angle to the cataclastic foliation. The cataclastic foliation is subparallel to the foliation in the adjacent country rocks. Cataclasis is evident in nearly every thin section, and is most extensive adjacent to the north and south contacts. It is manifested mainly as mortar structure, but also as protomylonite (Higgins, 1971). Locally, mylonite zones as much as a foot thick can be seen; observed zones generally strike northeast, at large angles to the regional foliation, and are related to the late north-northeastward-trending faults in the area.

In addition to dominant plagioclase and quartz, the rock contains sparse biotite, chlorite, and hornblende as mafic minerals and a small percentage of microcline (table III-17). Total mafic minerals rarely exceed five percent. Green hornblende occurs only at the eastern end of Burntside Lake. Biotite, which is brown to greenish brown, is present in the northern part of the body but is partially altered to chlorite near the northern contact; it is completely altered to chlorite in the outcrops along the south shore of Burntside Lake. Plagioclase is altered in those rocks that contain chlorite. Accessory or alteration minerals not listed specifically in Table III-17 include zircon, apatite, rutile, allanite, epidote, calcite, and leucoxene.

A chemical analysis of a sample of the gneiss collected by Grout (1926, p. 29) from the portage from Burntside Lake to Little Long Lake (see fig. III-30) is given below (table III-18). The rock is similar chemically as well as mineralogically to the trondhjemite gneiss in the Northern Light Gneiss (Goldich and others, in press).

The rock is interpreted to have consolidated as a hornblende-biotite trondhjemite, which subsequently was partly retrograded, probably virtually contemporaneously with the cataclastic deformation.

The age of the biotite schist and amphibolite country rock relative to the Ely Greenstone, Knife Lake Group, and Newton Lake Formation is not known directly. The compositions of the metamorphic rocks, however, are consistent with an interpretation that they represent amphibolite-facies equivalents of a part of the volcanic-sedimentary sequence. The trondhjemite is cut by hornblende lamprophyres that occur on Miller island and along the south shore of Burntside Lake. It is considered to be an early phase of Algoman igneous activity.

Table III-18. Chemical analysis, in weight percent, of Burntside Granite Gneiss of Grout. Analyst: Douglas Manuel.

SiO_2	68.54
Al_2O_3	17.89
Fe_2O_3	1.77
FeO	0.52
MgO	1.22
CaO	4.02
Na_2O	5.14
K_2O	1.05
H_2O+	0.46
H_2O-	0.18
TiO_2	0.20
S	0.06
Total	101.05

SAGANAGA BATHOLITH

Gilbert N. Hanson[1]

The Saganaga batholith, along the International boundary in the vicinity of Saganaga Lake, is a rather homogeneous, composite intrusive body that has maximum dimensions at the surface of about 14 by 20 miles. Formerly, the rocks comprising the batholith were called Saganaga Granite (A. Winchell *in* N. H. Winchell, 1888), but further study (Goldich and others, in press) has shown that the designation Saganaga Tonalite is more appropriate.

The batholith and the rocks in adjacent areas have been investigated intermittently for nearly a century. Following the earlier reconnaissance investigations by A. Winchell (*in* N. H. Winchell, 1888), Grout (1929, 1936) mapped the Saganaga batholith and adjacent rocks, emphasizing its petrographic and structural features. At about the same time, Tanton (1931, 1938) mapped the Canadian part of the area. Somewhat later, Gruner (1941) mapped the area along the western side of the batholith, as part of a detailed study of the stratigraphy and structure of the Knife Lake Group. The results of these investigations were summarized by Grout and others (1951), and formed the basis for the geochronologic studies by Goldich and others (1961). More recently, Harris (1968) mapped the northern part of the batholith and adjacent areas in Ontario, Morey and others (1969, Minn. Geol. Survey, unpub. open-file map) mapped the Long Island Lake quadrangle, along the southern margin of the batholith, and S. S. Goldich and I carried out petrologic, structural, and geochronologic studies in the Saganaga Lake-Northern Light Lake area (Hanson and others, 1971b; Goldich and others, in press).

GEOLOGIC SETTING

The Saganaga Tonalite intrudes greenstone and the Northern Light Gneiss of Goldich and others (1961, p. 44) and is overlain along the western margin by metaconglomerate and associated metasedimentary rocks of the Knife Lake Group (fig. III-32). It is overlapped locally along the southern margin by rocks of the Middle Precambrian Animikie Group. The Animikie Group was intruded by mafic sills that are referred to as Logan sills (Lawson, 1893, p. 48) or Logan intrusions (Grout and others, 1959), and by the Duluth Complex, both Late Precambrian in age.

The Saganaga Tonalite has been considered a typical Laurentian intrusive body (Grout and others, 1951), for it intrudes greenstone, previously called Keewatin and equated with the Ely Greenstone, and is overlain by the Knife Lake Group. As discussed earlier (Sims, this chapter), however, the age of the greenstone it intrudes is equivocal and the conglomerate, presumed originally to be at the base of the Knife Lake Group, is believed to be stratigraphically higher in the succession (Gruner, 1941); accordingly, the Minnesota Geological Survey no longer considers the term "Laurentian" applicable to the Vermilion district (Morey and others, 1970).

The major greenstone body on the north side of the batholith is separated from it at most places by a high-angle fault that roughly follows the north shore of Saganaga Lake (fig. III-32). The body consists of steeply-dipping mafic metavolcanic rocks, many of which are pillowed flows, and lesser felsic metavolcanic rocks (Harris, 1968). As shown on Figure III-8, the body is tentatively correlated by the Minnesota Geological Survey with the formally designated Newton Lake Formation.

A lesser body of greenstone occurs on the south side of the batholith. It consists dominantly of metabasalt and metadiabase, which pass southward into hornblende andesite porphyry (Weiblen and others, 1971; Gruner, 1941, pl. 1). Adjacent to the contact with the Saganaga Tonalite, these rocks are metamorphosed to amphibolite-facies assemblages (Weiblen and others, 1971), but the contact zone is complicated by a fault that strikes northwestward. Possibly, this body is a part of the Knife Lake Group, as suggested by Sims (this chapter), but it may be older.

The Northern Light Gneiss of Goldich and others (1961), a foliated trondhjemitic gneiss, lies east of the Saganaga batholith and is intruded by the Saganaga Tonalite. The gneiss is dominantly a fine-grained quartz-rich leucotonalite, which contains lenses and layers of amphibolite and lesser amounts of metarhyodacite and metarhyolite; it is considered to represent an original volcanic pile of lavas and possibly pyroclastic materials (Goldich and others, in press).

Conglomerate and associated sedimentary rocks of the Knife Lake Group (Gruner, 1941) overlap the western margin of the batholith. At the well known locality in Cache Bay of Saganaga Lake (fig. III-32), a conglomerate containing boulders of the Saganaga Tonalite rests directly on the batholith. The conglomerate probably is equivalent to outcrops at Ogishkemuncie Lake, which is the type locality of the Ogishke conglomerate of A. Winchell (*in* N. H. Winchell, 1887). Overlying the conglomerate stratigraphically on the United States side of the International boundary are graywacke, slate, and tuff, aggregating about 2,000 feet in thickness, which grade upward into younger conglomerate, graywacke, and agglomerate (Gruner, 1941, p. 1604-1605; McLimans, this chapter). The conglomerate and associated strata along the western margin of the batholith dip westward at an angle of about 70°. Gruner (1941) has shown conclusively that the conglomerate is not a basal conglomerate of the Knife Lake Group, and that it was not deposited on a major unconformable surface, as inferred by Clements (1903).

[1] Research was supported by NSF grant GA 575.

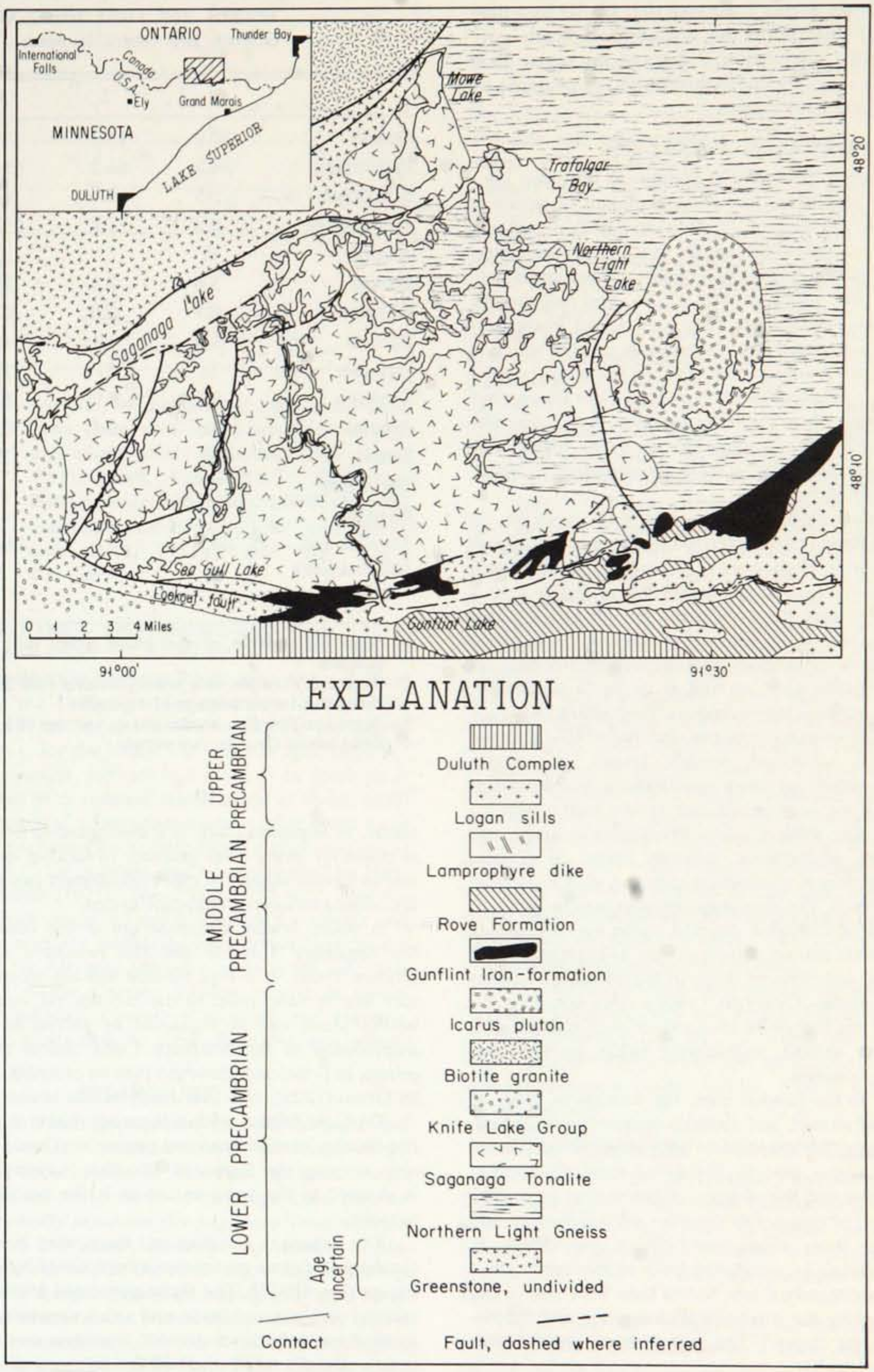


Figure III-32. Generalized geologic map of Saganaga Lake-Northern Light Lake area (modified from Goldich and others, in press, by addition of data on Long Island Lake quadrangle (Morey and others, 1969).

The Northern Light Gneiss and the Saganaga Tonalite were intruded in the Icarus Lake area (fig. III-32) by a pluton composed of syenodiorite and granodiorite, which is informally called the Icarus pluton. Both of the major rock types are alkalic, and are characterized by the presence of aegerine-augite and hornblende.

SAGANAGA TONALITE

The Saganaga batholith is a composite intrusion containing several rock types, the most abundant of which is gray tonalite. Where sheared, the tonalite is pink or red and more felsic. A younger, gray tonalite that has intrusive contacts with the normal Saganaga Tonalite occurs on Horseshoe Island (T. 48° 14′ N., R. 90° 52′ W.) and neighboring areas, but its full extent is not known. A border phase having the general composition of hornblende diorite, which is as much as several hundred feet thick, is present along the northeast margin adjacent to the Northern Light Gneiss, but is generally absent along the other margins. A red, fluorite-bearing granodiorite and pegmatite, on Gold Island, and dikes ranging in composition from lamprophyre to aplite cut the main batholith rocks.

The dominant (normal) phase is a gray, medium- to coarse-grained tonalite characterized by large quartz aggregates that resemble phenocrysts. The quartz "eyes," commonly about 1 cm across, are aggregates of grains 1 to 2 mm in diameter that have different optical orientations. Quartz also occurs as an interstitial mineral. The quartz "eyes" probably are a primary igneous feature, for they occur in small tonalite dikes as well as in the larger bodies, and the quartz contains few inclusions. Plagioclase ($An_{20\text{-}28}$) constitutes about 60 percent of the rock (table III-19); most is non-perthitic, unaltered, weakly zoned, and poorly twinned. Antiperthite increases in abundance as the borders of the batholith are approached, and is very well developed in the more highly sheared parts. Microcline is sparse and occurs as small, antiperthitic anhedral grains, as discrete zones at the borders of plagioclase grains, and as interstitial grains. Hornblende, the dominant ferromagnesian mineral, forms euhedral or subhedral discrete grains that are altered slightly to chlorite and epidote or occurs as aggregates with biotite, epidote, and chlorite. Some of the hornblende grains contain relict augite. Generally, augite is confined to the eastern part of the batholith, in the vicinity of Icarus Lake. Apatite, epidote, sphene, and opaque oxides are the common accessory minerals.

Peripheral to the central core, the tonalite is more or less sheared and altered, and contains more microcline than the normal phase. The increase in microcline is interpreted as having resulted in part by exsolution from plagioclase, aided by shearing and the addition of water, and in part by the introduction of microcline together with carbonate and quartz into the shear zones (see Goldich and others, in press). The shearing is manifested by a reduction in grain size from approximately 4 mm to less than 0.02 mm in the groundmass and by the mechanical elongation and flattening of the quartz "eyes." The "eyes" have length/width ratios of as much as 7:1.

A fine- to medium-grained hornblende-biotite tonalite intrudes the sheared tonalite in the vicinity of Horseshoe Island, in Saganaga Lake. It is distinguished from the normal phase by being finer grained, in lacking quartz "eyes," and in having somewhat more hornblende (see table III-19). It contains numerous mafic inclusions.

Table III-19. Modes, in volume percent, of the Saganaga Tonalite and Gold Island Granite (after Goldich and others, in press).

	1	2	3	4
Quartz	20.9	15.3	2.1	11.4
Plagioclase	64.2	64.2	61.5	54.6
Microcline	1.5	3.4	1.6	20.2
Hornblende	6.5	11.4	25.8	Tr
Augite	Tr			
Biotite	3.6	2.3	1.4	6.0
Chlorite	0.9	0.9	0.9	1.4
Epidote	1.5	1.1	4.5	0.4
Opaques	Tr	Tr	Tr	1.6
Sphene	0.3	0.2	1.1	
Apatite	Tr	0.1	0.5	
Zircon	Tr	Tr	Tr	Tr
Carbonate	Tr			0.8
Fluorite				3.6
An Content (Michel-Levy method)	24	24	26	10

1—Saganaga Tonalite, normal phase, Trout Bay; average of 20 samples
2—Saganaga Tonalite, late phase, Leaning Pine Bay and Horseshoe Island area; average of 5 samples
3—Saganaga Tonalite, border phase; average of 9 samples
4—Gold Island Granite; one sample

A mafic border phase occurs at the contact between the Saganaga Tonalite and the Northern Light Gneiss (Harris, 1968). It is well foliated and has an internal structure that is subparallel to the contact. At most places the border phase can be explained by partial assimilation of amphibolite in the Northern Light Gneiss (Goldich and others, in press), but its origin may be complex, as suggested by Grout (1929), who first described the border rocks.

On Gold Island, within Saganaga Lake, a sheared fluorite-bearing granodiorite and pegmatite (Grout, 1929) probably intrudes the Saganaga Tonalite. Apparently the rock is sheared to the same extent as is the main mass of the tonalite.

The Saganaga Tonalite and the nearby Northern Light Gneiss and Icarus pluton have been dated by several techniques (fig. III-33). The techniques used include the K-Ar method using hornblende and mica separates, the Rb-Sr method using both whole-rock isochrons and mineral separates, and the U-Pb method for zircon and sphene (Goldich and others, 1961; Anderson, 1965, unpub. Ph.D. thesis, Univ. Minn.; Hanson, 1968; Tilton and Grünenfelder,

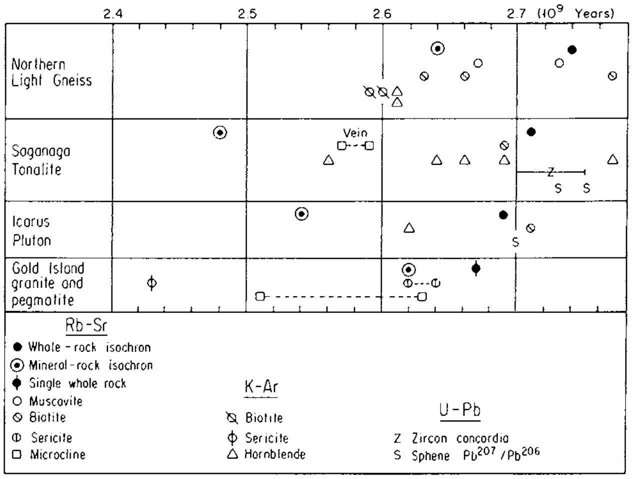


Figure III-33. Summary of radiometric ages in the Saganaga Lake area (compiled from Goldich and others, 1961; Anderson, 1965, unpub. Ph.D. thesis, Univ. Minn.; Tilton and Grünenfelder, 1968; Hart and Davis, 1969; Catanzaro and Hanson, 1971; and Hanson and others, 1971b). Rb^{87} $\lambda\beta = 1.39 \times 10^{-11} yr^{-1}$.

1968; Hart and Davis, 1969; Hanson and others, 1971b; Catanzaro and Hanson, 1971). As can be seen in Figure III-33, the whole-rock isochron Rb-Sr ages ($\lambda\beta = 1.39 \times 10^{-11}$ yr^{-1}), the U-Pb ages for sphene, and the U-Pb ages for zircon are in agreement, suggesting an age of 2,700-2,750 m.y. for the rocks. The mineral ages, however, are generally younger, perhaps as a result of an epeirogenic event associated with regional stabilization at about 2,600 m.y. or as a result of a low-grade metamorphic event since 2,700 m.y. A prehnite-pumpellyite-facies metamorphism at about 1,600 m.y., for example, has been suggested by Hanson and Malhotra (1971). The 2,700-2,750 m.y. ages for Rb-Sr whole-rock isochrons and U-Pb ages for sphene and zircon in this area are similar to ages determined by the same methods for Algoman intrusive rocks to the west, that is, the Giants Range Granite, Linden syenite of Grout (1926), and Vermilion Granite (Catanzaro and Hanson, 1971; Prince and Hanson, in press; Peterman and others, in press).

STRUCTURE

Major faults bound the batholith on both the north and south margins, and lesser ones, some of which are shown on Figure III-34, occur within it. The fault along the north margin follows the north shore of Saganaga Lake (Harris, 1968), and generally separates the Saganaga Tonalite from greenstone; at a few places, however, the tonalite is preserved on the north side of the fault. According to Harris, the breccia zone along the fault, consisting of broken metavolcanic rocks, tonalite, and peridotite cemented by quartz and calcite, is as much as 1,000 feet wide. The fault along the southern margin, previously named the Lookout fault (Sims and others, 1969), trends northwestward and is steeply inclined. A breccia zone consisting of highly sheared metabasalt in a matrix of quartz and calcite occurs along the fault. In the vicinity of Sea Gull Lake (fig. III-32), the fault separates tonalite from greenstone, but to the east (Morey and others, 1969, Minn. Geol. Survey, unpub. open-file map, Long Island Lake quadrangle) the fault transects the greenstone and separates amphibolite on the Saganaga Tonalite side of the fault from greenstone (Weiblen and others, 1971) on the other side. The faults that have been mapped within the batholith are based largely on observed shear zones and to a lesser extent on topographic lineaments (Goldich and others, in press). Strong cataclasis associated with the faults produced a cataclastic foliation and lineation in the tonalite. Along the faults that bound the north and south margins, the cataclastic foliation is steeply inclined and the lineation plunges 15° to 30° E. Within the batholith, the cataclastic foliation parallels the shear zones and generally has a northward strike and gentle dips to the east (fig. III-34). Primary igneous foliation and lineation are best developed in the contact zone, and are well developed along the northeastern margin where the structure is generally conformable to that in the Northern Light Gneiss.

Grout (1936) interpreted the foliation and lineation in the Saganaga Tonalite as primary flow structures, and inferred that the batholith and the Northern Light Gneiss had been tilted about 70° westward subsequent to deposition of the conglomerate at Cache Bay. With this interpretation, the eastern part of the Saganaga batholith would represent the "roots" of the batholith, uplifted from a depth of some 15 miles. Goldich and others (in press), however, interpreted the internal structure in the sheared tonalite as secondary and as having formed mainly by shearing along the several faults. They concluded that the batholith rose approximately vertically, and after crystallization the tonalite and the Northern Light Gneiss continued moving upward as a diapiric mass. Faults formed along the contacts between the diapiric mass and adjacent greenstone, as well as within the batholith. The western part of the batholith, at least, was unroofed, and shed detritus to the west, forming the so-called Ogishke conglomerate. Subsequently, these sediments were folded and faulted (Gruner, 1941), in part perhaps by the continued rise of the diapiric mass.

The time of the faulting is not known and indeed may differ for particular sets. An altered but undeformed lamprophyre dike that cuts the fault along the north shore of Saganaga Lake has a K-Ar age on biotite of about 1,750 m.y. (Hanson and others, 1971b). This is a minimum age for the faulting.

GEOCHEMISTRY AND PETROGENESIS

On the basis of the geochemistry and petrography of the Saganaga Tonalite, Hanson and Goldich (in press) suggested that the magma was derived by partial melting of either eclogite or amphibolite of basaltic composition at mantle depths. The basis for this interpretation is summarized below.

The low Sr^{87}/Sr^{86} initial ratio for the Saganaga Tonalite of $0.7009 \pm .0002$, based on a whole-rock isochron intercept and an apatite analysis (table III-21), eliminates the possibility of a source in pre-existing granitic crustal rock, for this ratio is low even for strontium derived from the mantle 2,700 m.y. ago. The magma, therefore, must have been derived by partial melting of short-lived material recently derived from the mantle, such as basalt or potassium-poor graywacke.

Partial melting of a graywacke as a source for tonalitic magmas is improbable. Generation of a tonalitic magma through partial melting of either wet or dry potassium-poor graywacke would require temperatures in excess of 1,000°

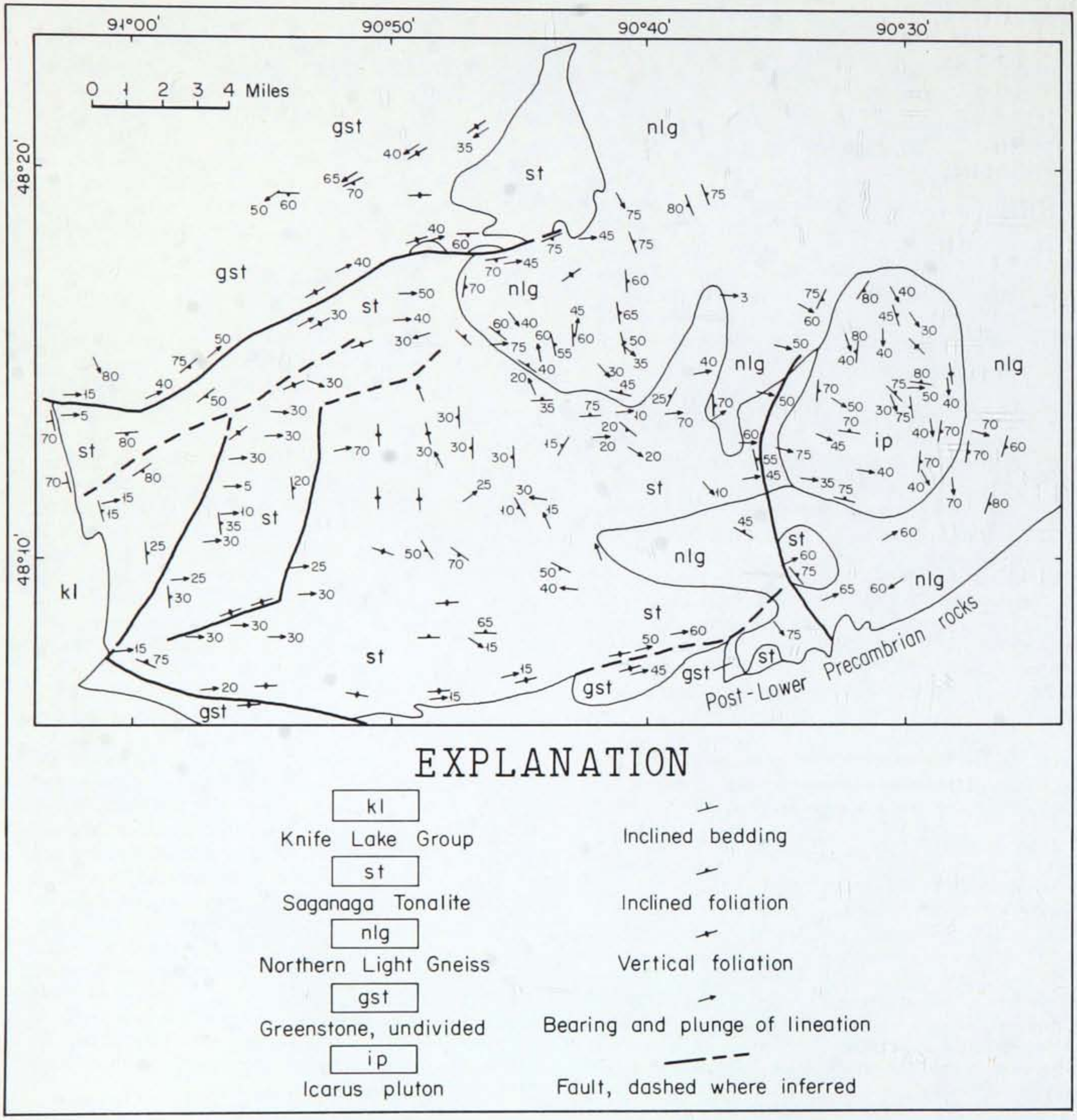


Figure III-34. Structural map of the Saganaga Lake-Northern Light Lake area (from Goldich and others, in press). Geology shown on Figure III-32.

Table III-20. Chemical analyses of the Saganaga Tonalite (after Goldich and others, in press). Analysts: N. H. Suhr, J. B. Bodkins, and S. S. Goldich.

	DH-6	DH-4	SH-23
SiO_2	55.8	63.3	67.0
Al_2O_3	17.7	19.0	17.0
TiO_2	0.46	0.28	0.28
Fe_2O_3	1.91	1.05	1.00
FeO	3.90	1.43	1.35
MnO	0.09	0.03	0.04
MgO	5.02	1.46	1.59
CaO	6.65	5.30	4.02
SrO	0.13	0.15	0.11
Na_2O	5.57	6.23	5.41
K_2O	1.21	0.90	1.24
Rb_2O	n.d.	0.001	0.002
BaO	n.d.	n.d.	0.06
P_2O_5	0.25	0.13	0.15
H_2O+	1.41	0.48	0.44
H_2O-	0.07	0.03	0.06
CO_2	0.10	0.03	0.01
	100.27	99.80	99.76

DH-6—Border phase, S side of Buck Island
DH-4—Quartz-eye tonalite, S end of Trout Bay
SH-23—Quartz-eye tonalite

Table III-21. Comparative data, in ppm, for the Saganaga Tonalite and average Archean basalt (after Hanson and Goldich, in press).

	Archean* Basalt	Saganaga Tonalite
K	2,100	10,000
Rb	5.9	25
Sr	175	900
Ba	70	600
K/Rb	360	400
Sr/Ba	2.5	1.5
K/Sr	12	11
$Sr^{87}/Sr^{86}{}_i$†		0.7009

* From Hart and others (1969)
† The uncertainty in the Sr^{87}/Sr^{86} initial ratio is 0.0002 (one standard deviation)

C (Lambert and Wyllie, 1970), and although the heat flow in Early Precambrian time probably was higher than at present, such temperatures probably were not attained at crustal levels. Also, most graywackes have potassium content in excess of that in the Saganaga Tonalite (see fig. 2, Hanson and Goldich, in press). Partial melting at crustal depths would, of course, enrich the melt in potassium relative to a low-potassium parent on the quartz-plagioclase side of the minimum in the quartz-plagioclase-potassium feldspar ternary system.

Partial melting of a rock of basaltic composition under somewhat wet conditions at pressures greater than 12 kilobars would produce dacite or rhyodacite as a low-temperature melting fraction (Green and Ringwood, 1968). To attain such a pressure, a very thick pile of volcanics is required, with either partial melting of the basal amphibolite or conversion of the basal basalts to quartz eclogite, which would subsequently sink into the mantle and melt at a great depth.

The Saganaga Tonalite is distinguished by a high Na_2O/K_2O ratio (tables III-20 and III-21), a relatively high K/Rb ratio (about 400), and a very low Rb/Sr ratio (0.028). A rare earth analysis of a composite sample by Haskin and others (1968) shows that, relative to other granitic rocks, the tonalite is strongly depleted, with a light rare earth abundance similar to andesite or ocean ridge basalt, and similar to chondrites. From the presently known distribution coefficients (Schnetzler and Philpotts, 1970), the rare earth data would suggest separation of the melt that yielded the Saganaga Tonalite from a residue relatively high in garnet. To explain the relatively low Rb/Sr ratios and high K/Rb ratios, which are similar to Archean basalts (table III-21), the residue from which the magma separated must have had a low hornblende and plagioclase content, for plagioclase would retain strontium as compared to rubidium and both plagioclase and hornblende would retain potassium relative to rubidium, giving a melt with higher Rb/Sr and lower K/Rb ratios relative to the original basalt composition. For the partial melting model of either amphibolite or quartz eclogite, the residue remaining after partial melting probably consisted mainly of garnet and pyroxene. Neither of these minerals retains K, Rb, or Sr, thus enriching the melt in these components in proportion to the ratios in the original basalt. The garnet would deplete the melt of heavy rare earths, and the pyroxene would have little effect on the relative abundances of the rare earth elements. As the magma formed and began to move upward, quartz, which would be a near-liquidus phase at depth, was rafted up with the viscous melt. Upon reaching shallower depths, plagioclase replaced quartz as the liquidus phase, as shown by the resorbed quartz phenocrysts which now occur as "eyes." At the time of emplacement and crystallization, the earlier precipitated clinopyroxene reacted to form hornblende.

VERMILION GRANITE-MIGMATITE MASSIF

D. L. Southwick

"The rock varies from coarse to fine-grained, and from nearly white to red. It is quite likely that within the general area are patches of massive rocks of different Archean dates. . . . The whole region is a forbidding granitic waste, and but little time has been spent on it."

N. H. Winchell, 1899,
The Geology of the Vermilion Lake plate:
Minnesota Geol. Survey Final Report,
vol. 4, p. 541.

The "forbidding granitic waste" so brusquely dismissed by N. H. Winchell is a large complex of Lower Precambrian granitic rocks which Grout (1923, 1925b) later called the Vermilion batholith. It underlies an area in Minnesota about 35 miles wide and 80 miles long and is the largest body of granitic rocks exposed in the state. This huge complex is bounded on the south by older metavolcanic and metasedimentary rocks of the Vermilion district (Clements, 1903; Sims and others, 1968b, 1970) and on the north by a similar belt that passes under Rainy Lake and the Kabetogama Peninsula. It crosses the International boundary east of Lac La Croix and continues northeastward into Ontario for an undetermined distance. A thick mantle of glacial drift obscures its termination on the west (fig. III-35).

Within this large area are at least four distinguishable intrusive rock types—hornblende quartz diorite, hornblende diorite, biotite granodiorite, and biotite granite—that are intimately related to a variety of amphibolitic and biotitic migmatites. The most widespread rock type is a generally

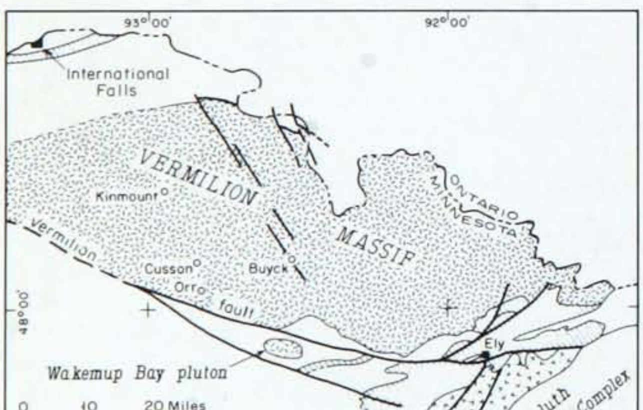


Figure III-35. Map showing regional setting of the Vermilion granite-migmatite massif. Pre-granite rocks north of the Vermilion fault and a subsidiary fault are mafic metavolcanic rocks (diagonal rule) and biotite schist (blank). A part of the Giants Range batholith is shown by check pattern.

massive, light-pink, medium- to coarse-grained biotite granite, which Grout (1923) named the Vermilion Granite. Except for late pegmatites it is the youngest rock in the complex.

Mapping and petrologic work completed to date indicate that this terrane has had a complex history of injection, metamorphism, anatexis, and metasomatism. Most of the granite appears to have formed from a melt; metasomatic processes, however, were important near the margins and within migmatite belts. The so-called Vermilion batholith contains much more migmatite than most batholiths in younger orogenic belts (Buddington, 1959; Hamilton and Myers, 1967), and is generally similar to migmatite-granite complexes in other Precambrian shields (Rankama, 1963; Mehnert, 1968; Anhaeusser and others, 1969). I have chosen to use "granite-migmatite massif" instead of "batholith" as a general term for the complex as a whole.

The rocks belonging to the Vermilion granite-migmatite massif are of Early Precambrian age. They were assigned to the Algoman intrusive episode by Grout (1925b). Goldich and others (1961) reported K-Ar and Rb-Sr ages between 2,300 and 2,500 m.y. for various phases of the complex. Recent radiometric studies by Peterman and Goldich (1970) have yielded a whole-rock Rb-Sr isochron for the Vermilion massif of $2{,}680 \pm 95$ m.y., with an initial Sr^{87}/Sr^{86} ratio of 0.7005 ± 0.0012.

This report is neither final nor complete. About 65 percent of the Vermilion massif has been mapped in semi-reconnaissance fashion as of this writing (January, 1971), and laboratory work has progressed only to the stage of routine petrography. Conclusions here expressed must be regarded as tentative, subject to modification as work on these rocks continues.

ROCK NOMENCLATURE

Perhaps no single rock type has been defined in more ways than granite. This unfortunate circumstance has been lamented by many geologists (see, for example, Marmo, 1967), and is ably discussed by Streckeisen (1967, p. 146-147; 166-167). Streckeisen's classification of the nonfeldspathoidal plutonic rocks (1967, p. 160), is a realistic approach to the problem of defining granite, and is reproduced in Figure III-36; field 3 contains the granites. Many other classification systems subdivide Streckeisen's field 3 into two parts (subfields 3a and 3b of figure III-36). In American literature, the terms "adamellite" or "quartz monzonite" have been favored for subfield 3b, whereas the term "granite" has been restricted to subfield 3a (see, for example Bateman and others, 1962). Studies of compilations of modal analyses show, however, that the commonest "granitic" rocks actually fall into subfield 3b (Chayes, 1957; Streckeisen, 1967, p. 221-227). This follows from the

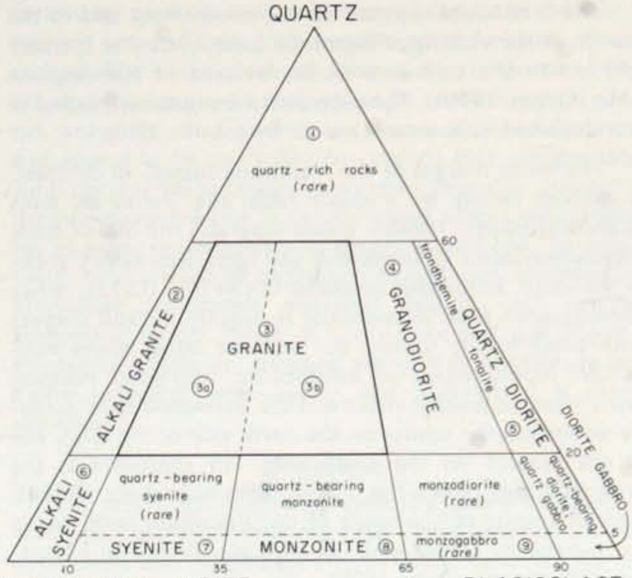


Figure III-36. Streckeisen's (1967) classification of non-feldspathoidal plutonic rocks (upper case names are the most common rock type. Fields based on modal mineral compositions.)

fact that subfield 3b contains the minimum-melting compositions in the chemical system quartz-albite-orthoclase-H_2O (Tuttle and Bowen, 1958). It seems reasonable, therefore, to use the term "granite" for the commonest of the granitic rocks, namely for both subfields 3a and 3b. This usage is widespread in Europe and is consistent with the clear definition framed by Mehnert (1968, p. 354): "Granite (is a) phanerocrystalline, massive rock consisting of quartz, potash feldspar, and sodic plagioclase (typically oligoclase) in nearly equal amounts, and a generally small amount (5-10%) of mafic minerals (biotite, hornblende, and others)."

Using this definition, the commonest rock in the Vermilion massif is granite. If other classification schemes are used, most of the Vermilion massif becomes adamellite or quartz monzonite. Because these rocks have been called granite for many years (A. N. Winchell *in* Winchell and others, 1899; 1900; Grout, 1923, 1925b, 1926; Goldich and others, 1961), it seems advisable to continue that usage.

Almost as troublesome as granite is the term "migmatite." The history of the term is discussed by Mehnert (1968, p. 1-8), and his definition is adopted here (1968, p. 355): "Migmatite (is a) megascopically composite rock consisting of two or more petrographically different parts. One is the country rock in a more or less metamorphic stage, the other is of pegmatitic, aplitic, granitic, or generally plutonic appearance."

REGIONAL SETTING

The Vermilion massif is composed largely of granitic rocks that are younger than the flanking belts of greenstone and schist. Except along part of the south side of the massif, the contacts are broadly gradational across marginal migmatitic zones from 1 to 4 miles wide. The north contact zone is very well exposed around Kabetogama Lake and in the channel south of Kettle Falls (fig. III-37). Biotite schist

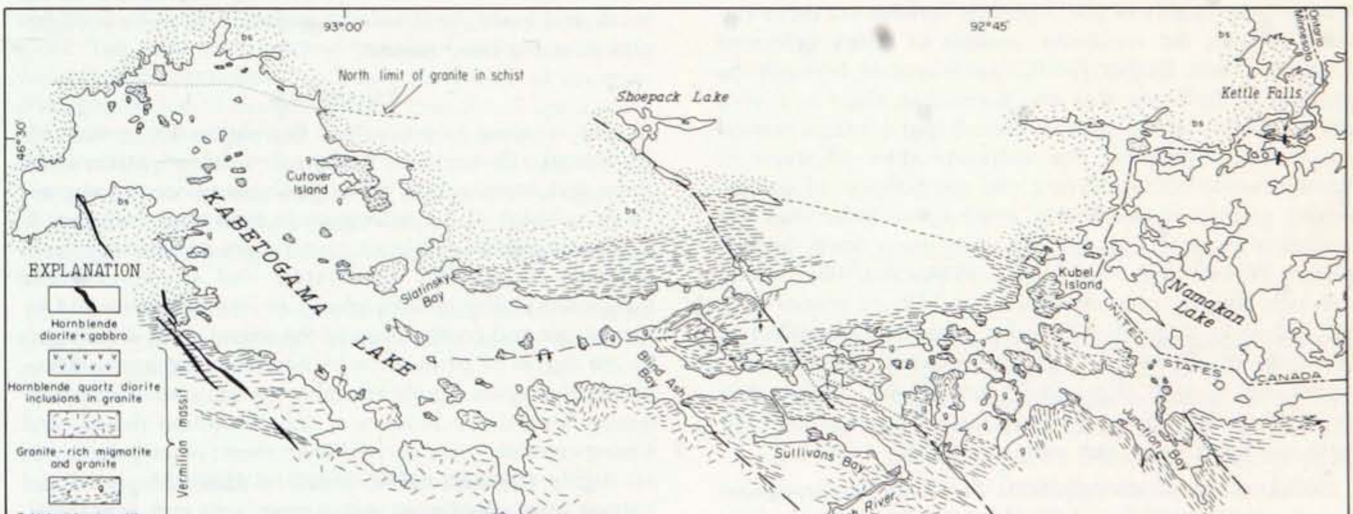


Figure III-37. Geologic map of part of the north contact zone of the Vermilion massif. (The line pattern in the migmatite schematically portrays the structural pattern and approximate ratio of paleosome to neosome.) Only the larger granitic bodies are shown in the schist north of Slatinsky Bay.

and metagraywacke of Knife Lake type (Morey and others, 1970) crop out over much of the Kabetogama Peninsula and along part of the north shore of Kabetogama Lake. These rocks typically have good relict bedding that strikes generally east and dips north, and also have one or more secondary foliations. They are composed principally of quartz, sodic plagioclase, and biotite, and contain local pale pink garnet. At roughly the latitude of Cutover Island in Kabetogama Lake, sillimanite becomes common in the more aluminous beds and conformable bodies of almost white granite and pegmatite begin to appear. These sill-like granite bodies range in thickness from a few centimeters to tens or hundreds of meters. Commonly the outermost margins of the granite bodies have a "frayed" appearance owing to the delicate interlayering of millimeter-thick biotite-rich septa with nearly as thin but coarser grained layers of quartz and feldspar.

South of Cutover Island the granite becomes more abundant, but schist remains the dominant rock. The largest, most homogeneous masses of granite increase in width and length; smaller sheets and stringers as thin as millimeter-scale permeate the schistose wall rocks. Approximately at the latitude of Slatinsky Bay, the proportions of granite and schist are roughly equal and much of the schist contains so many quartzofeldspathic stringers that the combined rock is gneissic. This terrane is characterized by an interlayering of granitic and schistose rocks on all scales, to form a gross stromataform migmatite.

Throughout the outer part of the contact zone most of the granitic layers are essentially conformable to inherent structures in the older schists and gneisses, and the structural pattern of the area is dominated by east-west trends. Further south, about the latitude of Blind Ash Bay and Sullivans Bay, this structural continuity is disrupted as the granitic fraction becomes transgressive. The rocks in this area consist roughly of 60-75 percent massive but inclusion-laden granite; the remainder consists of highly deformed migmatite. Still further south, approximately between the latitudes of Sullivans Bay and Kinmount, there is a wide belt of fairly uniform pinkish granite that contains numerous biotitic inclusions. The inclusions show all stages of transformation to the texture and composition of the enclosing granite; normally they make up no more than five percent of the total rock but in certain linear zones the proportion of inclusions is as high as 50 percent. These inclusion-rich zones commonly grade into belts of granite-poor but definitely migmatized biotite schist and amphibolite. The belts of schist-rich migmatite exposed between Kinmount and Cusson (figs. III-35 and III-40) contain only about 10 to 25 percent granitic material and grade both north and south into much more granite-rich terrane.

Belts of schist-rich migmatite are common throughout the western part of the Vermilion massif. Relatively little migmatite occurs east of a line through Little Vermilion, Echo, and Picket Lakes (see fig. III-40) except within a mile or two of the contacts. The large area of the massif lying south of Lac La Croix is almost wholly a monotonous light grayish-pink granite whose uniformity is interrupted only by scattered zones of nebulitic inclusions.

The gradational contact at the easternmost end of the massif, in the vicinity of Basswood Lake, is similar in many respects to the contact zone in the area of Kabetogama Lake (Green, 1970a). The older rocks are metamorphosed to the amphibolite facies and are cut by granitic dikes.

The south margin of the Vermilion massif, in contrast, is marked locally by a major fault, the Vermilion fault (Sims and others, 1968b), which separates the massif from greenschist-facies metavolcanic and metasedimentary rocks to the south. In the area southeast of Orr (fig. III-35), rather homogeneous pink leucogranite is directly in fault contact with older biotite schists, but at most other places migmatitic biotite schist or amphibolite intervenes between fairly massive granitic rocks and the Vermilion fault. Granite permeates the schists on the north side of the fault, but is not present on the south side. An exception is the Wakemup Bay pluton (fig. III-35; Sims and others, 1968b), a small granite pluton south of the Vermilion fault that is surrounded by biotite schist of the Lake Vermilion Formation (Morey and others, 1970).

THE MIGMATITIC ROCKS

In discussing migmatitic rocks it is convenient to adopt the terminology of Dietrich and Mehnert (1961). They referred to the "older part of a composite rock (*i.e.*, the remaining or pre-existing part)" as the paleosome, and to the "younger part of a composite rock (for example, the injected, exuded, or metasomatically introduced material)" as the neosome. In the Vermilion massif, the various biotite schists, gneisses, and amphibolites that form sheets, slabs, or isolated inclusions within granitic material collectively constitute the paleosome, whereas the granitic material itself constitutes the neosome. On a regional scale the paleosome is mainly biotite schist in northern parts of the massif and biotite schist interlayered with amphibolite in the south and east; the dominant neosome is white to light-pink coarse-grained granite.

Paleosomes

For mapping purposes those migmatites that contain 25-75 percent paleosome are called schist-rich migmatites, and those that contain 5-25 percent paleosome are called granite-rich. Much of the paleosome in both types is a gray to dark gray biotite-rich rock that ranges in structure from schistose to gneissic. The texture and structure of the paleosome in any specific place are in part determined by the texture and composition of the parent rock, and in part by the degree of permeation by neosomatic material. Thus, some paleosomes closely resemble the schist and metagraywacke beyond the borders of the Vermilion massif, and almost certainly were derived from them; other paleosomes are highly replaced and/or veined by granitic material and are not so readily linked with known wall rock lithologies.

Characteristically the paleosomes are well foliated and lineated. The chief exceptions to this are the essentially massive nebulitic inclusions in granite-rich migmatite whose compositions approach that of the granite that contains them. Texturally, the paleosomes are granoblastic and medium to coarse grained. Those that are extensively permeated by neosome tend to be somewhat coarser than those that are not, and commonly contain porphyroblasts of microcline. The commonest paleosomes are composed of biotite, quartz, and sodic plagioclase in variable ratios. Garnet, muscovite, and chlorite are abundant in certain layers and appear to be controlled by original bulk composition. Amphibolitic rocks comprise no more than five percent of the paleosome fraction in the northwest part of the complex but increase in abundance to the south and east. Typically these rocks consist of blue-green hornblende, intermediate plagioclase (commonly andesine), epidote, and variable amounts of quartz. North of Crane Lake and near Buyck a paleosome containing subequal amounts of hornblende and biotite, together with quartz and plagioclase, is fairly abundant.

Neosomes

The schist-rich migmatites typically have a neosome of white to light-pink coarse-grained leucogranite. Quartz, microcline, and plagioclase (fig. III-38A; table III-22) are the major minerals together with traces of biotite. The texture is generally coarse allotriomorphic-granular, but varies widely. Vague zones and clots of very coarse quartz and feldspar commonly occur within somewhat finer grained rock, and subhedral to anhedral megacrysts of microcline are abundant locally.

The contacts between neosome and paleosome commonly are wavy and are marked by zones a few mm thick that are anomalously rich in biotite. In places where paleosome and neosome are intimately intermingled on the scale of a few mm or cm, the whole mass has a braided appearance with mafic and feldspathic wisps intertwined.

Perthitic microcline is characteristic of the neosome. Some very thin neosome stringers contain microcline crystals that are larger than the width of the stringer; these crystals penetrate the paleosome on both sides without disturbing its foliation, indicating that they grew by replacement. The ratio of microcline to quartz and plagioclase is variable in stringers less than 2 or 3 cm thick, but the mineral composition of stringers thicker than 2 or 3 cm is quite constant and averages near the minimum melting composition of the "granite" system (Tuttle and Bowen, 1958). The compositional inhomogeneity and general coarseness of the thinnest neosome stringers argue in favor of a replacement origin rather than a magmatic *lit-par-lit* origin for them.

The neosome of the granite-rich migmatite is indistinguishable from the major pink granite, and is described in a later section.

MASSIVE GRANITIC ROCKS OLDER THAN THE MAJOR PINK GRANITE

Away from the contact zones and large areas of schist-rich migmatite, the Vermilion massif is predominantly igneous-looking, massive, light grayish-pink granite. However, there are relatively minor quantities of coarse intrusive rocks that are older than the major pink granite, and these are discussed in this section.

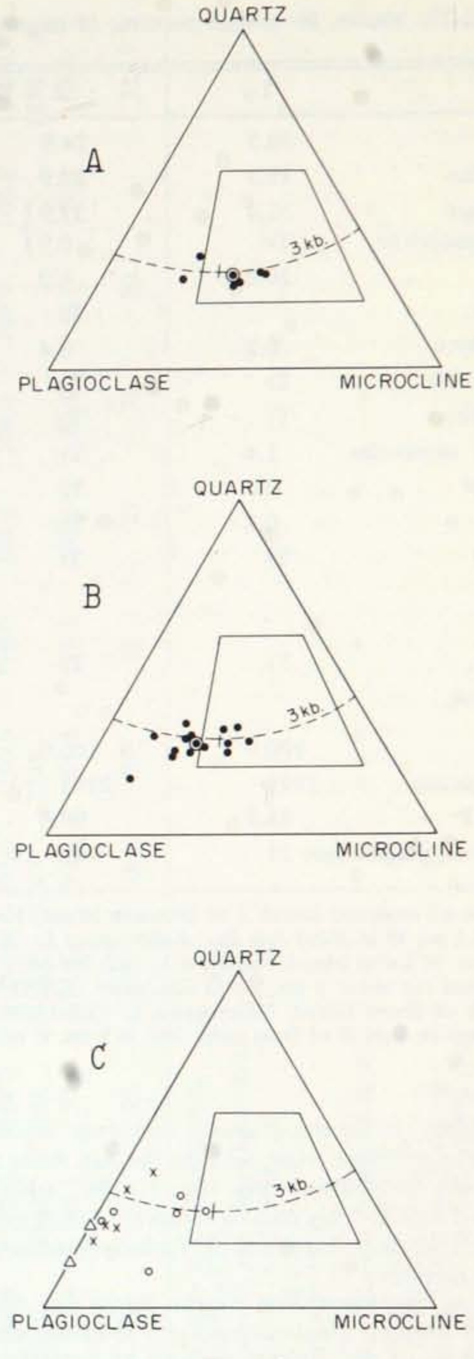


Figure III-38. Modal quartz-microcline-plagioclase compositions of granitic rocks in the Vermilion massif. The diagrams are based on the data in tables III-22, III-23, and III-24. The dashed line is the normative quartz-feldspar boundary in the system Q-Ab-Or-H_2O at 3 kb. P_{H_2O} (Tuttle and Bowen, 1958, p. 56). The isobaric minimum melting composition is shown with a short cross-line. The trapezoidal area outlines the granite field of Streckeisen's (1967) rock classification. A, migmatite neosomes; B, major pink granite; C, early plutonic phases.

Table III-22. Modes, in volume percent, of migmatite neosomes in the Vermilion massif.

	1	2	3	4	5	6
Quartz	30.3	24.9	27.7	27.0	27.4	26.3
Microcline	19.3	34.9	38.0	20.5	39.5	31.4
Plagioclase	38.4	37.9	30.5	51.1	27.9	40.4
Sec. muscovite	Tr	0.9	Tr	Tr	Tr	Tr
Biotite	10.2	1.0	1.2	1.0	1.8	0.6
Chlorite		Tr	Tr	Tr	Tr	Tr
Granophyre	0.2	0.4	Tr	Tr	Tr	Tr
Microgranite	Tr	Tr	Tr	Tr	Tr	0.2
Myrmekite	Tr	Tr	Tr	Tr	Tr	Tr
Primary muscovite	1.4	Tr	2.2	0.4	3.4	1.1
Opaques	0.1	Tr	Tr		Tr	
Apatite	0.1	Tr	Tr	Tr	Tr	
Zircon	Tr	Tr	Tr	Tr	Tr	Tr
Sphene			0.1			Tr
Rutile					Tr	
Epidote	Tr	Tr	0.3			
Carbonate					Tr	Tr
Total	100.0	100.0	100.0	100.0	100.0	100.0
No. of points	1999	2195	2205	2262	2311	2525
Q+K+P	88.0	98.6	96.2	98.6	94.8	98.1
An content, plagioclase	21	22	24	22	31	21

1: W end of unnamed island S of Bowman Island, Kabetogama L. (KS-29-161a)
2: Bay 0.3 mi. W of Blind Ash Bay, Kabetogama L. (KS-44-384a)
3: SW cor. of Kubel Island, Namakan L. (KS-74b-602)
4: Railroad cut about 2 mi. SE of Kinmount (KS-20-102)
5: E side of Sheep Island, Kabetogama L. (KS-33-200)
6: Outcrop in field S of State hwy. 217, 0.5 mi. E of Galvin Road (KS-18-79)

Medium- to coarse-grained, dark-gray biotite-hornblende quartz diorite crops out near the Ash River about 2 miles south of Sullivans Bay (fig. III-38C; table III-23, columns 1 and 2). This rock is massive, of uniform mineralogic composition, and has a typical hypidiomorphic-granular texture.

Cutting the hornblende quartz diorite are numerous dikes and irregular masses of light-gray to almost white biotite granodiorite (fig. III-38C; table III-22, columns 3 to 8). This rock ranges in composition from granite to quartz diorite but is mostly granodiorite; texturally, some bodies are medium grained and uniform but others vary widely and rapidly from medium grained to pegmatitic. Petrographically the granodiorite is characterized by non-perthitic or weakly perthitic microcline, very dark greenish-brown biotite, and widespread traces of primary muscovite.

Massive, uniform gray granodiorite petrographically similar to that near Ash River is extensively exposed in the complex of bays and islands at the west end of Namakan Lake, and also northwest of Ely, near the south edge of the massif. At all localities the hornblende quartz diorite and biotite granodiorite are cut by numerous dikes of medium- to very coarse-grained light-pink biotite granite which is petrographically identical to the major Vermilion Granite.

Gray granodiorite forms dikes, stringers, and irregular masses that are intimately involved with paleosomatic schists and amphibolites in the migmatite belt that extends from near Cusson eastward to Buyck (see fig. III-40). In this area, many thin granodiorite stringers are concordant to layering in the paleosome and were folded along with it, but crosscutting relations are common as well. Both granodiorite and paleosome are invaded on all scales by the ubiquitous pink granite.

In the Buyck area, there is evidence for two periods of migmatite formation. The gray biotite granodiorite locally cuts an earlier pegmatoid neosome that has been ptygmatically folded and boudinaged along with the paleosome. The granodiorite itself is cut by the major pink biotite granite which clearly grades into the neosomes of migmatites in the north contact zone of the massif. It appears, then, that some migmatite neosomes are older than the gray granodiorite, and some are younger.

Around Burntside Lake near Ely (fig. III-35) are exposures of a light-colored granitoid gneiss of trondhjemitic

Table III-23. Modes, in volume percent, of massive plutonic rocks older than the major pink granite.

	Dark hbl qtz diorite		Gray granodiorite						Burntside Granite Gneiss of Grout (1926)				Layered hbl diorite
	1	2	3	4	5	6	7	8	9	10	11	12	13	14
Quartz	8.5	16.4	26.4	9.8	28.0	31.2	26.7	24.1	21.8	25.6	34.4	41.4	23.9	0.1
Microcline			3.1	18.5	26.9	16.6	18.6	1.5	2.0	1.9	3.3	7.1	5.6	2.5
Plagioclase	53.7	28.9	58.4	57.8	40.3	43.3	47.5	65.9	72.3	69.9	58.7	50.5	66.9	37.0
Sec. muscovite	2.4	21.8	2.7	Tr	Tr	Tr	Tr	Tr	Tr					23.4
Hornblende	25.4	21.6							0.7					29.1
Biotite	8.1	7.5	8.9	9.0	4.6	6.6	6.0	8.1	3.5	1.7	2.3	Tr	Tr	Tr
Chlorite					Tr	0.1	Tr	Tr		0.9	0.6	0.8	2.9	3.7
Granophyre				Tr	Tr	Tr								
Microgranite			Tr	Tr	Tr	Tr	0.4							
Myrmekite				Tr	Tr	Tr	Tr							
Primary muscovite				4.8	0.2	1.7	0.8	0.1						
Opaques	1.4	1.6	0.2	Tr	Tr	Tr		Tr	0.7	Tr	0.5	0.2	0.3	1.8
Apatite	0.2	0.6	0.2	Tr	Tr	Tr	Tr	0.3	Tr	Tr	0.1	Tr	0.1	0.8
Zircon	0.1	Tr	Tr	Tr	Tr	Tr	Tr	Tr	Tr		Tr			0.1
Sphene	0.1	1.2	0.1			0.1				Tr	0.1		0.1	0.8
Rutile	0.1										Tr			Tr
Epidote		0.1	Tr		Tr	0.1			Tr	Tr	Tr		0.2	0.7
Carbonate				0.1				Tr						Tr
Total	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0
No. of points	1939	1823	1790	2148	2211	2133	2240	2295						1959
Q+K+P	64.6	67.1	90.6	86.1	95.2	91.4	92.8	91.5	96.1	97.4	96.4	99.0	96.4	63.0
An content, plagioclase	34	44	23	22	23	28	23	28	25	23	23	20		31

Analyses 9-13 by M. G. Mudrey, Jr.; remainder by D. L. Southwick.
1: Ash River Trail, SW¼ sec. 10, 68N/20W (KS-3-3)
2: Large Knob S of jct. of Gannon Creek with Ash River (KS-49-419)
3: Large Knob in S part of sec. 9, 68N/19W (KS-52-441a)
4: W side of Williams Island, Namakan L. (KS-72-581)
5: W side of Moose Bay, Namakan L. (KS-53-452a)
6: N side of Namakan Island, Namakan L. (KS-73-595)
7: Outcrop near Ash River just N of jct. of Gannon Creek (KS-48-418)
8: Railroad cut about 1 mi. SE of Kinmount (KS-21-103)
9: E end of Burntside L., SW¼ sec. 8, 63N/13W (Ely 48A)
10: S shoreline of Burntside L., SE¼ sec. 13, 63N/13W (ENW 146)
11: N shoreline of Burntside L., S¼ cor., sec. 8, 63N/13W (ENW 170)
12: Echo Trail N of Burntside L., cen. sec. 8, 63N/12W (ENW 728)
13: Echo Trail N of Burntside L., NE¼ sec. 1, 63N/12W (ENW 757)
14: Knob in north-central part of sec. 32, 65N/17W (CL-91-1299)

composition (tables III-23 and III-25; fig. III-38) which Grout (1926) named the Burntside Granite Gneiss. This gneiss is older than the main pink granite of the Vermilion massif, and may represent an earlier plutonic episode. The Burntside Granite Gneiss is discussed in detail elsewhere in this chapter.

Within a few miles of the south contact of the massif are a few isolated bodies of a peculiar hornblende diorite composed of dark-green hornblende, extensively sericitized plagioclase, and minor microcline, with abundant accessory sphene and magnetite (table III-23, column 14). The feldspar crystals range in size from 2 to 8 mm, are nearly equant, and are subhedral. They are size-graded within layers 2-3 feet thick; the larger crystals consistently occur on the same side of each layer. In one outcrop in the north-central part of section 32, T. 65 N., R. 17 W., the diorite layering is parallel to bedding in adjacent amphibolite and biotite schist; all three rocks are cut by veins of the ubiquitous light pink granite. It is possible, therefore, that the diorite may have occurred as gravity-differentiated sills in the older supracrustal rocks. Faintly layered rocks of biotite-bearing hornblende diorite composition occur as large inclusions within pink granite north of Picket Lake. Although their general aspect is igneous, these dioritic rocks could be interpreted equally well as highly recrystallized volcanogenic paragneisses.

The relationship between these layered diorites and the layered shonkinite described by Grout (1925b) northeast of Ely is uncertain.

MAJOR PINK GRANITE AND THE NEOSOME OF GRANITE-RICH MIGMATITE

Coarse-grained light-pink biotite granite (Vermilion Granite of Grout, 1923) is the major rock type of the interior of the Vermilion massif. It is extremely uniform in the eastern half of the massif, where inclusions and even pegmatitic phases are rare, but is variable in both texture and composition in the western half, where it permeates belts of schist-rich migmatite and is the major constituent of granite-rich migmatite (fig. III-40). The most common pink granite is equant and hypidiomorphic-granular, with crystals in the size range of 5 to 10 mm. Weakly porphyritic phases containing microcline megacrysts only slightly coarser than the enclosing groundmass also are fairly common. The granite contains subequal amounts of quartz, plagioclase (An_{20}-An_{30}), and perthitic microcline, and has a small percentage of red-brown biotite (table III-24; fig. III-38B). Low-grade metamorphism has affected it locally to differing degrees, resulting in variable sericitization of the plagioclase and chloritization of the biotite.

The granite-rich migmatite in the western part of the massif consists of 75-95 percent pink granite and 5-25 percent paleosomatic inclusions. As in most other granitic terranes, the amphibolitic inclusions are more resistant to assimilation than the pelitic ones, and retain their identity in places where most of the biotite schist inclusions have been converted essentially to gray biotite granite. Many outcrops in this area, though dominantly granite, are blotchy or marble-like because of the numerous highly digested, nebulitic inclusions they contain.

Irregular patches and indistinct dikes of pegmatite occur in almost every large outcrop in the western part of the massif. These range in width from a few cm to tens of meters; normally the texture at the contacts grades from granitoid to pegmatitic over a few cm or tens of cm. Sharp-walled, straight pegmatite dikes also are common. These appear to be late-stage fracture fillings, whereas the irregular, gradationally bounded pegmatite masses are of more obscure origin.

Northwest of Echo Lake, some of the pegmatites contain anomalously large amounts of coarse-grained primary magnetite. These pegmatites were explored as possible iron deposits in the early 1920's, but were found to be too small and variable for economic exploitation. They have been described in detail by Grout (1926).

A light-colored granite gneiss is exposed along the south shore of Namakan Lake, roughly from the head of Hammer Bay westward to the vicinity of Long Slough (see topographic map of Hale Bay 7.5-minute quadrangle for localities). Grout (1925b, p. 472-473) described this gneiss, and interpreted it as border phase in the north contact zone of the main Vermilion Granite. Although Grout did not specify the origin of the "border phase," his description suggests that it formed while partly crystallized granite magma flowed upward and was dragged along wall-rock contacts.

There can be no doubt that this gneiss is genetically related to the major pink granite, for as Grout noted (1925b, p. 475), it grades southward into structurally isotropic granite on the west side of the deep bay in sec. 36, T. 69 N., R. 18 W. Most of the gneiss is uniform and has a composition that approximates granite. Here and there, however, there are distinct layers a few tens of cm thick that are more biotitic than the rest of the gneiss and are similar in form, scale, and attitude to metagraywacke beds in wall rock schists that occur only a short distance to the northeast. Furthermore, this gneiss grades westward along strike into typical stromataform migmatite that contains a paleosome of biotite schist.

The microtexture of the granitic gneiss on the south shore of Namakan Lake differs significantly from that of typical granite. Oriented flakes of intense deep-brown biotite (or its alteration products) are more or less free-floating in a medium-grained, dimensionally isotropic groundmass of quartz and feldspar. The texture of the quartz and feldspar is allotriomorphic-granular and rather equant; the mineral grains interlock in a sutured mosaic that could readily be interpreted as metamorphic in origin. Large poikilitic grains of perthitic microcline contain ragged inclusions of biotite and plagioclase and appear to have grown by replacement. Some biotite flakes are slightly bent and some quartz grains exhibit wavy extinction, but in general there is no evidence of penetrative cataclasis.

The southward gradation of the gneiss into normal granite is accompanied by changes in the microtextures. The interlocking mosaic of quartz and feldspar gives way to more typically "igneous" hypidiomorphic-granular texture, and the preferred orientation of biotite disappears.

Table III-24. Modes, in volume percent, of major pink biotite granite.

	1	2	3	4	5	6	7	8	9	10	11	12	13	14
Quartz	26.9	31.7	27.3	26.0	30.3	29.2	32.2	27.2	17.7	24.2	28.0	24.3	22.5	23.6
Microcline	31.4	31.5	35.9	24.4	28.5	12.1	18.0	20.3	7.4	22.1	21.2	18.0	18.5	31.2
Plagioclase	36.8	34.0	32.1	43.4	38.7	57.7	47.4	45.0	69.6	47.4	47.3	51.9	49.1	37.7
Sec. muscovite	1.6		0.5	Tr	Tr	Tr	Tr	Tr	Tr	Tr	Tr	Tr	Tr	Tr
Biotite	2.7	2.3	3.9	4.7	1.1	0.3		3.4	4.7	5.9	1.9	1.0	8.3	5.8
Chlorite	0.4		Tr			0.2	Tr	0.7				1.9		
Granophyre	Tr		Tr	0.2										
Microgranite	Tr	Tr	Tr	Tr										
Myrmekite	Tr	0.4	Tr	Tr										
Primary muscovite		Tr	0.2	Tr	Tr	Tr	0.9	0.5	Tr	1.1	0.4	1.8	1.1	1.6
Opaques	0.2	0.1	0.1	1.0	1.2	0.4	Tr	0.6	Tr	Tr	0.6	1.0	0.1	0.1
Apatite	Tr	Tr	Tr	0.3	Tr	Tr		0.6	Tr	Tr	0.4	Tr	0.2	Tr
Zircon	Tr		Tr	Tr	Tr	Tr		0.4	Tr	0.1	Tr	Tr	Tr	
Sphene								1.3	Tr	Tr	Tr	Tr	0.2	
Epidote		Tr		Tr	0.2	0.1	1.5	Tr	0.6	Tr	0.2	0.1	Tr	Tr
Carbonate								Tr					Tr	
Total (Vol. %)	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0
No. of points	2002	1976	2018	2043	946	1171	748			1385				
Q+K+P	96.7	97.2	95.8	93.8	97.5	99.0	97.6	92.5	94.7	93.7	96.5	94.2	90.1	92.5
An content, plagioclase	21	22	22	22	16	16	12	15	24	20	17	12	15	18

(Columns 1-4: perfectly massive, unsheared granites from Kabetogama-Ash River area; columns 5-14: somewhat sheared granites in and near Vermilion fault zone at south edge of massif. Analysis 7 is of strongly cataclasized granite from the fault zone. Analyses 1-4 by D. L. Southwick; remainder by M. G. Mudrey, Jr.)

1. Otc. 500 ft. N of Ash River Trail, S½sec. 6, 68N/19W (KS-14-56)
2. Roadcut on U.S. Hwy. 53, 1.5 mi. W of St. Louis Co. Rd. 122 (KS-13a-50)
3. Otc. in swamp near Daley Bay, Kabetogama L., SW¼sec. 5, 68N/20W (KS-35-228)
4. Large otc. on SE side of Ash River, NE¼sec. 5, 68N/19W (KS-47-417)
5. S end of North Arm of Burntside L., SW¼sec. 10, 63N/13W (ENW 12)
6. N shoreline of Burntside L., SW¼ sec. 12, 63N/13W (ENW 166B)
7. E end of Burntside L., SW part sec. 8, 63N/12W (ENW 582 B)
8. Fenske L. along Echo Trail, cen. sec. 30, 64N/12W (ENW 646A)
9. N of Burntside L., NW¼sec. 7, 64N/12W (ENW 925)
10. Two mi. W of Crab L., SE¼sec. 16, 63N/14W (CrL 38 B)
11. One mi. W of Crab L., NW¼sec. 14, 63N/14W (CrL 57)
12. N of L. Vermilion, SE part sec. 26, 63N/15W (CL 39A)
13. Four mi. W of Crab L., NW¼sec. 21, 63N/14W (CL 59-1)
14. W side of Trout Bay, L. Vermilion, NW¼sec. 26, 64N/16W (SPI-43)

The possible relict bedding and the textural evidence favoring solid-state rather than magmatic crystallization of the feldspars may indicate that the gneiss on the south shore of Namakan Lake formed by replacement of wall rock near the margin of the Vermilion massif. There is no compelling evidence that it formed in response to magmatic flowage, and virtually no reason to invoke post-crystallization shearing. If indeed this gneiss formed by replacement, it is not necessary to conclude that the entire volume of the Vermilion massif also did. Sharp-walled dikes of pink granite in earlier quartz diorites and granodiorites are hard to explain as other than magmatic features, and the great compositional uniformity of the granite throughout the eastern half of the massif reflects a degree of petrologic homogenization that is hard to explain without recourse to a magma.

SUMMARY OF IGNEOUS EVENTS

The sequence of igneous events in the Vermilion massif is summarized in Figure III-39. The sequence along the right-hand path of the diagram is clearly magmatic, with the successively younger members sharply cutting across the older. The left-hand path, involving the formation of migmatites, is not as certainly magmatic. Textural and compositional evidence has already been given to support a replacement origin for some of the thinnest neosome stringers. The intimate interlayering of schistose and granitic components in some of the migmatites seems mechanically incompatible with a mechanism of magmatic injection, and perhaps is better explained by selective replacement, metamorphic differentiation, or incomplete anatexis.

CHEMICAL COMPOSITION

Chemical data on rocks in the massif are sparse and not well suited to petrogenetic interpretation. The only chemical analyses available were first published in the mid-1920's by F. F. Grout (1925b, 1926). These analyses, repeated here in Table III-25, are of samples chosen to represent typical phases of the massif as interpreted by Grout, whose views differ from my interpretations in several respects. Analyses 1 and 2 of Table III-25 are of the major pink granite (as used in this paper) or the Vermilion Granite of Grout (1923). Analysis 3 is referred to by Grout (1925b) as a "biotitic phase of granite"; probably it represents highly digested schlieren in granite-rich migmatite. Analysis 4 is of a magnetite-rich granite pegmatite that sharply cuts major pink granite; analysis 5 is of typical Burntside Granite Gneiss (as used by Grout, 1925b).

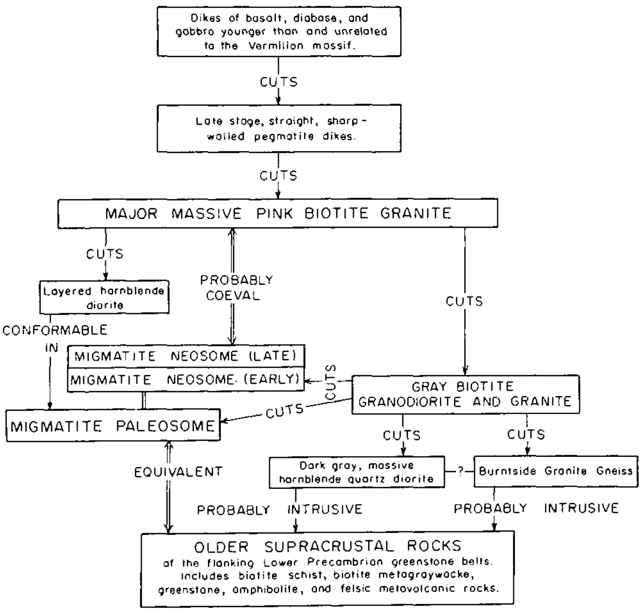


Figure III-39. Sequence of igneous events, Vermilion massif.

Table III-25. Chemical analyses of rocks from the Vermilion massif (after F. F. Grout, 1925b, 1926).

	(1)	(2)	(3)	(4)	(5)
SiO_2	72.06	71.73	69.01	58.72	68.54
Al_2O_3	16.00	14.76	16.92	10.81	17.89
Fe_2O_3	0.46	0.58	0.55	12.84	1.77
FeO	0.72	1.35	1.43	6.80	0.52
MgO	0.97	0.62	1.28	0.94	1.22
CaO	0.86	1.18	2.67	0.82	4.02
Na_2O	4.56	3.58	4.59	3.59	5.14
K_2O	3.54	4.63	2.55	4.89	1.05
H_2O+	0.39	0.64	0.41	0.34	0.46
H_2O-	0.05	0.20	Tr	0.16	0.18
TiO_2	0.12	0.53	0.23	0.42	0.20
P_2O_5	0.09	0.14	0.11	0.20	Tr
MnO	0.06	0.03	0.03	0.08	
ZrO_2	0.03	0.06	0.04	0.02	0.00
CO_2	0.10				
S			0.04		0.06
FeS_2	0.09	0.06		Tr	
Cr_2O_3	0.02	0.02	0.01		0.00
BaO	0.12	0.14	0.03	0.07	
SrO					
TOTAL	100.24	100.25	99.90	100.70	101.05

(1) Vermilion Granite, SE of Pelican L., 64N/19W, St. Louis Co. Analyst: F. F. Grout (Grout, 1925b, p. 483, table 6, no. 2)
(2) Vermilion Granite, NW part of Sand Point L., sec. 12, 68N/17W, St. Louis Co. Analyst: F. F. Grout (*Ibid.*, no. 1)
(3) Biotitic phase of granite, Gun L., sec. 14, 65N/12W, St. Louis Co. Analyst: F. F. Grout (*Ibid.*, no. 15)
(4) Magnetite-rich granitic pegmatite, sec. 3, 66N/17W, St. Louis Co. Analyst: F. F. Grout (*Ibid.*, no. 16)
(5) Burntside Granite Gneiss, portage from Burntside L. to Little Long L., St. Louis Co. Analyst: Douglas Manuel (Grout, 1926, p. 29)

INTERNAL STRUCTURE

The major structural features within the Vermilion massif are summarized in Figure III-40. A pattern of gently-plunging, large open folds is revealed by the orientation of foliation in the migmatitic western part of the complex. This fold system also is indicated by the pattern of anomalies on aeromagnetic maps (U.S. Geological Survey, 1968; Zietz and Kirby, 1970).

Two points merit special emphasis with respect to these large folds. First of all, they trend essentially east-west, parallel to the major structural lines in older rocks lying both north and south of the massif. Secondly, even though the folds are mapped from scattered paleosomatic inclusions within a vast area that is chiefly granite, the individual measurements of foliation and lineation are remarkably consistent. Small crinkle axes or lines of oriented mafic minerals (hornblende and/or biotite) plunge consistently east or west within a given major fold, even in relatively small chunks of foliated schist or gneiss "floating" in granite. Linear structures having plunges of similar magnitude and orientation occur throughout older rocks in the Kabetogama Peninsula and the Vermilion district (Sims and others, 1968b). The structural consistency of migmatite paleosomes in the western part of the Vermilion massif suggests that the granite fraction, whatever its origin, was emplaced in a manner that did not completely disrupt structural trends in pre-existing rocks.

Small-scale folds of great geometric complexity abound within schist-rich migmatite, especially along the north margin of the massif. The most highly folded rocks consist of interlayered felsic and mafic units a few centimeters to a few meters thick. The folds change rapidly in shape and symmetry over short distances and invariably exhibit abrupt pinching and swelling of the constituent layers. They show all of the "wild" characteristics typical of folds in migmatites (Mehnert, 1968, p. 24), yet maintain rather consistent plunges within fairly large regions.

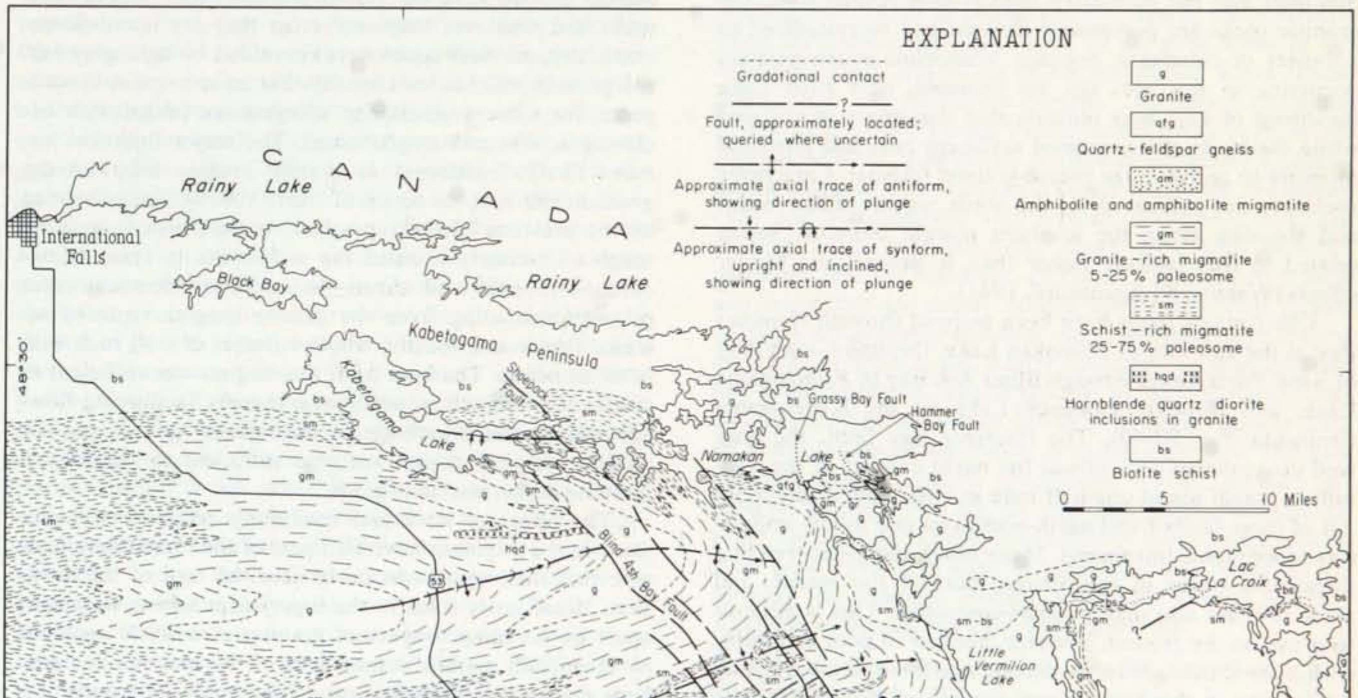


Figure III-40. Simplified geologic map of that part of the Vermilion massif in the International Falls two-degree sheet. (The form lines portray in a general way the structure of the migmatitic parts of the complex.)

Ptygmatic folds in relatively thin veins of quartz and quartz-feldspar pegmatite also are widespread. The complex form of the ptygmatic and other folds in the migmatite clearly indicates that the rocks were deformed while in a soft, mobile state. The apparent consistency of the axial plunges, however, indicates that they retained some strength and were not perfectly plastic.

Another common structure in schist-rich migmatite is boudinage. Typically the boudined layers are thin granitic neosomes in dominantly paleosomatic rocks. The orientations of boudin axes are difficult to measure in most of the flat, smooth outcrops in this glaciated region, but where they can be observed they plunge parallel to the major fold axes.

Fracturing of the Vermilion massif seems to have occurred later than the folding and was more or less independent of the early fold system. The Vermilion fault (Sims and others, 1968b) lies at or near the southern boundary of the Vermilion massif for many miles and is the most important fracture in the area (pl. 1). Within 4 to 5 miles of this fault and the subsidiary fault system related to it, the granitic rocks are pervasively sheared and recrystallized to a variety of cataclastic gneisses. Microcline augen gneisses occurring in this area (as, for example, near First Lake northwest of Ely) may indicate that shearing was initiated while the rocks still possessed sufficient heat and chemical mobility to recrystallize coarse-grained feldspar. Cataclastic rocks are not present along the north contact of the massif and shearing along the southern margin therefore seems related to true faulting rather than to protoclastic border effects (Waters and Krauskopf, 1941).

Other major faults have been mapped through Hammer Bay at the east end of Namakan Lake, through Grassy Bay of Sand Point Lake, through Blind Ash Bay of Kabetogama Lake, and through Shoepack Lake on the Kabetogama Peninsula (fig. III-40). The Hammer Bay fault, the only well documented one, offsets the north contact of the Vermilion massif about one-half mile in a left-lateral direction. All of these faults trend north-northwest and follow impressive topographic lineaments. Many other northwest-trending lineaments occur in the western part of the massif, and some of these also may be fault-controlled. This is difficult to prove on the ground, however, because of poor exposures within the lineament zones. Strong north-northeast-trending lineaments in the granitic rocks north of Ely (as, for example, through Angleworm Lake) are in part fault-controlled also (P. K. Sims, 1971, oral comm.), but most of this area has not yet been mapped geologically.

CONCLUSIONS AND SPECULATIONS

Light-pink biotite granite is the major rock type in the Vermilion massif. Sharp-walled dikes of the granite cut earlier massive hornblende quartz diorite and biotite granodiorite, and, in the western part of the massif, the granite is markedly migmatitic. The compositional homogeneity of the granite (away from the margins and the inclusion-laden western part) and its clear crosscutting relationships to earlier intrusive rocks favor a magmatic origin.

In the marginal and migmatitic zones, however, processes other than simple magmatism have operated. The intimate centimeter-scale interlayering of granitic neosome with paleosome in some of the schist-rich migmatites seems mechanically incompatible with the idea of *lit-par-lit* injection, and is better explained by a replacement process, or perhaps locally by anatexis. Textural evidence for replacement growth of microcline in these thin-layered migmatites is further evidence against simple injection. The granitic gneiss on the south shore of Namakan Lake, which grades southward into typical pink granite, has a sutured mosaic texture that can be interpreted as metasomatic in origin and contains layers that can be interpreted as relict bedding. It is possible that this gneiss is bedded wall rock that has been transformed to granite composition by outmigration of feldspar components from the large body of granitic magma to the south.

In summary, the following simplified model is proposed for the history of the Vermilion massif. First, small, local concordant plutons of hornblende quartz diorite and hornblende diorite invaded Lower Precambrian supracrustal rocks and whatever subjacent crust they lay upon. Somewhat later, all these rocks were reinvaded by light-gray biotite granodiorite; at least locally this intrusion was synorogenic, for some granodiorite stringers are folded with enclosing schists and amphibolites. The major injection (or, more likely, injections) of granite magma followed the granodiorite and, in terms of sheer volume, overwhelmed all the previous igneous episodes. At the contacts, granitic magma intimately invaded the wall rocks as lensoid concordant tongues and sheets. In addition, chemical components emanating from the granite magma replaced selected layers and, locally, whole volumes of wall rock with felsic minerals. The heat from the magma was sufficient to reduce the strength of wall rocks, thereby facilitating flowage folding and boudinage in the injected and chemically mobile marginal zones, and was sufficient to cause local increase in metamorphic grade.

The apparent structural continuity retained by inclusions in the migmatitic western part of the massif may indicate that this region was fairly near the roof of the intrusion. Broad early folds in the superjacent schists were split apart by invading tongues of granite, extensively replaced by chemically mobile constituents of the magma, and perhaps even partially melted on a local scale. The entire injection-replacement process was relatively quiescent, however, so that remnant blocks of the earlier rocks were not extensively rotated from their original orientations. It is likely that some broad-scale reorientation of earlier structural trends may have taken place as magma slowly pushed its way into the older rocks, but this process did not disrupt the consistent plunge of mineral lineations in the inclusions. Rather, the granite seems to have passively engulfed an earlier fold system.

The absence of inclusions in the eastern part of the massif, and the consequent absence of decipherable pregranite structure, probably mean that the roof of the complex was well above the present erosion surface in that area.

If the Vermilion massif is primarily an intrusive complex that involved much chemical replacement at its contacts, what can be concluded about the source of the magma? Clearly there is not enough cogenetic mafic rock available at the present level of exposure to account for the vast volume of granite by classical magmatic differentiation (Bowen, 1928), as Grout (1925b) attempted to do. A possible alternative is to call upon partial fusion of older volcanic and clastic rocks like those now occurring in the Lower Precambrian greenstone belts flanking the granitic massif. Four objections can be raised to this hypothesis, however. First, the grade of metamorphism in the greenstone belts is generally low (Sims and others, 1968b; Ontario Dept. of Mines, 1966) and rises above greenschist facies only within a mile or so of the granite contact zone. This is well shown on the north side of Kabetogama Lake where sillimanite-garnet schists are closely confined to the vicinity of the outermost granite injections. Less than a mile away from the contact zone the schists consist of quartz, sodic plagioclase, and biotite with scattered but locally abundant garnet; interbedded are thin layers (former mafic tuffs?) composed of chlorite, actinolite, epidote, and sodic plagioclase. It is difficult to imagine partial melting on the scale required to form the Vermilion massif without elevating the metamorphic grade of the associated greenstone belts above the upper greenschist facies. Indeed, the narrowness of the high-grade zone indicates that the metamorphism is more likely the *result* of granite plutonism rather than the cause of it. A second objection to partial melting is the relative paucity of mafic material associated with the granite. If rocks having the alkali content of Knife Lake-type metagraywacke and schist were partially melted to form granite (see chemical analyses in Grout, 1926, p. 19), it is clear that a large volume of residual minerals (roughly 45 percent of the total mass) would not be consumed. Although there are numerous paleosomatic remnants in the western part of the massif, they are not present in sufficient volume to account for the whole massif by partial fusion. A third objection is found in the sequence of intrusion. The earliest intrusive rocks in the massif are hornblende quartz diorite and diorite; these are followed in turn by biotite granodiorite and biotite granite. If the massif is to be explained by partial fusion, the *earliest* rock should be granite of approximately minimum-melting composition (Tuttle and Bowen, 1958), followed successively by more refractory rocks of more mafic compositions (Presnall, 1969). Finally, there is no clear evidence that Archean volcanic and sedimentary rocks ever were voluminous away from the present greenstone belts. Anhaeusser and others (1969) argued that Archean volcanic and clastic rocks may have accumulated in relatively narrow fault-controlled linear troughs that now are the greenstone belts, and may never have completely covered the areas between. If so, there may never have been enough supracrustal material from which to distill the Vermilion massif by partial fusion.

Two other hypotheses for the generation of large Precambrian granite-migmatite complexes can be considered. First, it is conceivable that the Vermilion massif and its brethren in ancient Precambrian terranes around the world could represent thoroughly transformed and remelted primeval continental crust, augmented by granitic magma derived from the differentiating mantle. This presupposes the existence of primeval granitic crust that formed during the initial differentiation of the earth's crust and mantle, and requires subsequent large-scale thermal events to cause the remobilization and reinjection. Primeval granitic crust has been notoriously difficult to identify in most Precambrian areas, but recent work by Condie and others (1970) in the Barberton Mountain Land of South Africa indicates the existence there of exposed granitic rocks which supplied detritus to sedimentary rocks as long ago as 3,400 to 4,000 m.y. A second possibility is the recent idea that granitic magmas may come as primary or derivative melts directly from the mantle, having formed in the root zones of andesitic volcanic arcs along major subduction zones (Dickinson, 1970, and references therein). This idea is intriguing in the case of the Lower Precambrian rocks because the greenstone belts, worldwide, contain calc-alkaline andesite-dacite volcanic suites and allied clastic rocks that are chemically and petrologically very similar to volcanic and sedimentary rocks in present-day island arcs. It is possible, therefore, that the vast tracts of Lower Precambrian granite and migmatite are deeply eroded infrastructure of former volcanic arc systems, and that the greenstone belts are the remnants of superjacent, nearly contemporaneous volcanic and sedimentary accumulations. If, in fact, granitic rocks can be generated by partial fusion of basaltic, oceanic crustal plates in subduction zones, as recent petrologic and geophysical studies suggest (Dickinson, 1970), perhaps we need no longer search for the primeval granitic continental crust upon which Lower Precambrian rocks accumulated. That crust may have been oceanic, and was "used up" in forming the andesitic volcanic suites and granite so characteristic of the nuclear Precambrian shields.

The work completed so far on the Vermilion massif does not establish either of these hypotheses in a concrete, irrefutable way. It is hoped that future work on trace element distributions, isotopic abundance, and chemical variation within this and other Precambrian massifs by my geochemically oriented colleagues will help us better understand the genesis of these complex rocks.

GIANTS RANGE BATHOLITH

P. K. Sims and S. Viswanathan

The Giants Range batholith, named from the low range of hills that lies immediately north of the Mesabi range, is the longest exposed body of granitic rocks in the state. It is known from scattered outcrops to extend from the vicinity of Grand Rapids northeastward to a point about 15 miles east of Ely, a distance of about 100 miles, where it disappears beneath the younger Duluth Complex (fig. III-41). Judged from geophysical data (Craddock and others, 1970; Zietz and Kirby, 1970), the batholith extends continuously beneath the glacial drift an equal distance southwestward from Grand Rapids. The exposed part varies in width from about 2 miles, west of Babbitt, to 22 miles in the area north of Chisholm. The southern margin of the batholith locally is overlapped by the Middle Precambrian Pokegama Quartzite and Biwabik Iron-formation, and scattered drill holes and gravity data (Ikola, 1968a and b) indicate that the granitic rocks continue in the subsurface beneath these rocks for distances of as much as several miles.

Where exposed, the batholith is composed chiefly of quartz-bearing granitic rocks ranging in composition from tonalite to granite, but includes scattered smaller masses of darker and generally older plutonic rocks and remnants of metamorphosed sedimentary and volcanic rocks. Rocks in the compositional range of granodiorite, adamellite, and granite predominate. The granitic rocks are in discrete masses or plutons, which are elongate in a northeasterly direction, parallel with the long dimension of the batholith and with the structure of the wall rocks. Contacts between plutons and wall rocks and between adjacent plutons generally are steep, subconformable, and sharp. Thin, discontinuous lenses and angular blocks of metamorphic rocks occur within the separate plutons and rarely between adjacent plutons, but constitute only a small percentage of the total volume of the batholith.

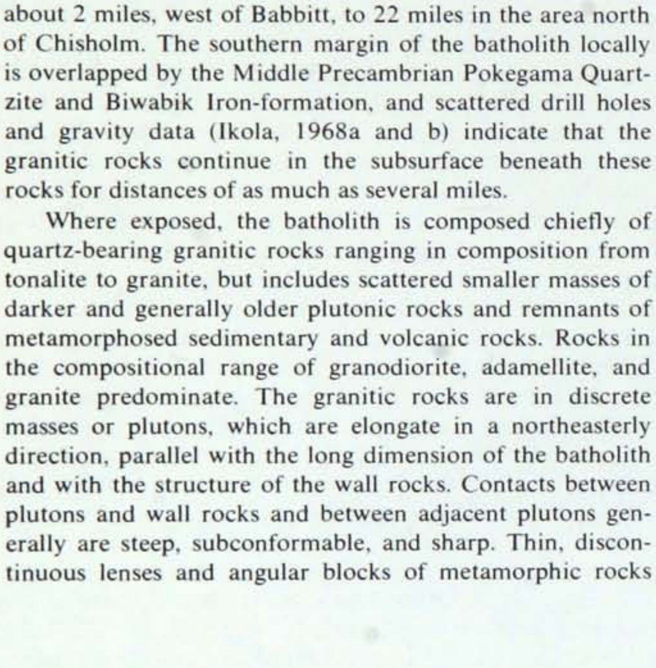


Figure III-41. Map showing approximate outline of the Giants Range batholith. The eastern and western parts are described separately in the text.

The granitic rocks comprising the batholith have been called Giants Range Granite (Allison, 1925; Sims and others, 1970), following the original formal designation by Spurr (1894). Recent mapping has shown that the Giants Range Granite can be subdivided on the basis of composition, texture, and intrusive relations or other evidence of relative age into several mappable lithologic units. Accordingly, on the following pages the granitic rocks that make up the Giants Range Granite are discussed in terms of the lithologic units that have been defined.

The country rocks of the batholith are metavolcanic and metasedimentary rocks of Early Precambrian age. Throughout the western part of its exposed length, the batholith is mainly in contact with metagraywacke and related rocks of the Lake Vermilion Formation, whereas in the eastern part it is mainly in contact with mafic metavolcanic rocks of the Ely Greenstone. For the most part, the contacts are sharp. Both metamorphic grade and intensity decrease systematically away from the contacts, from middle amphibolite facies to greenschist facies, the normal regional metamorphic grade. The thermal aureole is narrow, commonly being no more than a few hundred feet wide, except in the area south of Tower where it attains a width of several thousand feet. In the area between Embarrass and the eastern termination of the batholith, the northern margin is bounded at many places by steeply-dipping faults, which bring the plutonic rocks into juxtaposition with greenschist-facies rocks.

The generally conformable structure of the plutons with the adjacent country rocks is interpreted as indicating that the igneous rocks are syntectonic, having been emplaced synchronously with regional deformation. They have the characteristics of rocks emplaced in the mesozone of Buddington (1959). Subsequent to crystallization, the rocks were extensively granulated, as a consequence of regional faulting and associated cataclasis that followed the main episode of folding during the Algoman orogeny.

The Giants Range batholith is considered a typical Algoman batholith (Lawson, 1913b; Grout and others, 1951; Goldich and others, 1961). Biotite- and muscovite-bearing granitic rocks from the western part have a Rb-Sr isochron age of $2{,}670 \pm 65$ m.y. (Prince and Hanson, in press). This compares with a Pb^{207}/Pb^{206} age for sphene of 2,730 m.y. (Catanzaro and Hanson, 1971), and a discordant U-Pb age for zircon concentrates of 2,700-2,750 m.y., depending somewhat on the model chosen for discordance (D. H. Anderson, 1965, unpub. Ph.D. thesis, Univ. Minn.). These data, which are summarized in Figure III-42, are interpreted by Prince and Hanson (in press) as indicating that the

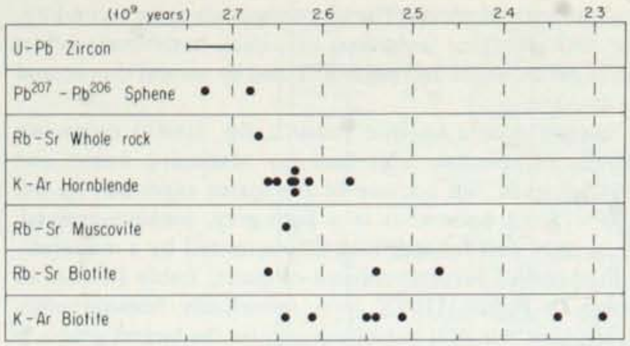


Figure III-42. Radiometric ages for the Giants Range Granite. (After Prince and Hanson, in press.)

plutonic rocks of the batholith crystallized about 2,700 m.y. ago. K-Ar ages for hornblende (Hanson, 1968) and biotite and Rb-Sr ages for biotite (Goldich and others, 1961) are younger; they range from 2,290 to 2,660 m.y. and cluster around 2,500-2,650 m.y., and reflect some post-crystallization event or events. Possibly, the younger ages partly reflect post-consolidation shearing related to the late Algoman episode of faulting, mentioned above, but they also may reflect the still younger deformation and low-grade metamorphism that is imprinted on the Middle Precambrian mafic dikes in the western part of the Vermilion district and adjacent areas (see Sims and Mudrey, this volume).

For purposes of discussion on the following pages, the batholith is divided into an eastern and a western part. This subdivision is in accord with Leith's earlier (1903) recognition that the predominant granite west of Mountain Iron is more gneissic than the granite to the east. Also, the larger plutons in the western part are characterized by biotite as the dominant mafic mineral, whereas those in the eastern part are characterized by dominant hornblende.

HISTORY OF INVESTIGATIONS

Following the earlier reconnaissance investigations by Grant (*in* Winchell and others, 1899) and Leith (1903), Allison (1925) described the petrography and structure of the rocks constituting the Giants Range batholith. Allison recognized a "border phase" of mafic rocks, an "intermediate phase" exposed over large areas, an "Embarrass phase" exposed between Embarrass and Birch Lake, and a "biotite-granite phase" exposed mainly between Chisholm and Mountain Iron. Later, Grout and Thiel (*in* Tyler and others, 1940) further subdivided the rocks of the batholith on the basis of heavy accessory minerals, and attempted to delineate the general areal extent of the various phases (see pl. 1 in Tyler and others, 1940). More recently, relatively detailed mapping in the western part of the batholith by S. Viswanathan (1971, unpub. Ph.D. thesis, Univ. of Minn.), in the Embarrass-Babbitt area by W. L. Griffin (1967, unpub. Ph.D. thesis, Univ. of Minn.; Griffin and Morey, 1969), and in the Gabbro Lake quadrangle by Green (Green and others, 1966; Green, 1970a) has delineated the approximate distribution of major rock types and has clarified many aspects of their petrology.

ROCK NOMENCLATURE

A modification of Streckeisen's (1967) classification is used to describe the granitic rocks of this area (fig. III-43). The classification is similar to that used by Southwick (this chapter, Vermilion massif), but differs from it mainly in having a major division at 30 percent quartz rather than 20 percent quartz. A discussion and rationale for the classification is given by Viswanathan (1971, *op. cit.*).

WESTERN PART

West of Mountain Iron, the Giants Range batholith is composed of several mappable units of granitic rocks that strike northeastward, subparallel to the regional structure. Some of the units constitute separate plutons, which range in composition from tonalite to granite; a few appear to have formed by static granitization. The granitic rocks are poorly exposed and, except locally, contacts between mappable units are rarely observed and for the most part are inferred from structural and geophysical data.

The granitic rocks in the western part of the batholith can be grouped according to composition, texture, and relative age into three intrusive series—tonalitic, granodioritic, and granitic—and a fourth granitoid series. Rocks of the tonalitic series are the oldest; they are cut by one unit of the granodioritic series. Rocks of the granitic series are inferred from regional relationships to be the youngest intrusive rocks. A satellitic phase of the granitic series cuts the granitoid rocks. The discussion of these rocks that follows is summarized from Viswanathan (1971, *op. cit.*).

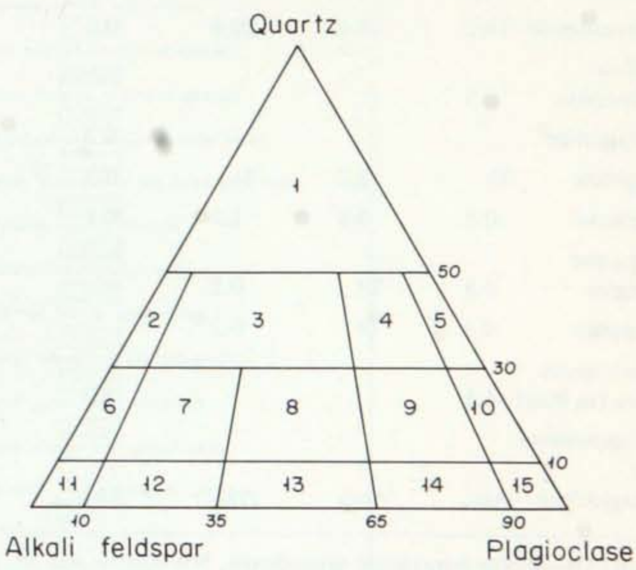


Figure III-43. Classification of plutonic rocks used in this section. Fields: 1, quartz-rich rocks (rare); 2, alkali granite; 3, granite; 4, quartz granodiorite; 5, quartz tonalite; 6, alkali syenogranite; 7, syenogranite; 8, adamellite; 9, granodiorite; 10, tonalite; 11, alkali syenite; 12, syenite; 13, monzonite; 14, syenodiorite; 15, diorite.

Tonalitic Series

Rocks of general tonalitic composition are exposed sporadically on the low hills north of the Mesabi range between Chisholm and Nashwauk (fig. III-44) and in the vicinity of Grand Rapids. Two mappable units have been distinguished, a biotite-hornblende tonalite and a biotite quartz tonalite. The age relations of the two units are not known.

Biotite-hornblende tonalite (unit 1 on fig. III-44) crops out discontinuously in a 1-mile-wide belt between Chisholm and Mahoning Lakes. Typically, the rock is a gray, medium-grained, well foliated tonalite or granodiorite (table III-26; fig. III-45). A layering is imparted by thin, alternating hornblende- and biotite-rich layers and quartz-feldspar layers. Sphene is conspicuous. The K-feldspar is a perthitic microcline that occurs either as large subhedral grains or as small interstitial grains. The biotite appears to be secondary. The rock contains inclusions of biotite-hornblende schist and is cut by aplite and pegmatite and by several diorite and basalt dikes.

Biotite quartz tonalite (unit 2, fig. III-44) crops out sporadically on low hills between Mahoning Lakes and Grand Rapids, but because of the sparse exposures its full extent is not known. It is a light-gray, medium-grained, moderately well foliated rock characterized by a moderately high (28-40 percent) content of quartz (table III-26). As shown in Figure III-45, it is remarkably homogeneous. Plagioclase, which is zoned, constitutes the largest grains in the rock. In the vicinity of Nashwauk, the rock contains amphibolite inclusions and numerous granitic veins; and near Grand Rapids, it contains inclusions of biotite-hornblende schist and is cut by biotite granite and pegmatite. At both localities, the granitic rocks as well as the tonalitic

Table III-26. Modes, in volume percent, of tonalitic series and granodioritic series.

	Biotite-hornblende tonalite unit (1)			Biotite quartz tonalite unit (2)	Foliated hornblende granodiorite unit (3)[5]			Foliated hornblende granodiorite unit (4)	
								Main body[1]	Leucogranodiorite phase[2]
	HBG-1B	HBG-3A	KTN-1A	Ave. of 4	STL-33	STL-34T	STL-36A	Ave. of 4	Ave. of 5
Quartz	13.2	9.1	12.3	34.5	20.3	25.4	34.5	17.5	25.3
K-spar	17.8	1.1	7.3	0.7	17.3	0.2	0.8	14.9	21.2
Plagioclase	47.2	65.0	55.7	54.8	40.4	64.2	57.3	57.6[3]	51.3[4]
Biotite	4.0	5.5	2.4	8.2	0.6	7.9	6.4	0.5	1.1
Hornblende	16.2	18.4	20.6	0.2	19.6	1.1	0.7	8.8	0.8
Clinopyroxene	0.3				1.2	1.2	Tr		
Muscovite				0.2					
Epidote	Tr	0.2	Tr	0.3	0.1	0.1	0.1	0.1	Tr
Sphene	0.6	0.8	1.3	0.1	1.1		0.2	0.1	Tr
Opaque oxides	0.3	Tr	0.2		0.4	0.4		Tr	Tr
Apatite	0.3	Tr	0.2		0.2	0.2	0.1	0.6	0.1
Ave. grain size (in mm)	1.5			2.0	3.0	1.5	1.5	2.0	
Composition of plagioclase	An_{16}	An_{20}	An_{18}	An_{17-20}	An_{20}	An_{20}	An_{20}	An_{9-15}	An_{8-12}

HBG-1B—Biotite-hornblende granodiorite, NW¼SE¼ sec. 19, 58N/20W, Hibbing 7.5-minute quadrangle
HBG-3A—Biotite-hornblende tonalite, NE¼NW¼ sec. 26, 58N/21W, Hibbing 7.5-minute quadrangle
KTN-1A—Biotite-hornblende granodiorite, NE¼SE¼ sec. 28, 58N/21W, Keewatin 7.5-minute quadrangle
STL-33—Hornblende granodiorite, SW¼SW¼ sec. 23, 59N/22W, Stingy Lake 7.5-minute quadrangle
STL-34T—Biotite tonalite, SW¼SW¼ sec. 26, 59N/22W, Stingy Lake 7.5-minute quadrangle
STL-36A—Biotite quartz tonalite, SE¼NE¼ sec. 34, 59N/22W, Stingy Lake 7.5-minute quadrangle

[1] Body constitutes core of an anticline, NW of Big Rice L.
[2] Occurs as distinct, cross-cutting dikes within the main body of hornblende granodiorite; designated unit 4a in Table III-29
[3] Plagioclase is zoned; core is An_{9-13}; rim is An_{8-10}
[4] Plagioclase is zoned; core is An_{9-12}; rim is An_{6-11}
[5] See Table III-29 for an additional mode of a typical phase

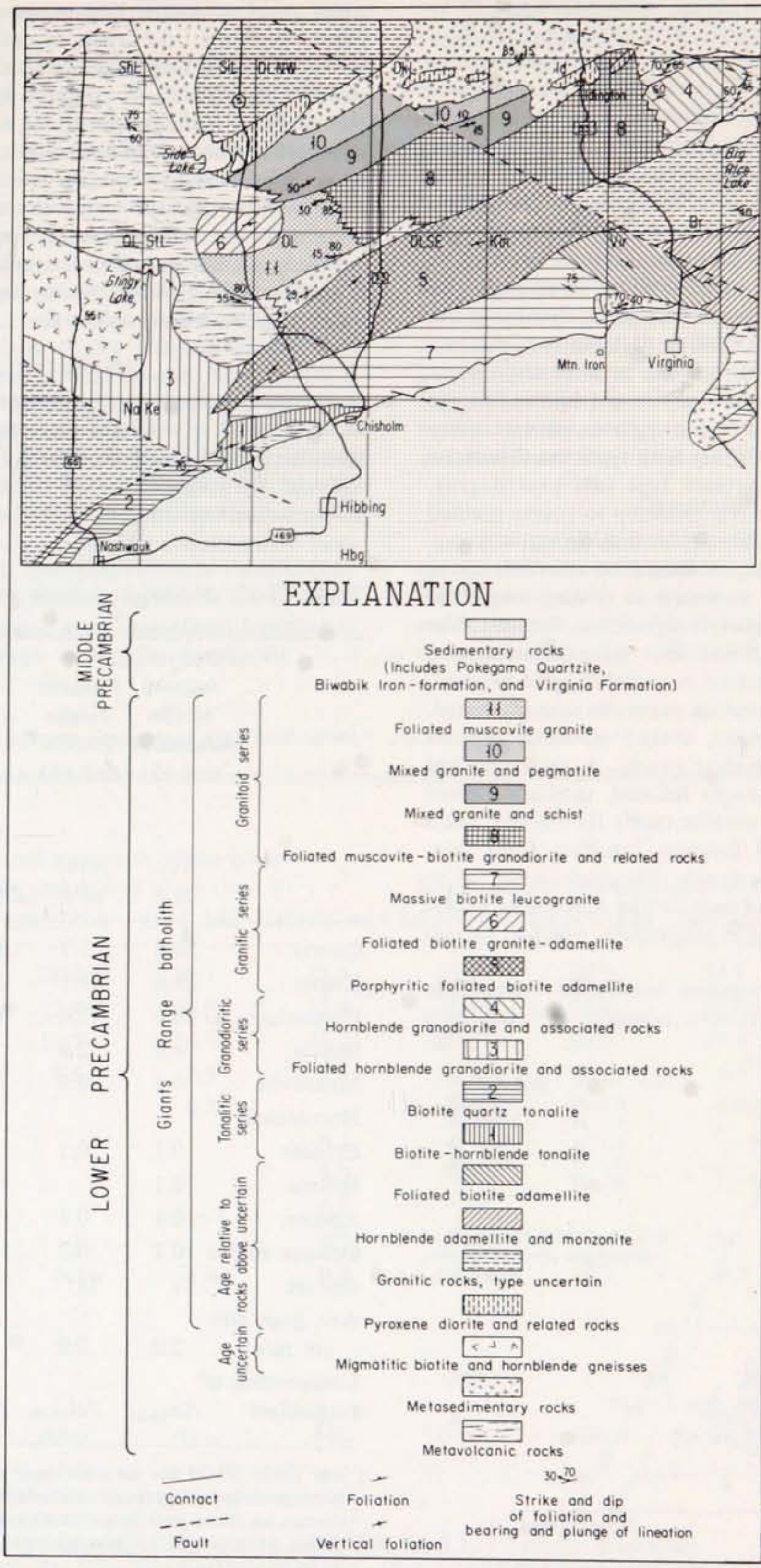


Figure III-44. Generalized geologic map of the western part of the Giants Range batholith (modified from Sims and others, 1970). Index to quadrangles: Br, Britt; DKL, Dark Lake; DLNW, Dewey Lake NW; DLSE, Dewey Lake SE; Hbg, Hibbing; Id, Idington; Ke, Keewatin; Kin, Kinney; Na, Nashwauk; OL, O'Leary; ShL, Sherry Lake; SiL, Side Lake; StL, Stingy Lake; Vir, Virginia.

rocks show evidence of late-stage alteration—biotite is chloritized, plagioclase is sericitized, and secondary muscovite is present.

Granodioritic Series

Two bodies that are composed dominantly of granodiorite, but are variable in composition, have been mapped (fig. III-44). One is northwest of Hibbing, in the Stingy Lake area, and the other is northwest of Big Rice Lake.

Foliated hornblende granodiorite and associated rocks (unit 3, fig. III-44) occupy an area of about 20 square miles northwest of Hibbing and are referred to informally as the Stingy Lake body. The granodioritic rocks intrude the layered mafic rocks to the northeast, but because of poor exposures their relation to other units is not known. In the same way, the age relations of the separate phases within the unit are obscure. Two distinct rock types constitute the Stingy Lake body. The dominant type is a pinkish-gray, equigranular, generally massive, medium- to coarse-grained hornblende granodiorite (table III-26 and fig. III-43) and, locally, adamellite that is characterized by the presence of black hornblende crystals as much as 6 mm long. The adamellite has a hypidiomorphic-granular texture. The hornblende is green and zoned, and some crystals have clinopyroxene cores; microcline is perthitic, grid-twinned, and largely interstitial; plagioclase occurs as strongly zoned, untwinned or carlsbad-twinned, altered subhedral crystals and as albite-twinned subhedral grains. A less abundant type is a pinkish-gray, strongly foliated, medium-grained biotite tonalite and quartz tonalite (table III-26) that has a conspicuous compositional layering resulting from thin, alternating mafic and felsic bands. The rock is similar to the biotite quartz tonalite of unit 2. The K-feldspar in this rock occurs mainly as blebs in plagioclase, which is strongly zoned concentrically.

The second body, composed mainly of hornblende granodiorite (unit 4, fig. III-44), occupies the core of a major anticline east of Idington and northwest of Big Rice Lake. Typically, the rock is a pinkish-gray, medium-grained, foliated hornblende granodiorite of uniform composition (fig. III-45). It contains sparse inclusions of amphibolite and is cut by abundant, irregular bodies of massive, aplitic leucogranodiorite (see table III-26). The leucogranodiorite locally contains inclusions of the foliated granodiorite. Adjacent to the contact with mafic metavolcanic rocks, on the north margin of the pluton, the granodiorite is strongly granulated; hornblende has broken down to smaller aggregates of chlorite, epidote, and sphene, and the plagioclase is recrystallized.

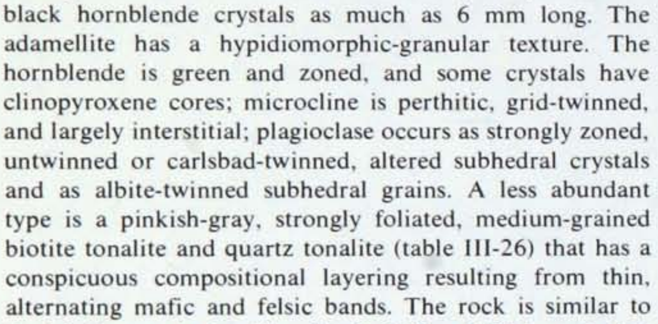


Figure III-45. Alkali feldspar-plagioclase-quartz variation diagram for plutonic rocks in western part of Giants Range batholith. See Figure III-44 for location of various geologic units referred to here.

Granitic Series

Intrusive rocks having the general composition of granite occur north of the Hibbing-Mountain Iron sector of the Mesabi range. Two of the units (5 and 7, fig. III-44) are adjacent and subparallel, and constitute the largest known pluton in the western part of the batholith. Satellitic bodies

Table III-27. Modes, in volume percent, of granitic series.

	Porphyritic, foliated biotite adamellite unit (5)	Foliated biotite granite unit (6)	Massive biotite leucogranite unit (7)		
			Main body[1]		Satellitic phases[3]
	Ave. of 4	Ave. of 4	DLSE 8	DLSE 18F[2]	Ave. of 3
Quartz	28.0	27.7	33.4	23.3	30.5
K-spar	29.6	31.1	28.5	17.5	32.0
Plagioclase	36.0	37.4	31.4	44.6	34.9
Biotite	4.7	2.4	5.7	8.6	1.6
Muscovite	0.5	0.5	0.2		1.0
Hornblende				5.6	
Epidote	0.1	0.2		0.1	Tr
Sphene	0.1		0.2	0.1	
Apatite	0.1	0.1	0.1	0.1	Tr
Opaque oxides	0.7	0.3	0.6	0.1	0.1
Garnet	Tr	Tr			
Ave. grain size (in mm)	2.0	2.0			<2.0
Composition of plagioclase	An_{14-18}	An_{16-18}	An_{20}	An_{20}	An_{13-16}

[1] See Table III-29 for an additional mode of a typical phase
[2] Representative of granodioritic phase
[3] Occurs as dikes and larger bodies in country rock screen between units 5 and 11, and in units 8 and 11, designated unit 7a in Table III-29

DLSE-8—Biotite granite, SE¼NE¼ sec. 15, 58N/20W, Dewey Lake SE 7.5-minute quadrangle
DLSE-18F—Hornblende-biotite granodiorite, SE¼NE¼ sec. 2, 58N/20W, Dewey Lake SE 7.5-minute quadrangle

of unit 7 transgress and are younger than the rocks constituting the granitoid series.

A body of porphyritic, foliated biotite adamellite and granite (unit 5, fig. III-44) occupies an area about 10 miles long and 4 miles wide in the low-lying plain north of the Giants Range, extending from the vicinity of Hibbing eastward to Mountain Iron. Typically, the rock is pink, porphyritic, moderately or strongly foliated, and coarse grained. Phenocrysts from 1-2 cm long and 0.5-1 cm wide of pink, slightly perthitic microcline occur in a gray, medium-grained matrix of yellowish-brown biotite, smoky quartz, plagioclase, and microcline (table III-27). The plagioclase is moderately or highly altered, and contains secondary muscovite, epidote, and calcite. The texture is hypidiomorphic-porphyritic. Small inclusions of schist are present locally, as are feldspathic veinlets.

To the south of unit 5 and parallel to it is an apparently conformable body of biotite leucogranite (unit 7, fig. III-44), which is exposed nearly continuously for a distance of 11 miles on the hills between Chisholm and Mountain Iron. The rocks cut the foliated biotite-hornblende tonalite (unit 1) west of Chisholm. The dominant phase is a pink, generally equigranular, massive, medium-grained biotite leucogranite; locally, the rock is slightly porphyritic. It is variable in composition (see fig. III-43), containing from 17 to 39 percent quartz, 28 to 70 percent microcline, 12 to 32 percent plagioclase, and one to eight percent chloritized biotite. Modes of typical felsic phases are given in Table III-27. More biotitic phases that have a composition of granodiorite occur locally; these are gray and distinctly more mafic than the main phase.

A third, separate body of granitic rocks (unit 6, fig. III-44) occurs south of Side Lake about 10 miles north of Chisholm. Contacts with adjacent rocks are not exposed, but judged from its internal structure it is subconformable to adjacent rocks to the north and south. The rock is dominantly a pink or pinkish-gray, equigranular, foliated biotite adamellite or granite (table III-27). The K-feldspar is a grid-twinned, perthitic microcline; quartz is anhedral, highly strained, and largely interstitial; the plagioclase is sericitized, and the biotite is chloritized and bleached.

Granitoid Series

Rocks ranging in composition from granodiorite to granite that are interpreted to have formed by static granitization occur north of the major body of intrusive granitic rocks (units 5 and 7, fig. III-44), between these rocks and the metasedimentary rocks of the Lake Vermilion Forma-

Table III-28. Modes, in volume percent, of granitoid series.[1]

	Foliated muscovite-biotite granodiorite and related rocks (unit 8)				Mixed granite and schist unit (9)[2]			Mixed granite-pegmatite unit (10)	Foliated muscovite granite unit (11)
	DKL-7	DKL-20A	DKL-38C	DLNW-44A	DLNW-10	DLNW-55B	Ave. of 4	Ave. of 3	Ave. of 4
Quartz	32.3	20.2	20.7	25.5	37.5	23.9	33.7	28.6	37.2
K-spar	21.4	12.4	20.7	15.3	26.7	24.1	24.6	24.1	30.3
Plagioclase	38.7	58.7	54.7	45.2	27.2	45.5	34.4	38.8	25.0
Biotite	4.4	8.6	2.3	6.3	6.7	6.4	6.3	6.1	1.1
Muscovite	2.1		0.2		1.5		0.9	1.1	6.2
Hornblende				6.2					
Epidote				1.0	0.2		0.1	0.1	Tr
Sphene				0.1					Tr
Apatite				0.3			Tr	0.2	0.1
Opaque oxides	1.2	0.1	1.4	0.2	0.2	0.2	0.2	0.4	0.3
Garnet									Tr
Ave. grain size (in mm)	2.0				1.0			1.0	2.0
Composition of plagioclase	An_{14}	An_{12}	An_{20}	An_{18}	An_{14}	An_{16}	An_{14-19}	An_{17-18}	An_{8-14}

[1] See Table III-29 for additional modes of typical phases
[2] Modes only of granitic rocks within unit
DKL-7—Muscovite-biotite granite, SE¼SE¼ sec. 20, 60N/19W, Dark L. 7.5-minute quadrangle
DKL-20A—Biotite granodiorite, SE¼SE¼ sec. 17, 60N/19W, Dark L. 7.5-minute quadrangle
DKL-38C—Biotite granodiorite, SE¼SE¼ sec. 23, 60N/20W, Dark L. 7.5-minute quadrangle
DLNW-44A—Hornblende-biotite granodiorite, SW¼NW¼ sec. 32, 60N/20W, Dewey L. NW 7.5-minute quadrangle
DLNW-10—Muscovite-biotite granite, SW¼SE¼ sec. 24, 60N/21W, Dewey L. NW 7.5-minute quadrangle
DLNW-55B—Biotite adamellite, NW¼SE¼ sec. 24, 60N/21W, Dewey L. NW 7.5-minute quadrangle

Table III-29. Partial chemical analyses and modes, in volume percent, of plutonic rocks, western part of Giants Range batholith (analyses by S. Viswanathan; method, X-ray fluorescence).

	Chemical composition, in weight percent												
	Tonalitic series		Granodioritic series			Granitic series				Granitoid series			
Map unit	1	2	3	4	4a[2]	5	6	7	7a[3]	8	9	10	11
SiO_2	57.30	66.01	65.30	67.22	75.63	69.33	70.10	75.82	72.83	67.30	72.17	74.61	74.75
Al_2O_3	20.20	16.60	15.60	16.20	12.10	15.10	16.10	12.30	15.00	17.40	14.00	12.10	13.90
TiO_2	0.60	0.46	0.43	0.22	0.05	0.32	0.25	0.10	0.07	0.32	0.37	0.22	0.03
FeO[1]	6.40	3.90	4.23	2.80	0.75	2.15	2.00	1.50	1.00	3.30	2.50	1.73	1.00
MnO	0.09	0.06	0.07	0.05	0.03	0.04	0.03	0.03	0.03	0.07	0.04	0.03	0.03
MgO	2.66	1.15	1.52	0.52	0.20	0.73	0.46	0.13	0.13	1.75	0.36	0.63	0.03
CaO	4.65	3.15	3.42	2.53	1.15	1.85	1.75	0.62	0.78	2.20	1.85	1.38	0.35
K_2O	3.00	2.00	4.10	2.80	3.20	4.45	4.63	5.90	5.35	1.65	4.05	4.10	5.15
Na_2O	4.57	5.82	4.80	6.24	5.38	5.15	3.81	2.39	3.94	5.48	3.72	4.18	2.94
Total	99.47	99.15	99.47	98.58	98.49	99.12	99.13	98.79	99.13	99.47	99.06	98.98	98.18
	Modes, in volume percent												
Quartz	13.2	28.5	14.6	20.0	32.7	20.5	31.5	38.7	32.4	30.3	31.0	35.7	39.8
K-spar	17.8	0.4	30.3	14.6	17.5	28.7	31.5	38.9	31.0	5.6	25.6	22.2	25.8
Plagioclase	47.2	63.6	43.3	59.8	46.7	44.9	34.0	20.7	33.1	54.6	36.6	37.3	25.1
Biotite	4.0	6.7		1.8	1.2	4.4	2.5	1.2	2.8	8.7	6.2	3.7	0.7
Muscovite						0.5	0.1	0.6	0.8		0.3	0.5	8.7
Hornblende	16.2	0.7	8.5	3.0	0.9					Tr			
Clino-pyroxene	0.3		2.1										
Epidote	Tr			0.3	0.6	0.2	0.2			0.2			Tr
Sphene	0.6	0.1	0.3	0.5	0.1	0.1				Tr			Tr
Apatite	0.3		0.1	Tr	0.1	0.1				0.3	0.1	0.5	Tr
Opaque oxides	0.3		0.6	Tr	0.2	0.7	0.1	0.2	Tr	0.1	0.3		
Garnet													Tr

[1] Total oxides, reported as FeO
[2] Aplitic leucogranodiorite that cuts main body of granodiorite in map unit 4
[3] Leucogranite, satellitic phase of unit 7 that cuts rocks in map unit 8

1. HBG-1B—Biotite-hornblende granodiorite, NW¼SE¼ sec. 19, 58N/20W, Hibbing 7.5-minute quadrangle
2. CSTE-1A—Biotite quartz tonalite, NW¼NW¼ sec. 4, 55N/25W, Cohasset East 7.5-minute quadrangle
3. STL-34—Hornblende adamellite, SW¼SW¼ sec. 26, 59N/26W, Stingy L. 7.5-minute quadrangle
4. BR-56—Biotite-hornblende granodiorite, NW¼NE¼ sec. 31, 61N/17W, Britt 7.5-minute quadrangle

4a. BR-49D—Leucogranodiorite, NE¼NE¼ sec. 32, 61N/17W, Britt 7.5-minute quadrangle

5. DL-6—Biotite adamellite, SW¼SE¼ sec. 30, 59N/20W, Dewey L. 7.5-minute quadrangle
6. DL-37—Biotite granite, NW¼NW¼ sec. 10, 59N/21W, Dewey L. 7.5-minute quadrangle
7. DLSE-4—Leucogranite, SE¼SW¼ sec. 5, 58N/19W, Dewey L. SE 7.5-minute quadrangle

7a. DLNW-45C—Leucogranite, SE¼NE¼ sec. 32, 60N/20W, Dewey L. NW 7.5-minute quadrangle

8. DLNW-45A—Biotite quartz tonalite, SE¼NE¼ sec. 32, 60N/20W, Dewey L. NW 7.5-minute quadrangle
9. DKL-18—Biotite granite, NW¼NE¼ sec. 9, 60N/19W, Dark L. 7.5-minute quadrangle
10. DLNW-17—Biotite granite, SW¼SE¼ sec. 7, 60N/20W, Dewey L. NW 7.5-minute quadrangle
11. DL-16—Muscovite granite, SW¼SW¼ sec. 6, 59N/20W, Dewey L. 7.5-minute quadrangle

tion. In general, the rocks contain more intercalated schist and pegmatite than the rocks described above, and are characterized by the presence of both muscovite and biotite. Commonly, the contacts between the schist inclusions and the granitoids are diffuse and gradational.

Foliated muscovite-biotite granodiorite and related rocks (unit 8, fig. III-44) constitute a complex mineralogic unit immediately north of unit 5. The principal phase is a pinkish-gray, gneissic, medium- to fine-grained muscovite- and biotite-bearing rock ranging in composition from tonalite to granite (table III-28 and fig. III-43); it is associated with aplite, leucogranite nearly identical to unit 7, and pegmatite, and contains small bodies of hornblende and biotite schist. Characteristically, microcline embays and appears to have replaced plagioclase, and microcline grains are rimmed by myrmekite. Antiperthitic plagioclase is untwinned, strongly zoned, and quite variable in grain size. Pinkish-gray, massive, fine-grained biotitic aplite cuts the gneissic granitic rocks; it has a cataclastic texture. In turn, the aplite is cut by dikes of leucogranite.

North of unit 8 is a unit about 2 miles wide that consists of mixed granite and schist (unit 9), in a ratio of about 7 to 3. The principal rock is a gray, equigranular, weakly foliated, fine-grained muscovite-biotite granite (table III-28) that is compositionally layered as a result of alternating biotite-rich and quartzofeldspathic layers. Antiperthitic and myrmekitic plagioclase are lacking. The granitoid rock grades transitionally into the schist (both biotite schist and hornblende-biotite schist) and both are cut by aplite, leucogranite, and pegmatite.

A unit (10, fig. III-44) characterized by abundant pegmatite lies between unit 9 and the Lake Vermilion Formation, and apparently is conformable with these rocks. It is composed dominantly of a gray, equigranular, weakly foliated, fine-grained muscovite-biotite adamellite or granite that is remarkably uniform in composition. Clots and veinlets of feldspar and biotite occur locally. Pegmatite occurs as segregations and veins, and constitutes 20 to 25 percent by volume of the unit. The microcline in the granitoid rock is perthitic (rods and veins) and lacks grid twinning.

A unit (11, fig. III-44) that appears transitional to unit 8 lies north of a screen of metasedimentary rocks that separates it from unit 5 of the granitic series. Typically, the rock is a light-gray, foliated, medium-grained biotite-muscovite granite, which locally is garnetiferous. A few large, angular amphibolite bodies occur in the granitic rock. Dikes of leucogranite cut the mass. A garnetiferous aplite is associated with the amphibolite bodies. Muscovite and garnet, where present, are disseminated uniformly through the rock.

Chemical Composition

Partial chemical analyses of representative samples from the major rock units and the principal satellitic phases are given in Table III-29. The rocks that constitute the batholith belong to the calc-alkaline series, as shown by Figure III-46. The alkali-lime index is 61, a value that separates the calc-alkaline series from the calcic series (Peacock, 1931). With respect to alkali-lime relationships, rocks from the granitic series are similar chemically to those of the granitoid series (fig. III-47). Both rock series fall within a rather narrow field that lies on the Na-K side of the mean for igneous rocks, as determined by Green and Poldervaart (1958). With respect to Na-Ca-K, the tonalitic rocks are similar to the volcanogenic metasedimentary rocks in the area, and both occupy a narrow field near the Na-Ca boundary.

Compared with the so-called granite averages (Nockolds, 1954), granitic rocks from the Giants Range batholith are lower in magnesia and iron and higher in soda—characteristics also shown by other Precambrian granitic rocks from the Canadian Shield (Reilly and Shaw, 1967).

The concentrations of significant trace elements together with critical ratios for the rocks of the Giants Range are given in Table III-30.

Petrogenesis

The field relations, petrology, and chemistry of the rocks in the batholith are consistent with a magmatic origin for the granodioritic and granitic series, a metasomatic origin for the granitoid series, and an anatectic origin for the tonalitic series.

Plots of the data on the major oxides (table III-29) indicate that the four intrusive rock series probably do not belong to a single, genetically related suite derived from a common magma. As shown in the Niggli variation diagrams in Figure III-48, there are no systematic trends common to all the rock series. The available data, however, suggest that both the granitic series and the granodioritic series have separate and distinct systematic trends. Points for the granitic series (adamellite-granite-leucogranite) of units 5, 6, 7, and 7a (see table III-29) and the granodioritic series (granodiorite-adamellite-leucogranodiorite) of units 3, 4, and 4a lie along separate curves (fig. III-48).

The trace element data support a magmatic origin for the granodioritic and granitic series and separate magmatic differentiation trends for the two series. For the granodioritic series, with increasing fractionation, Ba (fig. III-47A), Sr (fig. III-49E), Zr, Ti (fig. III-47C), K/Rb (fig. III-47D), Ba/Rb (fig. III-50B), Ba/Sr, Ca/Sr, Ti/Zr, K/Pb, and Ti/Fe decrease, whereas Rb (fig. III-50A), Pb, and K/Ba increase. K/Sr (fig. III-49F) and Rb/Sr remain nearly constant. The high Ba/Rb (fig. III-48B) and K/Rb (fig. III-49D) ratios and low K/Ba (fig. III-47B), K/Sr (fig. III-49F), and Rb/Sr ratios indicate that the granodioritic series was derived from unfractionated liquids. For the granitic series, with increasing fractionation, Ba (fig. III-49A), Sr, Zr, Ti (fig. III-49C), Pb, K/Rb (fig. III-49D), Ba/Rb (fig. III-50B), Ba/Sr, Ca/Sr, Ti/Zr, and Ti/Fe decrease, whereas Rb (fig. III-50A), K/Sr (fig. III-49F), Rb/Sr, K/Pb, and K/Ba (fig. III-49B) increase. Judged from the low Ba/Rb (fig. III-50B) and K/Rb (fig. III-49D) ratios and high K/Ba (fig. III-49B), K/Sr, and Rb/Sr ratios, the granitic series was formed from highly fractionated magmas, which culminated with the development of small, further fractionated, satellitic leucogranite bodies. The rocks of both series have chemical characteristics consistent with derivation from the lower crust or upper mantle. The granodioritic series represents more primitive, less fractionated magmas than the granitic series.

Although the two-mica granitic rocks of the granitoid series are comparable in major element chemistry to the rocks of the granitic series, they have distinctly different trace element patterns. Their Ba (fig. III-49A), Sr, Ti (fig. III-49C), and Zn contents and Ba/Rb (fig. III-50B), Ti/Fe, and Zn/Fe ratios are higher, and their K/Rb (fig. III-49D), K/Ba (fig. III-49B), Ca/Sr, and Rb/Sr ratios are lower than those in the granitic series. Also, the Rb, Ba, Pb, Zr, Mn, and Zn (fig. III-51A) contents and Mn/Fe, Zn/Fe (fig. III-51B), and Zn/Pb ratios of biotite in the rocks are significantly lower than those in biotites from rocks in the granitic series. The trace element concentrations and ratios in the granitoid series are similar to those in the volcanogenic metasedimentary rocks (see Viswanathan, 1971, *op. cit.*); accordingly, we conclude that the trace element pattern in the granitoid series has been inherited largely from these rocks, and that they formed from country rocks through metasomatism.

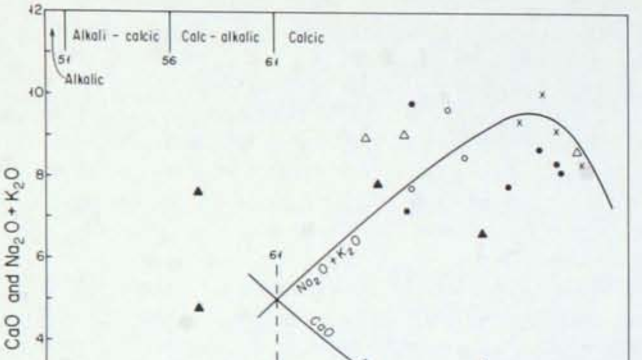


Figure III-46. Peacock variation diagram for plutonic rocks, western part of the Giants Range batholith (△, granodioritic series; O, granitic series, main bodies; x, granitic series, leucocratic satellitic phases; •, granitoid series; ▲, tonalitic series).

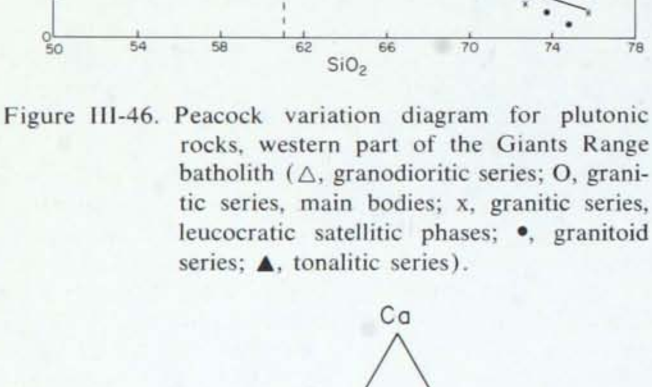


Figure III-47. Na-K-Ca variation diagram for plutonic rocks, western part of Giants Range batholith. See Figure III-46 for explanation of symbols.

The rocks of the tonalitic series have Rb (fig. III-50A), Ba (fig. III-49A), Pb, Sr (fig. III-49E), Zr, and Nb contents and K/Rb (fig. III-47D), K/Ba (fig. III-49B), Ba/Rb, Rb/Sr, Zr/Nb, Ti/Fe, Mn/Fe, and Zn/Fe ratios similar to those of the volcanogenic metasedimentary rocks. In the same way, the Rb, Ba, Pb, Zr, Nb, Mn, and Zn (fig. III-51A) contents and K/Ba, Ba/Rb, K/Pb, Ti/Zr, Ti/Nb, Ti/Fe, Mn/Fe, Zn/Fe (fig. III-51B), and Zn/Pb ratios of biotite from these rocks are similar to those in biotite from the metasedimentary rocks. From these close similarities in trace element chemistry, we infer an anatectic origin for these rocks.

To relate the chemical data on the granitic rocks from the western part of the batholith to experimental data, normative compositions (see Viswanathan, 1971, *op. cit.*, for method of computation) are plotted on Figure III-52, which shows the cotectics in the system Q-Ab-Or-H_2O at 500 and 5,000 atm P_{H_2O} (after Tuttle and Bowen, 1958; Luth and others, 1964) and a dashed line showing the loci of the ternary eutectic (or minimum) for this system as a function of increasing P_{H_2O} (increasing toward Ab) (Luth and others, 1964). Projected from the An corner of the Ab-An-Or-Q tetrahedron as a dotted line are also shown the loci of minimum temperature melts as a function of the Ab/An ratio in the system Ab-An-Or-Q-H_2O (from von Platen, 1965) at 2,000 atm P_{H_2O}. Three values for the Ab/An ratio are shown, 3, 5, and infinity.

Rocks of the granitoid series show considerable scatter, although four samples are closely grouped above the quartz-feldspar cotectic at 5,000 atm P_{H_2O}. In contrast, all but one of the samples from the granitic series plot outside this cotectic and toward the Ab-Or side. Samples from the more mafic granodioritic and tonalitic series also plot on or outside of the cotectic at 5,000 atm P_{H_2O}. The rather close grouping of samples from the three intrusive series suggests that each crystallized at moderately high water pressures. In contrast, samples from the granitoid series suggest that this series crystallized at somewhat lower water pressures.

In considering the genesis of the granitoid series, an origin through partial melting of the volcanogenic metasedimentary rocks should be evaluated. Assuming an average Ab/An ratio for the original metamorphic rocks of 2 or less, the first melt to form (see fig. III-52) should contain more Q and for the most part more Or at 2,000 atm P_{H_2O} (and probably at higher pressures) than the observed compositions of the rocks of the granitoid series. Hence, it is not likely that the granitoid series represents an initial melt derived from the partial melting of the metasedimentary rocks. As shown in Figure III-53, progressive potash feldspathization of volcanogenic metasedimentary rocks could account for their present composition.

The rocks that have been dated (Prince and Hanson, in press) are biotite- and muscovite-bearing granites from the granitoid series. The initial Sr^{87}/Sr^{86} ratio for these rocks, 0.7009 ± 0.0004, is indicative of derivation from a source having a low Rb/Sr ratio, such as the mantle. If these rocks are metasomatites, as suggested by us, the original volcanogenic metasedimentary rocks must have had a short residence time in the crust. This seems consistent with the model suggested in this volume that the volcanic-sedimentary-granitic cycle in the greenstone-granite complexes has a time span of only 50 to 100 million years.

Oxygen Isotope Data

$\delta(O^{18}/O^{16})_{SMOW}$ ratios of quartz from the granitic rocks of the batholith and associated volcanogenic metamorphosed sedimentary rocks are given in Table III-31. The values for the rocks of the granodioritic and granitic series, which range from 9.2 to 9.9 permil, are characteristic of granitic rocks of magmatic origin (Taylor, 1968), and support a magmatic origin for these rocks.

Table III-30. Average trace element concentrations, in ppm, and ratios of granitic rocks in Giants Range batholith (analysis by S. Viswanathan; method, X-ray fluorescence[1]).

	[2]Tonalitic series (3)	Granodioritic series (3)	Granitic series Main bodies (4)	Granitic series Satellitic phases (3)	Granitoid series (6)
				Trace elements (ppm)	
Rb	75	123	279	312	291
Ba	980	1293	520	313	972
Pb	65	54	59	70	62
Sr	527	1138	304	100	468
Zr	237	290	225	108	216
Nb	40	32	34	39	41
Ti	2820	1423	1573	425	1857
Mn	437	380	298	223	283
Zn	62	50	42	12	71
			Ratios		
K/Rb	246	260	141	148	131
K/Ba	22	22	92	156	60
Ba/Rb	12.3	10.5	1.9	1	3.3
K/Pb	293	585	686	655	573
Ca/Sr	52	14	38	73	17
Rb/Sr	0.14	0.11	0.9	3.1	0.7
K/Sr	36	24	138	452	194
Ba/Sr	2.95	1.1	1.7	3.1	2.1
Ti/Zr	11.6	4.9	7	3.9	7.3
Ti/Nb	71	66	46	11	38
Zr/Nb	5.9	11	6.5	2.9	6.5
1000 Ti/Fe	92	51	43	40	64
1000 Mn/Fe	13	13	18	23	15
1000 Zn/Fe	2	2.8	2.7	2	4.3
Zn/Pb	1	1.2	0.8	0.2	1.2

[1] Data on standards used, precision, and ranges given in Viswanathan (1971, unpub. Ph.D. thesis, Univ. Minn.)
[2] Numbers in parentheses indicate number of samples analyzed

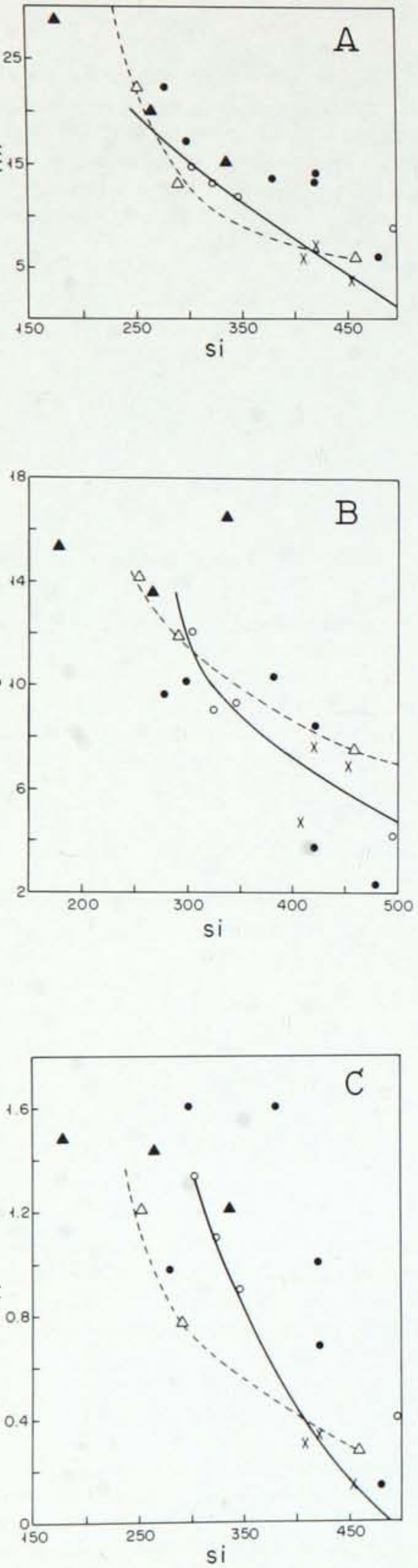


Figure III-48. Niggli variation diagrams. A, fm versus si; B, c versus si; C, ti versus si. See Figure III-46 for explanation of symbols.

The $\delta(O^{18})_{Quartz}$ values (10.7-11.4 permil) for rocks of the granitoid series are significantly higher than the values for the magmatic rocks, and are intermediate between the ratios for magmatic rocks and those for the associated metasedimentary rocks. These data are consistent with a metasomatic origin for the granitoid series, in which the O^{18}-rich metasedimentary rocks were modified by lighter fluids of a magmatic source.

The $\delta(O^{18})_{Quartz}$ values for rocks in the tonalitic series (9.4 to 10.1 permil) also are typical of magmatic rocks, but are equally consistent with an anatectic origin, as suggested by the trace element data. For a discussion of this problem the reader is referred to Viswanathan (1971, *op. cit.*).

The available data on the leucocratic phases of the magmatic granodioritic and granitic series indicate that these late-stage, satellitic phases have $\delta(O^{18})_{Quartz}$ values nearly identical to the rocks in the main bodies. Inasmuch as some of the leucocratic phases cut the O^{18}-rich metasedimentary country rocks, they provide a tool to test the extent of oxygen isotopic exchange between the melts and metasedimentary country rock. Analyses of a satellitic body of leucogranite that is an offshoot of unit 7 (fig. III-44) gave a value of $\delta(O^{18})_{Quartz}$ of 9.2 permil, whereas the adjacent metasedimentary country rock gave a value of 12.4 permil. These results clearly indicate that there was little oxygen communication between the melt and its O^{18}-rich country rock, and suggest that the oxygen isotopic compositions of the rocks were affected very little by supergene or other heavy aqueous fluids. Very likely, the batholith was emplaced in a moderately dry environment, as compared with some of the younger, higher level plutonic bodies (see Shieh and Taylor, 1969).

EASTERN PART

The eastern part of the Giants Range batholith includes the granitic rocks that extend from the vicinity of Virginia to the eastern exposed extremity about 15 miles east of Ely, where they are overlain and metamorphosed by the Keweenawan Duluth Complex. Judged from gravity data (Ikola, 1968b, 1970), the granitic rocks extend beneath the Duluth Complex at least as far east as Forest Center, a distance of about 15 miles. Exposures are excellent on the narrow ridge of hills that forms the Giants Range and in the area northeast of Babbitt. In contrast, exposures are sparse in the lowland north of the Giants Range and consist generally of small, flat, glaciated surfaces. Two areas within the eastern part of the batholith have been mapped recently—the Embarrass-Babbitt area, north of the East Mesabi district (Griffin, 1967, *op. cit.*, 1969; Griffin and Morey, 1969), and the Gabbro Lake 15-minute quadrangle, at the eastern exposed extremity of the batholith (Green and others, 1966; Green, 1970a). The remaining areas have been examined in a reconnaissance manner for preparation of the Hibbing (Sims and others, 1970) and Two Harbors Sheets (Green and others, Minn. Geological Survey, unpub. map) of the Geologic Map of Minnesota.

General Features

The eastern part of the batholith is composed dominantly of medium- or coarse-grained, commonly porphyritic,

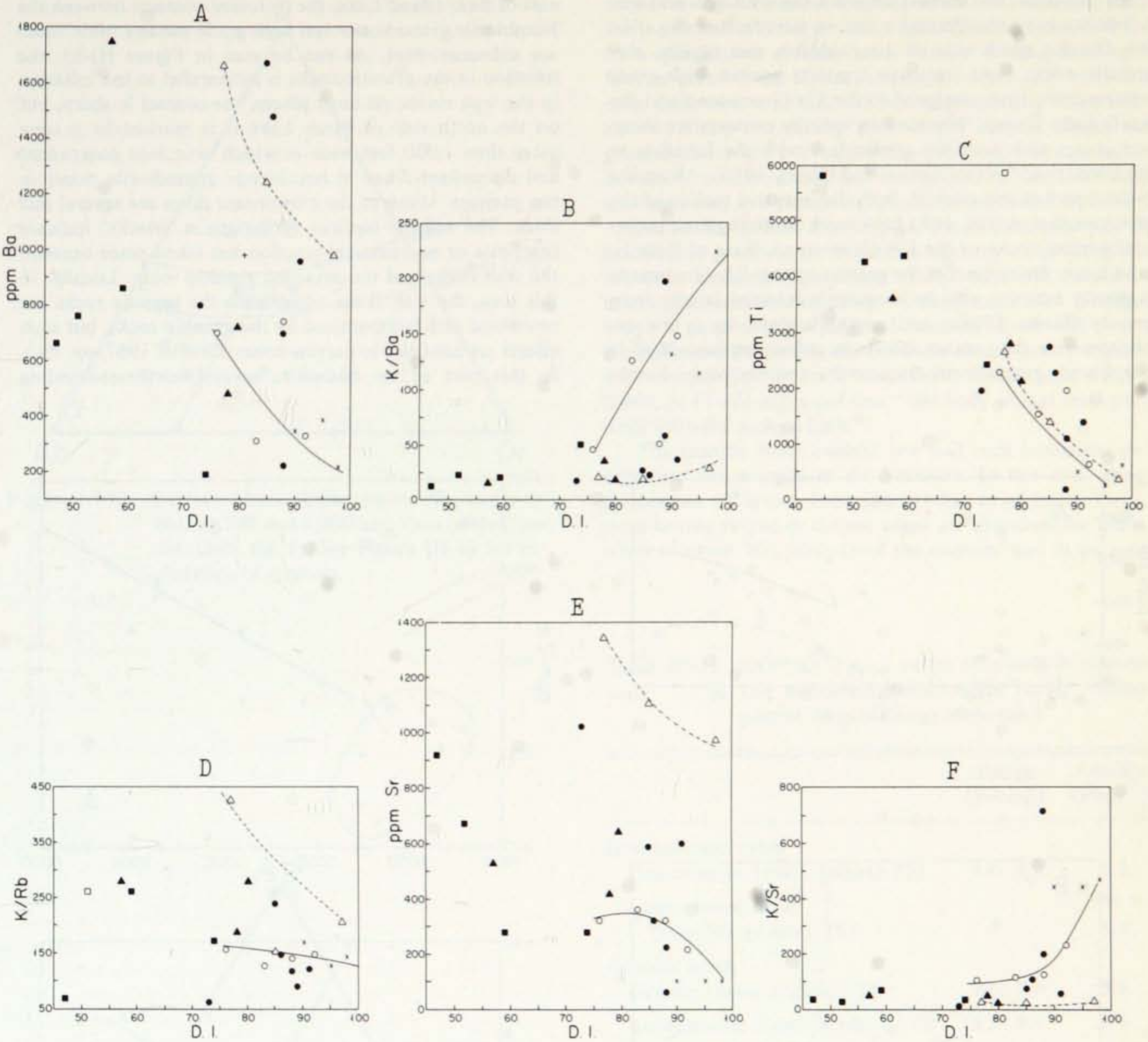


Figure III-49. Variation diagrams for whole rocks. A, Ba versus differentiation index; B, K/Ba versus differentiation index; C, Ti versus differentiation index; D, K/Rb versus differentiation index; E, Sr versus differentiation index; F, K/Sr versus differentiation index. The differentiation index is the sum of normative Q, Or, and Ab (Thornton and Tuttle, 1960). ■, metasedimentary rocks. See Figure III-46 for explanation of other symbols.

hornblende-bearing plutonic rocks that range in composition from adamellite to diorite and are dominantly adamellite and granodiorite. The rocks comprise the intermediate phase of Allison (1925), and with an increase in quartz content grade locally into the Embarrass phase (Leith, 1903; Allison, 1925). The quartz content is less than that in "typical" granitic rocks, which have close to the eutectic amount of approximately 30-35 percent quartz (Chayes, 1951), and in some rocks is less than 10 percent. For the most part, the various facies appear to be intergradational, and do not show clear intrusive relations to one another. Accordingly, the rocks probably constitute a heterogeneous pluton that is compositionally complex. Dikes and small bodies of fine- to medium-grained biotite granodiorite cut the hornblende adamellite-granodiorite, as do small dikes of pegmatite and aplite.

In this area, the contact between the batholith and the wall rocks is in part intrusive and in part faulted (fig. III-54). On the north side of the batholith east of Ely, the granitic rocks have intrusive contacts against high-grade metamorphic rocks assigned to the Ely Greenstone and the Knife Lake Group. Where observed, the contacts are sharp and clean, and generally concordant with the foliation in the country rocks (see Green and others, 1966). Along the eastern part of this contact, both the intrusive rocks and the metamorphosed wall rocks have been faulted against greenschist-facies rocks of the Ely Greenstone. West of Bear Island Lake, the contact of the granite against Ely Greenstone is poorly exposed, and its location is inferred largely from gravity (Ikola, 1968a) and magnetic data. In a few exposures near the contact, however, subconcordant dikes of hornblende granodiorite transect the country rocks. South-east of Bear Island Lake, the intrusive contacts between the hornblende granodiorite and high-grade metamorphic rocks are subconcordant. As can be seen in Figure III-55, the foliation in the granitic rocks is subparallel to the foliation in the wall rocks. At most places, the contact is sharp, but on the north side of Birch Lake it is marked by a zone more than 1,000 feet wide in which abundant concordant and discordant dikes of hornblende granodiorite occur in the gneisses. Many of the concordant dikes are several feet thick. The angular outlines of the gneiss "blocks" indicate that little or no chemical reaction has taken place between the wall rocks and the invading granitic rocks. Locally, in this area, the wall rocks adjacent to the granitic rocks are coarsened and feldspathized by the granitic rocks, but such effects are confined to narrow zones (Griffin, 1967, *op. cit.*). In this part of the batholith, several northeast-trending

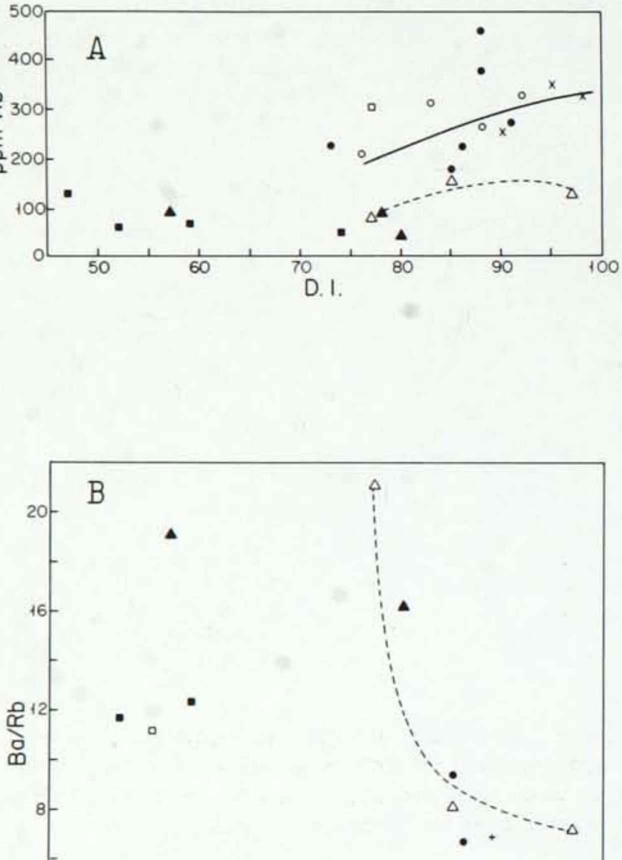


Figure III-50. Variation diagrams for whole rocks. A, Rb versus differentiation index; B, Ba/Rb versus differentiation index (see fig. III-46 for explanation of symbols).

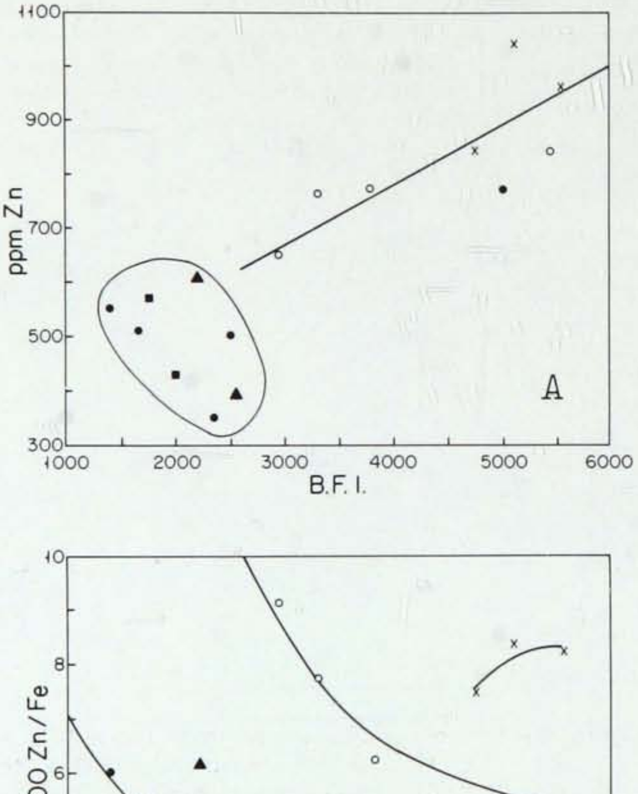


Figure III-51. Variation diagrams for biotite. A, Zn versus biotite fractionation index; B, 1000 Zn/Fe versus biotite fractionation index. (The biotite fractionation index is the amount of Mn, in ppm, in biotite.) See Figure III-46 for explanation of symbols.

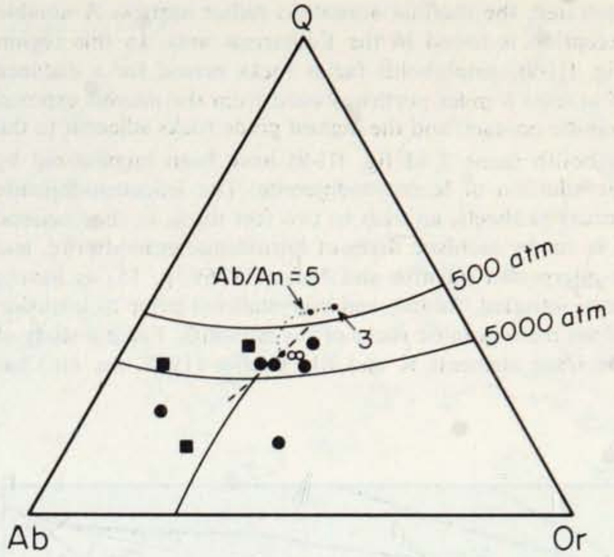


Figure III-52. Q-Ab-Or plane of the system Ab-An-Or-Q-H_2O at 500 and 5,000 atm P_{H_2O} (after Condie, 1969, fig. 8). See Figure III-46 for explanation of symbols.

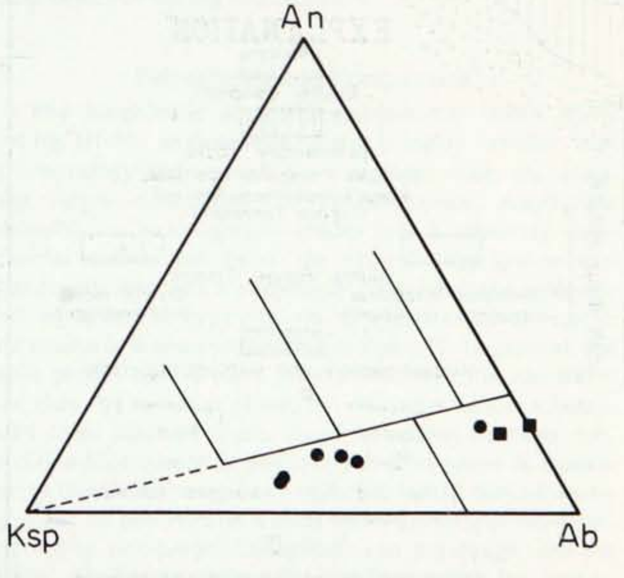


Figure III-53. K-feldspar-Ab-An variation diagram, showing rocks of the granitoid series and their presumed parent rocks (after O'Connor, 1965). •, granitoid series; ■, volcanogenic metasedimentary rocks.

faults having left-lateral displacements cut and offset both the wall rocks and the granitic rocks. The faults are marked by a conspicuous red alteration and by milky quartz veinlets; adjacent to the faults, the granitic rocks are cataclastically deformed, with little or no accompanying recrystallization. Along the eastern part of the Mesabi range, the granitic rocks are overlapped by the Middle Precambrian Biwabik Iron-formation, and northeast of Birch Lake they are transected and metamorphosed by the Duluth Complex.

In the Embarrass area (fig. III-54), a satellitic body of hornblende granodiorite occurs in country rocks along the northeast-trending Waasa fault. The granodiorite has a pronounced northwest-trending, steeply-inclined foliation which is parallel to the regional foliation in the wall rocks; this foliation, in turn, is cut by shear zones, now silicified, that are parallel to the Waasa fault. Griffin and Morey (1969, p. 15-16) suggested that "this body was at least partially intruded along a fault."

The granitic rocks contain few wall rock inclusions except near the margins of the batholith. In one area along the contact on Spruce Lake (fig. III-55), amphibolite inclusions having frayed or diffuse edges are abundant for a distance of about 200 yards from the contact; and in an area

Table III-31. $\delta(O^{18}/O^{16})_{SMOW}$ ratios of quartz in granitic and associated metamorphic rocks, western part of Giants Range batholith.[1]

	Range (Permil)	Average (Permil)
Granodioritic series		
Granodiorite (main bodies) (2)	9.4- 9.9	9.7
Leucogranodiorite (satellitic phases) (1)		9.8
Granitic series		
Granite (main bodies) (2)	9.5- 9.6	9.6
Leucogranite (satellitic phases) (2)	9.2- 9.6	9.4
Granitoid series		
Granite (5)	10.7-11.4	10.9
Aplogranite (1)		10.2
Partially granitized volcanogenic metasedimentary rocks (3)	10.4-12.5	11.1
Tonalitic series		
Quartz tonalite (2)	9.4-10.1	9.8
K-feldspathized tonalite (1)		12.1
Volcanogenic metasedimentary rocks (4)	11.1-12.9	12.2
Migmatitic gneiss (5)	8.9- 9.9	9.5

[1] Analyses by S. Viswanathan, in E. C. Perry's laboratory, Univ. Minn. Number in parentheses after rock name is number of samples analyzed

near the east end of the Giants Range (NE¼ sec. 18, T. 60 N., R. 12 W.) there are ovoid inclusions of greenstone, some containing relict pillows (Griffin, 1967, *op. cit.*). Southeast of South Farm Lake, in the Gabbro Lake quadrangle, local, large masses of mafic gneiss and granoblastic rocks occur in the dominant hornblende adamellite (Green, 1970a, p. 52).

In general, the wall rocks adjacent to and included within the batholith are medium- or high-grade metamorphic rocks that have mineral assemblages characteristic of the amphibolite facies (Turner, 1968). As discussed above, the contacts are locally complicated by faulting and related alteration effects, but where intrusive contacts can be demonstrated, the thermal aureole is rather narrow. A notable exception is found in the Embarrass area. In this region (fig. III-9), amphibolite-facies rocks extend for a distance of at least 8 miles northwestward from the nearest exposed granitic contact, and the highest grade rocks adjacent to the batholith (zone 3 of fig. III-9) have been migmatized by the addition of leucotrondhjemite. The leucotrondhjemite occurs as sheets, an inch to two feet thick, in the gneisses; it is cut by aschistic dikes of hornblende granodiorite, and is interpreted (Griffin and Morey, 1969, p. 15) as having been intruded, folded, and recrystallized prior to intrusion of the main granitic rocks of the batholith. From a study of the trace elements K and Rb, Griffin (1967, *op. cit.*) has

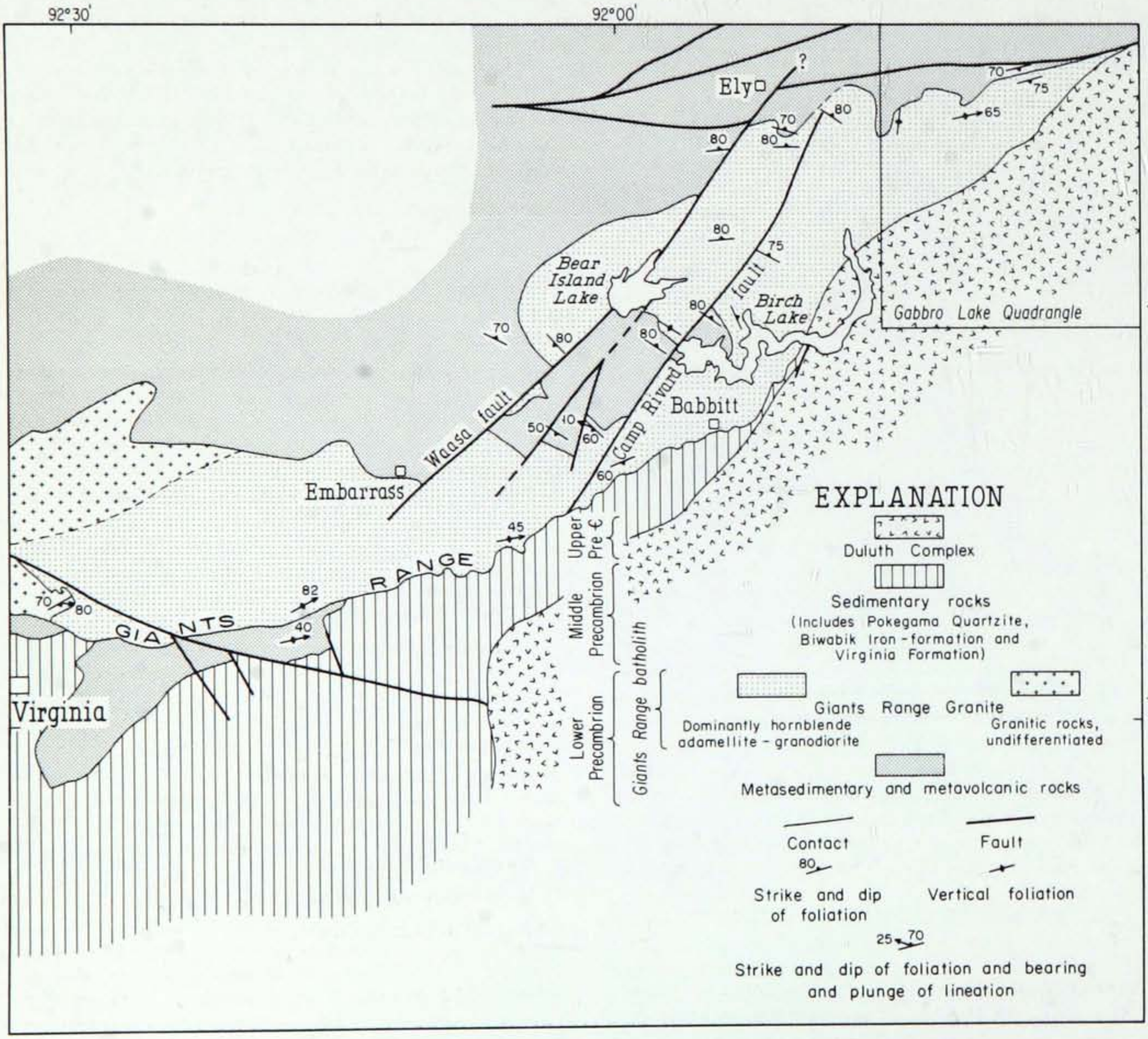


Figure III-54. Geologic sketch map of the eastern part of the Giants Range batholith. Compiled by P. K. Sims, 1971, from the following sources: Green and others (1966); Sims and others (1968b, 1970).

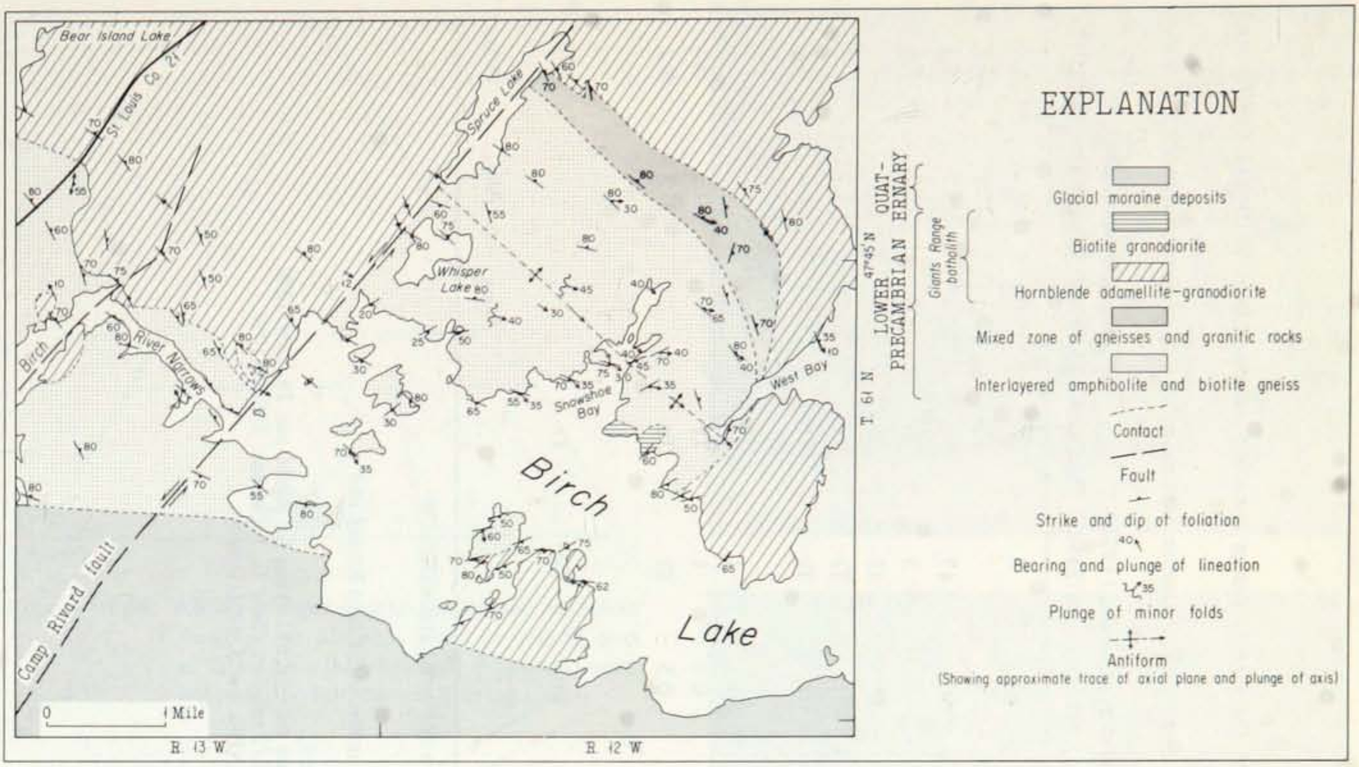


Figure III-55. Geologic map of Birch Lake area. Modified from W. L. Griffin (1967, unpub. Ph.D. thesis, Univ. Minn.).

concluded that the leucotrondhjemite is related to the batholithic rocks and either was derived from the Ely Greenstone by a process of partial melting or was derived through modification of granitic magma with the wall rocks before or during emplacement.

Petrography and Composition

The hornblende adamellite-granodiorite (table III-32 and fig. III-56), as described herein, is highly variable, both in mineralogy and texture. On a regional scale, the dominant facies—a medium- or coarse-grained, porphyritic adamellite or granodiorite—grades into a relatively equigranular facies that lacks the typical large microcline phenocrysts and which commonly contains mafic schlieren and segregations. Typically, this phase is relatively mafic, and quartz is inconspicuous megascopically. In general, the mafic phases have a more pronounced foliation and lineation than the common phase, but the latter differs substantially from place to place. Another variant contains conspicuous blue quartz as phenocrysts. This phase is equivalent to the Embarrass granite of Leith (1903). Similar variations can be observed on a local scale; gradations from porphyritic to non-porphyritic phases can be abrupt, and the quartz content can differ substantially across a few inches.

In the Gabbro Lake 15-minute quadrangle, Green (1970a) distinguished two facies, which he informally named the Farm Lake and Clear Lake facies from type areas in the quadrangle. The Farm Lake facies constitutes most of the batholith within the limits of the Gabbro Lake quadrangle, and is the common facies in adjacent areas, including the Embarrass-Babbitt area mapped by Griffin (1967, *op. cit.*). The Clear Lake facies occurs in a wedge-shaped area, about 6 miles long, in the easternmost extremity of the batholith (Green and others, 1966); it is comparable to the mafic phases described by Griffin and Morey (1969, p. 30-31).

The Farm Lake facies in the Gabbro Lake area (Green, 1970a, p. 54-56) is mainly hornblende adamellite and monzonite (table III-32). The most abundant type is pink, porphyritic, and medium grained (fig. III-57A); typically, it has a primary (flow) foliation and generally a lineation formed by alignment of equally distributed hornblende prisms and microcline phenocrysts. The primary foliation and lineation are fairly consistent in attitude over areas a mile or more across. The rock has a hypidiomorphic-granular texture. Microcline phenocrysts generally range in cross section from ½ × 1 cm to 1 × 2 cm; the quartz content generally is 15 to 20 percent, with a maximum of about 25 percent. Locally, plagioclase also occurs as phenocrysts, as does quartz and hornblende.

In the Gabbro Lake area, non-porphyritic varieties of the Farm Lake facies (Green, 1970a, p. 55-56) are intimately intermixed with the porphyritic rocks, without clear intrusive relationships. They tend to have less quartz than the typical facies, and they contain substantial amounts of intermediate and mafic schlieren, segregations, and xenoliths (?). These mafic zones have sharp to diffuse contacts with the surrounding porphyritic and non-porphyritic granitic rocks.

Table III-32. Modes, in volume percent, of typical plutonic rocks in the eastern part of the Giants Range batholith (after Green, 1970 and Griffin and Morey, 1969).

	Hornblende adamellite-granodiorite												
	Farm Lake facies of Green (1970)						Clear Lake facies of Green (1970)				Biotite granodiorite		
	M-7272	M-7504	M-7599	M-12530a	M-12531a	M-12885	M-7300	M-7460	M-7459	M-7669	M-12704	M-12898	M-12957b
Quartz	10	25	Tr	17.3	12.7	5.6	3	15	3	28	25.9	24.4	25.6
K-spar	34	30	4	14.4	14.2	19.0	30	15		30	20.3	25.2	26.1
Plagioclase	40	34	74	62.4	61.8	58.3	50	50	47	35	48.5	55.7	45.1
Biotite	1	Tr	5			5.4		2	3	6	4.0	1.5	8.1
Hornblende	10	6	10		9.8	8.8	15	15	45				
Chlorite	2	4	Tr	5.4	Tr		Tr	1	Tr	Tr		1.7	
Muscovite	Tr	Tr	Tr				Tr	Tr	Tr	Tr	1.2	1.8	4.1
Epidote	1.5	Tr		0.2	1.1		Tr	1			0.3	0.9	0.9
Apatite	Tr	0.5	2			0.5	0.5	Tr	1				
Sphene	1	Tr	2			0.8	0.5	0.5	1	Tr		0.2	0.2
Opaques	Tr	0.5	3	0.3	Tr	0.8	1	0.5	Tr	Tr		0.2	0.2
Allanite	Tr												
Monazite			Tr										
Zircon	Tr		Tr				Tr	Tr	Tr	Tr			
Composition of plagioclase				An_{16}	An_{16}	An_{17}					An_{23}	An_{24}	An_{24}

M-7272—Pink, medium- to coarse-gr., flow-foliated, porphyritic hornblende-adamellite; S shore of Farm Lake, 62N/11W
M-7504—Medium-gr., porphyritic adamellite; SE¼ sec. 1, 62N/11W
M-7599—Medium-gr., weakly porphyritic biotite-hornblende diorite; SE¼ sec. 29, 62N/11W
M-12530a—Pink, coarse-gr., porphyritic granodiorite; adjacent to St. Louis Co. hwy. 21, near junction with Camp Rivard rd.
M-12531a—Pink, medium-gr., weakly porphyritic granodiorite; One Pine L. area
M-12885—Gray, medium- to coarse-gr., flow-foliated, porphyritic syenodiorite; north of Birch L.
M-7300—Medium-gr., allotriomorphic-granular hornblende monzonite; N. Kawishiwi River, one mile SE of Uranus L., NW¼ sec. 26, 63N/10W
M-7460—Medium-gr. hornblende granodiorite with mortar texture; S. Kawishiwi River, about 2/3 mi. S of Clear L., NE¼ sec. 5, 62N/10W
M-7459—Medium-gr. hornblende diorite; island in bay of S. Kawishiwi River, S of Clear L., NE¼ sec. 5, 62N/10W
M-7669—Fine- to medium-gr. biotite adamellite, SW cor. Bruin L., N½ sec. 18, 62N/10W
M-12704—Biotite granodiorite from small intrusive body in hornblende granodiorite, on Birch L.
M-12898—Biotite granodiorite dike in hornblende granodiorite
M-12957b—Biotite granodiorite dike in gneisses, parallel to NE shearing

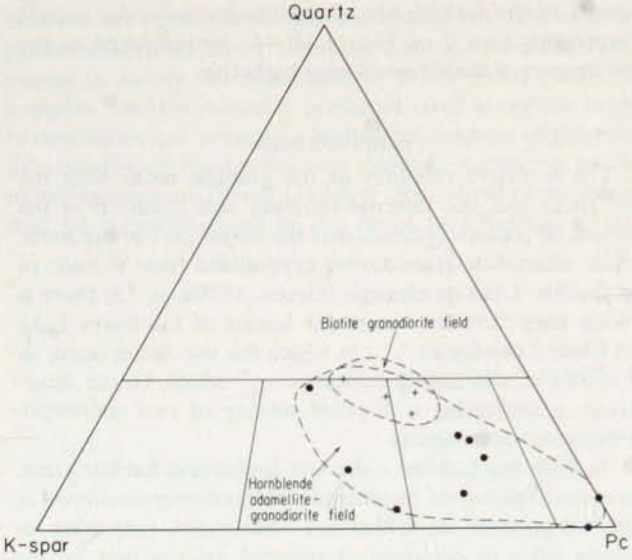


Figure III-56. Alkali feldspar-quartz-plagioclase variation diagram for plutonic rocks in eastern part of Giants Range batholith. +, biotite-bearing rocks; •, hornblende-bearing rocks.

In the Embarrass-Babbitt area (Griffin and Morey, 1969, p. 29-30), rocks equivalent to the Farm Lake facies are dominantly granodioritic in composition; more mafic facies are syenodiorite or diorite (table III-32).

The Clear Lake facies of Green (1970a, p. 57-59) lacks distinct microcline phenocrysts, is finer grained, and has less quartz than the Farm Lake facies (table III-32). It resembles typical syenites. According to Green, the rock is pink, fine to medium grained and has an allotriomorphic-granular texture (fig. III-57B). Plagioclase of albite to sodic oligoclase composition commonly occurs as phenocrysts, and the microcline is finely microperthitic. Augite occurs in some mafic facies, and commonly is rimmed by hornblende. The Clear Lake facies has varying degrees of foliation and lineation given by aligned hornblende. The Clear Lake facies is substantially more inhomogeneous than the modes in Table III-32 would indicate. In addition to variations in quartz and feldspar content, there are many local variations in mafic mineral content, giving abundant mafic and intermediate varieties, some of which are sufficiently large to be mapped at a scale of 1:31,680 (see Green and others, 1966).

The biotite granodiorite that cuts the hornblende adamellite-granodiorite is homogeneous, both compositionally and texturally (table III-32 and fig. III-56). It is a pink or pinkish-gray, fine- to medium-grained, equigranular, massive to weakly foliated rock that has a small range in modal composition. Presently available mapping indicates that it is widely distributed in the eastern part of the batholith as dikes, sills, and small irregular plutons that cut all facies of the hornblende adamellite-granodiorite. The largest mapped body in the Gabbro Lake quadrangle (Green and others, 1966) has approximate dimensions of 5,000 × 2,000 feet. It also occurs sporadically in the wall rocks, especially in the Birch Lake area (fig. III-55; Griffin and Morey, 1969).

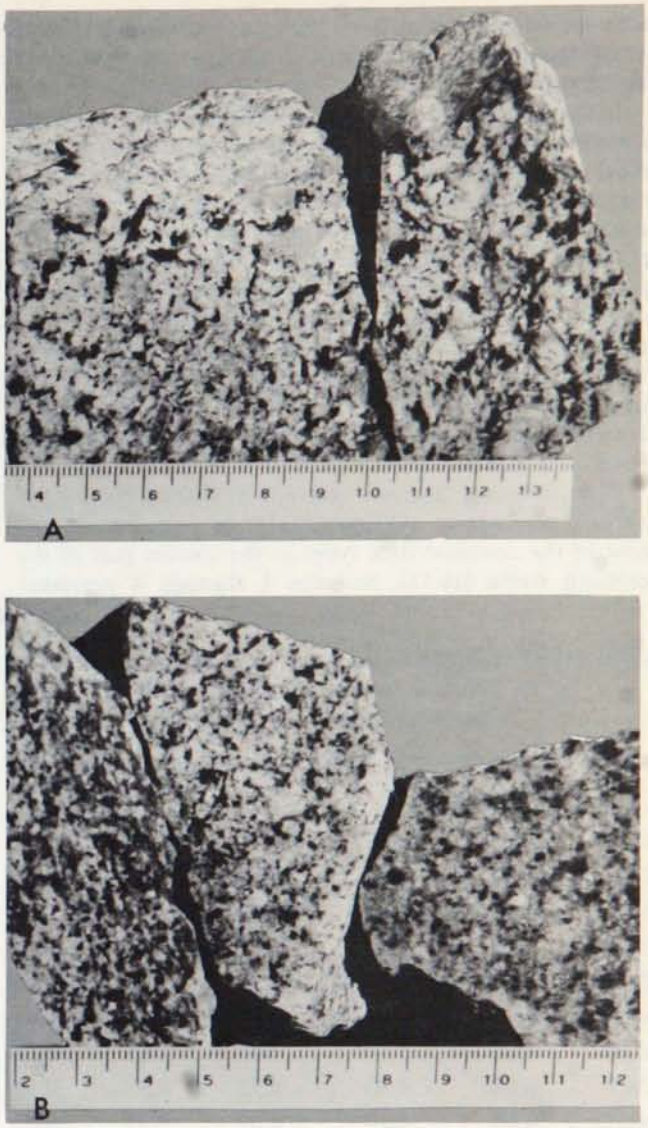


Figure III-57. Textures characteristic of igneous rocks in eastern part of Giants Range batholith (scale in cms). A, Farm Lake facies; B, Clear Lake facies (from Green, 1970a).

Aplite and pegmatite dikes cut all the phases described above. According to Green (1970a, p. 60), ". . . many of the aplites are younger than at least some of the folding and shearing, but some have been broken and pulled apart after solidification. Rarely, pegmatites are cut by aplites, but in most outcrops the two appear to be more or less contemporaneous." The aplites are allotriomorphic-granular and fine- to medium-grained. They contain about equal proportions of quartz, oligoclase, and microcline, and have some hornblende and/or biotite. The pegmatites are generally granitic or adamellitic in composition, and contain quartz, albite, and microcline perthite as major components and biotite, secondary sericite, opaque oxides, zircon, and apatite as lesser components.

In the Gabbro Lake area (Green, 1970a, p. 60), a variety of mafic and intermediate dikes, from about 10 inches to 5 feet thick, cut the Farm Lake and Clear Lake facies of the batholith. Green interpreted them as cogenetic with the batholithic rocks inasmuch as many of them are cut by aplites and pegmatites. Interestingly, these rocks—which are mainly hornblende diorite and syenodiorite—and the red granitic rocks described from the western part of the Gabbro Lake quadrangle, near the Fernberg road (Green, 1970a, p. 61-62), resemble the small syenite and related lamprophyric bodies in the western part of the Vermilion district (see Sims and Mudrey, this chapter), and may be coeval with them.

New chemical analyses are not available, but earlier analyses obtained by Allison (1925) are representative of some of the common rock types in the eastern part of the batholith (table III-33). Samples 1 through 4 represent variants of the hornblende adamellite-granodiorite; sample 5 represents unit 7 of Figure III-44, distinguished in the western part of the Giants Range batholith.

Table III-33. Chemical analyses, in weight percent, of plutonic rocks in eastern part of Giants Range batholith (after Allison, 1925, table 2).

	1	2	3	4	5
SiO_2	54.57	60.42	66.31	72.39	71.45
TiO_2	0.20	0.74	0.49	0.49	0.13
Al_2O_3	17.91	12.28	12.70	13.39	12.99
Fe_2O_3	2.68	5.57	1.98	0.18	2.44
FeO	4.80	4.32	2.66	1.44	1.78
MnO	0.14	0.06	0.04	0.01	0.04
MgO	3.86	3.40	3.01	1.61	0.84
CaO	6.01	4.95	4.77	0.92	0.65
Na_2O	4.69	4.50	5.03	5.35	3.03
K_2O	2.34	2.02	2.33	2.42	4.79
H_2O+	1.68	1.42	0.75	1.22	1.24
H_2O-	0.06	0.08	0.05	0.02	0.10
CO_2	0	0	n.d.	n.d.	n.d.
ZrO_2	0.07	0.07	0.06	0.15	0.10
P_2O_5	0.11	0.20	0.13	0.10	0.04
S	0.04	0.02	Tr	0	Tr
Cr_2O_3	0	0	0.01	0	0.02
BaO	0.11	0.07	0	0	0
Total	99.27	100.12	100.32	99.69	99.64

Analyses 1 and 3 by I. S. Allison; 2, 4, and 5 by R. J. Leonard

1. Monzonite, NW¼ sec. 7, 61N/12W (a facies of hornblende adamellite-granodiorite)
2. Hornblende-granite (granodiorite), Hinsdale quarry, sec. 17, 59N/17W
3. Hornblende-biotite granite (granodiorite), sec. 28, 59N/17W
4. Embarrass granite (granodiorite), from Embarrass station, NW¼ sec. 25, 60N/15W
5. Biotite granite, sec. 31, 59N/18W (unit 7 of Fig. III-44)

Emplacement

The intrusive relations of the granitic rocks with the wall rocks and the internal foliation and lineation in the margins of plutons indicate that the major part of the hornblende adamellite-granodiorite crystallized from a melt. In the Gabbro Lake quadrangle (Green, 1970a, p. 53) there is a wide zone between the major bodies of the Farm Lake and Clear Lake facies ". . . in which the two facies occur in an intimate, alternating mixture . . ." which Green interpreted as indicating a physical mixing of two contemporaneous viscous magmas.

In both the Gabbro Lake and Embarrass-Babbitt areas, dikes and veinlets of hornblende adamellite-granodiorite in adjacent gneisses are folded and boudinaged, indicating intrusion prior to cessation of regional deformation. In the Embarrass-Babbitt area, there is some evidence that granite emplacement was controlled locally along the margins of the batholith by northeast-trending faults, implying emplacement at relatively shallow depths. These and other relations indicate that the granitic rocks are late syntectonic.

Metamorphism by Duluth Complex

Green (1970a, p. 84) has described the metamorphic effects of the Duluth Complex on the granitic rocks in the northeastern part of the batholith. He stated, "In some specimens the rock is recrystallized enough to change the primary hypidiomorphic texture of the leucocratic minerals to xenomorphic, although phenocrysts are preserved; in others, some of the newly formed ferromagnesian minerals are poikiloblastic. The plagioclase (oligoclase) typically contains antiperthitic blocks and patches of microcline; it appears as though the feldspars have reacted with each other as the temperature was raised, followed by exsolution on cooling. The K-feldspar (either orthoclase or microcline) is microperthitic; myrmekite is found between the two feldspars in some areas, and in others a replacement intergrowth is found with corroded plagioclase cores partly replaced by orthoclase in optical continuity. The quartz is poikiloblastic in one sample. One specimen only one or two hundred feet from the contact in the South Kawishiwi River shows the typical primary ferromagnesian assemblage hornblende-biotite-magnetite, but the hornblende is pale and contains tiny magnetite grains. In the three other samples studied the ferromagnesian assemblage is now pale hornblende (with magnetite grains)-augite-hypersthene-biotite-magnetite. The hypersthene contains exsolution lamellae and is probably inverted from pigeonite. Some of the hypersthene is partly altered by retrograde reactions to a mixture of biotite and actinolite. In one granite outcrop (the small hill just north of the Spruce road northwest of Filson Creek in NE¼ sec. 24, T. 62 N., R. 11 W.) a trace of copper sulfides has been introduced into the rock, most probably from the gabbro that once immediately overlay it.

"The mineral assemblages in the contact aureole of the Duluth Complex are characteristic of the transitional zone between the amphibolite, the hornblende-hornfels, and the pyroxene-hornfels facies. Many of the rocks have too many phases to satisfy the requirements of the phase rule, and probably contain minerals produced over a certain range of conditions and preserved in the rock because of the relative rapidity of the heating and cooling. According to the recent estimate of Turner (1968, p. 366), these rocks would thus have recrystallized in the range of 600-675° C and under pressures of roughly 1.5 to 2.5 kb, equivalent to a depth of from 15,000 to 30,000 feet. This overburden would have consisted principally of the Middle Precambrian sediments that were still preserved by Keweenawan time, plus the Keweenawan lavas that had been erupted by the time of this particular phase of intrusion of the Duluth Complex, plus the already intruded part of the complex. . . ."

SYENITIC PLUTONS AND ASSOCIATED LAMPROPHYRES

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

In addition to the major batholiths, there are many small bodies of general syenitic composition that intrude the supracrustal rocks in the Vermilion district and adjacent areas to the west. Lamprophyres are spatially associated with some of these bodies. Previously, the larger and better known syenitic bodies in this group, such as the Snowbank and Kekekabic stocks, were considered satellitic intrusions of the Giants Range Granite; and some of the smaller, more mafic bodies were considered early border phases of the Vermilion Granite (Grout, 1925b). Instead, these bodies seem to comprise a distinct, separate family of syenitic rocks that are genetically and temporally related, for the rocks have many features in common and differ from those in the major batholiths in mineralogy, texture, and structure, as well as in size.

SNOWBANK AND KEKEKABIC STOCKS

Two small composite stocks of general syenodiorite composition occur in the eastern part of the Vermilion district, in northern Lake County. The larger one, a roughly circular body having dimensions of about 5 by 4 miles, is exposed on the shores and islands of Snowbank Lake (fig. III-58). The smaller one, an elliptical body about 4 miles long and a maximum of 1½ miles wide, is exposed in the vicinity of Kekekabic Lake (fig. III-60), near the eastern end of the district.

The stocks were first described by U.S. Grant in reports of the Minnesota Geological and Natural History Survey, and one of the reports (*in* Winchell, 1893b) remains a useful reference to the geology of the Kekekabic stock. A later paper by Stark (1927) described the internal structure of the Kekekabic stock. Subsequent to the early studies, the Snowbank stock was mapped by Sanders (1929), who emphasized the petrography of the body, and later by Balk and Grout (1934), who emphasized its structure and mode of emplacement. More recently, Hanson and Gast (1967) studied the effects of thermal metamorphism by the Duluth Complex on mineral ages of the rocks in the Snowbank stock.

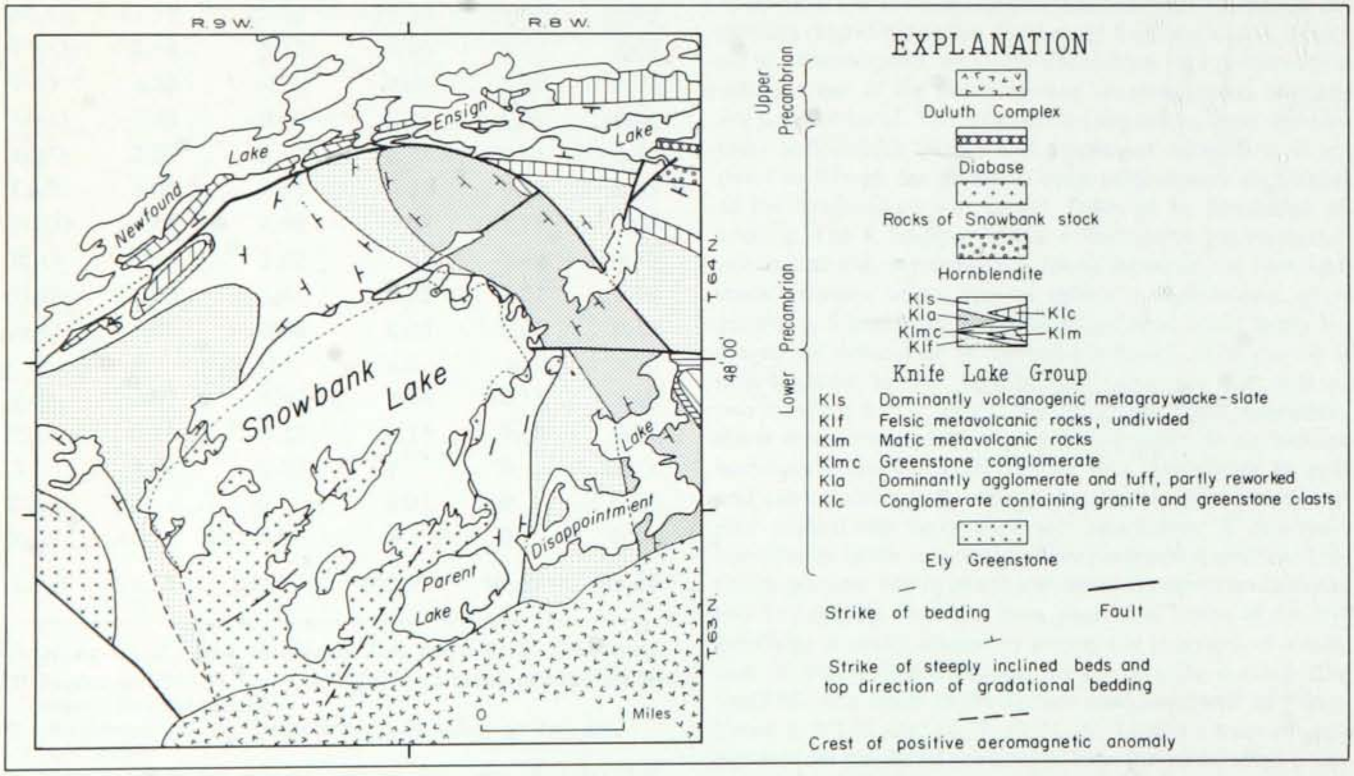


Figure III-58. Generalized geologic map of Snowbank Lake area. Compiled by P. K. Sims, 1971, from the following sources; Gruner, 1941, pl. 1; Sanders, 1929; Balk and Grout, 1934; Green, 1970a, pl. 1; Gibson, 1934, unpub. Ph.D. thesis, Univ. Minn.; Dutton, 1931, unpub. Ph.D. thesis, Univ. Minn.

Geologic Setting and Age

Both the Snowbank and Kekekabic stocks were emplaced in rocks of the Knife Lake Group. The Snowbank stock (fig. III-58) cuts an interlayered, dominantly volcanogenic succession of slate, graywacke, tuff, agglomerate, and conglomerate (Gruner, 1941, pl. 1), which is interpreted as being a part of the older succession of the Knife Lake Group (see fig. III-8). These rocks are steeply inclined and folded on dominant northwest-trending axes. The Kekekabic stock is interpreted to cut rocks, on the other hand, belonging to the younger succession of the Knife Lake Group; this part of the group contains epiclastic deposits derived from the Saganaga Tonalite as well as substantial quantities of volcaniclastic debris. These rocks are folded on northeast-trending axes; probably, this generation of folding is younger than the northwest-trending folds in the Snowbank Lake area.

The Snowbank stock is interpreted from Rb-Sr and K-Ar ages on biotite from the stock and from K-Ar ages on muscovite from associated pegmatite in the adjacent country rock (table III-34; Hanson and Gast, 1967, p. 1122) to have been emplaced about 2,700 m.y. ago. The Kekekabic stock has not been dated, but presumably also is about the same age.

The Snowbank and Kekekabic stocks, as well as the adjacent rocks, were intruded by the Duluth Complex about 1,050 m.y. ago (Hanson and Gast, 1967). Hanson and Gast (1967, p. 1122-1123) postulated that a wedge of the complex covered the Snowbank stock (fig. III-59), and as a consequence, the mineral ages of the stock were profoundly altered and the mineralogy was modified. Within 1.2 miles of the contact at the present surface, the Rb-Sr and K-Ar ages of the biotite from rocks in the 2,700 m.y.-old stock are between 1,000 and 1,200 m.y., indicating that the biotites lost essentially all their radiogenic daughters during the time of the contact metamorphism. The biotite ages increase rather regularly outward away from the contact, and at a distance of about 3 miles from the contact they retain their approximate original age.

Snowbank Stock

The Snowbank stock is a composite intrusive body that ranges in composition from diorite to granodiorite. Sanders (1929, fig. 1) distinguished and mapped four main facies,

Table III-34. Rb-Sr and K-Ar mineral ages for the Snowbank stock (after Hanson and Gast, 1967).

Field No.	Sample	D[1]	Rb-Sr age in m.y.[2]	K-Ar age in m.y.
M 7058	Biotite	3.1	2,720	2,590
M 7045A	Muscovite	3.1	2,640	2,620
M 7046	do., in pegmatite	2.7	2,710	

[1] Distance in mi. from contact with Duluth Complex
[2] $Rb^{87}\lambda_\beta = 1.39 \times 10^{-1}\ yr^{-1}$

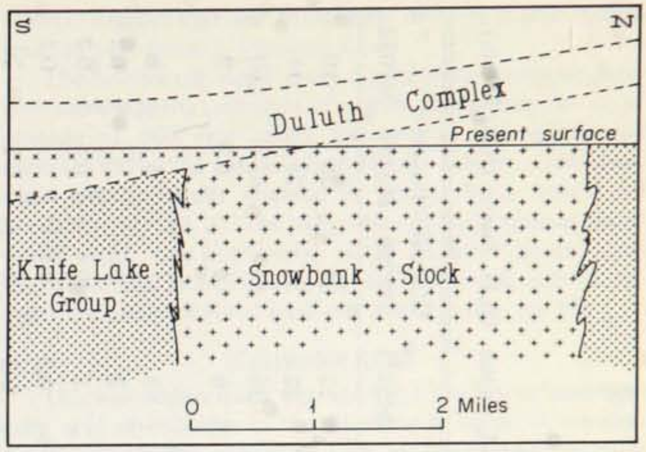


Figure III-59. Section showing postulated position of Duluth Complex over Snowbank stock in Keweenawan time (after Hanson and Gast, 1967).

from oldest to youngest: a porphyritic augite syenite; a medium-grained syenite, which he called the Snowbank syenite; a fine-grained rock which he called the Round Lake syenite; and the Snowbank granite. He showed that the three syenitic phases constitute distinct, separate intrusive bodies within the main, inner part of the stock, whereas the younger granite occurs mainly along the margins and as small satellitic bodies in the country rocks.

In their study, Hanson and Gast (1967) distinguished two principal facies: (1) a gray to dark pinkish-gray, medium- to coarse-grained, porphyritic syenodiorite and diorite; and (2) a pink, fine- to medium-grained granodiorite, but they did not map their distribution. Representative modes of the two facies are given in Table III-35, and chemical analyses of a syenitic and a granitic sample are given in Table III-36. The granodiorite contains several mafic inclusions. The major minerals in both facies are plagioclase (An_{15-30}), K-feldspar (orthoclase and microcline), augite, hornblende, biotite, opaque iron oxides, and sphene; quartz is uncommon. The nature and composition of the minerals in the pluton differ somewhat, depending on their spatial relationship to the contact with the Duluth Complex. As the contact is approached, microcline gives way to orthoclase, the albite content of the K-feldspar increases, and hornblende is replaced by augite, although hornblende remains stable to within about 6 feet of the contact. Apparently, the composition of the plagioclase was not affected by the thermal metamorphism, but within the innermost 125 feet the twinning and extinction are wavy and indistinct. The biotite, which is light to dark brown, does not change color as the contact is approached, but within about 600 feet of the contact the chlorite that commonly is interlayered with it largely disappears, and some of the biotite converts to pyroxene.

The stock contains several lamprophyre dikes, interpreted to be cogenetic with the stock (Balk and Grout, 1934; S. W. Sundeen, 1936, unpub. Ph.D. thesis, Univ. Minn.), and is cut by a few thin diabase dikes that probably are Late Precambrian in age.

Table III-35. Modes, in volume percent, of representative rocks from the Snowbank stock (after G. N. Hanson, 1964, unpub. Ph.D. thesis, Univ. Minn.).

	Syenodioritic facies										Granodioritic facies			
	BC7442	M5237	M5238a	M5240a	M5243	M5244	M5245	M5250	M5254	M7011	M5231	M5236	M5252	M7032
Quartz									0.1	0.3	10	21	26	10
K-spar		20	14	13	25	5	13	18		18	22	21	26	11
Plagioclase	61	47	71	55	44	62	60	65[1]	83	69	56	51	41	60
Amphibole	4	0.6	7	3	0.3	12	11	8	2	4	6	4	0.8	9
Pyroxene	12	17		15	18	8		3	2	2				
Biotite-chlorite	16	10	4	9	7	4	4	3	10	4	3	1	5	6
Sphene	0.3	2	1	1	1	2	0.7	1		0.5	2	0.4		0.3
Magnetite-ilmenite	5	2	1	2	3	2	0.7	1	0.3	1	1	0.4	0.4	0.7
Apatite	1	2	1	1	1	2	0.5	0.3		1	0.2	0.1		0.2
Epidote	0.2	0.4	0.3	1	0.5	4	10	0.3	1	0.5	0.2	1	0.1	2
Rutile				0.1										
Kyanite	0.2													
An content of plagioclase	32	17	20	20	20	20	20	20	23	16	20	15	24	30

[1] Includes K-spar

BC7442—Gray, med.-gr. diorite
M5237—Light gray to pink, med.- to coarse-gr. syenodiorite
M5238a—Pink, med.- to coarse-gr. syenodiorite
M5240a—Orange and black, med.-gr. syenodiorite
M5243—Pink and gray, med.- to coarse-gr. syenodiorite
M5244—Pink mottled with dark gray, med.-gr. diorite; mafic minerals occur in discontinuous patches
M5245—Pink, med.-gr., equigranular syenodiorite
M5250—Pink mottled with dark gray, med.-gr., equigranular syenodiorite
M5254—Pink mottled with dark gray, foliated syenodiorite
M7011—Pink, med.-gr., homogeneous syenodiorite
M5231—Pink mottled with dark gray, med.-gr., equigranular granodiorite
M5236—Pink, fine- to med.-gr., homogeneous granodiorite
M5252—Pink, med.-gr., homogeneous adamellite
M7032—Pink mottled with dark gray, med.-gr., homogeneous granodiorite

Table III-36. Chemical analyses, in weight percent, of selected rocks from the Snowbank and Kekekabic stocks.

	Snowbank stock		Kekekabic stock	
	1	2	3	4
SiO_2	64.10	69.50	67.42	66.84
Al_2O_3	16.70	16.86	15.88	18.22
Fe_2O_3	1.54	0.50	1.37	2.27
FeO	1.81	0.64	1.14	0.20
MgO	1.86	1.45	1.43	0.81
CaO	3.25	1.70	3.49	3.31
Na_2O	5.35	4.58	6.42	5.14
K_2O	3.77	3.94	2.65	2.80
H_2O+	0.43	0.42	0.05	0.46
H_2O-	0.08	0.20		
TiO_2	0.46	0.53		
P_2O_5	0.22		0.07	Tr
MnO	0.05			
Total	99.62	100.32	99.92	100.05

1: Fine syenite, from area between Snowbank L. and Round (Parent) L.; analyst, S. W. Sundeen. Source: Sundeen, S. W., 1936, unpub. Ph.D. thesis, Univ. Minn.
2: Snowbank granite, from S side of large island in sec. 35, 64N/9W, 15 ft. from contact with syenite; analyst, C. W. Sanders: Ref.: Sanders, 1929
3: Granite porphyry, Kekekabic L., Cook Co.; analysts, J. A. Dodge and C. F. Sidener. Ref.: Grant, 1893, p. 41
4: Normal granite, Kekekabic L., Cook Co.; analysts, J. A. Dodge and C. F. Sidener. Ref.: Grant, 1893, p. 41

In a study of its internal structure, Balk and Grout (1934) showed that the stock has an irregular but steeply-dipping foliation that is subparallel to the outer contact and a steeply-inclined lineation. Both foliation and lineation are better developed in the marginal portions than in the inner parts of the body, and are interpreted as having resulted from dominantly upward flowage of the magma. The stock is partly discordant to the regional structure of the country rocks, especially along the northwest side of Snowbank Lake where the foliation in slate of the Knife Lake Group is approximately normal to the contact (fig. III-58). Immediately adjacent to the pluton, however, the structure in the country rocks is conformable to that in the stock itself —the foliation is steeply inclined and the lineation plunges at angles greater than 50° (see Balk and Grout, 1934, fig. 3). This concordance of structure along the immediate contact was interpreted by Balk and Grout as having resulted from crowding of the wall rocks by forceful intrusion. They stated (p. 636), "The several structures clearly show that the mass rose as an elliptical chimney, crowding aside its walls so that the schists were deformed . . . for about a mile from the igneous rock." The observations of Balk and Grout (1934) and the analysis of the regional structure by Gruner (1941), suggest that the Snowbank stock is a late syntectonic or post-tectonic intrusive body.

The Snowbank stock gives a moderate, irregular positive aeromagnetic anomaly of about 1,000 gammas at an altitude of 500 feet above the ground (Meuschke and others, 1963). The maximum anomaly—about 1,500 gammas—overlies rocks mapped as Round Lake syenite by Sanders (1929, pl. 3). The magnetic anomaly extends about a mile south from the overlap of the Duluth Complex onto the Snowbank stock, indicating that the stock extends beneath the complex for at least that distance (fig. III-59).

Kekekabic Stock

The Kekekabic stock (fig. III-60) is similar mineralogically and chemically to the Snowbank stock. It consists dominantly of pink or pinkish-gray, medium- to fine-grained

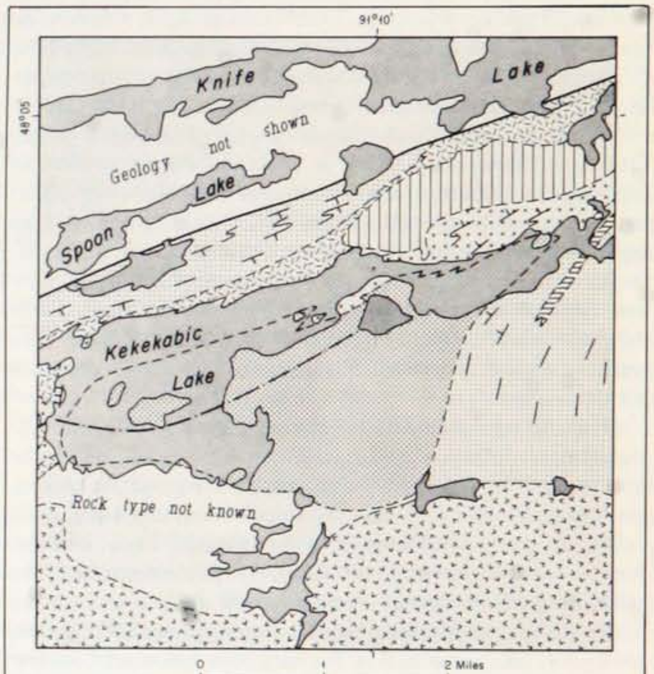


Figure III-60. Geologic map of Kekekabic Lake area (modified from Gruner, 1941, pl. 1 by P. K. Sims, 1971).

foliated syenodiorite, called granite by Grant (*in* Winchell, 1893b) and Stark (1927). A less common facies is a porphyritic syenodiorite, which cuts the "normal granite" locally and has sharp dikelike contacts. According to Stark (1927, p. 725), however, porphyritic facies also occur within the dominant massive granite and appear to grade transitionally into it. Less common rocks have more mafic compositions. An average mode of 30 specimens of the so-called "normal granite" (Stark, 1927, p. 726) follows: quartz, 7 percent; orthoclase, 11 percent; microperthite, 43 percent; oligoclase, 25 percent; clinopyroxene, 11 percent; hornblende, 2 percent; accessory minerals, 1 percent. The accessory minerals include microcline, apatite, sphene, zircon, magnetite, pyrite, and biotite. Chemical analyses of the "normal" and porphyritic facies are given in Table III-36. Probably, the microperthite listed in the modes is antiperthite, for a mineral analysis given by Grant (*in* Winchell, 1893b, p. 44) shows 6.23 percent Na_2O and 3.05 percent K_2O; this would be consistent with the chemical analyses listed in Table III-35. Plagioclase, commonly zoned, forms the phenocrysts. The augite occurs as stubby euhedral prisms; it is very dark green, and in transmitted light is described as bottle green (Grant *in* Winchell, 1893b, p. 45). Zoning is common, with the cores being nearly colorless and the rims light green. An analysis (Grant *in* Winchell, 1893b, p. 48) of the augite follows (in weight percent): SiO_2, 53.19 percent; Al_2O_3, 2.38 percent; Fe_2O_3, 9.25 percent; FeO, 5.15 percent; MgO, 9.43 percent; CaO, 17.81 percent; Na_2O, 2.63 percent; K_2O, 0.38 percent; H_2O (total), 0.01 percent. The quartz is mainly interstitial, and commonly shows undulatory extinction.

Both the equigranular and the porphyritic facies have a conspicuous foliation and lineation given by aligned feldspar and pyrobole crystals and by schlieren and inclusions (see Stark, 1927, fig. 2). The foliation trends eastward, subparallel to the long dimension of the elliptical body, and the lineation is steeply inclined. Stark (1927) interpreted the foliation and lineation as a primary flow structure. Because of the general conformity of the internal structure of the stock with the structure in the adjacent schistose country rocks, he concluded that the magma was emplaced during the folding of the wall rocks. Accordingly, the body probably is a syntectonic or a late-syntectonic intrusion.

The Kekekabic stock gives a moderate linear, positive aeromagnetic anomaly of about 500 gammas at an altitude of 500 feet above ground level. The crest of the anomaly is shown on Figure III-60.

LAMPROPHYRE BODIES IN EASTERN PART OF VERMILION DISTRICT

Lamprophyres of several types have been described (Sundeen, 1936, *op. cit.*) from the eastern part of the district in association with the Saganaga batholith and the Snowbank stock. The descriptions that follow are taken mainly from Sundeen's thesis.

Saganaga Lake Area

Lamprophyre bodies are common in the Saganaga batholith and probably also in the adjacent country rocks. They generally occur as dikes that have sharp boundaries and lack chilled margins, and which range in size from stringers less than an inch thick and a few feet long to one body that is 60 feet thick and at least 6 miles long. Most trend north-northwestward and are remarkably straight and steeply inclined. Sundeen reported that one dike transects foliated "granite" of the batholith and in turn is cut by pegmatite, which appears to grade into the "granite." He (1936, *op. cit.*, p. 14) interpreted this relationship as indicating ". . . that the enclosing rock was still a viscous mush when the dike formed." Because of relationships such as this and the close spatial association of the bodies with the batholith, he believed that the lamprophyres and batholithic rocks are cogenetic.

Most dikes are hornblendic lamprophyres having either granular or porphyritic texture. Hornblende forms most of the phenocrysts and also occurs in the groundmass. Other minerals that form phenocrysts are orthoclase, plagioclase, sphene, and apatite, all of which tend to be euhedral. Most of the feldspar and quartz is interstitial to the phenocrysts. In the distinctly porphyritic rocks, the hornblende crystals are as much as 4 mm in diameter and typically are twinned and zoned. All the rocks are altered, but to different degrees.

Augite-bearing lamprophyres and biotitic lamprophyres are less common. In one variety, biotite also forms phenocrysts as much as 6 mm long. Hornblende is abundant in some of the bodies. The groundmass consists mainly of plagioclase, orthoclase, and quartz, in varying proportions, and opaque iron oxides, apatite, sphene, and pyrite. The common alteration products are carbonate minerals, chlorite, white mica, leucoxene, epidote, and clay minerals. Four bodies examined by Sundeen contain olivine as well as augite.

Snowbank Lake Area

In the Snowbank Lake area, lamprophyre bodies are equally common in the Snowbank stock and in the surrounding country rocks of the Knife Lake Group (Sundeen, 1936, *op. cit.*). For the most part they form dikes of diverse orientation which tend to be straight in the stock and highly irregular in the country rock. Sundeen described several interesting occurrences of dikes from both the stock and the country rocks that are fragmented by the rocks which they intrude. One dike described from the stock consists of aligned fragments, commonly with rounded, blunt ends, that are separated by several inches of the enclosing syenite; evidence for faulting is lacking, and Sundeen concluded (*op. cit.*, p. 51) that ". . . the surrounding syenite has flowed into the space between the separated ends of the fragments." He described a similar occurrence (*op. cit.*, p. 51) from a dike that cuts a metamorphosed conglomerate of the Knife Lake Group—the dike cuts the country rock but in turn is fragmented and apparently intruded by the conglomerate (see photograph in Grout, 1933b, p. 216). Sundeen interpreted the lamprophyres to be cogenetic with the plutonic rocks constituting the Snowbank stock.

As at Saganaga Lake, hornblendic lamprophyres are the dominant type; biotitic and augitic lamprophyres are rare. The hornblendic lamprophyres are principally greenish-black, fine-grained rocks that have a granular or a microporphyritic texture. Hornblende of at least two generations is present. In the distinctly porphyritic rocks, orthoclase forms the groundmass and corrodes the hornblende. Accessory minerals are opaque iron oxides and pyrite, some of which have rims of hematite. Alteration is moderate. Hornblende is altered to biotite, chlorite, and epidote; other alteration products include leucoxene, clay minerals, sericite, and hematite.

Age

A lamprophyre related to the Snowbank stock that cuts metasedimentary rocks of the Knife Lake Group has a Rb-Sr age for biotite (Hanson and Gast, 1967) of 1,560 m.y. ($Rb^{87} = 1.38 \times 10^{-11} years^{-1}$). Comparison of this age with the ages determined on rocks of the Snowbank stock itself, which were lowered by the thermal metamorphism caused by the intrusion of the Duluth Complex, would indicate that the lamprophyre is approximately 2,600 m.y. old (see Hanson, 1968, p. 14). This age is approximately equivalent to the age determined for the Snowbank stock by Hanson and Gast (1967).

One of the lamprophyres that cuts the Saganaga Tonalite has a K-Ar age for biotite of 1,755 m.y. (Hanson and others, 1971b, p. 1115). Hanson and others (1971b) were uncertain as to "Whether the age is the time of intrusion of the dike or the time of later alteration . . ." but apparently, they believed that the lamprophyres are substantially younger than the Saganaga Tonalite.

SYENITIC ROCKS AND RELATED LAMPROPHYRES IN WESTERN PART OF VERMILION DISTRICT

Small bodies of syenitic rocks, lamprophyres, and lesser granitic rocks are common in the western part of the Vermilion district and adjacent areas to the west (fig. III-61). They intrude the Lake Vermilion Formation and the Ely Greenstone, and less commonly the metasedimentary-metavolcanic rocks north of the Vermilion fault.

Syenitic Rocks

The syenitic rocks form small, generally discordant plutons that range in composition from diorite to syenite. Several have been mapped recently (Hibbing Sheet, Geologic Map of Minnesota, under designation "Intrusive rocks") between longitude 92° W. and 93° W., and several others are inferred from their characteristic positive magnetic anomalies to be present in drift-covered areas. Probably the eastern part of the granitic pluton east of Big Fork and southwest of Deer Lake also belongs to this family of rocks, for a sample collected by John Berkley from an isolated exposure in the southeastern part of T. 61 N., R. 25 W. is syenite. It contains aperthitic alkali feldspar, green clinopyroxene, biotite, and sphene—minerals characteristic of the syenitic family. Three of the bodies in the eastern part of the Hibbing Sheet that have been studied are described briefly below.

Pluton Southwest of Lost Lake

A small, roughly circular pluton of syenitic rocks, about a square mile in areal extent, is exposed adjacent to State Highway 1 about 3½ miles southwest of Lost Lake (loc. 1, fig. III-61). The pluton is discordant to the metagraywacke-slate and tuff of the Lake Vermilion Formation, and has a narrow, inconspicuous metamorphic aureole.

The pluton is heterogeneous and differs from place to place in mineralogy, texture, and structure. The rocks in the core are dominantly syenite that differs through gradual changes in the amount and proportion of the two main minerals, clinopyroxene and feldspar (table III-37). Most of these rocks are massive and medium grained. An apparent local facies is conspicuously porphyritic, and contains tabular feldspar crystals as much as an inch long in a pyroxene-feldspar matrix. In this facies, the feldspar crystals are crudely aligned, producing a fair foliation. Pegmatite containing small miarolitic cavities occurs in local patches, and appears to grade into the syenite. At least two bodies of medium-gray pyroxene-biotite lamprophyre are known within the pluton, but exposures are insufficient to determine their size and shape.

Border facies of the pluton are more heterogeneous, and are distinguished from the main facies by being notably finer grained and generally more quartzose. Common facies are leucocratic syenogranite, containing small amounts of biotite and chlorite, and biotite-pyroxene monzonite (see table III-37). Small, angular inclusions of the country rocks are common in the border facies. One dike about three-fourths of a mile south of the pluton, which apparently is an offshoot of the pluton, is a diorite with an ophitic texture.

The major minerals in the rocks within the core are microcline microperthite, plagioclase that commonly is antiperthitic, and clinopyroxene (figs. III-62A and B). The accessory minerals are quartz, sphene, apatite, and opaque iron oxides. Both feldspars are distinctly pink. The dominant feldspar is microcline microperthite of the patchy variety, which contains about 25-35 percent plagioclase. It tends to form large, optically continuous grains that surround and include all other minerals except quartz. Commonly, grain boundaries of the microperthite are ragged. Plagioclase grains are smaller than the microcline grains and more euhedral. The clinopyroxene is distinctly green and moderately pleochroic, ranging from yellow-green to brilliant green. It forms subhedral, generally poikilitic crystals that tend to cluster, commonly forming a glomeroporphyritic texture. The quartz is fine grained and interstitial. The sphene occurs as euhedral crystals, generally closely associated with pyroxene. Locally, crystals of both sphene and apatite are nearly as coarse as are crystals of pyroxene. The rocks are slightly inequigranular, and textures range from xenomorphic-granular to hypidiomorphic-granular or porphyritic. Except for local, slightly strained quartz there is no evidence for metamorphism or deformation subsequent to crystallization.

Rocks within the border facies contain the same minerals as the main facies and in addition have biotite, chlorite, hornblende, and calcite (see table III-37). The larger plagioclase grains in these rocks tend to be concentrically zoned and have more albitic rims; the clinopyroxene is both zoned and twinned, the cores being very pale-green pyroxene and the rims a medium green variety comparable to the pyroxene in the inner parts of the pluton. In several sections, the plagioclase in the cores of zoned grains appears broken, whereas the rims are not. The hornblende is a bluish-green variety and the biotite is an olive-green variety. Calcite is interstitial, and apparently formed contemporaneously with hornblende, biotite, and chlorite, which may be alteration products of pyroxene. The rocks in the border facies appear aplitic in hand specimen, and have a granoblastic, inequigranular texture.

Daisy Bay Pluton

A small, irregular pluton composed of massive syenodiorite and adamellite is exposed on the peninsula between Daisy and Frazer bays of Lake Vermilion (loc. 2, fig. III-

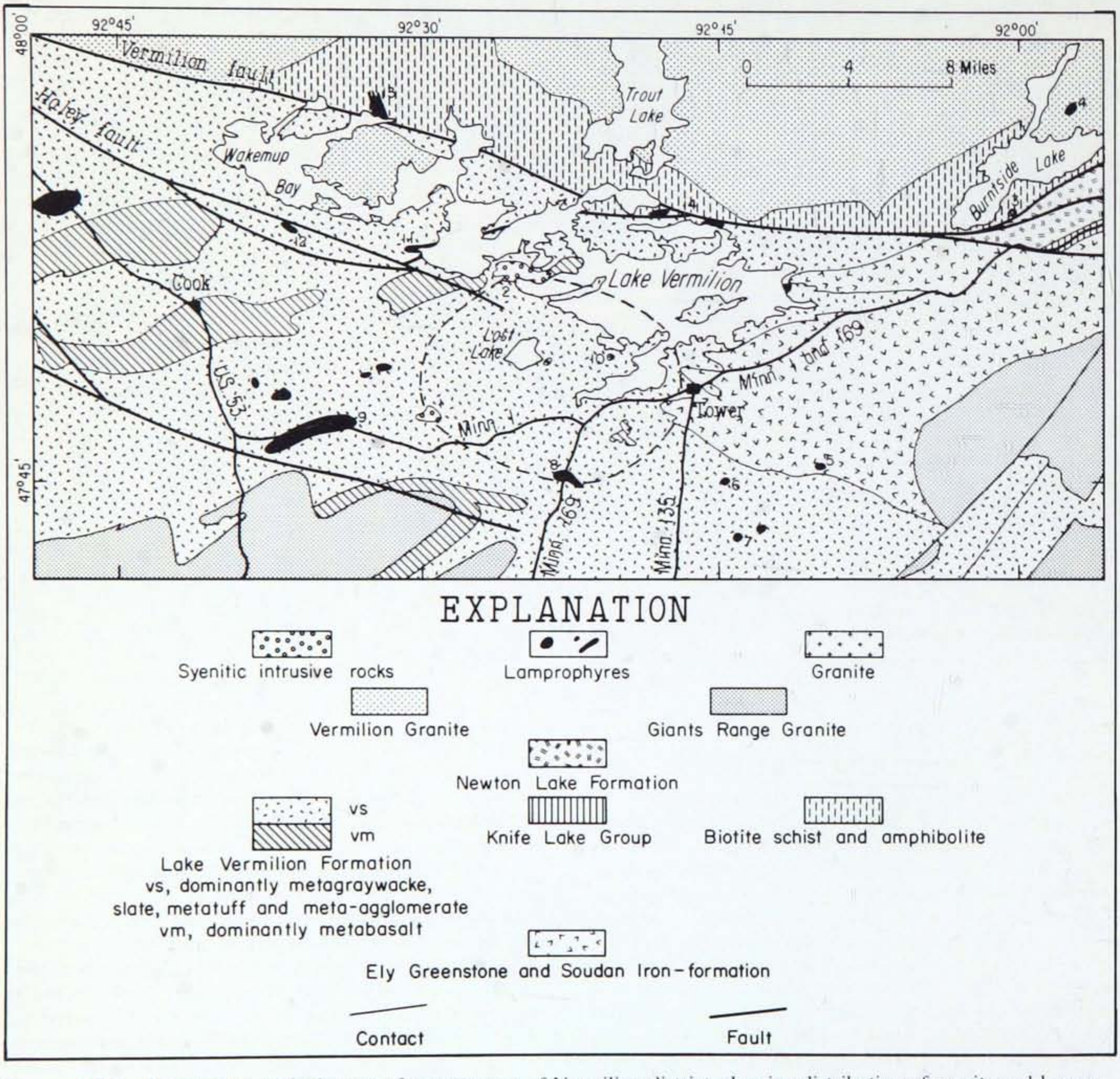


Figure III-61. Generalized geologic map of western part of Vermilion district, showing distribution of syenite and lamprophyre bodies relative to magnetic anomaly at Lost Lake. Modified from Geologic Map of Minnesota, Hibbing Sheet, 1970, with additions by P. K. Sims, 1971. Numbers refer to plutons mentioned in text. Outline of magnetic anomaly indicated by dashed line.

Table III-37. Approximate modes, in volume percent, of representative samples of small syenitic and granitic plutons in western part of Vermilion district (locations of plutons shown on figure III-61).

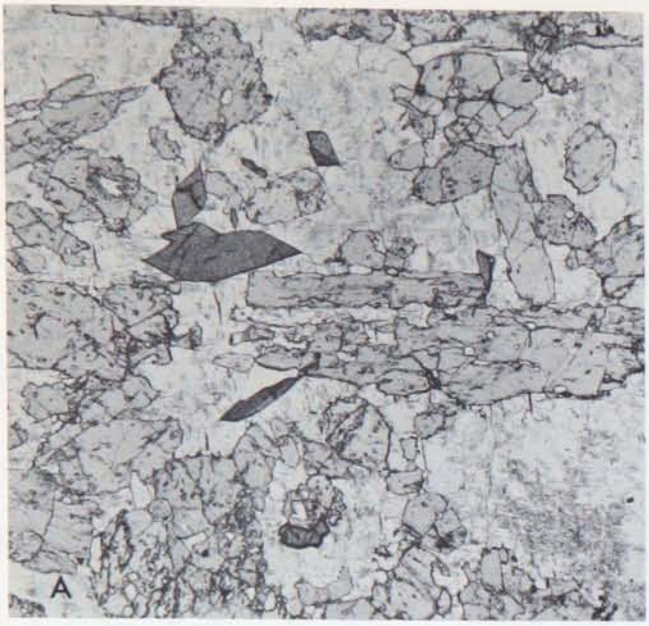
	Pluton southwest of Lost Lake					Daisy Bay pluton		Outlet Bay pluton	Everetts Bay pluton
	LL-219A	LL-224	LL-218A	ASE-6C	LL-201	LL-12	LL-13	Crl 84	T248
Quartz	7	3	21	5	1.5	2	19	2	24
Plagioclase	18	3	13	15	54.5	38	38	49	50
K-spar	53	42	63	25		10	24	11	16.5
Biotite			2	30		19		10	2.5
Muscovite							1		
Hornblende					8.5		13	27.5	
Clinopyroxene	20	42		10	30	23			
Epidote					Tr	2	1	Tr	2
Chlorite	Tr	3			3	2	2	Tr	3
Calcite				10		Tr	Tr		Tr
Apatite	Tr	4		2		1		Tr	Tr
Pyrite					2.5				
Opaque iron oxides	Tr	Tr	1	Tr		Tr	1		Tr
Sphene	2	3		3		3	1	0.5	2
Hematite	Tr								
Zircon				Tr	Tr	Tr			
Total	100	100	100	100	100	100	100	100	100
An content of plagioclase	n.d.	n.d.	An_{8-9}	An_{14-15}	n.d.	n.d.	An_{10}	$An_{27\ ca.}$	An_{20-15}

LL-219A—Syenite from core facies; K-spar is microperthite
LL-224—Syenite from core facies; K-spar is microperthite
LL-218A—Syenogranite from border facies; K-spar is microperthite
ASE-6C—Monzonite from border facies
LL-201—Diorite dike, 3/4 mi. S of pluton
LL-12—Syenodiorite
LL-13—Adamellite, in part granulated; most abundant facies
Cr1-84—Syenodiorite, in part granulated
T-284—Granodiorite

61). The pluton is about 2 miles long in an east-west direction and a maximum of three-fourths of a mile wide. Although subconcordant on a gross scale, it is discordant in detail. The syenitic rocks contain scattered small inclusions of the country rocks, and are cut by locally abundant faults and fractures, some of which have conspicuous slickenside striae. Immediately adjacent to the pluton, the metabasalt and metagraywacke country rocks are slightly prograded.

The pluton consists mainly of gray or pink, medium-grained, nearly massive adamellite, which contains hornblende as the principal mafic mineral (see table III-37). The hornblende is bluish-green and mottled, and occurs as ragged, subhedral, poikilitic crystals, and probably formed pseudomorphously after clinopyroxene. A fibrous actinolitic variety is intergrown with the hornblende and also forms discrete crystals. The plagioclase forms subhedral crystals of about the same size as the hornblende, although some crystals are as much as 12 mm long. K-feldspar occurs mainly as fine-grained crystals in the interstices. Quartz is sparse and tends to be concentrated in patches; it forms "eyes" as much as 5 mm in diameter. Sphene, opaque iron oxides, and apatite are the principal accessory minerals. In the sections examined, the K-feldspar and quartz are strongly granulated, and the original hypidiomorphic-granular texture is partly destroyed by cataclasis. Probably as a consequence of the cataclastic deformation, the plagioclase is moderately altered to epidote, calcite, and a white mica, and the amphibole is altered somewhat to epidote.

A less common facies, a gray diorite, contains both biotite and clinopyroxene as ferromagnesian minerals. Some of the plagioclase is antiperthitic, and contains irregular patches of microcline. The biotite is a pale-green variety. The pyroxene is pale green, slightly pleochroic, and poikilitic, with inclusions of biotite, quartz, plagioclase, sphene, and apatite. Sphene and apatite are the principal accessory minerals. The texture of the rock is hypidiomorphic-granular.



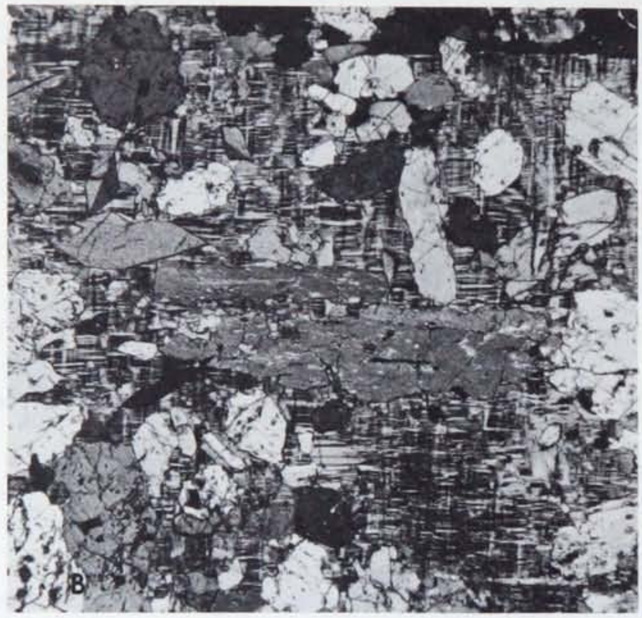


Figure III-62. Photomicrographs of syenite and adamellite from small bodies in western part of Vermilion district. A, syenite from small pluton southwest of Lost Lake. Corroded clinopyroxene (medium gray) is poikilitically enclosed within microcline microperthite (light gray). Note euhedral sphene crystals. Plane polarized light. X20. B, same view, crossed nicols. Note optical continuity of microcline. C, adamellite from Everetts Bay pluton. Partly resorbed, zoned plagioclase and chloritized biotite (dark gray) are surrounded by microcline microperthite (light gray) and quartz (white). Crossed nicols. X20. Photomicrographs by R. B. Taylor.

Outlet Bay Pluton

A small body of syenodiorite, approximately 2,500 feet long and 1,200 feet wide, occurs on the east side of Outlet Bay, at the southern end of Burntside Lake (loc. 3, fig. III-61). It intrudes folded biotite gneiss and amphibolite, but the nature of the contact is not known because of poor exposures. The southeastern side of the body is truncated by the Vermilion fault.

The pluton is composed of a pinkish-gray, medium-grained biotite-hornblende syenodiorite (see table III-37) that has a pronounced vertical lineation given by aligned amphibole crystals and a fair foliation given by tabular feldspar crystals. The principal minerals are plagioclase (An_{25}), hornblende of two generations, K-feldspar, and biotite. Microcline and sparse quartz occur as interstitial minerals. The plagioclase forms subhedral grains of varying sizes and has a weak concentric zoning. The hornblende is green and occurs in aggregates that form a glomeroporphyritic texture; it is rimmed by and replaced by a green actinolitic hornblende, and both are partly altered to chlorite. The biotite is a distinctly green variety. It occurs as ragged, subhedral grains which contain scattered small grains of epidote and sphene. The minerals are altered considerably and partly granulated. The plagioclase is clouded by a white mica and epidote, and the albite twinning is largely obscured; the biotite is bent and has a shadowy extinction. Possibly the later generation of amphibole is related to this alteration.

Probably the foliation and lineation are primary structures resulting from magmatic flowage, but because of the severe cataclasis and extensive alteration, which obscure the primary texture, this is uncertain. The alteration is interpreted as having accompanied shearing related to the Vermilion fault.

Adamellite-Granodiorite

Three small plutons of granitic rocks that differ markedly from the Vermilion Granite occur near the south shore of Lake Vermilion, west of Tower. The larger one, south of Everetts Bay (loc. 10, fig. III-61), is 1,500 feet long in a north-northwest direction and 750 feet wide. It is discordant to the structure in the metagraywacke country rock. Immediately adjacent to the pluton, the biotite in the metagraywacke is slightly coarser grained than that away from the body.

The plutons are composed of variable, massive, pink, medium-grained adamellite or granodiorite (table III-37). The rocks are slightly inequigranular. Plagioclase, the most abundant mineral, is zoned concentrically (see fig. III-62C), and has cores of about An_{20} and rims of about An_{15}. Many crystals are ragged, and probably are partly resorbed. Some plagioclase grains have scattered patches of microcline. Microcline occurs both as large, optically continuous crystals that surround and include plagioclase and as an interstitial mineral. The larger crystals are microperthite of bleb type. Quartz occurs as large, interstitial, amoeboid grains showing undulatory extinction. Biotite forms relatively small, ragged grains; it varies in color from light olive to dark yellowish brown. Aside from scattered tiny flakes of white mica on the plagioclase and some chloritization of the biotite, the minerals are little altered. All sections show some cataclasis, probably resulting from the late, regional faulting.

Judged from the massive internal structure, discordant contacts, and pronounced zoning in the plagioclase, the plutons are post-tectonic or late tectonic and were emplaced at relatively shallow depths.

Lamprophyres

The lamprophyres that are spatially associated with the syenitic plutons occur both as dikes and as crudely elliptical plutons. The plutons are as much as a mile in diameter. Where observed, contacts with the country rocks are distinct and sharp. Inasmuch as one of the better exposed lamprophyre bodies, the Dead River pluton, is described in the following section (Geldon, this chapter), regional differences in the mineralogy and textures of the rocks are emphasized here.

The lamprophyres are melanocratic or mesotype rocks that contain euhedral ferromagnesian phenocrysts in a microgranular groundmass that tends to have a panidiomorphic-granular texture. Hornblende and clinopyroxene are the dominant phenocrysts; both plagioclase and K-feldspar, in varying proportions and amounts, and ferromagnesian minerals occur in the groundmass. Apatite, calcite, sphene, and opaque iron oxides are characteristic minor minerals, and locally are abundant. The lamprophyres can be classed mainly as metavogesite and metaspessartite.

The lamprophyres have textures indicative of a complex crystallization history and subsequent metamorphism. The least modified rocks can be illustrated by a sample from the Dead River pluton, near Burntside Lake (loc. 4, fig. III-61). In this rock, clinopyroxene typically constitutes the phenocrysts in a groundmass composed mainly of plagioclase, biotite, and clinopyroxene (fig. III-63A). The phenocrysts are euhedral, have sector twinning, are concentrically zoned, and are poikilitic. Some pyroxene phenocrysts have partial rims of green hornblende, which are interpreted as being of late magmatic origin.

More commonly, the pyroxene phenocrysts or glomeroporphyritic clusters are replaced by continuous rims of a dark-green hornblende, typically accompanied by homoaxial cores of the same hornblende (fig. III-63B). An example is the Hay Lake pluton (loc. 7, fig. III-61), which has been described by Griffin (1967, unpub. Ph.D. thesis, Univ. Minn.). Euhedral phenocrysts of the same hornblende also occur in the rock. In the Hay Lake pluton, the groundmass contains another variety of hornblende, which is a paler green and forms nonaligned prisms that surround and replace the monocrystalline hornblende rims on pyroxene. Greenish-brown biotite also occurs in the groundmass; it replaces pseudomorphs of hornblende after pyroxene. Judged from the textural relationships, the hornblende rims and large euhedra of the same hornblende are products of magmatic crystallization, and both the pale-green hornblende and the biotite in the groundmass probably are metamorphic minerals.

A further stage in the magmatic reaction of pyroxene to hornblende is indicated in Fig. III-63C. Although the outlines of the pyroxene euhedra remain in this rock, each

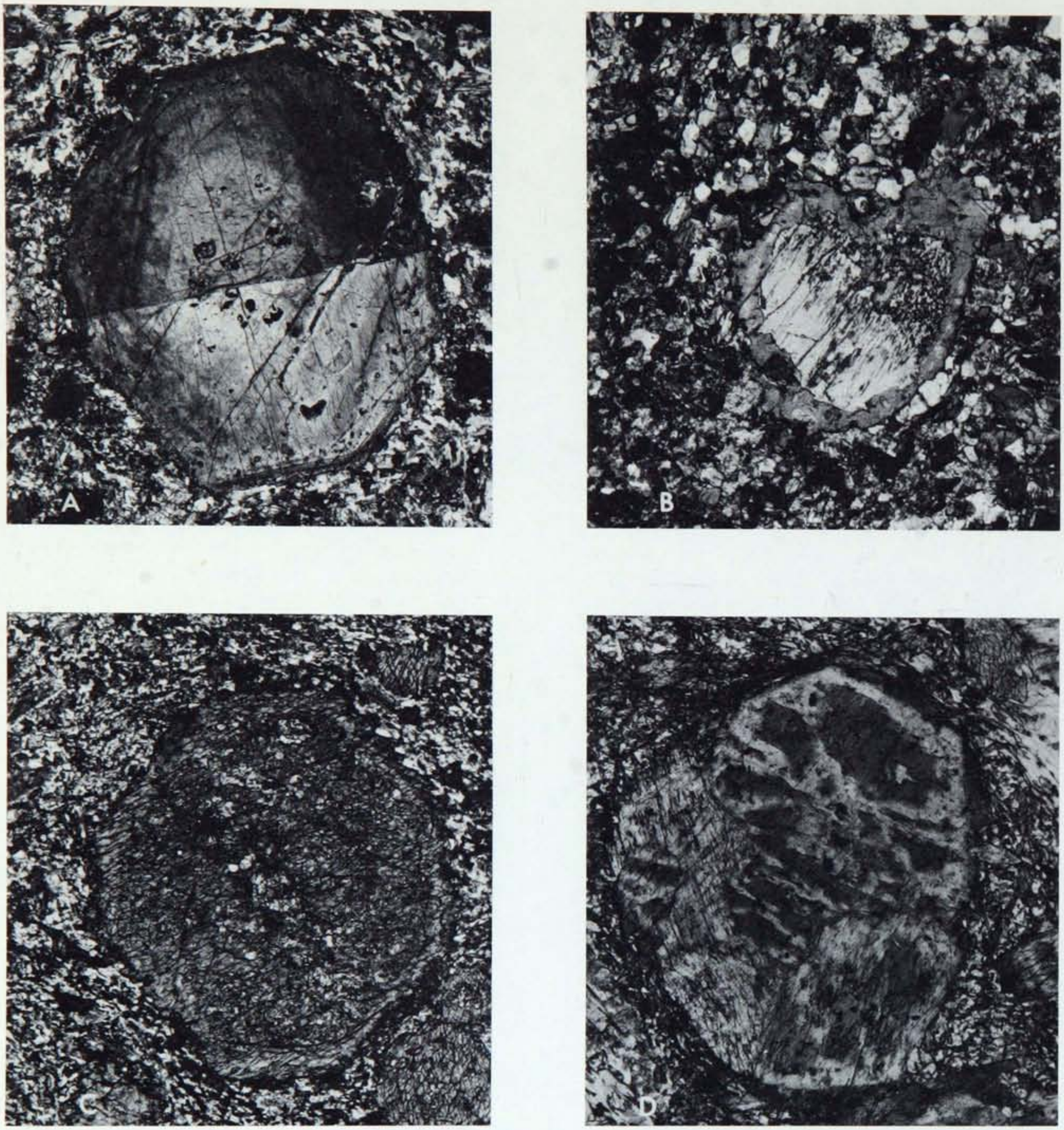


Figure III-63. Photomicrographs of lamprophyres, showing the two-stage replacement of clinopyroxene phenocrysts by hornblende. A, Dead River pluton (loc. 4, fig. III-61). Fresh, slightly corroded clinopyroxene phenocryst in fine-grained groundmass of plagioclase, pyroxene, and biotite (dark gray). Crossed nicols. X15. B, vogesite from Hay Lake pluton (loc. 7, fig. III-61). Partly altered clinopyroxene phenocryst with rim of dark green magmatic hornblende. Groundmass is dominantly plagioclase (light green), microcline, and hornblende (dark gray). Crossed nicols. X65. C, vogesite from small pluton to north of Hay Lake pluton, Embarrass quadrangle. Clinopyroxene phenocryst is almost entirely replaced by dark-green magmatic hornblende. All hornblende is homoaxial. Groundmass is microcline, plagioclase, and hornblende. Plane polarized light. X50. D, lamprophyre from locality 6 (fig. III-61). Glomeroporphyritic clump of clinopyroxene grains is replaced by two generations of hornblende. Dark outer rim and dark inner parts are green magmatic hornblende; light parts are pale-green metamorphic hornblende. All hornblende is homoaxial. Plane polarized light. X65. Photomicrographs by R. B. Taylor.

phenocryst consists of a single crystal of poikilitic hornblende containing a few shreds of the original pyroxene. The pseudomorphs may consist of a single variety of hornblende or of two or more varieties.

In many bodies, as for example the Kugler Township and Pike River plutons (locs. 6 and 8 respectively, fig. III-61), the phenocrysts consist entirely of hornblende, generally of two types. Judged from the crystal outlines, the hornblende probably is pseudomorphous after pyroxene. In the Kugler Township body, the outer parts of each pseudomorph and irregular patches within each consist of a dark-green hornblende; the remainder is a pale-green hornblende (fig. III-63D). All the hornblende in each pseudomorph is optically continuous. Small euhedra of the light-green hornblende are scattered through the groundmass, in part in poikilitic patches.

In more intensely altered samples, the original porphyritic texture is partly obliterated, and the ferromagnesian minerals are dominantly metamorphic in origin. A foliation produced by recrystallization may mask the original fabric and produce a hypidiomorphic- or xenomorphic-granular texture. In these rocks, the hornblende pseudomorphs commonly are overgrown by actinolitic hornblende, and epidote, chlorite, sphene, opaque iron oxides, and other secondary minerals are common. Apparently, the opaque iron oxides result in part from the retrograde alteration of clinopyroxene, probably in a late stage of the magmatic history.

The lamprophyres within the Vermilion granite-migmatite massif on the north side of the Haley and Vermilion faults (fig. III-61), differ from those in the supracrustal rocks in being coarser grained, more nearly equigranular, and more strongly foliated. At locality 12 (fig. III-61), for example, some of the rock within the sill-like body has a porphyritic texture characteristic of lamprophyres, but other parts are medium grained and essentially equigranular. Apparently, the minerals in the mesostasis of these rocks have been recrystallized to approximately the same grain size as the phenocrysts, producing a granular texture. In the body that was studied, hornblende is the dominant ferromagnesian mineral and plagioclase exceeds K-feldspar. Earlier, Grout (1925b, 1926) noted several such bodies peripheral to the Vermilion batholith, and called the rocks shonkinites; he considered them early border facies genetically related to the Vermilion Granite.

Some of the differences in mineral assemblage, texture, and structure of the lamprophyres can be attributed to metamorphism under different pressure-temperature conditions, but possibly some also reflect slight differences in the relative ages of the bodies. In general, the lamprophyres that are most altered and deformed are in greenschist- or low-amphibolite facies country rocks; commonly, these rocks show a retrogression of their igneous mineralogy to lower-grade metamorphic assemblages, and many have a fair foliation. Possibly, the lamprophyres in the Vermilion granite-migmatite massif, north of the Vermilion fault, were emplaced in moderately warm country rocks, and subsequently were recrystallized under moderate temperature-pressure conditions. Some of these are cut by dikes of pink leucogranite that possibly is a late facies of the Vermilion Granite.

Because the lamprophyres are widespread in the Vermilion district, they are potentially a useful tool for determining the relative ages of fold generations. If the bodies are essentially contemporaneous in age, which seems possible, they should serve as a time marker with which to compare structural and metamorphic events.

Possible Buried Pluton near Lost Lake

Many of the syenitic plutons and lamprophyre bodies in the western part of the Vermilion district are spatially associated with and possibly related to a larger, subjacent igneous mass. The presence of a major buried igneous body is inferred from a distinctive, positive aeromagnetic anomaly in the vicinity of Lost Lake, which cannot be accounted for by the exposed country rocks (fig. III-64). The anomaly is a broad, dome-shaped, positive feature that has an amplitude of about 1,300 gammas at a flight elevation of 400 feet and of about 1,100 gammas at an altitude of 1,000 feet above ground level (Bath and others, 1965). It is comparable in amplitude to the anomalies overlying the Snowbank

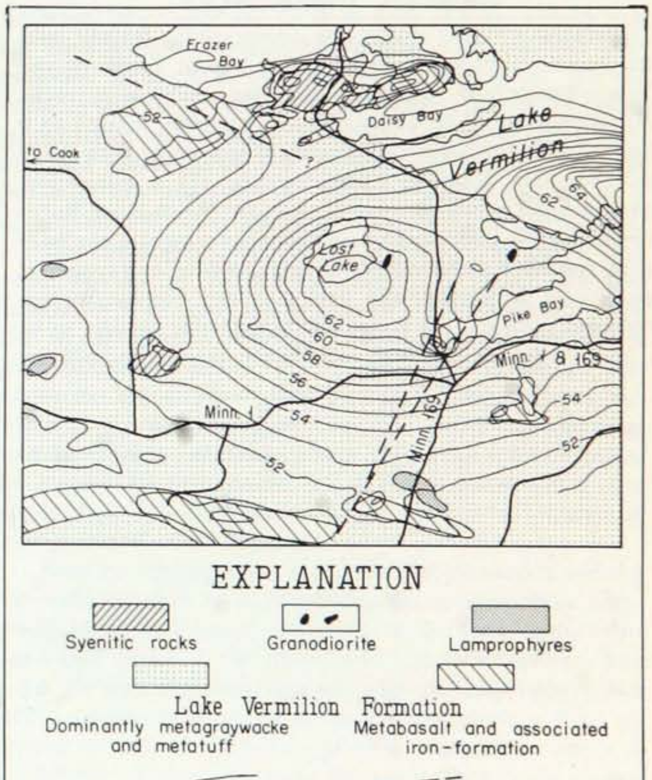


Figure III-64. Map of Lost Lake area, showing relation of syenitic, granitic, and lamprophyric bodies to magnetic anomaly (compiled by P. K. Sims, 1971).

and Kekekabic stocks. It has a width in a north-south direction of about 8 miles and a length in an east-west direction of at least 9 miles. The eastern limit of the anomaly is obscured by the anomalies given by the strongly magnetic Soudan Iron-formation in the area north of Tower. Judged from the magnetic gradients, the depth to the top of the magnetic body is about a mile.

The syenitic pluton southwest of Lost Lake and the Daisy Bay pluton coincide with small positive magnetic anomalies superposed on the broad anomaly (fig. III-64). Undoubtedly, the small anomaly having steep magnetic gradients at the west end of Pike Bay is produced by a shallow, buried syenitic body, an apophysis from the main body.

The granitic bodies between Lost Lake and Everetts Bay do not produce magnetic anomalies, but are inferred also to be related to the buried pluton because of their spatial association, textural evidence for their being high-level intrusions, and their dissimilarity to known facies of the Vermilion Granite.

GENETIC INFERENCES

Because of their distinctive composition, texture, and structure and their dissimilarity to known facies of the Giants Range and Vermilion Granites, the syenitic rocks are interpreted to be cogenetic and probably related to a distinct, separate magmatic episode. The syenitic rocks have the following common characteristics: (1) a distinctive, green augitic pyroxene and a green or bluish-green hornblende are the dominant ferromagnesian minerals, the feldspars approach mesoperthite in composition, and quartz is sparse and paragenetically late; (2) both plagioclase and pyroxene typically are zoned; (3) hornblende appears to be a late magmatic mineral, commonly having formed by pseudomorphous replacement of clinopyroxene; (4) the rocks have relatively low silica, high Fe_2O_3/FeO ratios, and relatively high lime and total alkali content; (5) the rocks are massive or have primary flow structures, and generally lack evidence for post-consolidation metamorphism and deformation other than shearing related to the late, regional faulting; and (6) the plutons are discordant on both a large and a small scale to structures in the wall rocks, indicating emplacement subsequent to or during late stages of regional deformation. In addition, the plutons from both areas produce moderate, positive magnetic anomalies. These syenitic rocks differ from those in the Linden pluton mainly in having lower lime and total alkalis, especially potash; also, the Linden pluton produces a moderately high gravity anomaly, whereas the plutons described above give neutral or negative gravity anomalies.

The syenitic bodies are interpreted from the characteristic zoning of plagioclase and pyroxene, the fine-grained border facies of at least one pluton—near Lost Lake—the presence of miarolitic cavities within pegmatite from the same pluton, and narrow thermal aureoles to have been emplaced at relatively shallow depths in moderately cool country rocks. Judged from the relatively high oxidation ratio and the presence of calcite, apparently as a late-stage magmatic mineral, the partial pressure of oxygen and the volatile content of the magma must have been relatively high.

Because of the close spatial relationship of the lamprophyres to the syenitic plutons and the evidence for a relatively high volatile content of the magma that yielded the syenitic rocks, the two rock types are interpreted to be cogenetic. Inasmuch as all the lamprophyres are metamorphosed at least mildly, they must have been emplaced relatively early in this magmatic episode, prior to cessation of the stresses that produced the folding in the region.

PETROLOGY OF THE LAMPROPHYRE PLUTON NEAR DEAD RIVER

Arthur L. Geldon

One of the better exposed lamprophyre bodies within the Vermilion district is located 5 miles northwest of Ely, on the north side of Burntside Lake between Shipman Bass Lake on the west and Dead River on the east (see fig. III-61). It is a crudely elliptical pluton about 0.6 miles long and 0.3 miles wide (fig. III-65) that intrudes folded and migmatized amphibolite and biotite schist of the Vermilion granite-migmatite massif, as defined by Southwick (this chapter). Where observed, contacts with the wall rocks are sharp and steeply inclined, and the wall rocks immediately adjacent to the contact appear to be slightly hornfelsed; there is no evidence for brecciation of the wall rock or for protoclastic deformation. The field relations clearly indicate that the pluton postdates metamorphism and deformation of the country rocks. However, it is transected by faults and joints related to the late Algoman fault system.

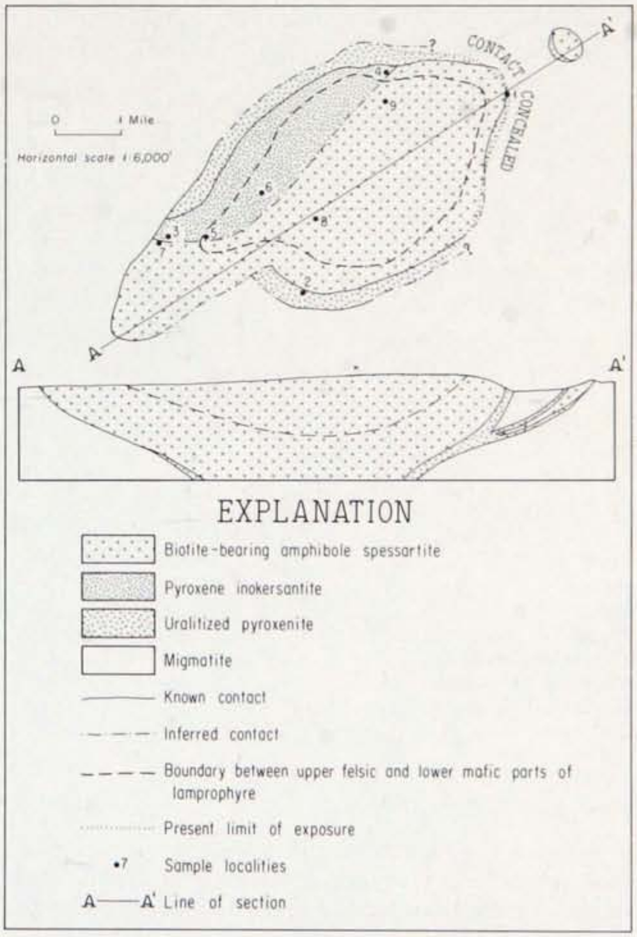


Figure III-65. Geologic map and section showing inferred structure of the Dead River pluton. Geology by Arthur L. Geldon, 1971.

Because of the lack of a uniform nomenclature for lamprophyres, the two dominant types in the pluton are defined herein. "Inokersantite" is used for a rock intermediate between kersantite and spessartite, as defined by Williams and others (1958), in which clinopyroxene and/or amphibole comprise from one- to two-thirds of the ferromagnesian minerals and K-feldspar comprises less than one-third of the total feldspar. "Spessartite" also has a groundmass of dioritic composition, but amphibole and pyroxene comprise more than two-thirds of the ferromagnesian minerals. Additional discussion of the scheme of nomenclature is given in Geldon (1972, unpub. M.S. thesis, Univ. Minn.).

DESCRIPTION OF PLUTON

The pluton has a narrow, discontinuous border zone of uralitized pyroxenite and an inner zone of lamprophyre (see fig. III-65). Approximately two-thirds of the exposed lamprophyre is biotite-bearing amphibole spessartite; the remaining one-third, comprising the western part of the body, is pyroxene inokersantite. The field relations and petrochemical data indicate that the spessartite formed through late-magmatic alteration of the inokersantite, possibly as a consequence of water that was introduced into the magma.

The lamprophyre has a crude subhorizontal cryptic layering marked by an upward increase in grain size and amount of K-feldspar and by change in the compositions of the ferromagnesian minerals—mainly an increase in the iron/magnesium ratio. The approximate boundary between the lower, more mafic part of the lamprophyre and the upper, more felsic part is shown on the geologic map (see fig. III-65). The lamprophyre has a weak foliation given by aligned feldspar grains that is better developed in the upper than in the lower part of the body. The foliation is interpreted as a flow structure.

Both the lamprophyre and uralitized pyroxenite are cut by numerous thin dikes of monzonitic or adamellitic composition, which contain xenocrysts of the same mafic minerals that occur in the pyroxenite and lamprophyre. The dikes are considered comagmatic with the lamprophyre, but could represent the late, pink leucogranite phase of the Vermilion Granite. Evidence supporting a cogenetic origin is as follows: (1) the dike rocks contain a more calcic plagioclase than the pink Vermilion leucogranite; and (2) they lack the cataclasis that typically pervades the Vermilion Granite.

PETROGRAPHY

Lamprophyre

The lamprophyre is melanocratic. It consists of euhedral diopsidic augite and/or hornblende phenocrysts and glomeroporphyritic clusters, as well as optically continuous flakes of biotite, in a fine-grained trachytic groundmass (table III-38 and fig. III-63A). The biotite grains partially

Table III-38. Modes, in volume percent, and grain size of selected rocks from Dead River pluton (locations of samples shown on figure III-65).

	Pyroxenite			Inokersantite			Spessartite			Felsic dikes
	1	2	3	4	5	6	7	8	9	10
Amphibole[1]	26.0	60.1	Tr	0.5			22.2	27.3	26.1	6.0
Clinopyroxene[2]	68.3	16.3	26.6	23.7	27.6	20.1	13.2	7.6	5.1	
Phlogopite-biotite[3]	1.3	12.6	20.6	23.0	18.4	11.9	12.7	12.0	14.6	
Plagioclase[4]		6.1	40.1	39.3	38.0	45.2	>31<40	34.9	36.5	>9<40
K-spar[5]		0.8	8.0	10.4	13.5	17.7	>9<18	15.5	15.0	>44<76
Muscovite		0.8								
Apatite		1.2	0.8	1.4	0.6	0.9	1.0	1.0	1.4	0.1
Opaques	3.6	0.4	3.9	1.7	1.9	4.2	0.1	0.4	0.2	
Sphene	0.3	1.1					0.4	0.6	0.3	1.0
Epidote[6]		0.5					0.8	0.1		7.3
Calcite	0.5	Tr					Tr			1.2
Zircon	Tr	Tr	Tr	Tr				Tr	Tr	
Allanite		0.1								
Picotite(?)							0.1	0.2	0.1	
Quartz		Tr	Tr					0.4	0.7	
	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0
Anorthite content of plagioclase		31	35	35	35	35	31	31	35	34
					Grain size (in mm)					
Pyroxene-amphibole phenocrysts	3-5	3-10	2-5	2-5	2-4	2-3	2-5	2-4	2-3	n.d.
Plagioclase laths	n.d.	n.d.	0.8	0.8	1.6-2.0	2.4-2.6	1.2-1.4	1.6-2.3	2.2	n.d.
Groundmass	n.d.	n.d.	0.1	0.1	0.2-0.3	0.3-0.4	0.1	0.2-0.3	0.3-0.4	n.d.

[1] Includes actinolite
[2] Includes talc
[3] Includes chlorite and pumpellyite
[4] Includes sericite
[5] Includes perthite
[6] Includes clinozoisite

1: Sample DRP-17a, from lower part of exposed uralitized pyroxenite
2: Sample DRP-1, from upper part of uralitized pyroxenite
3: Sample DRP-9, from lower stratigraphic zone of pyroxene inokersantite
4: Sample DRP-GS-3, from lower stratigraphic zone of pyroxene inokersantite
5: Sample DRP-12a, from upper stratigraphic zone of pyroxene inokersantite
6: Sample DRP-13, from upper stratigraphic zone of pyroxene inokersantite
7: Sample DRP-7a, from lower stratigraphic zone of biotite-bearing amphibole spessartite
8: Sample DRP-3, from upper stratigraphic zone of biotite-bearing amphibole spessartite
9: Sample DRP-175-6G, from upper stratigraphic zone of biotite-bearing amphibole spessartite
10: Sample DRP-17c, monzonitic dike rock that cuts uralitized pyroxenite

envelop the augite and hornblende phenocrysts, and are ophitically intergrown with andesine of the groundmass. Subordinate to andesine in the mesostasis are biotite, diopsidic augite or hornblende, and microperthitic K-feldspar; the latter is interstitial. Typically, the alignment of andesine laths and associated minerals is deflected around the larger phenocrysts. Biotite in the groundmass occurs in radial clusters as much as a mm long surrounding aggregates of magnetite, sphene and, rarely, apatite. Reaction rims of ferromagnesian minerals around opaque minerals are common.

The pyroxene inokersantite is a mottled gray rock characterized in part by dark-gray phenocrysts. In the lower part of the body, the groundmass is white on weathered surfaces and gray on fresh surfaces, whereas in the upper part it is pink on weathered surfaces. There is an increase upward in K-feldspar content and a concomitant decrease in biotite content, as can be seen by the modes in Table III-38. Potassium feldspar content increases from 8 percent in the lowermost sample (3) to 18 percent in the uppermost sample (6); biotite decreases in the same samples from 21 to 12 percent. The biotite is a red-brown variety. Diopsidic augite, which varies in amount but shows no systematic compositional trend (table III-39 and fig. III-65), is commonly zoned from yellowish gray in the core to olive gray in the rim. Where present, hornblende is restricted to the margins of pyroxene grains. Reaction rims of partly altered clinopyroxene and biotite, which may be altered along its margins to green biotite, occur around aggregates of iron oxides, sphene and, rarely, apatite. These aggregates are commonly dispersed through the groundmass, but occur also in the margins of the augite phenocrysts. Rare accessory minerals include zircon, quartz, pyrite, and ilmenite.

The biotite-bearing amphibole spessartite differs megascopically from the inokersantite in that the phenocrysts appear dark green on freshly broken surfaces because of the abundant hornblende. Variations in mineral abundance from lower to higher stratigraphic parts of the body differ from those in the inokersantite. In the spessartite, the total amount of hornblende and augite decreases upward from 45 to 31 percent whereas biotite shows no systematic variation. In general, the amount of biotite, andesine, and K-feldspar at any particular stratigraphic position in the lamprophyre body is less in the spessartite than in the inokersantite. The diopsidic augite (table III-39 and fig. III-66) is faintly pleochroic and zoned, both gradationally and in an oscillatory manner from pale greenish yellow to light green. Picotite (?) occurs as aligned schiller in the pyroxene. The augite is more intensely uralitized than that in the inokersantite; it is replaced along grain margins and cleavages, and where replacement is complete hornblende forms pseudomorphs after clinopyroxene. The biotite is brown and is more intensely altered to chlorite and pumpellyite than that in the inokersantite. In this rock, widely dispersed aggregates of iron oxides, amphibole, and sphene have reaction rims of hornblende that may be succeeded outward by biotite. Accessory minerals are more abundant than in the inokersantite, and include apatite, magnetite, titanomagnetite, pyrite, sphene, epidote, calcite, zircon, and quartz (table III-38). Apatite, sphene, and zircon commonly occur in inclusions in biotite. Epidote is spatially associated with the biotite.

Uralitized Pyroxenite

The uralitized pyroxenite is a dark-green, medium-grained, nearly equigranular rock composed dominantly of clinopyroxene and hornblende (table III-38). It has an automorphic-granular texture interpreted as indicating a cumulus origin. The clinopyroxene is diopside (table III-39 and fig. III-66) that is substantially altered to hornblende, with the intensity of uralitization increasing upward stratigraphically. The uralite is similar in habit to that in the spessartite, but complete pseudomorphous replacement is not as common. Phlogopite (table III-39 and fig. III-66) is partly interleaved with muscovite and altered to chlorite and pumpellyite; it increases upward in the layer, both in amount and in proportion to clinopyroxene and hornblende. Andesine occurs as peculiar rounded clusters of sericitized grains. Potassium feldspar is sparse and interstitial to the ferromagnesian minerals, as is quartz. The other accessory minerals—subhedral sphene, euhedral and partially corroded magnetite, pyrite, apatite, calcite, epidote, quartz, and allanite—occur in the interstices of the rock; apatite, zircon, and magnetite also occur as poikilitic inclusions in the ferromagnesian minerals.

Paragenesis

The paragenesis of the primary magmatic minerals, as determined from textural evidence, is shown in Figure III-67.

Low-grade Metamorphism

The lamprophyre is partly altered to low-grade mineral assemblages. Commonly, thin, generally northwest-trending veinlets containing the assemblage chlorite-epidote-calcite transect the lamprophyre; adjacent to the veinlets, hornblende or pyroxene is altered to actinolite. Also, oval clusters, a few mm in diameter, of chlorite, epidote, actinolite, opaque oxides, and relict clinopyroxene occur sporadically in the rocks; pumpellyite occurs as small laths in chloritized biotite or phlogopite. The retrogressive alteration minerals associated with the veinlets probably are related to the late Algoman faulting in the region, for the veinlets are subparallel to and are especially abundant adjacent to observed faults of this system. The other alteration minerals may also be related to this event, but they could equally well be related to a post-Algoman retrograde metamorphic event, as suggested earlier by Hanson and Malhotra (1971).

PETROCHEMISTRY

The ferromagnesian minerals in the pluton show a progressive enrichment in iron relative to magnesium as crystallization proceeded, as shown in Table III-39 and Figure III-66. The earliest clinopyroxene to form—in the border zone—was diopside, whereas the later pyroxene—in the lamprophyre—was diopsidic augite. Similar iron enrichment is found from the cores to the rims of individual grains in each of the rock types. The differences in composition for individual grains of clinopyroxene are shown by

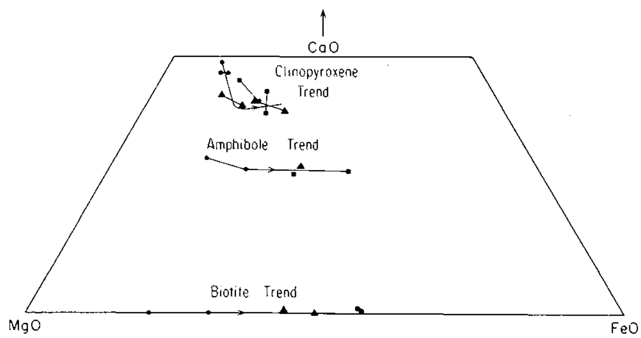


Figure III-66. Molar ratio CaO-MgO-FeO of clinopyroxene, amphibole and biotite in the Dead River pluton. •, pyroxenite; ▲, inokersantite; ▪, spessartite. Tie lines for clinopyroxene are for core and rim of grains.

Table III-39. Chemical analyses, in weight percent, of clinopyroxene, amphibole, and biotite and phlogopite from Dead River pluton (analyses by electron microprobe; locations of samples shown on Figure III-65; sample numbers same as on Table III-38).

	Clinopyroxene							Amphibole			Biotite and phlogopite		
	1	4 Rim	4 Core	7 Rim	7 Core	6 Rim	6 Core	1	7	9	1	7	9
SiO_2	53.62	47.70	49.66	48.97	48.70	51.30	51.63	45.22	45.22	46.76	38.71	38.42	36.80
TiO_2	0.10	1.66	0.46	1.60	1.63	1.32	1.61	1.77	1.55	1.30	3.39	2.97	2.91
Al_2O_3	1.63	4.33	1.93	3.47	3.50	2.63	1.94	10.61	6.42	8.52	12.56	14.60	14.38
Cr_2O_3	0.09	0.08	0.34	0.00	0.11	0.00	0.00	0.34	0.00	0.00	0.50	0.00	0.01
FeO	5.35	10.44	7.86	11.56	8.06	14.44	11.21	8.01	17.17	19.80	10.49	17.68	22.13
MnO	0.09	0.31	0.00	0.59	0.25	0.37	0.14	0.08	0.29	0.24	0.00	0.00	0.04
MgO	14.99	15.68	17.08	14.16	14.30	12.36	14.39	15.96	12.56	8.89	22.72	10.66	9.79
CaO	23.89	20.37	22.00	20.03	21.82	18.92	20.18	12.48	11.65	10.44	0.05	0.02	0.00
Na_2O	0.09	0.24	0.42	0.00	1.50	0.00	0.00	2.71	2.72	0.50	0.00	0.94	1.03
K_2O	0.05	0.05	0.01	0.31	0.11	0.00	0.00	0.79	0.71	1.34	7.90	9.16	9.27
	100.00	100.86	99.76	100.76	99.98	99.63	99.10	97.97	98.29	97.79	96.32	95.45	96.36

Number of Ions

	(Based on 6 oxygens)							(Based on 23 oxygens)			(Based on 22 oxygens)		
	1	4 Rim	4 Core	7 Rim	7 Core	6 Rim	6 Core	1	7	9	1	7	9
Si	1.973 } 2.00	1.788 } 1.98	1.864 } 1.95	1.843 } 2.00	1.834 } 1.99	1.915 } 2.00	1.916 } 2.00	6.528 } 8.00	6.797 } 7.94	7.026 } 8.00	5.564 } 7.69	5.846 } 8.00	5.643 } 8.00
Al	0.027	0.191	0.085	0.154	0.155	0.085	0.085	1.472	1.138	0.974	2.128	2.154	2.357
Al	0.044					0.031		0.324		0.535		0.465	0.243
Ti	0.003	0.047	0.013	0.045	0.046	0.037	0.045	0.192	0.175	0.147	0.366	0.340	0.336
Fe	0.165	0.327	0.247	0.364	0.254	0.451	0.348	0.967 } 4.97	2.158 } 5.18	2.488 } 5.19	1.261 } 6.55	2.250 } 5.47	2.838 } 5.66
Mg	0.822	0.876	0.955	0.794	0.802	0.704	0.796	3.434	2.814	1.991			0.005
Mn	0.006 } 1.99	0.010 } 2.10	} 2.14	0.019 } 2.05	0.008 } 2.11	0.012 } 2.06	0.004 } 2.00	0.010	0.037	0.031	4.867	2.418	2.237
Cr	0.003	0.002	0.010		0.003			0.039			0.057		0.001
Ca	0.942	0.818	0.885	0.808	0.880	0.757	0.802	1.930	1.876	1.681	0.008	0.003	
Na	0.006	0.017	0.031		0.110			0.759 } 2.83	0.793 } 2.81	0.146 } 2.06	} 1.46	0.277 } 2.06	0.306 } 2.12
K	0.002	0.002		0.015	0.005			0.145	0.136	0.257	1.449	1.778	1.814

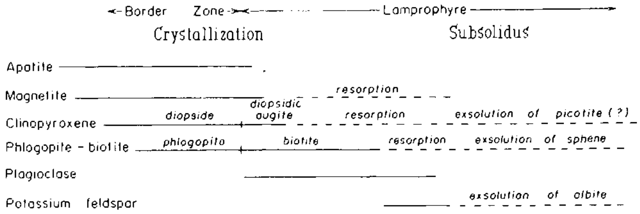


Figure III-67. Paragenesis of primary magmatic minerals, Dead River pluton.

the tie lines in Figure III-66. Interestingly, at each stratigraphic level within the lamprophyre the augite of the spessartite contains relatively more iron than the augite of the inokersantite.

In the same way, the dominant amphibole in the rock, hornblende, shows an increase in the iron/magnesium ratio from the pyroxenite to the lamprophyre. The available data suggest—in contrast to the clinopyroxenes—that the compositions of the hornblendes in both the inokersantite and spessartite are approximately the same at each stratigraphic level. Apparently, total alkalis in the hornblendes decrease slightly with the differentiation trend, but potassium increases.

The brown micas show a marked enrichment in iron as crystallization progressed, from phlogopite in the pyroxenite to biotite in the lamprophyre. The aluminum/silica ratio remained nearly constant in successively younger micas. Probably, the reddish-brown biotite in the inokersantite has a higher ferric/ferrous ratio than the brown biotite in the spessartite, for there are no appreciable differences in titanium content of the two. The available analyses indicate that the biotite in the spessartite is more enriched in iron than the biotite in the inokersantite.

Bulk chemical compositions of six rocks from the pluton are given in Table III-40. Aside from having a higher content of water, the spessartite at each stratigraphic level has approximately the same composition as the inokersantite. The alteration, therefore, cuts across the cryptic layering. The rocks have a distinctly alkaline affinity (fig. III-68), and clearly differ chemically from the Lower Precambrian volcanic rocks of the region, which have tholeiitic and calc-alkaline trends (see Sims, this chapter). They differ from typical olivine basalt (Turner and Verhoogen, 1960, p. 168) only in having a higher content of alkalis and volatiles. A plot of iron enrichment versus silica (fig. III-69) follows a trend similar to the Skaergaard liquids or Nockold's tholeiitic trend. This suggests either a constant composition or decreasing oxygen fugacity as crystallization progressed (Roeder and Osborn, 1966). The latter seems more probable inasmuch as crystallization started initially at high oxygen fugacity, as indicated by early crystallization of magnetite. The liquid line of descent was controlled primarily by removal of clinopyroxene (fig. III-70), with the initial crystallization of the pyroxenite layer.

Table III-40. Chemical analyses, in weight percent, of selected rocks from Dead River pluton.[1] (Locations of samples shown on Figure III-65.)

	Uralitized Pyroxenite		Inokersantite		Spessartite	
	1	2	4	6	7	9
SiO_2	48.18	45.74	49.92	51.63	50.63	51.73
TiO_2	.70	1.70	1.26	1.41	1.35	1.04
Al_2O_3	3.83	7.93	15.21	15.64	14.49	15.68
FeO	10.64	10.88	9.39	10.03	9.75	10.13
MnO	.15	.17	.05	.08	.14	.10
MgO	14.37	13.90	6.87	4.81	6.52	4.91
CaO	19.69	12.87	8.44	8.50	9.12	7.30
Na_2O	.72	2.00	3.02	3.12	3.61	3.16
K_2O	.31	1.90	3.61	3.58	3.12	4.01
P_2O_5	.00	.52	.64	.42	.28	.64
Cr_2O_3	.23	.08	.09	.07	.04	.03
S	.16	.26	.03	.05	.04	.02
CO_2	.18	.02				
ZrO_2 + REE	.05	.08	.05			.05
H_2O	.53	1.78	.69	.35	.85	1.10
TOTAL	99.74	99.83	99.27	99.69	99.94	99.90
Less S = O	.08	.13	.02	.02	.02	.01
TOTAL	99.66	99.70	99.25	99.67	99.92	99.89

[1] Compositions were calculated from modal percentages of rocks and compositions of analyzed minerals and should be considered approximate. S, CO_2, and REE are taken for pure pyrite, calcite, and zircon. Total iron reported as FeO. H_2O estimated by difference for analyzed minerals. Calculations by A. Geldon. Sample numbers are same as on Tables III-38 and III-39.

CONCLUSIONS

The lamprophyre and pyroxenite are comagmatic, as indicated by the field relations and the petrographic and chemical data. Probably the adamellitic and monzonitic dikes are comagmatic also, but this remains equivocal. A possible source for the rocks is an olivine basalt magma that was contaminated by crustal material.

The pluton was emplaced in a semi-crystalline state and cooled rapidly. Initially high oxygen fugacity, as indicated by early crystallization of magnetite, decreased somewhat during crystallization. Resorption textures, reaction rims, mineral zoning, and the occurrence of ferromagnesian minerals in both the phenocrysts and the mesostasis imply that crystallization began at some intermediate depth and concluded at a shallower depth. The trachytic texture of the groundmass indicates that the magma was moving during crystallization.

The liquid line of descent was controlled primarily by the removal of clinopyroxene, as can be seen in Figure III-70. Reaction rims, zoning in pyroxene crystals, and mantling of opaques by biotite protected early-formed phases from resorption by liquids with which they were not in equilibrium.

Formation of a pyroxenite layer resulted from crystallization of magnesian clinopyroxene (diopside) and magnetite at depth in a rising magma, and from gravitational settling along the sides and bottom of the magma chamber. Subsequent crystallization and settling of phlogopite concurrently with diopside yielded the phlogopite-bearing pyroxenite.

The lamprophyre formed as the liquid became more viscous and crystal settling ceased to be effective. Imperfect settling of augite, biotite, and magnetite in the crystal mush resulted in successively lesser amounts of ferromagnesian minerals in the lamprophyre away from the phlogopite-bearing pyroxenite layer. Eventually, as magnetite and clinopyroxene were partially resorbed by the liquid, plagioclase began to crystallize ophitically with the crystallizing biotite. Simultaneous crystallization of biotite, plagioclase, and K-feldspar resulted in formation of much of the interstitial groundmass. The final stages of crystallization are marked by partial resorption of biotite and by crystallization of interstitial K-feldspar and minor plagioclase.

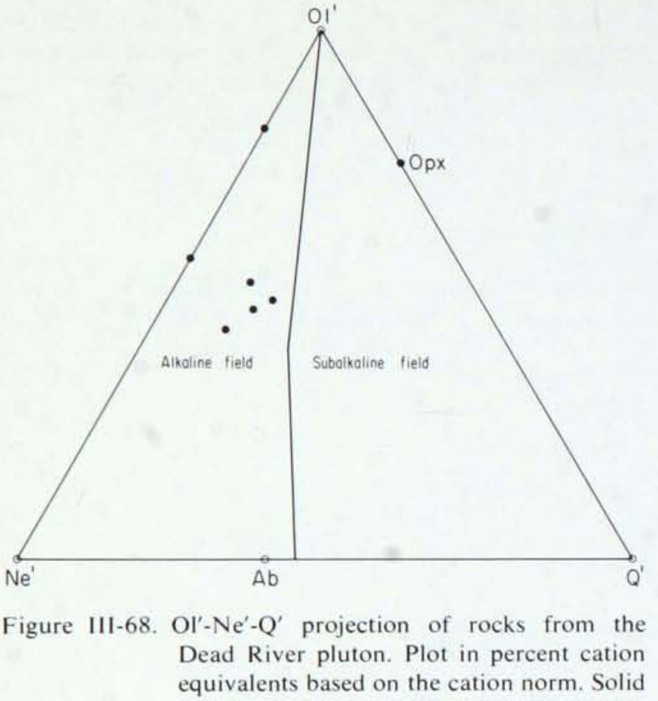


Figure III-68. Ol'-Ne'-Q' projection of rocks from the Dead River pluton. Plot in percent cation equivalents based on the cation norm. Solid line is dividing line for alkaline and subalkaline rocks proposed by Irvine and Baragar (1971, fig. 4).

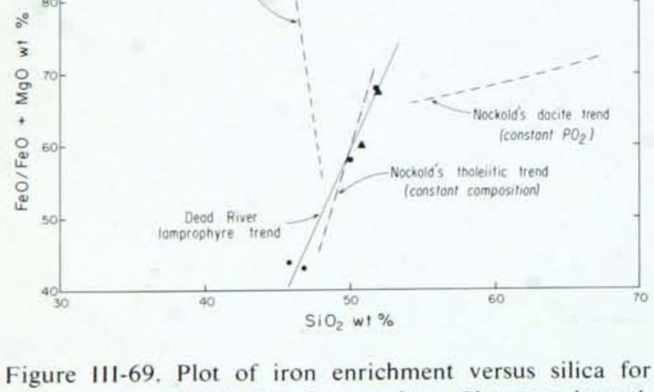


Figure III-69. Plot of iron enrichment versus silica for Dead River lamprophyre, Skaergaard trend, Nockold's tholeiitic trend, and Nockold's dacite trend.

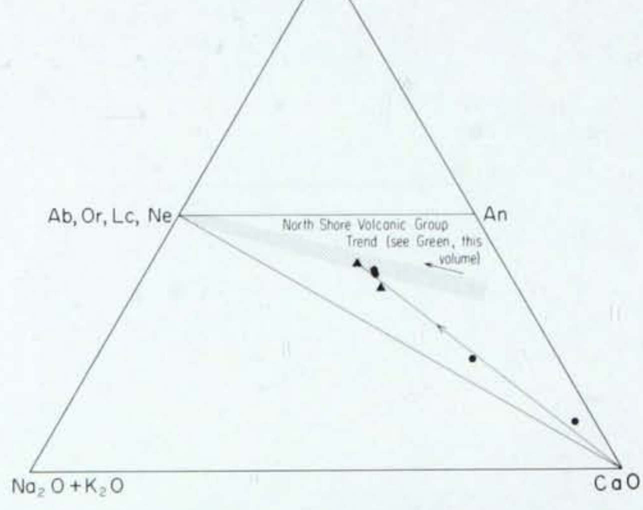


Figure III-70. Molar ratio Al_2O_3-CaO-(Na_2O+K_2O) for analyses from Dead River pluton and North Shore Volcanic Group. See Figure III-66 for explanation of symbols.

Liquid immiscibility may be responsible for the dike rocks and the rounded clusters of plagioclase in the phlogopite-bearing pyroxenite. This is suggested by equilibration instead of resorption of xenocrysts trapped in the dike rocks, by differences in composition between the lamprophyre and dike rocks that are similar to those noted by Roedder and Weiblen (1971, p. 521) for immiscibility in lunar rocks, and by field relations between the dike rocks and the remainder of the pluton. If liquid immiscibility is a valid assumption, separation of the liquid occurred early in the crystallization history of the pluton. Crystallization of the immiscible liquid occurred after crystallization of the ferromagnesian minerals trapped in the dikes.

Alteration of pyroxene to hornblende and a more iron-rich pyroxene, of biotite to chlorite and a more iron-rich and more ferrous biotite, and of plagioclase to sericite occurred while the rocks were still hot, and possibly before formation of the K-feldspar. The iron required for formation of the more iron-rich pyroxene and biotite resulted from the breakdown of biotite and magnetite; potassium and aluminum released from the dissolved biotite were incorporated into the hornblende; a residuum yielded the epidote. An external source for the water is suggested by the crosscutting relationship of the alteration to the cryptic layering of the lamprophyre body and by the intense uralitization of the marginal pyroxenite layer. Meteoric water, entering the pluton through fractures in the country rock, could have produced the observed pattern of alteration (see Taylor, 1968, p. 7855-7856).

The late-stage metamorphic assemblages of (1) chlorite+calcite+epidote±actinolite, and (2) chlorite+pumpellyite are consistent with at least one period of low-grade metamorphism subsequent to complete crystallization.

LINDEN PLUTON

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

The Linden pluton, at the western edge of St. Louis County, near the Itasca-Koochiching County line, is a moderately small, probably composite body of alkali-lime syenite. So far as known, it is unique in Minnesota. The body was first described by Grout (1926, p. 46-48), who called it the Linden syenite. The body is elliptical, has a maximum diameter in a northwest-southeast direction of about 7 miles, and occupies an area of about 30 square miles. Both the pluton and the adjacent country rocks are poorly exposed, and the contact is inferred mainly from aeromagnetic (Bath and others, 1965) and gravity (Ikola, 1968a) data. Known exposures of igneous rocks are mainly in the outer part of the pluton at the following localities: (1) bottom sec. 35, T. 63 N., R. 21 W.; (2) NW¼ sec. 1, T. 62 N., R. 21 W.; bottom sec. 7, T. 62 N., R. 20 W.; and north-central part sec. 23, T. 62 N., R. 21 W.

The country rock is exposed only on the west side of the pluton, and is pillowed metabasalt and metadiabase. These rocks have typical greenschist-facies assemblages except within a few hundred feet of the contact where they are prograded to amphibolite facies. Greenstone and intermediate-felsic metavolcanics are inferred to constitute the country rock for other parts of the pluton (Sims and others, 1970). Along the western margin, the foliation of the amphibolite is subparallel to the contact, at a marked angle to the easterly trend of the regional foliation, suggesting that the pluton was forcibly emplaced.

Judged from the known outcrops, the pluton has a pronounced vertical foliation and lineation. The foliation, given by tabular crystals, compositional layering, and concordant schlieren, is subparallel to the inferred contact; the lineation is given by aligned, elongate minerals, particularly pyroxenes, and, locally, by rod-shaped inclusions. At one locality on the south edge of the hill in sec. 35, T. 63 N., R. 21 W. the pyroxene is not strongly lineated, but instead occurs as radiating fibers.

Knowledge of the aeromagnetic and gravity expression of the pluton is limited to regional surveys. The airborne magnetometer survey (Bath and others, 1965), which was flown at an altitude of 1,000 feet above ground level, indicates that the margins of the pluton have a positive magnetic anomaly of 750-1,000 gammas; a steep magnetic gradient marks the outer edge of the body. The core is a magnetic low. The intrusion also is expressed by a positive gravity anomaly of about 8 milligals (Ikola, 1968a). Crude calculations of the gravity anomaly indicate that the core must consist of relatively mafic rocks.

The exposed rocks are pink to grayish-pink, medium-grained, foliated syenitic rocks that vary in composition, mainly through changes in the amounts and proportions of feldspar and pyroxene. Typically, oriented pyroxene is dispersed rather evenly through the rock (fig. III-31C), but locally it is aggregated into elongate lenses parallel to the foliation; as pyroxene increases in amount, the rocks become darker and tend to have a more pronounced foliation. Coarse pink pegmatite containing abundant perthite commonly cuts the rocks. At places, joint surfaces are incrusted with a thin coating of a light bluish-green amphibole.

Locally, the dominant intrusive rocks contain schlieren or larger rod-like conformable inclusions of more mafic rocks. Two of these bodies, associated with the syenitic rocks at the locality in sec. 35, T. 63 N., R. 21 W., were studied and are described briefly below.

Microcline microperthite and clinopyroxene are the dominant minerals in the rock. Plagioclase, biotite, apatite, and sphene are less common and variable in amount, and ilmenite and quartz are minor constituents. Some hematite is dispersed through the microperthite grains. The microperthite and plagioclase are slightly altered to clay minerals. Modes, given in weight percent, are listed in Table III-41. The rocks have a hypidiomorphic-granular texture, typical of rocks that crystallized at moderate depths in the crust. Cataclasis, expressed mainly as mortar structure, affected some of the rocks. The microcline microperthite occurs as large, subhedral, commonly twinned grains as much as 10 mm long (fig. III-31D). For the most part it is an uncommon variety in which the plagioclase occurs in irregular, somewhat feathery patches that under crossed nicols extinguish together. In some grains, the patches trend parallel to one another across the grain. Less commonly the microperthite is the bleb variety. Possibly the perthite characterized by the feathery texture resulted from exsolution as a consequence of the cataclasis. If so, it would postdate the bleb variety of exsolution.

The clinopyroxene, which varies inversely in amount with the microperthite, is a bright-green, strongly pleochroic augitic variety that generally occurs as subhedral or euhedral grains. It has a 2V of about 70° and the pleochroic formula: X = bright green, Y = brownish green, and Z = moderate light green. Its indices of refraction are, $\alpha = 1.720 \pm .004$, $\beta = 1.728 \pm .006$, and $\gamma = 1.740 \pm .004$ (Grout, 1926, p. 46). The clinopyroxene, as determined by microprobe analysis (table III-42), has the approximate formula (Wo_{44}, $Fs_{12.5}$, Hy_{28}, Ac_{13}, $JA_{2.5}$).

The biotite is a yellow-brown to golden-brown variety that contains roughly equal amounts of FeO and MgO, uncommonly low amounts of alumina, and moderately high K_2O. For the most part it is intergrown with or closely associated with pyroxene. Sphene occurs in beautiful wedge-shaped crystals as much as 4 mm long. Apatite forms subrounded, commonly elongate grains.

Approximate chemical compositions of the rocks were calculated from modes and analyses of minerals (table III-41). Judged from available microprobe analyses, mineral

Table III-41. Approximate modes, calculated chemical compositions[1], and a chemical analysis[2] (in weight percent) of exposed rocks in the Linden pluton.

No.	GNW-7A	MSW-2A	GNW-7-2	GNW-7B	17[2]
			Chemical composition		
SiO_2	60.3	56.0	54.8	55.0	60.21
Al_2O_3	15.3	12.8	7.8	12.3	16.28
FeO[3]	3.3	7.4	8.5	5.9	(Fe_2O_3) 2.49
					(FeO) 1.62
MgO	1.7	3.1	5.3	3.8	2.21
CaO	4.8	6.4	13.3	8.8	4.76
Na_2O	3.7	3.8	3.5	3.2	3.78
K_2O	9.3	7.6	4.8	8.1	6.32
TiO_2	1.0	2.1	.9	.6	.56
P_2O_5	.4	.5	1.2	2.3	.23
H_2O+					.10
H_2O-					.19
CO_2					.47
ZrO_2					.12
Cr_2O_3					.01
MnO					.08
BaO					.30
Total	99.8	99.7	100.1	100.0	99.76
			Modes (volume percent)		
Pyroxene	16.8	25.9	57.1	27.1	
K-spar	77.6	56.2	37.1	56.3	
Plagioclase	1.9	5.6	.4	.9	
Biotite		4.1	.4	9.0	
Apatite	.8	1.1	2.7	5.2	
Sphene	2.9	3.7	2.3	1.5	
Ilmenite		1.6			
Quartz	Tr				
Hematite (secondary)		1.8			
Total	100.0	100.0	100.0	100.0	
Specific gravity[4]	2.7	2.8	3.0	2.8	

[1] Mineral analyses used in making calculations are by electron microprobe (see Table III-42)
[2] Analysis by Grout (1926, p. 48); specific gravity is 2.739; mode not determined
[3] Probe determination gives total iron, reported as FeO
[4] Calculated from assumed mineral densities
GNW-7A, Taken from outcrops at S edge of hill, N side of county hwy.; is lighter-colored facies in exposed area
MSW-2A, Taken from dump of small adit driven into conspicuous outcrop on hill at bend in state hwy. 1, south-central part sec. 7, 62N/20W
GNW-7-2, Taken 400 ft. W of GNW-7A
GNW-7B, Dark facies from same outcrop as GNW-7A
17, Probably from same outcrop area as sample MSW-2A

Table III-42. Approximate analyses of selected rock-forming minerals, Linden pluton. (Analyses by electron microprobe; analyst, David Sinclair.) Sample A of pyroxene analyzed by Grout (1926, Table 7).

	Biotite GNW-7-B	Biotite MSW-3	Microcline microperthite[1] (Ave. of 3)	Sphene MSW-3	Ilmenite MSW-2A	Pyroxene (Ave. of 3)	A
SiO_2	44.0	43.0	61.5	31.0		53.0	46.85
Al_2O_3	11.0	13.0	19.0	3.5		1.0	2.50
							(Fe_2O_3) 11.40
FeO[2]	18.0	14.0	1.0	3.0	46.0	14.8	(FeO) 6.72
MgO	14.0	16.2	0.5		1.0	8.6	8.17
CaO			0.5	29.0		18.8	17.85
Na_2O	1.0	1.0	1.5	3.3		3.5	2.49
K_2O	10.5	11.5	15.8			.5	.50
TiO_2	.5	.5		31.0	53.0	.5	2.13
MnO							.31
H_2O+							.92
Total	99.0	99.2	99.8	100.8	100.0	100.7	99.84

[1] Analysis represents K-spar phase ($Or_{92}Ab_8$) and does not include exsolved albite in the microcline microperthite; microperthite is estimated to have composition (Or_{72} Ab_{26} An_2), which was used in calculating rock compositions (Table III-41)
[2] Probe determination gives total iron, reported as FeO

compositions are remarkably consistent from one sample to another and, accordingly, average compositions generally were used in making the calculations. The composition of microperthite used in calculating rock compositions is based on microprobe analysis of the K-feldspar phase and an estimate of the amount of exsolved albite. A chemical analysis of one whole-rock sample previously published by Grout (1926, p. 48), is included for comparison. The mode of Grout's sample is not known. The calculated compositions are believed to be accurate within five percent for the major oxides. The rocks have a high content of lime and total alkali, and can be classed as alkali-lime syenites.

One of the two bodies tentatively considered to be inclusions in the syenite—from sec. 35, T. 63 N., R. 21 W.—is a diorite believed to be a cognate inclusion. The rock contains a pale-green, concentrically zoned clinopyroxene, yellowish-brown to reddish-brown biotite, concentrically zoned plagioclase, iron sulfide, and microcline microperthite. The clinopyroxene is partly altered to green biotite (?) and rimmed by uralite. The microperthite is paragenetically late and surrounds the earlier minerals. The other presumed inclusion also has the composition of diorite, and is composed of about 80 percent hornblende, 15 percent plagioclase, and 5 percent biotite. The hornblende is partly altered to biotite and chlorite. This rock may be a modified amphibolite, derived from the country rock.

The strong discordance of the internal structure of the Linden pluton to the regional structure and the lack of a superposed metamorphism and deformation, other than a mild cataclasis, indicate that the internal structure resulted from nearly vertical (upward) flowage of magma. Accordingly, the pluton is considered post-tectonic and younger than the Vermilion and Giants Range batholiths.

The Linden pluton resembles the Icarus pluton from the Saganaga Lake-Northern Light Lake area, Ontario (Kavanaugh, 1969, unpub. M.S. thesis, State Univ. N.Y. at Stony Brook; Goldich and others, in press) in structure and mineralogy. Both the Linden and the Icarus plutons have been dated at about 2,700 m.y. (Hanson and others, 1971b; Catanzaro and Hanson, 1971).

RAINY LAKE AREA

Richard W. Ojakangas

The Rainy Lake area is in the western part of an east-northeast-trending metavolcanic-metasedimentary sequence more than 200 miles long and 6 to 20 miles wide that lies adjacent to and astride the International boundary. It is bounded on the south by the Vermilion granite-migmatite massif (see Southwick, this chapter) and on the north by another massif of granitic rocks of probably similar age. Primary zircon ages and whole-rock Rb-Sr ages from intrusive, metasedimentary, and metavolcanic rocks in the Canadian segment of Rainy Lake range from 2,690 to 2,750 m.y. (Hart and Davis, 1969); the Vermilion Granite has a whole-rock Rb-Sr isochron age of about 2,700 m.y. (Peterman and Goldich, 1970). Remapping of the United States segment of Rainy Lake was undertaken in 1968 to help resolve the Coutchiching question as well as to determine other aspects of the geology, including the mineral potential.

ROCK UNITS

The main rock type in the Rainy Lake area, and between Rainy Lake and the Vermilion Granite to the south, is biotite schist. Of greater interest, however, are the complexly interbedded rocks that comprise a belt of greenstone along the southern edge of the lake. The belt is about 2 miles wide in the United States, but widens to more than 6 miles to the northeast in Canada. It underlies the major islands—Dryweed, Grindstone, and Grassy—many small islands, Neil Point, and most of the south shore of the lake between Jackfish Bay and Black Bay (fig. III-71). All rock types in the belt are schistose and many are sheared, making recognition of primary rocks difficult.

Small bodies and dikes of granitic rocks intrude the metamorphic rocks and are most abundant in the northwestern and southeastern parts of the area. Some of these

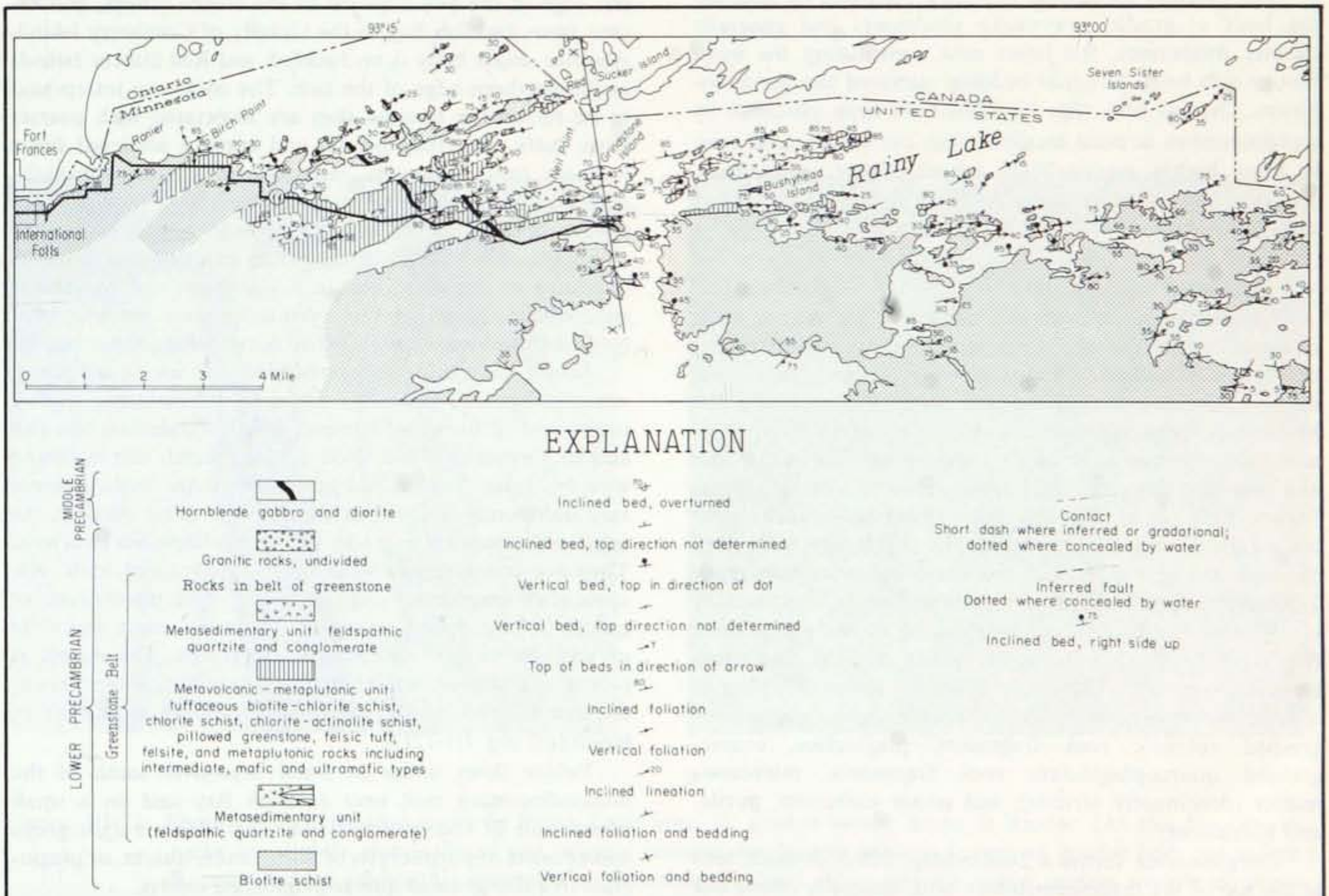


Figure III-71. Geologic map of western part of Rainy Lake area, Minnesota. Geology by R. W. Ojakangas (1968).

previously have been called Laurentian granites (Lawson, 1913a; Hart and Davis, 1969; Peterman and Goldich, 1970). Late mafic dikes cut all the rocks in the region.

Exposures are generally limited to the lake shores, islands, and shores of the Kabetogama Peninsula. For several miles to the south, large muskeg swamps obscure the geology.

Biotite Schist

Biotite schist occurs on both sides of the belt of greenstone as medium or dark-gray beds that are generally from 3 to 12 inches thick but are as much as 4 feet thick. An excellent foliation, generally parallel to bedding, is given by well aligned biotite. The biotite schist north of the belt of greenstone is fine to medium grained, with finely recrystallized biotite, quartz, and plagioclase and scattered larger original grains of quartz and plagioclase; garnet is rare. The schist to the south generally is coarser grained than that to the north. Original grain shapes and sizes have been obliterated by recrystallization. Most samples are composed of 50-60 percent plagioclase, 20-40 percent quartz, and 10-15 percent biotite. Minor minerals include chlorite, muscovite, garnet, sillimanite, staurolite, hornblende, tourmaline, epidote, apatite, and pyrite. Southwick (this chapter) describes related biotite schists farther to the south in greater detail.

The original succession probably consisted of alternating beds of graded graywacke sandstones and generally thinner mudstones, the latter now constituting the more biotite-rich beds. Original bedding survived the metamorphism. Graded beds (fig. III-72A) have been obscured by metamorphism at most localities, but can be detected even in some highly recrystallized schists. Other sedimentary features, including concretions, flame structures, and load casts on the bottoms of a few beds, are rare.

Metasedimentary Unit in the Belt of Greenstone

Feldspathic quartzite and conglomerate constitute a clastic succession within the predominant volcanic rocks of the belt. Feldspathic quartzite is the principal clastic rock, forming a nearly continuous unit about 10 miles long and half a mile wide. Quartzite and conglomerate occupy the same stratigraphic position 17 miles to the east in Canada, and continue for another 25 miles eastward (Ontario Dept. Mines, 1966, Map 2115; Merritt, 1934); these rocks have been called Seine Conglomerate. In the Rainy Lake area the beds are nearly vertical, and the outcrop width—about 3,000 feet—approximates the total thickness. Cross-bedding is common, but is obscured by shearing in many exposures (fig. III-72B). Despite the shearing, the original clastic texture is preserved. The rock contains about 50 percent quartz; the remainder consists of felsic to intermediate, fine-grained volcanic rock fragments, plagioclase, coarse-grained quartz-plagioclase rock fragments, micaceous matrix (dominantly sericite), and minor carbonate, pyrite, and K-feldspar.

Conglomerate forms a 2-mile-long, 600-foot-thick lens at the top of the metasedimentary unit, generally above but also interbedded with the feldspathic quartzite. It forms the southern part of the metasedimentary unit on Neil Point. Among the well rounded clasts (fig. III-72C), which are as much as 10 inches in diameter, granite is dominant and feldspar porphyry, quartz-feldspar porphyry, white quartzite, gray chert, and biotite schist are present. The matrix is dominantly sericite and biotite with framework grains of felsic volcanic rock fragments, quartz, and plagioclase. Another conglomerate, not shown on the map, is associated with greenschist on a point and an island in the western part of Jackfish Bay; it has a minimum exposed width of 250 feet. Pebbles as much as 6 inches in diameter are similar in lithology to those described above, but vein quartz is dominant.

Metavolcanic-Metaplutonic Unit in the Belt of Greenstone

A unit consisting of numerous rock types, including tuffaceous schists, felsic tuff, felsite, greenschist, pillowed greenstone, iron-formation, felsic hypabyssal rocks, metadiorite, metagabbro, peridotite, and anorthosite, is interbedded with the metasedimentary unit. Shearing is pervasive in all the rocks, and positive identification of rock types generally requires microscopic study. Chlorite, amphibole, and epidote impart a green color to the rocks.

Tuffaceous schists, consisting of thin intercalated and alternating fine-grained biotite-rich, chlorite-rich, and sericite-rich beds and laminae, occur in a zone near the southern edge of the belt adjacent to the biotite schists, and extend from Jackfish Bay to the vicinity of Cranberry Island. Another major body is on Jackfish and Red Sucker Islands at the northern edge of the belt. The rocks are interpreted to be tuffaceous because they are associated with coarser felsic tuffs, described below, and contain scattered felsic volcanic rock fragments, large quartz grains (volcanic?), and large plagioclase grains (volcanic?) in a recrystallized matrix of fine quartz and plagioclase, which makes up about two-thirds of the rock. Micas constitute as much as one-third of the rock; biotite is dominant and chlorite is subordinate. Much of the chlorite appears to have been formed from biotite, and sericite occurs along shear planes.

Minor felsic tuffs are interbedded with undivided greenschists at several widely spaced localities, including a small island east of Bushyhead Island, small islands near the east end of Dryweed Island, Red Sucker Island, the mainland west of Grassy Island, and just north of the metasedimentary unit south of Jackfish Bay. At the latter locality, the felsic tuffs appear to grade into the feldspathic quartzite. They are composed dominantly of fine-grained felsic volcanic rock fragments, many of which have phenocrysts of quartz or plagioclase, lesser quartz grains, some definitely of volcanic origin, and plagioclase grains. The matrix is biotite and subordinate chlorite and sericite. Exposures at the eastern end of Dryweed Island appear to consist of lapilli tuff (fig. III-72D).

Felsite flows occur in small exposures south of the metasedimentary unit near Jackfish Bay and on a small island west of Cranberry Island. The felsite is light green and consists of phenocrysts of hornblende, quartz, or plagioclase in a fine-grained quartz-plagioclase matrix.

Fine-grained, dense meta-andesite or metabasalt, commonly with deformed pillows (fig. III-73B), constitutes a

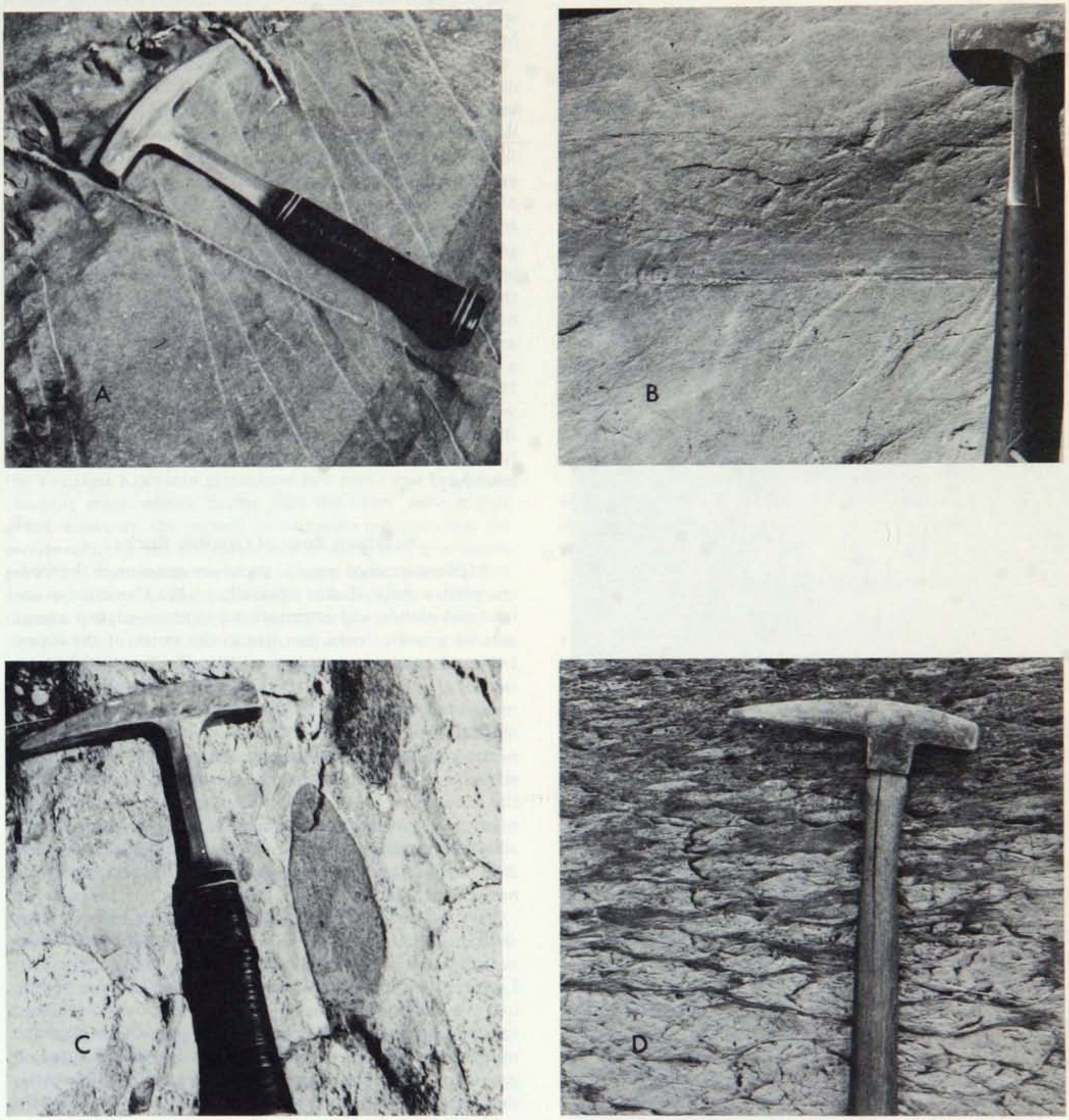


Figure III-72. Photographs of bedded rocks in Rainy Lake area. A, graded biotite schist at Ranier. (At this locality, the rocks are slightly recrystallized and original textures are largely preserved; tops of graded beds are to left.) B, cross-bedded feldspathic quartzite on Grindstone Island. Thin sericitic schist crosses middle of photograph. C, conglomerate on Neil Point. Note abundant granitic clasts. Biotitic schist pebble to right of hammer. D, sheared felsic lapilli tuff at east end of Dryweed Island.

poorly exposed 1,200 foot-wide unit along the shoreline east of Birch Point, near the western end of Rainy Lake. This unit was not mapped previously. Other much smaller exposures are present on small islands east and northeast of Dryweed Island, on the southern part of Grassy Island, and on a small island southeast of Grassy Island. Pillowed greenstone also is present to the northeast in Canada (Harris, 1969).

Rocks mapped as greenschist consist of thinly bedded, fine-grained chlorite schist, chlorite-actinolite schist, and actinolite schist (figs. III-73A and B). This rock type is common throughout the belt of greenstone; it occurs both north and south of the metasedimentary unit. It is most

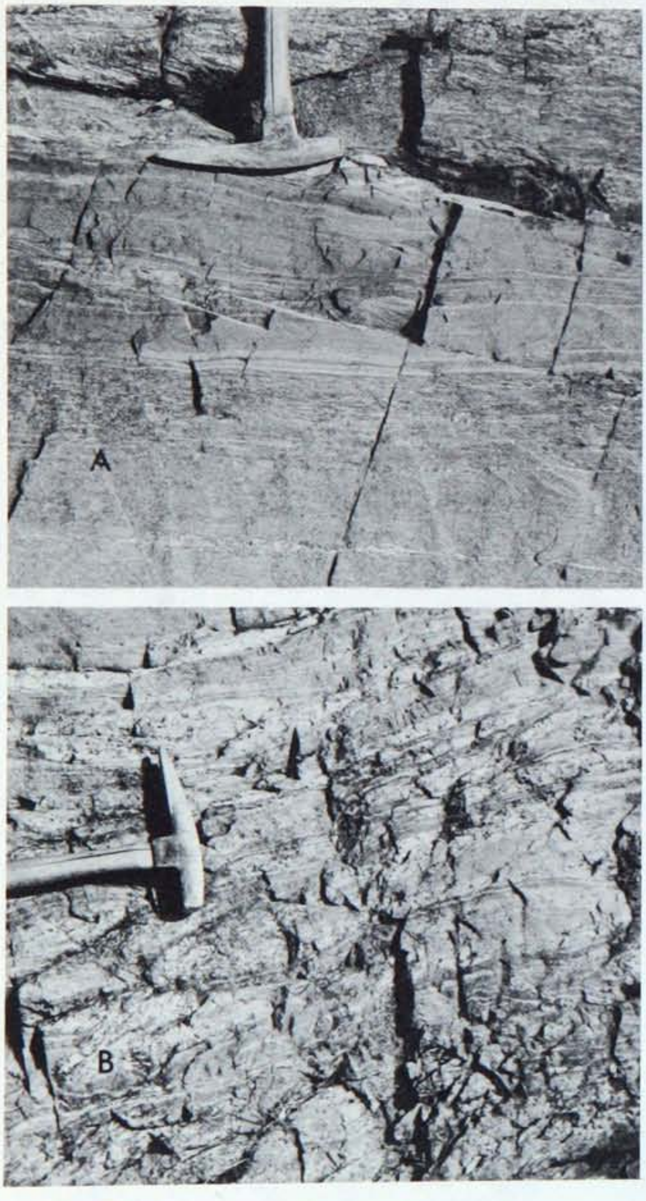


Figure III-73. Photographs of greenschists in Rainy Lake area. A, thinly laminated greenschist south of Jackfish Bay. B, deformed pillows in belt of greenstone south of Jackfish Bay.

easily seen in the area south of Jackfish Bay where it appears to grade upward into felsic tuff. Some samples have scattered large sand size grains of quartz, plagioclase, and felsic volcanic rock fragments, suggestive of a tuffaceous origin. The highly sheared nature of many greenschists obscures their original textures; some may be sheared flows or hypabyssal rocks.

Sheared and altered metaplutonic rocks, which are not distinguished separately in Figure III-71, constitute the bedrock of small islands, especially in the eastern part of the belt. Felsic, probably hypabyssal, intrusions occur as light-green, massive rocks that are intercalated with fine-grained greenschists, which may be highly sheared facies of these rocks, on Cranberry Island, islands near Cranberry Island, and on the Fox Islands. Metadiorites and metagabbros, composed of actinolitic amphibole, chlorite, and plagioclase, occur south of Jackfish Bay, on the northernmost island on the Minnesota side of the International boundary, on a few small islands north of Dryweed Island, and on a small island south of Grassy Island. A small island a few hundred feet off the northeast corner of Grindstone Island is composed entirely of serpentinized peridotite, which contains abundant olivine and some pyroxene. Anorthosite comprises the one small island of the Seven Sister Islands that is on the United States side of the International boundary.

Northern Zone of Granitic Rocks

Medium-grained granitic rocks are common in the western part of Rainy Lake, especially on the Canadian mainland and islands, and constitute the southern edge of a large area of granitic rocks that lies to the north of the Rainy Lake volcanic-sedimentary sequence. On the United States side of the International boundary, gray, biotitic granitic rocks occur on several of the small islands. They contain abundant, conformable biotite schist inclusions that have a nearly vertical foliation. Because the rocks generally occupy entire islands, contacts with the country rocks, mainly biotite schist, are not exposed. The granitic rocks have the general composition of granodiorite, and are composed of about 50 percent plagioclase, 10-20 percent microcline, and 20 percent quartz; biotite, chlorite, epidote, and sphene are minor constituents.

Within the map area (fig. III-71), a large body of gray, sheared tonalite occurs on Grassy Island and on the adjacent islands and mainland, and is intrusive into greenschists. Lawson (1913a) mapped this rock as two distinct bodies, one of Laurentian and one of Algoman age, but no evidence for this interpretation was found during my study. Near the margins of this intrusion are chlorite- and hornblende-rich granitic rocks, which Cram (1932) interpreted as differentiation products; they are interpreted here, however, as hybrid rocks resulting from the assimilation of greenschist country rock. The tonalite contains 70 percent plagioclase (albite-oligoclase), 10-20 percent quartz, 3-10 percent biotite, and minor epidote, chlorite, carbonate, sphene, and hornblende. The darker "hybrid" rocks contain more biotite, hornblende, chlorite, and epidote than the lighter colored granitic rocks.

Granites and Pegmatites Related to the Vermilion Granite

Fine- to medium-grained granite and coarse pegmatite dikes and sills, related to the Vermilion batholith, occur in biotite schists in the southern part of Saginaw Bay, and increase in abundance southward. These rocks are described elsewhere in this chapter by Southwick.

Late Mafic Dikes

Two well exposed mafic dikes, from 150 to 250 feet wide, which are the youngest rocks in the area, are present on the islands in the western part of Rainy Lake (fig. III-71), and several additional dikes are exposed sporadically in the vicinity of International Falls. These dikes strike N. 30-40° W. and are vertical. They appear to be aligned with similar dikes in Canada and southward in Minnesota, which are exposed intermittently across a distance of more than 30 miles. The dikes are described by Sims and Mudrey in the chapter on Middle Precambrian rocks.

METAMORPHISM

All the sedimentary and volcanic rocks as well as the older mafic intrusions have undergone regional dynamothermal metamorphism. Lawson (1913a, p. 27) used the grade of metamorphism as a criterion to distinguish between the so-called Keewatin greenstone belt rocks and the Coutchiching mica schists, saying that the latter were higher grade. However, the present investigation indicates that the biotite schist on the northwest side of the belt of greenstone probably was metamorphosed under approximately the same conditions as the rocks in the belt itself, with different metamorphic assemblages reflecting original differences in composition. The biotite schists adjacent to the southern margin of the belt are slightly higher grade than the rocks in the belt, but the differences are slight. Therefore, there is no evidence in this area for an older metamorphism which affected only the Coutchiching, as implied by Lawson.

The rocks in the belt of greenstone show abundant evidence of low-grade metamorphism. Bedding is preserved except where obliterated by pervasive shearing; original detrital grains (plagioclase, quartz, and felsic volcanic rock fragments) are abundant in the metasedimentary and tuffaceous rocks, and primary igneous textures persist in the metavolcanic and metaplutonic rocks. The rocks have mineralogies characteristic of the upper greenschist facies; they contain chlorite, quartz, albite, actinolite, muscovite, biotite, epidote, and calcite. Much of the chlorite is an alteration product of biotite, indicating some retrograde metamorphism.

The biotite schist on the north side of the belt, on the United States side of the International boundary, has similar low-grade mineral assemblages. The best preserved graded beds and other sedimentary structures in the area are in this unit. Scattered large detrital grains of quartz and plagioclase are present in a fine-grained matrix of recrystallized biotite, quartz, and plagioclase. The mineral assemblage—biotite, chlorite, quartz, epidote, magnetite, and plagioclase (albite?)—fits the upper greenschist facies, as does the mineralogy of other rocks in the greenstone belt. Small garnet crystals were noted in one sample from the village of Ranier; the exposure is only 1 to 2 miles south of a granitic body in Ontario and may indicate a local approach to amphibolite metamorphic facies.

With rare local exceptions, the biotite schist south of the belt is completely recrystallized. Although original grain outlines are rarely preserved, bedding is generally reflected by differences in grain size and biotite content. In these rocks one finds plagioclase, quartz, biotite, muscovite, garnet, sillimanite, staurolite, hornblende, and epidote, an assemblage characteristic of the amphibolite facies. There is a definite southward increase in grain size and in metamorphic grade. Plagioclase changes from albite (within 1½ miles of the belt) to oligoclase (1½ to 6 miles south) to calcic oligoclase and andesine still farther south. Garnet crystals increase from 0.10 mm in diameter near the belt to as much as 5.0 mm 1 mile south of it. Sillimanite appears about 6 miles south of the greenstone. These textural and mineralogical changes from north to south are attributed to the thermal effects of the Vermilion Granite. Schists of higher metamorphic grade near the Vermilion Granite are described by Southwick (this chapter).

Original compositional differences in adjacent beds of the sedimentary rocks are shown by modal variations in garnet and sillimanite in individual beds and laminae within beds, by uncommon hornblende-bearing beds, and uncommon muscovite-rich biotite-free beds. Presumably, the biotite-rich parts of the schist formed largely from original clayey detritus, which was present in the graywacke sandstones as well as in the intercalated original mudstones.

STRUCTURAL INTERPRETATION

The major structural grain of the region is east-northeast, and is given by major fold axes, faults, shear zones, bedding, and foliation. Most of the bedded rocks are inclined steeply; the only exception is in the Brule Narrows to Saginaw Bay area, where gentle open folds are present. Lineations—mainly elongated minerals, schistosity-bedding intersections, and minor fold axes—generally plunge 30°-50° ENE. Exceptions occur in a half-mile wide zone of biotite schist just north of the belt of greenstone, where lineations plunge 15°-40° WSW., and in the Saginaw Bay area, where they plunge gently both to the east-northeast and west-southwest. Both northwest- and northeast-trending lineaments, some of which represent minor faults, can be seen on aerial photographs.

Lawson (1913a) interpreted the belt of greenstone and associated rocks to be synclinal, with the quartzite unit (which he called Huronian) in the core and conglomerate (also called Huronian), greenstone (called Keewatin), and biotite schist (called Coutchiching) forming successively older units both north and south of the quartzite (fig. III-74A). Grout (1925a), noting that all the cross-beds in the quartzite face southward, interpreted the quartzite as lensing out near the eroded crest of the anticline and being absent on the north flank (fig. III-74B). It is noteworthy, however, that the northern conglomerate unit mapped by both Lawson and Grout appears to consist of a few minor lenses, and should not be assumed to be the same unit that is present to the south on Neil Point. Grout, in a discussion of a paper by Tanton (see Tanton, 1927, p. 743), conceded

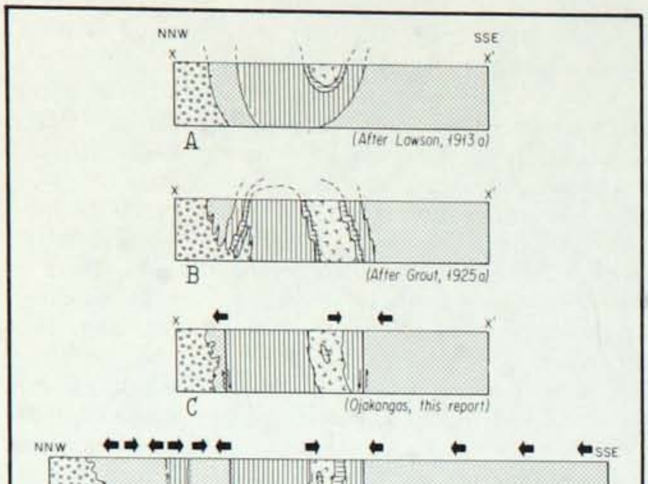


Figure III-74. Diagrammatic geologic sections showing structural interpretations of Lawson (1913a), Grout (1925a), and this report. Open arrows indicate direction of stratigraphic tops. Line of section X-X′ shown on Figure III-71.

that rocks which he mapped as conglomerate on the north shore of Grassy Island instead were formed by granite intrusion. Furthermore, my investigation has shown that greenschist occurs stratigraphically below the feldspathic quartzite and conglomerate; thus, these latter rock types are interbedded with greenschist, and cannot be younger rocks deposited upon a major unconformity, as suggested by Lawson.

Data obtained during my study permit a new interpretation involving both folding and faulting (fig. III-74C). The geologic map and the section show that nearly all stratigraphic tops in the belt of greenstone on Rainy Lake face southward, indicating a lack of major folding within the belt itself. The biotite schists that lie south of the greenstone unit face northward, as previously noted by Merritt (1934, p. 356), in an area several miles to the east, but at the same stratigraphic position. Also, most biotite schists just north of the greenstones face northward. Folding alone cannot account for these relationships; a major longitudinal fault along the southern boundary of the belt, parallel to the strike of the rock units, seems necessary, and a fault along the northern boundary of the greenstone belt apparently also is required.

Evidence for a major fault along the southern boundary of the belt includes widespread crinkling, abundant quartz veins and pods, silicification, slickensides, and cliffs (fault scarps?). Canadian geologic maps show a major fault on line with this one that extends more than 125 miles to the east (Ontario Dept. of Mines Maps 2065 and 2115, 1965 and 1966). Hawley (1930) and Merritt (1934), both of whom studied this fault zone east of Rainy Lake, concluded that the greenstone block moved eastward relative to the biotite schist block to the south, and that the south block moved upward as well. Their evidence for direction of movement appears to be equivocal, and direct evidence of the direction of movement was not obtained during my investigation. However, all the evidence taken together suggests that it is more likely that the southern block moved upward rather than downward relative to the greenstone block.

The rocks north of the belt appear to be more tightly folded than elsewhere in the area (fig. III-74D). Many graded beds indicate the presence of a synclinal axis near the middle of the 4,000-foot-wide biotite schist belt. The next unit to the north-northwest, a 2,000-foot-wide pillowed greenstone unit, is stratigraphically beneath this biotite schist. The biotite schist immediately north-northwest of the pillowed greenstone unit contains at least one minor fold axis, based on stratigraphic tops. Ontario Dept. of Mines Map 2115 (1966) shows a synclinal axis half a mile north of the mouth of Rainy River, approximately parallel to the folds on the American side of the International boundary. The axes of the Rice Bay and Bear Passage anticlines, both in the northern part of Rainy Lake, also are subparallel to this trend.

The southern biotite schist unit is anticlinal, with a major axis located about 7 miles south of the belt of greenstone, beyond the southern boundary of Figure III-71. With local exceptions, abundant tops in this 7-mile-wide interval face northward. This north flank of a major anticline is logically the south flank of a major syncline, and its magnitude suggests that the north flank of the syncline is several miles across. The rocks in the 2-mile-wide interval comprising the belt of greenstone and the biotite schists north of the major fault generally face southward, and probably form the north flank of the major syncline. If this gross structural interpretation is valid, the major fault zone on the south side of the belt must be located near the axis of a major syncline.

Younger granitic batholiths are situated both north and south of this area, and it seems that the volcanic-sedimentary accumulations between the batholiths would have subsided relative to the rising batholiths. It can be assumed that faulting in the region probably was initiated during the late stages of the emplacement of the batholiths, and that the biotite schist block south of the major fault moved upward relative to the greenstone block on the north side. In the same way, an upward relative movement is suggested for the northern block of biotite schist.

STRATIGRAPHIC POSITION OF THE COUTCHICHING

Past Work

The age relationship of the greenstone and associated rocks to the adjacent biotite schists has been controversial since Lawson (1913a) first mapped Rainy Lake and established the Coutchiching series as a stratigraphic section beneath the Keewatin greenstone, which he had defined earlier in the Lake of the Woods area. The biotite schists of this study constitute Lawson's Coutchiching series, which he named for Coutchiching Rapids on the Rainy River at the point where it leaves Rainy Lake. This locality is now largely underwater because of the dam across the river between International Falls and Fort Frances. Lawson defined three areas of Coutchiching mica schists: 1) the biotite schist that encircles the granitic gneiss of the Rice Bay anticline in the Canadian part of Rainy Lake; 2) an anticlinal belt that extends through Bear Passage, also on Rainy Lake in Canada; and 3) the Minnesota shore of Rainy Lake and its eastward extension into Canada. A part of his Bear Passage belt is equivalent to the northernmost belt of biotite schist of this report (fig. III-71). Lawson based his major stratigraphic and structural conclusions on the relationships of rocks in the Rice Bay and Bear Passage areas; he found no proof of their age relationships along the American shore, but stated that there the higher metamorphic-grade Coutchiching mica schists are overlain by Keewatin greenstones of lower metamorphic grade.

Lawson's conclusions evoked either support or heated criticism from other geologists. H. V. Winchell and Grant (*in* Winchell, 1895) recognized the Coutchiching as a valid unit. Grant (*in* Winchell and others, 1899, p. 176) said that the term Coutchiching seemed to have been correctly applied. N. H. Winchell (*in* Winchell and Grant, 1900, p. 10), however, stated, "The stratigraphic value of such a term as Coutchiching is nil, when it is applied to the chronologic scheme. It can only express a greater nearness to an accidental and local center of metamorphism or to an igneous protrusion."

A special stratigraphic commission on the Lake Superior region reported its viewpoint on the ages of these units in a single paragraph (Adams and others, 1905): "In the Rainy Lake district the party observed the relations of the several formations along one line of section at the east end of Shoal lake and at a number of other localities. The party is satisfied that along the line of section most closely studied the relations are clear and distinct. The Coutchiching schists form the highest formation. These are a series of micaceous schists graduating downward into green hornblendic and chloritic schists, here mapped by Lawson as Keewatin, which pass into a conglomerate known as the Shoal Lake conglomerate. This conglomerate lies upon an area of green schists and granites known as the Bad Vermilion granites. It holds numerous large well-rolled fragments of the underlying rocks, and forms the base of a sedimentary series. It is certain that in this line of section the Coutchiching is stratigraphically higher than the chloritic schists and conglomerates mapped as Keewatin. On the south side of Rat-root (Jackfish) bay there is also a great conglomerate belt, the dominant fragments of which consist of greenschists and greenstone, but which also contain much granite. The party did not visit the main belts coloured by Lawson as Keewatin on the Rainy Lake map, constituting a large part of the northern and central parts of Rainy Lake. These, however, had been visited by Van Hise in a previous year, and he regards these areas as largely similar to the greenschist areas intruded by granite at Bad Vermilion Lake, where the schists and granites are the source of the pebbles and boulders of the conglomerate."

Lawson was granted permission by the Geological Survey of Canada to restudy the area in 1911, and in 1913 he published a memoir essentially corroborating his earlier conclusions. He stated (p. 10): "But the International Commission on Geological Nomenclature and the United States Geological Survey deny the existence of such a series of rocks below the Keewatin. Neither of these authorities has, however, so far as their published utterances indicate, visited Rice Bay. The facts which I have recited are open for verification to any geologist and the area is easily accessible. Would it not be well for these eminent authorities before wiping out the Coutchiching series utterly, to examine the Rice Bay section?"

On page 13, he added: "The facts here recited in regard to this line of contact, particularly near the railway on the shores of Bear passage and the south end of Redgut bay, taken in connexion with the relations of the Coutchiching to the granite, appear to me to prove conclusively the superposition of the Keewatin upon the rocks mapped by me as Coutchiching in the report of 1887. I invite the attention of the International Committee and of the U.S. Geological Survey to this section and challenge them in view of the facts there apparent and easily accessible, to deny the relations of the Keewatin and Coutchiching as I mapped and described them a quarter of a century ago. The fact that these eminent authorities have denied *in toto* the existence of the Coutchiching series as a constituent member of the Archaean below the Keewatin, without any attempt to verify the very explicit statement of the evidence in regard to this section contained in the report of 1887 places them in a curious light from the point of view of scientific method. The evidence above set forth as to the superposition of the Keewatin upon the Coutchiching is practically the same as that published in 1887. In the course of the work of the past field season this has been supplemented by other observations which support the conclusions then arrived at."

Grout (1925a), using cross-bedding, graded beds, and drag folds as evidence for stratigraphic tops, concluded that at all the Minnesota localities the Coutchiching schists are younger than the Keewatin greenstones. He interpreted the belt of greenstone in the Rainy Lake area as comprising the core of an anticline, with younger biotite schists on either side (fig. III-74B). He minimized the significance of the Coutchiching schists beneath the Keewatin hornblende schists at Rice Bay, asserting that they were only 500 feet thick.

Tanton (1927) remained unconvinced of Grout's structural interpretation and agreed with Lawson's stratigraphy. Bruce (1927) also sided with Lawson, whereas Hawley (1930) concluded that the metasedimentary rocks south of the belt of greenstone are "post-Keewatin." Merritt (1934) reasoned that the metasedimentary rocks are younger than the greenstones; his report includes an excellent resumé of previous work, including much work on the Canadian side of the International boundary. Pettijohn (1937, p. 162-163) deduced that the Coutchiching is largely "post-Keewatin," except probably for the greenstones of Rice Bay and Bear Passage. Grout and others (1951), in their summary article on the Precambrian of Minnesota, stated that no Coutchiching is recognized in Minnesota, and contended that the mica schist along the Minnesota shore is younger than the Keewatin greenstones.

More recently, Yardley and others (1959) placed the Keewatin greenstone above the Coutchiching series of Lawson. Peterman (1959, unpub. M.S. thesis, Univ. Minn.) agreed with Lawson's stratigraphy, and placed the thickness of the Coutchiching at 7,000 feet. Goldich and others (1961) reviewed the problem and concluded that the Coutchiching is a thick, valid unit and that the greenstones of Rice Bay and Bear Passage do indeed correlate with Lawson's original Keewatin greenstones on Lake of the Woods. They further suggested that if the Ely Greenstone of the Vermilion district is "Keewatin" in age, then the Coutchiching probably also exists in Minnesota.

Attempts to resolve the problem by radiometric dating have not been successful, but have added much valuable information on the geologic history of the region (Hart and Davis, 1969; Peterman and Goldich, 1970).

Recent Work

The field relationships at Rice Bay and Bear Passage appear to support Lawson's argument that metasedimentary rocks (Coutchiching) lie beneath metavolcanic rocks at these localities. A recent preliminary geologic map by Harris (1969) shows that the relationships are more complicated than shown on Lawson's map, but that Lawson's stratigraphy probably is generally correct.

Lawson (1888) stated that the contact of the two units is unconformable, but later (1913a) indicated that the contact instead appears transitional. Transitional contacts between biotite schist and greenschist also have been observed during my investigation at a few localities along both the southern and northern boundaries of the belt. The tuffaceous schist unit of the belt apparently is a transitional rock type, but too few top indicators were found in this schist unit to substantiate this relationship.

Because of the complications caused by faulting, the available stratigraphic top information does not clearly resolve the age relationships of the biotite schist units and the rocks in the belt of greenstone. However, if the gross structural interpretation proposed here is accepted—*i.e.*, the metavolcanic-metasedimentary sequence is folded into a major syncline that is cut by a major longitudinal fault near its axis, and the greenstone block is downfaulted relative to the biotite schist to the south—it seems likely that the greenstone and interbedded rocks are younger than the biotite schist, and that Lawson's interpretation is generally correct. The narrow unit of pillowed greenstone near Birch Point (fig. III-71), north of the main belt of greenstone, is progressively younger to the south, as is the biotite schist unit that lies just north of the main belt. This greenstone unit may represent an early phase of mafic volcanism that interrupted deposition of clastic detritus.

Lawson assumed that all the metasedimentary rocks and greenstones belonged to the same two units, Coutchiching and Keewatin, and apparently did not consider the possibility of repetitions of rock types within the rock column. Although writing specifically about the Vermilion district to the south, Pettijohn (1937, p. 159), cautioned that the presence of "non-Keewatin" basaltic flows and tuffs in the column makes a close re-examination of Archean terranes necessary before the age relationships of the greenstones and metasedimentary rocks can be determined. The interpretation proposed here suggests that there are two greenstone units (fig. III-74D). The pillowed greenstone unit near Birch Point apparently is overlain by the biotite schist zone that adjoins the main greenstone belt on its northern margin. And within the main belt of greenstone itself, the greenschist sequence is interrupted by a thick feldspathic quartzite and conglomerate. However, all the sedimentary and volcanic units appear to be roughly contemporaneous, forming a single volcanic-sedimentary accumulation. Some of the units even may have been lateral equivalents of one another prior to the folding and faulting.

SOURCES AND SEDIMENTATION

The rocks that constitute the belt of greenstone in the Rainy Lake area include tuffs, flows, and associated sediments with volcanic components, indicating a major period of volcanism. The total thickness of these rocks in Minnesota is estimated to be about 8,000 feet. The rocks range in composition from mafic to felsic. The presence of thin bedding and lamination in the tuffaceous rocks indicates deposition in quiet water below wave base. The coarse conglomerates and large-scale cross-bedding in the quartzite suggest that an episode of high velocity currents (fluvial?) interrupted the volcanism.

The biotite schists appear to be metamorphosed equivalents of mudstones and graded graywackes. The sedimentary textures and structures, especially the bedding characteristics, are similar to those in younger turbidite sequences described in the literature. An estimated minimum thickness of the schist unit, based on stratigraphic top data south of the belt of greenstone, is 12,000 feet, and the actual thickness may be much more. Lawson (1888, p. 101) estimated a thickness of about 24,000 feet, but he used a different structural interpretation and did not utilize sedimentary structures as top indicators.

The biotite schists in the area extend, with some interruptions, for at least 50 miles to the south (Sims and others, 1970), where they pass gradually into lower-grade metagraywacke and slate. The latter constitute a substantial part of the Lake Vermilion Formation (Morey and others, 1970), and have been determined (Ojakangas, 1972 and this chapter) to have a largely volcanogenic provenance. Unfortunately, the biotite schists in the Rainy Lake area are, in general, too thoroughly recrystallized for their provenance to be determined by petrographic studies. Their source and their detailed relationship to the lower-grade rocks to the south remain equivocal.

Lawson (1888, p. 85) proposed a granitic basement for the sedimentary basin, and cited as evidence the presence of granitic boulders in the conglomerates. He suggested that the basement had been obliterated by subsequent plutonic fusion. Bass (1961), on the other hand, has proposed that the granitic boulders in Lower Precambrian volcanic-sedimentary rocks are from masses emplaced during the same major orogeny as that in which the conglomerates were deformed and intruded. This interpretation also may be applicable to the Rainy Lake area, and is supported by the data of Hart and Davis (1969), who concluded that the sedimentary rocks in the Coutchiching are no more than 50-100 m.y. older than the igneous rocks which intrude them. A similar interpretation has been proposed for the granitic clasts in conglomerates of the Vermilion district (McLimans, this chapter; Ojakangas, 1972 and this chapter).

The volcanic-sedimentary assemblage in the Rainy Lake area appears to be similar to typical volcanic-sedimentary accumulations of the Canadian Shield, as determined by Goodwin (1968a) and Goodwin and Shklanka (1967). However, it appears probable that the volcanic rocks in the Rainy Lake area overlie the main sedimentary succession.

MINERAL DEPOSITS IN LOWER PRECAMBRIAN ROCKS, NORTHERN MINNESOTA

P. K. Sims

The Lower Precambrian rocks of northern Minnesota have been a valuable source of iron ores and have yielded a small amount of gold. Ironically, it was the search for gold which led to the discovery of high-grade hematite ores in the Vermilion district in the middle of the 19th century. Mining started at the Soudan mine in 1882, and the first shipments of ore were made in 1884, eight years before production started on the more important Mesabi range to the south. Production from the Vermilion district continued until the 1960's, mainly from underground mines. At that time, taconite had achieved maturity as a source of iron, and the high-grade, natural hematite ores were no longer competitive. The last underground mine in the Vermilion district closed in 1964. Currently, greenstone belts in the Lower Precambrian sequence are being explored for base metal sulfides (Sims and others, 1969), and may be the basis for a resurgence of mining activity.

IRON ORE DEPOSITS

The Vermilion district has been a significant source of high-grade hematite ores. From 1884 until 1967, when the last shipment was made, 98,399,000 gross tons of direct-shipping ore and 5,354,000 gross tons of gravity concentrates were shipped from it (Alm and Trethewey, 1970). The ores differ from other iron ores that have been mined in the state in being massive, hard hematite generally containing 60 percent or more iron and very little chemically combined water. Production came mainly from the Soudan mine, in the western part of the district, and the Chandler, Pioneer, Sibley, and Zenith mines at Ely (see Reid, 1956, p. 111). Although small concentrations of massive hematite occur sporadically elsewhere in the iron-formations of the Vermilion district, mineable concentrations constitute much less than one percent of the total volume of iron-formation. Because the deposits at Soudan and Ely are dissimilar geologically, they are discussed separately below.

Soudan Mine

The Soudan mine yielded about 15.5 million long tons of iron ore before being closed in 1962. The ore was entirely of "lump" grade, being a massive, exceptionally hard, bluish-gray hematite that was sold at premium prices. Because of the economic and scientific interest in the geology and the ores, the mine was studied in detail by Klinger (1956; 1960, unpub. Ph.D. dissert., Univ. Wisconsin) a few years prior to its closing. These reports provide much of the data given below.

Several small, separate and distinct ore bodies were mined at Soudan. They are narrow, lenticular or somewhat sinuous deposits that occur locally within jaspilite or jasper bodies in the Soudan Iron-formation (fig. III-75). They are as much as 100 feet thick and have strike lengths of 50 to 1,000 feet and dip lengths of 50 to 2,500 feet. The shapes and sizes of the separate ore bodies differ markedly from level to level in the mine, as illustrated by Figure III-76, and extensive drilling was required prior to development of each successively deeper level.

The ore is a hard, massive hematite that contains small quantities of quartz, chlorite, apatite and, locally, pyrite, chalcopyrite and other copper minerals. A small part of the ore extracted was brecciated. At places, banding in the jaspilite host rock persists into the ore. For the most part, the hematite occurs as aggregates of randomly oriented plates from 0.01 to 0.05 mm in diameter. Martite is commonly present and magnetite occurs locally. The chlorite is a dark green, iron-rich variety which occurs mainly in vugs in the ore. The apatite tends to be associated with the chlorite. The average composition of the several ore bodies in the mine (Klinger, 1960, *op. cit.*) ranges from 63 to 66 percent Fe, 0.08 to 0.25 percent P, 0.4 to 2.0 percent Al_2O_3, and 2 to 8 percent SiO_2.

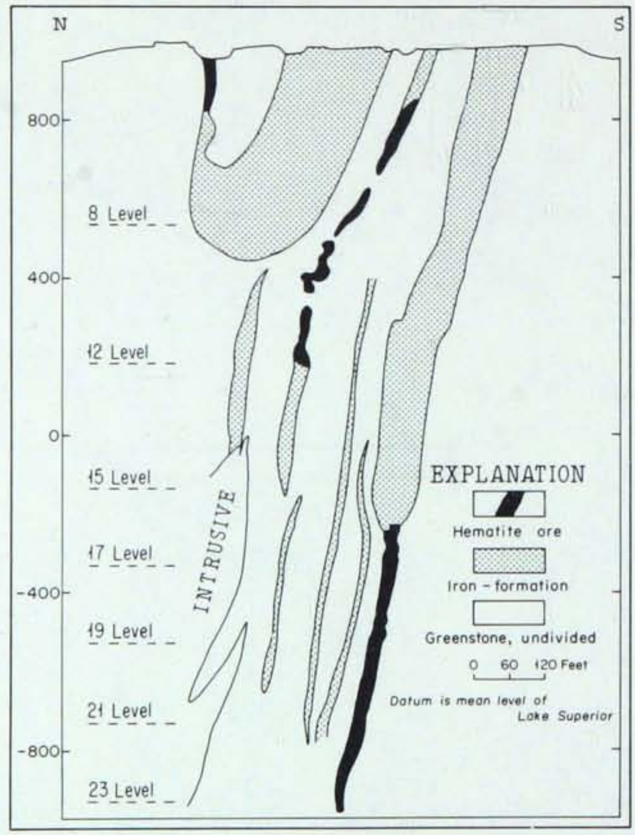


Figure III-75. Generalized section of Soudan mine, section 175 W. (after Klinger, 1956).

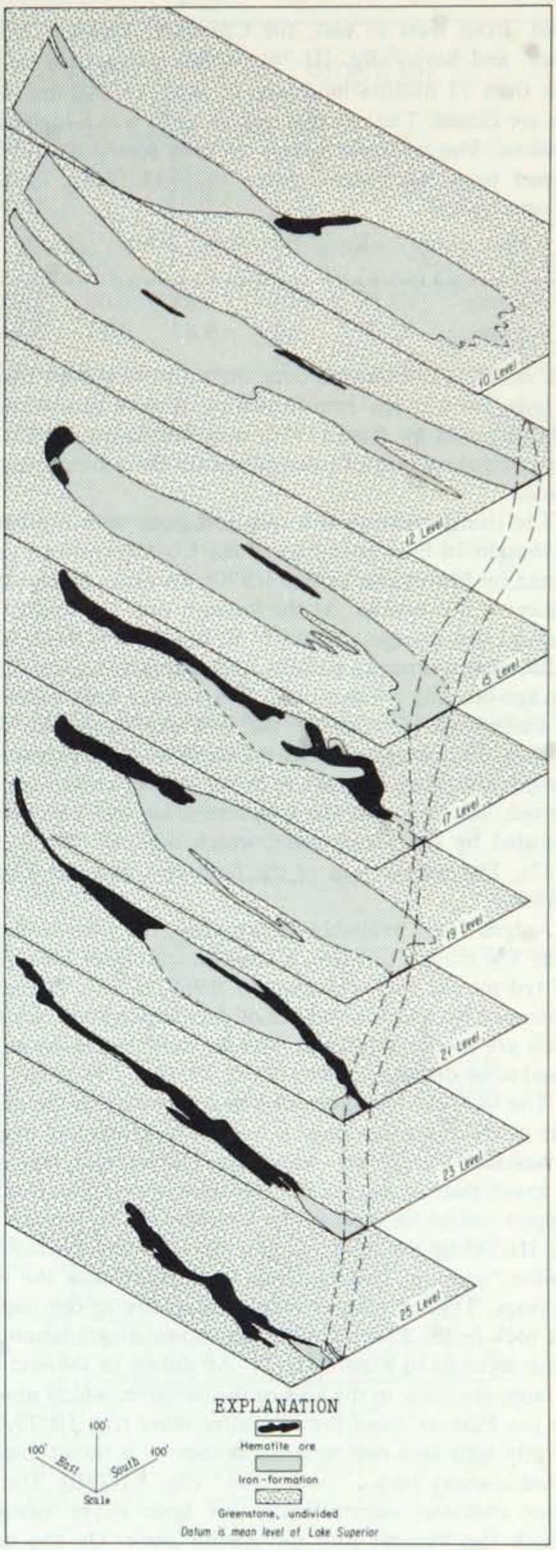


Figure III-76. Isometric block diagram of Shaft vein-651 ore bodies, Soudan mine (after Klinger, 1960, unpub. Ph.D. dissert., Univ. Wisconsin). Ore shown in black.

Although traces of sulfide minerals are widely distributed in the iron-formations, chalcopyrite and associated pyrite are moderately common in one of the ore bodies in the western part of the mine. The chalcopyrite forms fracture fillings in both iron-formation and ore and locally cements brecciated hematite. Native copper occurs sparsely as coatings on joint surfaces in ore, as fracture fillings, and as vug linings. The chalcopyrite and most, if not all, of the pyrite are later paragenetically than the hematite.

The ore bodies are interpreted from relict structures in the ores as well as from gross geometric relationships to have been formed by replacement of pre-existing beds of iron-formation. At places, replacement was complete from wall to wall, especially where the original chert-oxide beds were relatively thin or "pinched," but more commonly it was incomplete, as shown in Figure III-76. Typically, the hematite ore grades into the jasper host rocks; in the gradational zone, the hematite ore contains remnants of jasper of variable size and shape. Contacts of ore against the schistose country rocks that are interbedded with the ferruginous cherts tend to be sharp.

The factors responsible for localizing the ores are imperfectly known. Judged from some of the small occurrences of ore, however, fracturing and brecciation associated with faulting was a significant factor in localizing hematite deposition. An excellent example can be seen on Tower hill, at and adjacent to the abandoned Lee mine. At this locality, massive hematite occurs within a wide fractured and brecciated zone along a fault cutting a body of jasper and jaspilite. In detail, the hematite embays and replaces both the original hematite and the jasper in the cherty host rock. Near the outer contact, distinctly granular, white or pinkish-white quartz occurs in the massive hematite, and represents unreplaced, recrystallized quartz relicts. In the same way, the jasper adjacent to the replacement contact is recrystallized to a white, granular quartz.

The rocks adjacent to the hematite ore bodies are altered to chloritic schist, "paint rock" and, possibly, sericitic schist. The chloritic schists commonly contain more than 20 percent FeO (Schwartz and Reid, 1955, p. 299), which is two to three times the amount in metabasaltic country rocks. Apparently, they were formed from greenstone by the introduction of ferrous iron and the selective removal of some lime, silica, and magnesia. The "paint rock," a soft, red or reddish-brown, highly altered material, appears to have formed by two processes, the oxidation of chloritic wall rocks (Schwartz and Reid, 1955, p. 300) and the introduction of ferric oxide (Klinger, 1960, *op. cit.*). The relative importance of the two processes is not known. Schwartz and Reid (1955) considered the sericite schists associated with some hematite ore bodies as possible products of the alteration of rhyolitic rocks or more basic greenstones. Subsequently, Klinger (1960, *op. cit.*, p. 78) noted that the sericite content of the siliceous and sericitic rocks of tuffaceous and sedimentary origin in the mine area is "independent of the occurrence of ore." Observations in broader areas (Sims and others, 1968b) indicate that sericite is a common product of regional metamorphism of the tuffaceous rocks interbedded with the ferruginous cherts in the Soudan Iron-formation. Accordingly, sericite is not diagnostic of wall rock alteration, and probably is dominantly of regional metamorphic origin.

Most writers (Gruner, 1926; Klinger, 1956) agree that the hematite ores at Soudan, as well as at other iron ore mines in the Vermilion district, are of hydrothermal origin. The observed persistence of jaspilite banding into the ores, the occurrence of jasper inclusions in the ores, and the absence of slump structures such as those found in the Mesabi district strongly favor this hypothesis. The absence of metamorphic structures and textures in the ores, pointed out by Gruner (1926), indicates that the hematite formed after regional metamorphism. Neither the age of the mineralization nor the source for the ores is known, however.

Mines in Ely Trough

The iron ore deposit at Ely occurs in an isolated lens of iron-formation about 1¾ miles long and a maximum of one-fourth mile wide (fig. III-77). It was developed in five mines, from west to east, the Chandler, Pioneer, Zenith, Sibley, and Savoy (fig. III-78), which produced a total of more than 77 million long tons of iron ore. All the mines now are closed. The ore that was shipped was a high-grade hematite. The analyses below of ores (dried at 212° F) shipped from the Pioneer mine in 1955 (Reid, 1956, p. 147) are typical:

Type of ore	Fe	P	SiO_2	Mn	Al_2O_3
Lump	63.71	0.054	4.83	0.09	2.63
Fine	57.71	0.059	8.85	0.11	5.84

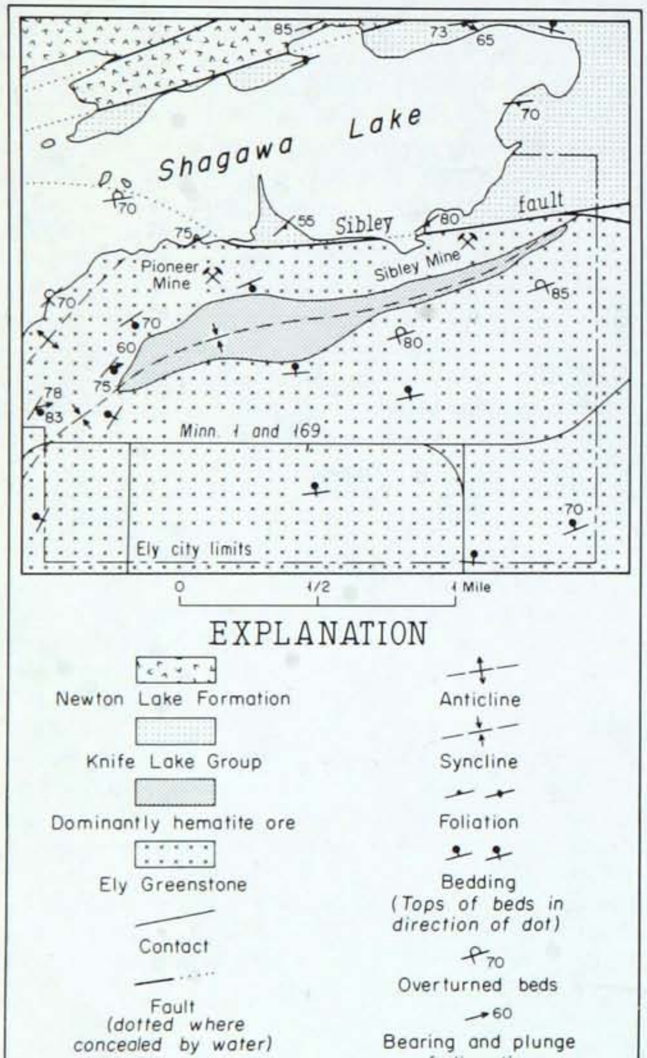


Figure III-77. Geologic map of Ely area. Geology by J. C. Green, 1970; base from Ely 7.5-minute quadrangle.

As a result of subsidence over the mine workings, neither the iron ore nor the iron-formation is now accessible for study. Reports by Reid (1956) and Machamer (1968) are the principal sources of information on the geology and ore deposits.

The iron-formation is a synclinal body within pillowed metabasalts in the upper part of the Ely Greenstone, as redefined by Morey and others (1970). Its exact stratigraphic position is not known. At the western end, the keel of the synclinal lens plunges about 45° E. (fig. III-78; Reid, 1956, p. 139), approximately parallel to the lineation given by the cleavage-bedding intersection. The factors responsible for the eastern termination of the lens are equivocal. Reid (1956, p. 142) suggested that the termination is primarily a stratigraphic pinch-out. W. P. Wolff (1969, written comm.) asserted, however, that the iron-formation and iron ore are truncated by the Sibley fault, which dips 60°-70° S. (fig. III-77). The intersection of the fault and iron ore plunges westward.

Judged from available descriptions, the iron-formation in the Ely trough consists of jaspilite and lesser lean jasper and red pyritic chert (Machamer, 1968, p. 13). The jasper is intruded by irregular bodies of dacitic porphyry, some of which are moderately large, and by more mafic dikes presumed to be diabasic gabbro (Reid, 1956, p. 138-139).

The iron ore—fragmental hematite which in the deeper parts of the mine workings is cemented to varying degrees by hematite or calcite—occurs as replacement bodies in the lower part of the folded iron-formation. The iron ore cropped out in the keel of the syncline at the western end (fig. III-78) of the body; it plunges eastward beneath the jaspilite "capping," and extends to depths below the mine workings. Typical relationships of iron ore to the jaspilite host rock in the Pioneer mine are shown diagrammatically in the sections in Figure III-79. As shown in the sections, the large ore body in the keel of the syncline, which plunges into the Pioneer from the Chandler mine (fig. III-79A) is abruptly split into two separate bodies by a broad mass of altered country rock—"paint rock" (fig. III-79B). The ore bodies continue intermittently on both limbs eastward through the Pioneer into the Zenith mine. On the north limb, the ore body attains a maximum width of about 400 feet in the Pioneer mine, then thins to a narrow stringer which pinches out within a distance of a few hundred feet into the Zenith mine (see Reid, 1956, fig. 19). The ore body on the south limb, although generally narrower, extends from the Pioneer into the Zenith and is continuous with the main ore body that was mined in the Zenith, Sibley, and

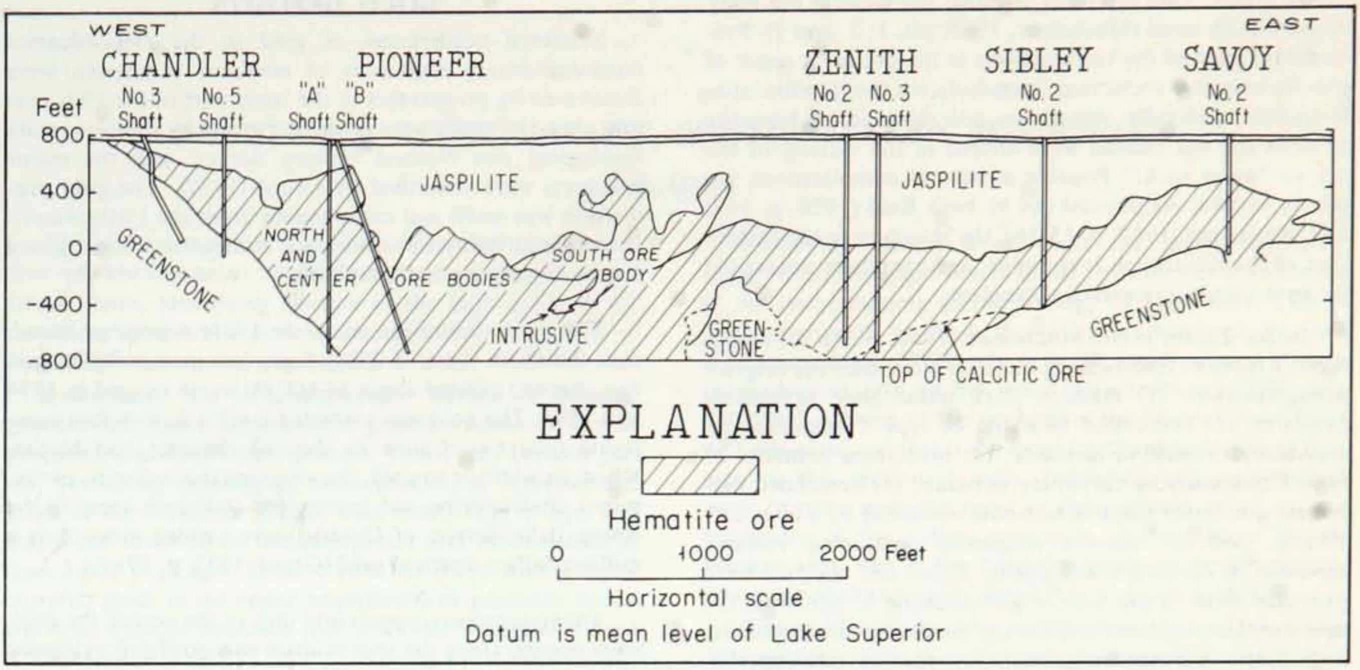


Figure III-78. Longitudinal projection of ore bodies in Ely trough. After Reid (1956, fig. 20); diagonally lined area, hematite ore.

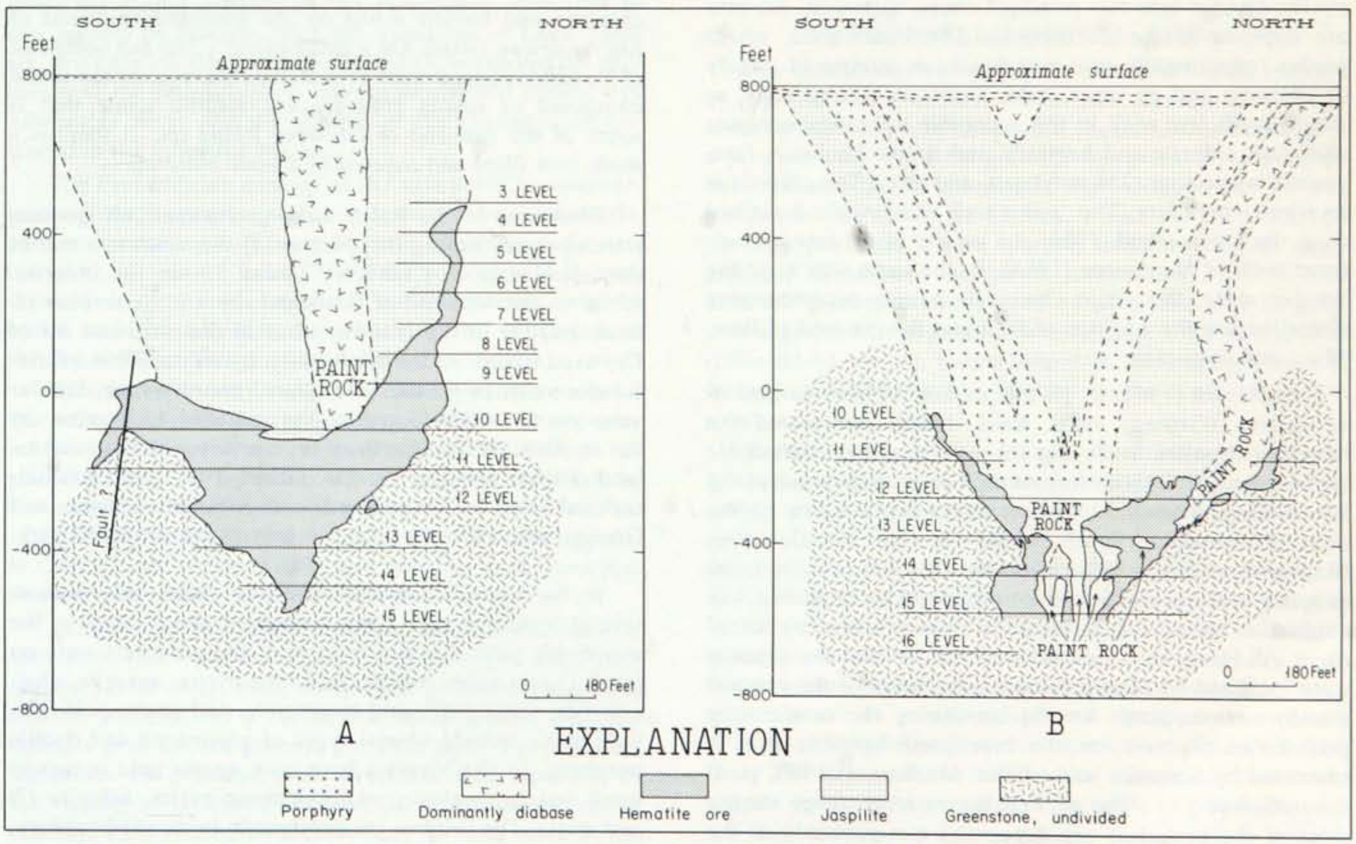


Figure III-79. Geologic sections of Pioneer mine (after Reid, 1956, figs. 21 and 22). A, section 300E; B, section 1900E.

Savoy mines. This ore body is called the Central ore body in the Zenith mine (Machamer, 1968, pls. 1, 2, and 3). Presumably, most of the interruptions in the ore are a result of pre-ore intrusive rocks that irregularly cut the jaspilite iron-formation; generally these were not replaced by hematite to form ore but instead were altered in the vicinity of the ore to "paint rock." Possible structural complications are poorly known. As pointed out by both Reid (1956, p. 144) and Machamer (1968, p. 15-16), the structure in the eastern part of the Ely trough is complex, and cannot be accounted for by a simple, symmetrical syncline.

In the Zenith mine, Machamer (1968, p. 20-30) recognized 5 types of iron-bearing material other than the original iron-formation: (1) massive, dark bluish-gray, crystalline hematite; (2) brecciated hematite of type 1 cemented by secondary crystalline hematite; (3) brecciated hematite of type 1 cemented by carbonate minerals; (4) brecciated carbonate and lesser massive hematite cemented by a later carbonate; and (5) massive magnetite containing variable amounts of carbonate and pyrite. Types one and two have provided most of the iron ore. In addition to hematite, the ores contain moderate amounts of magnetite, hausmannite, and pyrite and sparse goethite, pyrrhotite, chalcopyrite, covellite, chalcocite, and native copper. Non-metallic gangue minerals, in addition to the carbonates calcite and dolomite, include quartz, chlorite, kaolinite, apatite, poorly crystalline phosphate minerals, topaz, tourmaline, and gypsum.

A halo of altered greenstone (mainly metabasalt), which can be divided into two principal zones, surrounds the iron ore deposits at the Zenith mine. An outer zone, which grades transitionally into metabasalt, is composed largely of chlorite; and an inner, more intensely altered zone, as much as 40 feet wide in the accessible workings, contains abundant chlorite and hematite and sparse kaolinite, fine-grained muscovite (2M polytype), and illite. The chlorite is an iron-rich variety. The "paint rock" commonly described from the mines (Reid, 1956) is a variety of the intensely altered rock of Machamer (1968). In the same way as at the Soudan mine, the major chemical changes accompanying alteration are the addition of iron and the removal of lime, silica, and magnesia.

Machamer (1968, p. 36-39), as well as most earlier investigators (Gruner, 1926; Reid, 1956), interpreted the hematite deposits as having formed by post-metamorphic replacement of the iron-formation, and perhaps some of the surrounding greenstone, by iron oxides deposited by hydrothermal solutions. From studies at the Zenith mine, Machamer (1968, p. 35) concluded that ". . . at any point in space and time, the first iron oxide to be deposited was magnetite; later it was oxidized to or surrounded by hematite. . . ." He further concluded (p. 38-39) that the deposits were localized by fracturing and brecciation of the original iron-formation. Later brecciation during the ore-forming period can account for the brecciated hematite that is cemented by hematite and calcite. Machamer (1968, p. 4) inferred that ". . . the general temperature range during most of the period of ore deposition was probably in the range 350° to 400° C."

GOLD DEPOSITS

Scattered occurrences of gold in the metavolcanic-metasedimentary sequences of northern Minnesota were discovered by prospectors in the latter part of the 19th century. Most of these were noted in reports by the Minnesota Geological and Natural History Survey, and the major prospects were described by Grout (1937). The gold production was small and came mainly from the Little American mine in Rainy Lake, which is within the area proposed for the Voyageurs National Park.

The Little American mine, on Little American Island, near the south shore of Rainy Lake (see map in Ojakangas, this chapter) yielded about $4,600.00 worth of gold in 1894 and 1895. The gold was extracted from a 4- to 6-foot composite quartz-vein zone in sheared chloritic and biotitic schist. In addition to gold, the vein contains ankerite, pyrite, minor chalcopyrite, and tourmaline. Adjacent areas in the Rainy Lake district of Ontario have yielded more than a million dollars worth of gold (Grout, 1937, p. 59).

Other prospects, apparently dug in the search for gold, were opened along the approximate east-northeast extension of the vein zone encountered in the Little American mine. A shallow shaft was sunk on a quartz vein on Big American Island, and an adit was driven on the south side of Bushyhead Island just above water level. The adit penetrated a 4-foot shear zone, which contains quartz and massive pyrite, as well as several subsidiary subparallel zones. Also, a shallow pit was sunk on a quartz-pyrite vein in mixed chloritic and biotitic schist on the mainland southeast of Big American Island. On a small island 1,200 feet southeast of Pedersons Island there is a shallow pit, and on the island composed of mixed chloritic and biotitic schist that is south of the east end of Dryweed Island (pl. 2) there is a shaft, now filled and covered by a small building.

Elsewhere in the Rainy Lake greenstone belt, gossans were observed during mapping at (1) the southwest end of the island east of Cranberry Island, (2) in the chloritic schist on the west end of the island south of Steamboat Island, and (3) in the chlorite schist at the northeast tip of Dryweed Island. At the last locality, pyrite occurs in quartz-siderite veins in sheared and altered country rock. Similar veins occur on the dump of the so-called Lyle mine, on the small island just north of the north tip of Dryweed Island. Earlier geologic reports (Grout, 1937) describe mineralized veins on Cranberry Island, Steamboat Island, and Grassy Island. Grassy Island is west of the proposed park.

In the Vermilion district, prospect shafts were sunk at several localities near Lake Vermilion about 1865 in the search for gold. For the most part, the prospects were on small quartz veins, which contained pyrite, ankerite, chalcopyrite, lesser rutile and tourmaline, and sparse gold. The wall rocks include several types of greenstone and dacitic porphyry. In the Virginia horn area, sparse gold occurs in small quartz veinlets containing some pyrite, ankerite (?) and albite. The veinlets are dominantly in dacitic porphyry, metabasalt, and meta-tuff(?).

MINNESOTA RIVER VALLEY, SOUTHWESTERN MINNESOTA

J. A. Grant

The Minnesota River Valley provides a tantalizing window into the Canadian Shield on the eastern margin of the Great Plains, tantalizing because of the high grade of the metamorphism, and especially because of the antiquity of the rocks there exposed (fig. III-80).

Essentially, this is a migmatitic terrane of granitic gneisses and lesser amphibolitic gneisses, commonly with pyroxene, and biotite-rich gneisses, which may contain garnet, cordierite, sillimanite, anthophyllite, or hypersthene. Some of the rocks are greater than 3,000 m.y. in age, and they were involved in metamorphism and deformation at least 2,600 m.y. ago. These events left rocks with a metamorphic grade in the upper amphibolite or granulite facies, and with a major structure that is similar throughout most of the exposed area.

Later small, dominantly mafic intrusions cut the older rocks, and conglomerate and quartzite of the Sioux Quartzite of Late Precambrian age locally overlie them.

Deep weathering of the gneisses formed a regolith, commonly about 100 feet thick, part of which was reworked to form Cretaceous deposits of sandstone and shale. Over this came the glacial deposits of the Pleistocene, discussed in this volume by Matsch. With the formation of Lake Agassiz, drainage via Glacial River Warren scoured out the precursor of the present valley, leaving an underfit present-day Minnesota River and the glimpse of the Precambrian discussed in the following pages.

The Precambrian geology of the valley will be summarized in four geographic segments from northwest to southeast, encompassing a distance of some 120 miles and a valley width of about 2 miles. Then, the geology will be synthesized in terms of structure, metamorphism, original rocks, and geochronology, emphasizing the coherence of the Precambrian history of the valley.

References to early reports on the crystalline rocks of the valley are given in Himmelberg (1968), and one may note, in addition, the contribution of Upham (1883) to the glacial geology of the valley and that of Goldich (1938) to the understanding of rock weathering. A review of later investigations, both petrologic and geochronologic, appears in Goldich and others (1970), and only the work most pertinent to this chapter will be noted here.

Lund (1956), in a report on a reconnaissance of the Precambrian rocks of the valley, divided them into three groups: (a) an older, basic complex of gabbroic and quartz dioritic gneisses; (b) a Minnesota valley granite series of younger granites and granitic gneisses; and (c) post-granite intrusions. He provided some structural data, especially in the vicinity of Granite Falls, and delineated most areas of outcrop. This yielded a valuable base for later work in the valley.

Following Lund's work came the geochronological study by Goldich and others (1961), which was a milestone in the understanding of the geology of Minnesota. This is now superseded, so far as the valley is concerned, by Goldich and others (1970), which is the major source of geochronological data for the valley. Himmelberg (1968) gave a detailed account of the geology of the Montevideo-Granite Falls area, and this is the basis for the description of that area in this section. Finally, published and unpublished work of mine provides most of the geologic detail on the remainder of the valley, especially in the Sacred Heart-Morton area, and is the basis for much of the interpretation given in the synopsis of the Precambrian geology.

The nomenclature for granitic rocks used in this paper differs from that used to describe the Vermilion and Giants Range batholiths. With respect to Figure III-36 (Southwick, this chapter), field 3b herein is called quartz monzonite and granite is restricted to field 3a.

Reviews of this section by G. R. Himmelberg, P. K. Sims, and S. S. Goldich are gratefully acknowledged.

ORTONVILLE-ODESSA AREA

The Ortonville-Odessa area is underlain almost entirely by granitic rocks, which extend into adjacent South Dakota. In the vicinity of Ortonville, the dominant lithology is a purplish-pink, medium-grained, equigranular to porphyritic, foliated quartz monzonite that contains two to five percent biotite (table III-43, nos. 1 and 2). This rock apparently postdates a rare, purple, medium-grained, equigranular hornblende monzonite (table III-43, no. 3), and both are permeated and veined by abundant pink to gray, coarse-grained to pegmatitic leucogranite with blue quartz (table III-43, no. 4). Emplacement of the leucogranite was in part controlled by the foliation in the dominant quartz monzonite.

To the southeast, near Odessa, the dominant rock type is similar to that near Ortonville, but is more commonly porphyritic and more gneissic, and contains as much as 10 percent biotite (table III-43, nos. 5 and 6). Especially southeast of Odessa there are abundant schlieren, lenses, or bands of biotite-quartz-plagioclase schist with concordant and discordant granitic veins. These rocks range from banded granofels to schists and gneisses, and are generally gray, fine to coarse grained, equigranular, and granoblastic. They may contain as additional phases clinopyroxene (Lund, 1956, p. 1481), orthopyroxene, garnet, or potassium feldspar. The compositions and banded nature of these rocks suggest a sedimentary origin, in the realm of graywacke.

Near Odessa, a few uralitized diabase dikes cut the structures in the rocks described above.

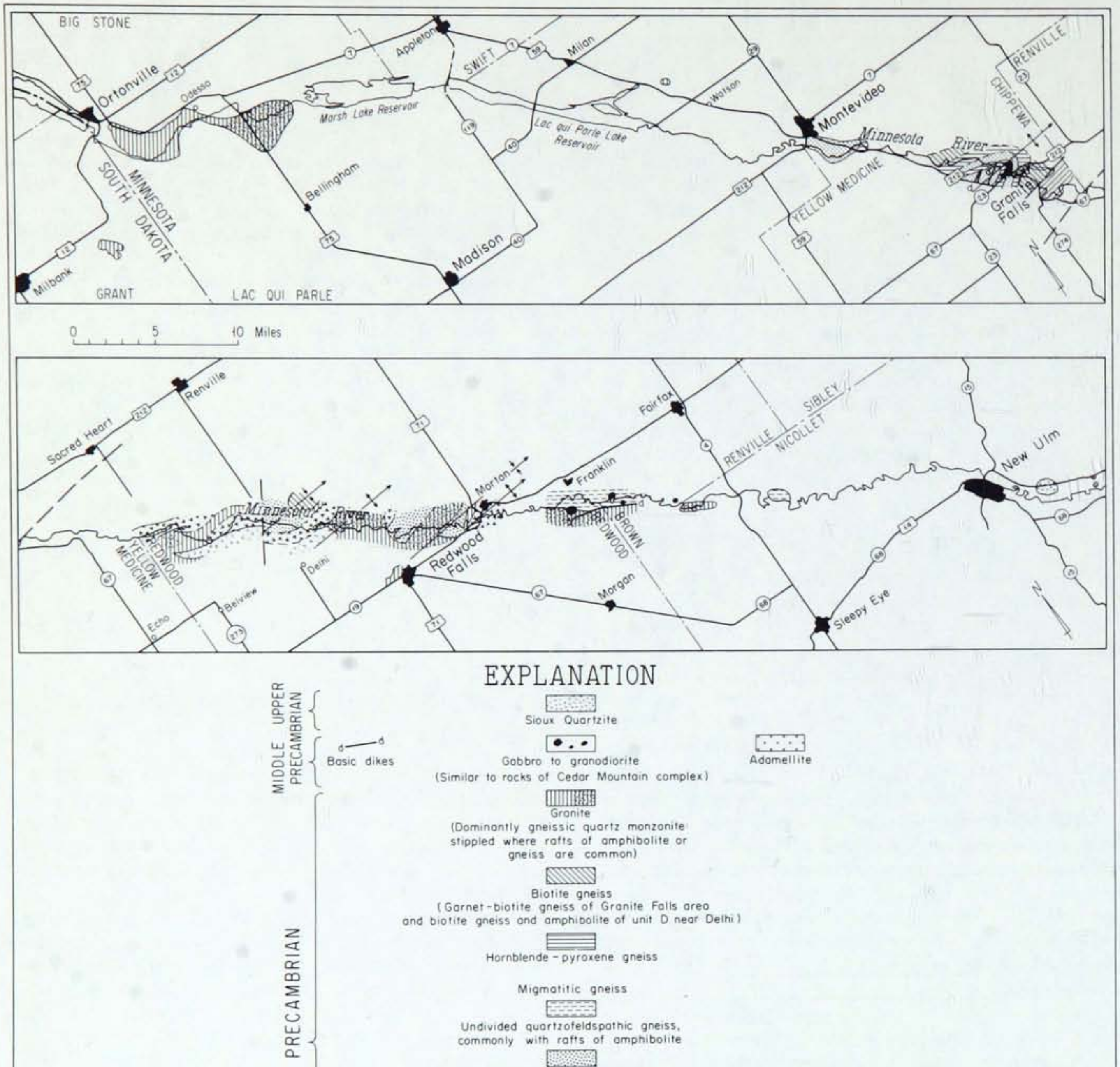


Figure III-80. Geologic map of Minnesota River Valley. Geology by J. A. Grant, 1965-70.

Table III-43. Modes, in volume percent, of rocks in the Ortonville-Odessa area.

	1	2	3	4	5	6	7a	7b	8
Quartz	30.1	21.1	5.4	33.9	27.7	27.8	9.5	35.9	32.5
K-feldspar	24.1	36.3	23.7	41.1	32.2	21.6		2.4	
Plagioclase	33.6	33.8	48.6	18.9	29.9	31.5	59.6	58.4	32.4
Myrmekite	5.1	2.4	2.4	5.7	4.8	5.3			Tr
Chlorite	Tr	Tr				Tr			
Muscovite		Tr		Tr		2.7			
Biotite	4.3	4.0	1.4	0.4	2.5	10.1	15.6	0.9	34.5
Actinolite									
Hornblende			14.3						
Orthopyroxene							14.1	2.4	
Garnet							0.6		
Epidote		Tr	Tr			Tr			
Sphene		Tr	Tr		0.7				
Zircon	Tr	Tr				Tr	Tr		Tr
Apatite	0.8	1.0	0.7		Tr	0.5			0.4
Opaques	2.0	1.4	3.5	Tr	2.2	0.5	0.6		0.2

1: Quartz monzonite (M14734), SE¼sec. 22, 121N/46W
2: Quartz monzonite (M14746), SW¼sec. 25, 121N/46W
3: Hornblende monzonite (M15033), SE¼sec. 22, 121N/46W
4: Leucogranite (M14735), SE¼sec. 22, 121N/46W
5: Quartz monzonite (M14737), NE¼sec. 30, 121N/46W
6: Biotite quartz monzonite (M14807), sec. 12, 120N/45W
7: Adjacent portions of gneiss (M15028), NW¼sec. 9, 120N/45W
8: Biotite-quartz-plagioclase schist (M14759), NW¼sec. 15, 120N/45W

Structure

In this and ensuing discussions of structure, B refers to major fold axes and to linear elements essentially parallel to them, and A refers to linear elements approximately at right angles to major fold axes (Moench and others, 1962, p. 40).

Foliation in the granitic rocks is given by biotite-rich schlieren, aligned feldspar megacrysts, and compositional banding. The latter is commonly accompanied by difference in grain size, probably associated with emplacement of the leucogranite. In particular, near Odessa the structures resemble those at Morton (fig. III-82A). In the metasedimentary rocks, schistosity or gneissosity is marked by parallelism of biotites and compositional differences.

A compilation of foliations from this area is shown in Figure III-81A. With one proviso, the data form a partial girdle in which maxima and minima are related to availability of outcrop as much as to structure. The estimated pole to this girdle, the B-axis, is N. 70° E., 32°. However, close to this axis appears a cluster of poles to foliation, all of which come from the same locality. Separate analysis of these outcrops, as well as those adjacent, suggests a local fold on an axis approximately S. 12° E., 6°. There is no local evidence of different ages for the two fold sets.

Although one might doubt the relevance of some of the compositional banding in the granitic rocks to folding (as opposed to control of the development of leucogranite by pre-existing structures including jointing), it seems impossible to view such outcrops as that of Figure III-82A without concluding that the dominant foliations in both granitic rocks and inclusions have been subjected to the same deformation.

Lineations are rare in the granitic rocks, but rodding, mineral elongation, and minor folds are found in the inclusions. There is too much scatter of the rare mineral lineations to be meaningful, but the minor fold axes cluster about N. 75° E., 34°. This is in close agreement with the B-axis from the foliation data (fig. III-81A), corroborating the impression that the structures in the granitic rocks and inclusions are congruent. Apparently the dominant quartz monzonite (and the earlier phase) and the inclusions have been subjected to the same deformation, whereas the leucogranite is synchronous with, or later than, this. Very similar relationships are found at the isolated outcrop at Watson Sag, some 30 miles down the valley.

Metamorphism

In the inclusions, the coexistence of quartz-antiperthitic plagioclase-biotite-orthopyroxene, with potassium feldspar or garnet, places these rocks in at least the transition between the amphibolite and granulite facies. These assemblages are duplicated in the valley to the southeast. The

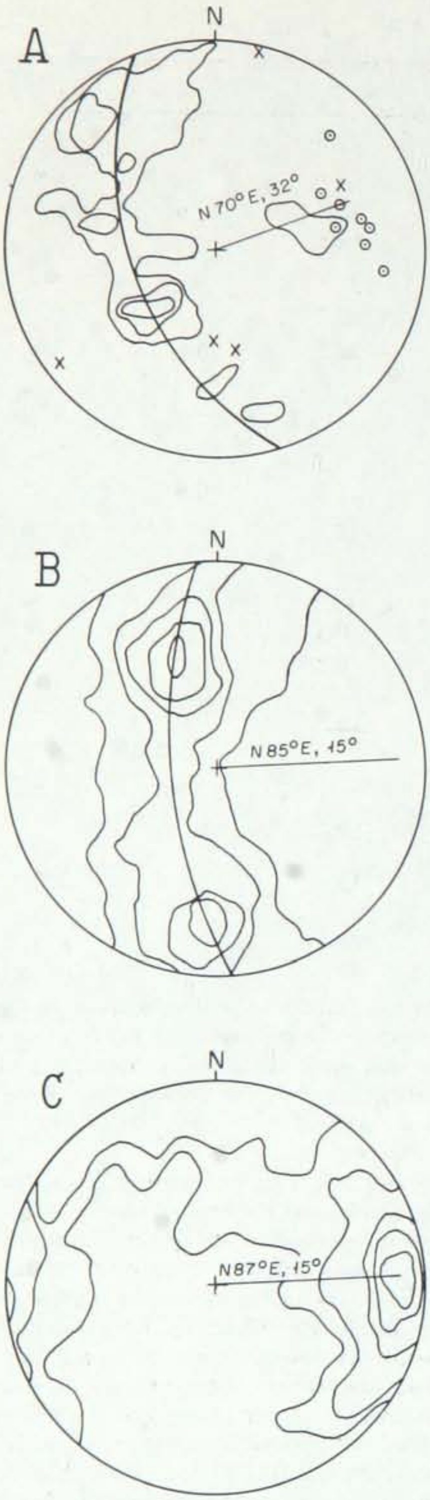


Figure III-81. Schmidt net diagrams. A, foliation in Ortonville-Odessa area. Contours 2.5, 5, and 7.5 percent. B, foliation in Montevideo-Granite Falls area (after Himmelberg, 1968, fig. 8); contours 0.14, 1, 3, 6, and 15 percent. C, lineation in northwest Granite Falls area (after Himmelberg, 1968, fig. 8); contours 0.26, 1, 5, 10, and 15 percent.

common assemblage of the granitic rocks, quartz-potassium feldspar-plagioclase-biotite ± hornblende is compatible with this, but it is uncertain whether the muscovite of sample 6 (table III-43) should be included in the granitic assemblage.

MONTEVIDEO-GRANITE FALLS AREA

The Montevideo-Granite Falls area is one of the two best exposed and most informative areas in the Minnesota River Valley. It was studied recently by Himmelberg (1968), and the following description is largely from his work.

The area is underlain by granitic gneiss, hornblende-pyroxene gneiss, and garnet-biotite gneiss, which generally constitute units 1,000 to 5,000 feet thick, although smaller-scale interlayering is common. Possibly a syncline exists between Montevideo and Granite Falls, but the major exposed structure is an eastward-plunging anticline at Granite Falls. Southeast of Granite Falls, the exposures end near an inferred major fault zone (see Austin and others, 1970).

The granitic gneiss is a pink to red, medium-grained, equigranular, leucocratic rock, which commonly has compositional layering resulting from alternations of biotite-rich and quartzofeldspathic layers. This layering is commonly paralleled by biotite plates and locally by flat quartz lenses. Relatively massive granite, granitic pegmatites, and bluish quartz lenses parallel and crosscut the foliation of the gneiss. Minor folding, warping, and incipient boudinage structures are common. Numerous hornblende-pyroxene gneiss layers and lenses of variable thickness are present, but no inclusions of garnet-biotite gneiss are known.

The granitic gneiss consists dominantly of quartz, microcline microperthite, antiperthitic plagioclase, and brown biotite, with local development of sparse garnet. The feldspars are clouded by opaque iron oxides and, in some of the red outcrops at Montevideo, by hematite. Representative modes of this and the other major rock types of the area are given in Table III-44.

The hornblende-pyroxene gneiss is a gray-black, medium-grained, equigranular rock that varies from a uniform amphibolite to a banded gneiss. The amphibolite occurs most commonly as layers or lenses in the granitic gneiss (the dominant mode of occurrence near Montevideo) and as discrete layers within the hornblende-pyroxene gneiss. Sharp conformable contacts with adjacent rocks are the rule, and the long axes of lenses parallel the foliation.

The extensive hornblende-pyroxene gneiss immediately south of Granite Falls is a gray gneiss that has a compositional banding resulting from different proportions of individual minerals. However, around the nose of the anticline northwest of Granite Falls, the unit consists of a heterogeneous interlayered series of mafic layers rich in hornblende and pyroxene but poor in quartz, quartzofeldspathic layers with subordinate hornblende and pyroxene, and layers of pegmatitic granitic gneiss.

Quartz lenses and veins are parallel to and crosscut the foliation. Structures such as minor folding, warping, and boudinage are rare in these rocks, but hornblende-pyroxene rodding is found locally.

The principal minerals in most of this gneiss are antiperthitic plagioclase, greenish-brown hornblende, orthopyroxene, pale green clinopyroxene, brown biotite, and opaque

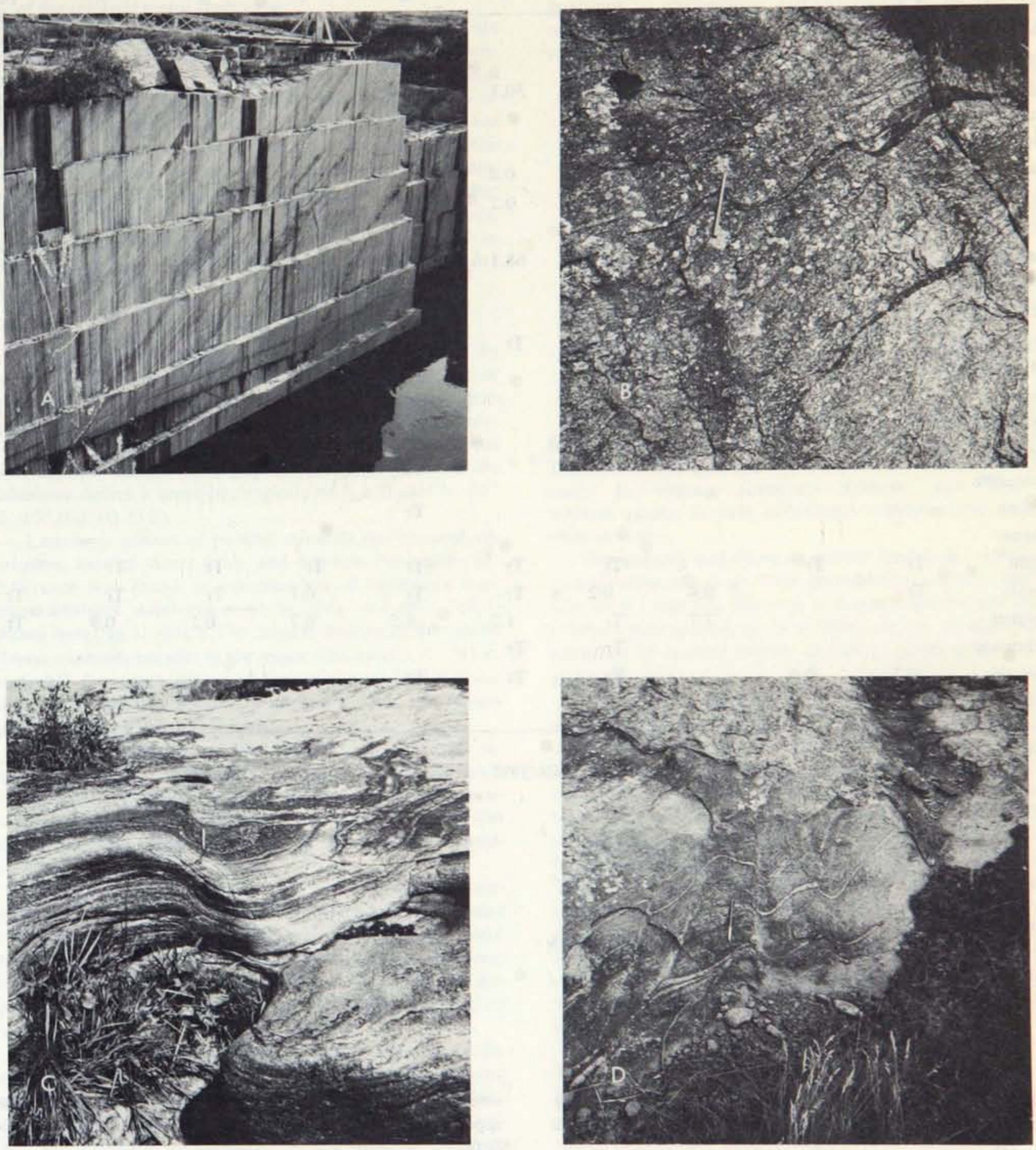


Figure III-82. Photographs of rocks in Minnesota River Valley. A, structure in gneissic granite containing inclusions, Odessa, NE¼ sec. 30, T. 121 N., R. 45 W. B, schlieren in quartzofeldspathic gneiss, with F_2 deformation; pencil lies N. 55° E. SE¼ sec. 32, T. 114 N., R. 36 W. C, compositional banding in quartzofeldspathic gneiss. NW¼ sec. 26, T. 114 N., R. 37 W. D, discrete granitic veining in quartzofeldspathic gneiss, with F_2 warping: note development of some granitic material subparallel to F_2 axial plane. NW¼ sec. 11, T. 113 N., R. 36 W.

Table III-44. Modes, in volume percent, of rocks in the Montevideo-Granite Falls area (after Himmelberg, 1968, tables 2, 3, and 4).

	1	2	3	4	5	6	7	8	9	10
Quartz	25.0	33.1	1.2			34.2	22.5	25.3	21.4	29.5
K-feldspar	15.9	39.5				1.1	0.2	0.3	0.6	
Plagioclase	54.0	26.1	65.4	41.1	30.1	50.4	59.8	50.3	42.6	44.5
Myrmekite										
Chlorite										Tr
Muscovite	Tr	Tr	Tr	Tr	0.4	Tr	Tr	Tr	Tr	Tr
Biotite	5.0	0.9	Tr	Tr	0.2	Tr	11.9	14.7	22.5	19.7
Actinolite				0.2						
Hornblende			0.5	36.4	68.1	5.0	1.2			
Cummingtonite				Tr						Tr
Anthophyllite										
Clinopyroxene			6.2	9.2	Tr	2.4	0.1			
Orthopyroxene			6.6	12.9			3.5	9.3	7.5	
Olivine										
Cordierite										
Garnet		Tr	12.0			0.2			4.5	6.3
Sillimanite										
Epidote						Tr				
Sphene										
Zircon	Tr	Tr		Tr	Tr	Tr	Tr	Tr	Tr	
Apatite	Tr		0.4	0.2	Tr	Tr	0.1	Tr	Tr	Tr
Opaques			7.7	Tr	1.2	6.5	6.7	0.1	0.9	Tr
Carbonate				Tr	Tr					
Hematite	0.1	0.4		Tr	Tr	Tr				
Pyrite			Tr						Tr	

Granitic gneiss:
1: (M8115)—Great Northern railroad cut; NW¼ sec. 28, 116N/39W
2: (M8116)—NW¼ sec. 28, 116N/39W
Hornblende-pyroxene gneiss:
3: (M8079)—NW¼ sec. 3, 115N/39W
4: (M8140, amphibolite lens)—NW¼ sec. 32, 116N/39W
5: (M8141, amphibolite lens)—NW¼ sec. 32, 116N/39W
6: (M8182)—Cen. of sec. 33, 116N/39W
7: (M8205)—Cen. of sec. 4, 115N/39W
Garnet-biotite gneiss:
8: (M8096)—At contact on Minn. Hwy. 67, sec. 3, 115N/39W
9: (M8219)—Minn. Hwy. 67, sec. 3, 115N/39W
10: (M8086)—NW¼ sec. 10, 115N/39W

oxides. Replacement of the amphibole and pyroxenes by cummingtonite, blue-green amphibole, and serpentine is common.

The garnet-biotite gneiss is a dark-gray, medium-grained, equigranular, well foliated gneiss, that has bands of light-gray, coarse-grained, granular gneiss, and a preferred orientation of biotite and prismatic orthopyroxene. The northern contact between these rocks and the hornblende-pyroxene gneiss is well exposed (on Minn. Hwy. 67), where it is conformable and essentially marked by the first appearance of garnet. The southern contact probably is a fault.

Quartz, antiperthitic plagioclase, reddish-brown biotite, and garnet are the major minerals, and orthopyroxene is common in the well foliated gneiss. There is a conspicuous lack of both clinopyroxene and hornblende.

A heterogeneous series of interlayered gneisses—leucogranitic gneiss, hornblende-pyroxene gneiss or amphibolite, hornblende-biotite-quartz-plagioclase gneiss, and hornblende-bearing granitic gneiss—crops out in the southeasternmost part of the area. The relatively fine-scale layering that characterizes these rocks may be related to the proximity of the inferred major fault zone.

Post-metamorphic mafic dikes as much as 75 feet wide cut the gneisses. Most are composed of dark-gray, medium-grained tholeiitic diabase having fine-grained margins. Replacement of augite by green hornblende is common. Hornblende andesite forms abundant, narrower dikes that are grayish-black and porphyritic-aphanitic and have phenocrysts of plagioclase or quartz. Black, aphanitic olivine diabase dikes are rare, and their relationships to the other mafic dikes are not known. At one locality, a pink, medium-grained biotite adamellite ("granite of section 28" in Goldich and others, 1970) intrudes a hornblende andesite dike.

Structure

The rocks described above have foliations defined by compositional banding, preferred orientation of planar or prismatic minerals, flat quartz lenses, and hornblende segregations. Although no differences in the attitudes of the several foliations have been found, some of the compositional banding may be pre-metamorphic in origin. Poles to the foliations define a great-circle girdle with a B-axis N. 85° E., 15° (fig. III-81B).

Lineations consist of parallel minerals and mineral aggregates, axes of minor folds, and boudins. No significant difference was found in compilations of lineations from three arbitrary subdivisions of the area, and only one is shown here (fig. III-81C). The range in bearing of B-maxima (linear elements parallel to the major fold axis) is N. 70° E. to S. 82° E., with a plunge of approximately 15°. A-lineations, with bearings essentially normal to this, are present locally (boudins and minor folds).

These data suggest a structure homogeneous with respect to all measured elements: a gently-plunging, inclined, cylindrical fold system, having at least monoclinic symmetry. The B-axis from foliation data (N. 85° E., 15°) and the B-axis from lineation data (N. 88° E., 15°) are considered equivalent.

Shear zones generally less than a foot wide are common, and may be marked by mylonite or other cataclastic rocks or stringers or lenses of granitic pegmatite. Most trend N. 35°-60° W., although a few trend northeast; mainly the dips are nearly vertical. Evidence for left-lateral displacement was noted on some of the northwest-trending shear zones.

A major, eastward-trending fault zone near the southeastern limit of outcrop is inferred from aeromagnetic and gravity data (Zietz and Kirby, 1970; Craddock and others, 1970) and from the tightly appressed compositional banding and evidence of cataclasis in the adjacent outcrops themselves.

All the mafic dikes cut the foliation of the metamorphic rocks, and dominantly occupy a nearly vertical fracture system trending approximately N. 55° E. The small shear zones mentioned above postdate the tholeiitic diabase, but predate the hornblende andesite dikes.

Metamorphism

A detailed study of the metamorphic petrology is given by Himmelberg and Phinney (1967). They listed (p. 329-330) twenty-two assemblages considered to represent a close approach to equilibrium, and the following are the maximum-phase assemblages with the corresponding numbers in the list given in parentheses:

1. quartz-potassium feldspar-plagioclase-biotite-hornblende-clinopyroxene-orthopyroxene-magnetite-ilmenite (2)
2. quartz-potassium feldspar-plagioclase-garnet-hornblende-clinopyroxene-magnetite-ilmenite (4)
3. quartz-plagioclase-hornblende-sphene-hematite (5)
4. quartz-plagioclase-garnet-hornblende-clinopyroxene-orthopyroxene-magnetite-ilmenite (6)
5. plagioclase-biotite-hornblende-clinopyroxene-orthopyroxene-hematite (12)
6. quartz-potassium feldspar-plagioclase-biotite-garnet-orthopyroxene-magnetite-ilmenite (17)
7. quartz-potassium feldspar-plagioclase-biotite-garnet-rutile-hematite (20)

The first five are from hornblende-pyroxene gneiss, 6 is from garnet-biotite gneiss, and 7 is from the granitic gneiss. The common assemblages from these three units are respectively 1, without potassium feldspar and with or without quartz, 6, with or without orthopyroxene, and 7, without rutile.

Himmelberg and Phinney (1967) found no empirically incompatible phases in these assemblages, regular distribution of Fe and Mg between coexisting hornblende-clinopyroxene-orthopyroxene, and only minor compositional variation in mineral grains, and for a given mineral in a given specimen.

Himmelberg and Phinney (1967) found no distribution of assemblages suggestive of metamorphic zoning, nor could any mineralogic isograds be drawn. The occurrence of pyroxene or hornblende at any locality appeared to be chemically controlled, rather than due to fluctuations in temperature or pressure. Thus, these assemblages are considered to be isofacial, and yet cannot be assigned to a hornblende-granulite, or pyroxene-granulite facies (see Turner, 1968, p. 186 and p. 320-336). They are simply assigned to the granulite facies, on the basis of (a) the lack of quartz-muscovite in rocks of appropriate composition, (b) the presence of orthopyroxene rather than Ca-free orthoamphibole and (c) the common occurrence of the typical assemblage plagioclase-hornblende-clinopyroxene-orthopyroxene (Turner, 1968, p. 320).

No earlier metamorphic or deformational events have been recognized here, although one might argue on the general basis that the low P_{H_2O} apparently required for granulite facies metamorphism would be most readily attained after either multiple metamorphism or metamorphism of long duration. Because of the pervasive effect of the metamorphism, the original nature of the three main gneiss units can only be conjectured from their bulk compositions and from the structural relations among them. The garnet-biotite gneiss is probably derived from a graywacke-like sedimentary protolith, the hornblende-pyroxene gneiss and the granitic gneiss from mafic and felsic protoliths, respectively. This is discussed further later in this paper.

Himmelberg and Phinney discussed the retrograde metamorphism (1967, p. 341-347), and suggested that it may be related in part to a discrete, later event.

SACRED HEART-MORTON AREA

After a hiatus of about ten miles, the other well exposed and informative section of the valley begins, extending from near the quarries south of Sacred Heart southeast to Morton, a distance of about 22 miles. In this area, it has been possible to delineate gross stratiformity in the migmatitic terrane, four units on the order of a few thousand feet thick having been mapped on the basis of lithologic similarity. In ascending order, the first three units are quartzofeldspathic gneisses, which, respectively, have abundant, common, and rare rafts of amphibolite, whereas the uppermost unit consists of biotite-rich gneisses and lesser amphibolite. Distinguished from these rocks is quartz monzonite, forming concordant and discordant bodies in the gneisses. Except for some of the discordant quartz monzonite, these rocks have been deformed into major folds, with wave lengths on the order of a few miles, and shallow eastward-plunging axes. From north to south, the main folds are a synclinorium and anticlinorium north of Delhi and a synclinorium and anticlinorium in the vicinity of Morton. The grade of metamorphism is upper amphibolite facies throughout, and no mineralogic isograds have been drawn here. Rarely, late mafic dikes are present.

The lowermost unit (A) consists of interlayered amphibolite and quartzofeldspathic gneiss. It is best exposed in the core of the Delhi anticlinorium, and has been traced northwestward on the south side of the valley into the vicinity of the Sacred Heart quartz monzonite. Where best exposed, there are two layers of amphibolite on the order of 100 feet thick, bounded by quartzofeldspathic gneiss with or without amphibolite rafts. The amphibolites are rarely more than a few tens of feet long, and quartzofeldspathic veins are both parallel to and across the layers, yielding a network-breccia which grades into quartzofeldspathic gneiss with amphibolite rafts.

The next unit (B), consisting of quartzofeldspathic gneiss with amphibolite rafts, is about 1,500 feet thick on the south side of the Delhi synclinorium, but to the northwest it is correlated with a lithologically and stratigraphically similar unit twice as thick, and this is tentatively correlated with the gneiss on the north side of the Sacred Heart pluton. To the south, this unit is again recognized in the core of the Morton anticlinorium, the correlation being based on the abundance of amphibolite rafts in the gneiss and the stratiform position below a unit lacking such inclusions. This, of course, is the type locality for the Morton Gneiss, and the point to be made here is that while this rock has many similarities with "Morton Gneiss" as mapped elsewhere in the valley (Lund, 1956), it is but one variant within one unit in a sequence of related migmatitic gneisses.

The third unit (C) is best developed around the Delhi synclinorium. It is a quartzofeldspathic gneiss, 1,500-3,000 feet thick, in which amphibolitic rafts are uncommon. Although similar rocks are found locally to the northwest, there is insufficient development of this rather uniform lithology to warrant defining additional subunits. To the southeast, as mentioned above, this is correlated with the rocks above and north of unit B, in the core of the Morton synclinorium.

The quartzofeldspathic gneiss, where it is monolithologic, is a gray to pale pink, medium- to coarse-grained biotite-quartz-plagioclase gneiss, locally containing minor green hornblende or potassium feldspar. Not uncommonly, plagioclase forms megacrysts, and mafic schlieren are present (fig. III-82B). Compositional banding, reflecting differences in the proportions of the major minerals, is locally well developed (fig. III-82C). Especially in the vicinity of the Sacred Heart pluton and from Delhi south to Morton, permeation by granitic material ranges from discrete veins (fig. III-82D) to a nebulitic structure, and masses of gneissic quartz monzonite can be found. Where permeation is most intense, as near North Redwood, it may be difficult to separate the quartzofeldspathic gneiss from gneissic quartz monzonite. Ideally at least, the latter is monolithologic within the quartz monzonite range, whereas the former has layers both within and outside this range. Representative modes of the quartzofeldspathic gneiss are given in Table III-45.

Where the amphibolites are least modified, they are black, medium-grained, granular or foliated rocks composed dominantly of green hornblende and plagioclase, with varying amounts of brown biotite and generally little or no quartz (see table III-45). Clinopyroxene is not uncommon, but only where biotite or quartz or both are scarce or absent. Orthopyroxene was found in only one thin section, from Morton. Commonly, the amphibolite is gneissic and layered, with black amphibolitic layers alternating with gray quartzofeldspathic gneiss (fig. III-83A) and veinlets of similar material locally crossing the amphibolite, to yield a network-breccia (fig. III-83B). (An even more clear-cut breccia occurs where amphibolite is in contact with, and veined by, quartz monzonite.) There is gradation from this type of interlayering to separation of the amphibolite into rafts in a matrix of quartzofeldspathic gneiss, apparently due to the relative competence of the amphibolite. Where both lithologies are major components of the rock, dilatation structure can be found, the less competent gneiss having flowed between blocks of amphibolite. Where the less competent lithology is dominant, the amphibolitic rafts tend to be lensoid and schlieric.

In the case of the most complex migmatites, where three major lithologies are involved, as at Morton, schlieric rafts of amphibolite or of quartzofeldspathic gneiss, commonly with the former appearing as a core to the latter, lie in a matrix of granitic gneiss (fig. III-83C). One gets the impression of a sequence of differential competence in a flowing medium, with amphibolite being more competent than the gray quartzofeldspathic gneiss, which is more competent than the granitic component.

The uppermost unit (D) is restricted to the core of the Delhi synclinorium. It lies conformably above unit C, with some interdigitation of the lithologies. There is a complexly folded association of biotite-rich gneisses and amphibolite, in which discontinuous amphibolite layers and disharmonic

Table III-45. Modes, in volume percent, of quartzofeldspathic gneiss and amphibolite in the Sacred Heart-Morton area.

	1	2	3	4	5	6a	6b	7a	7b	7c	8a	8b	9a	9b
Quartz	3.1	1.7			23.5		26.9		35.9	29.4	32.1	22.6	34.6	26.8
K-feldspar		Tr	Tr		1.4		2.5		0.2	21.8	17.8	39.4	1.0	56.4
Plagioclase	30.9	33.5	34.0	37.0	56.5	35.1	61.5	29.0	59.8	38.9	42.7	26.6	63.2	14.2
Myrmekite					0.2					5.8	2.9	6.4	0.2	1.4
Chlorite			Tr		0.4		Tr			Tr	0.6	2.0		
Muscovite					0.2				Tr	Tr	0.6	0.4	0.6	1.2
Biotite			2.4		17.7		5.0	4.6	2.7	3.6	2.9	2.6		
Actinolite				3.5				Tr						
Hornblende	65.3	63.5	62.1	31.1		59.7	2.7	62.1	0.2					
Cummingtonite				2.8										
Anthophyllite														
Clinopyroxene	Tr			12.3		5.2	1.4	1.5						
Orthopyroxene				9.2										
Olivine														
Cordierite														
Garnet														
Sillimanite														
Epidote					Tr		Tr		Tr					
Sphene					Tr	Tr	Tr	0.6	Tr	Tr	Tr	Tr		
Zircon					Tr				Tr		Tr	Tr		
Apatite			0.4		0.2		Tr	Tr	0.2		Tr	Tr		
Opaques	0.7	1.3	1.1	4.1	Tr	Tr	Tr	1.8	1.0	0.6	0.4	Tr	0.4	
Carbonate					Tr			0.4	Tr					

1. Amphibolite (M14161)—SW¼sec. 21, 114N/37W
2. Amphibolite (M14136)—SW¼sec. 29, 114N/36W
3. Amphibolite (M14248)—NW¼sec. 18, 113N/35W
4. Amphibolite (M14553)—Morton, NW¼sec. 31, 113N/34W
5. Gray quartzofeldspathic gneiss (M14140)—SE¼ sec. 32, 114N/36W
6. Amphibolite (6a), veined by quartzofeldspathic gneiss (6b) (M 14968-9)—SW¼sec. 24, 114N/37W
7. Adjacent amphibolite (7a), quartzofeldspathic gneiss (7b), and quartz monzonitic gneiss (7c) (M14603-5)—Morton, SE¼sec. 31, 113N/34W
8. Adjacent bands in quartzofeldspathic gneiss (M14290)—East cen., sec. 19, 113N/35W
9. Adjacent bands in quartzofeldspathic gneiss (M14150)—West cen., sec. 7, 114N/37W

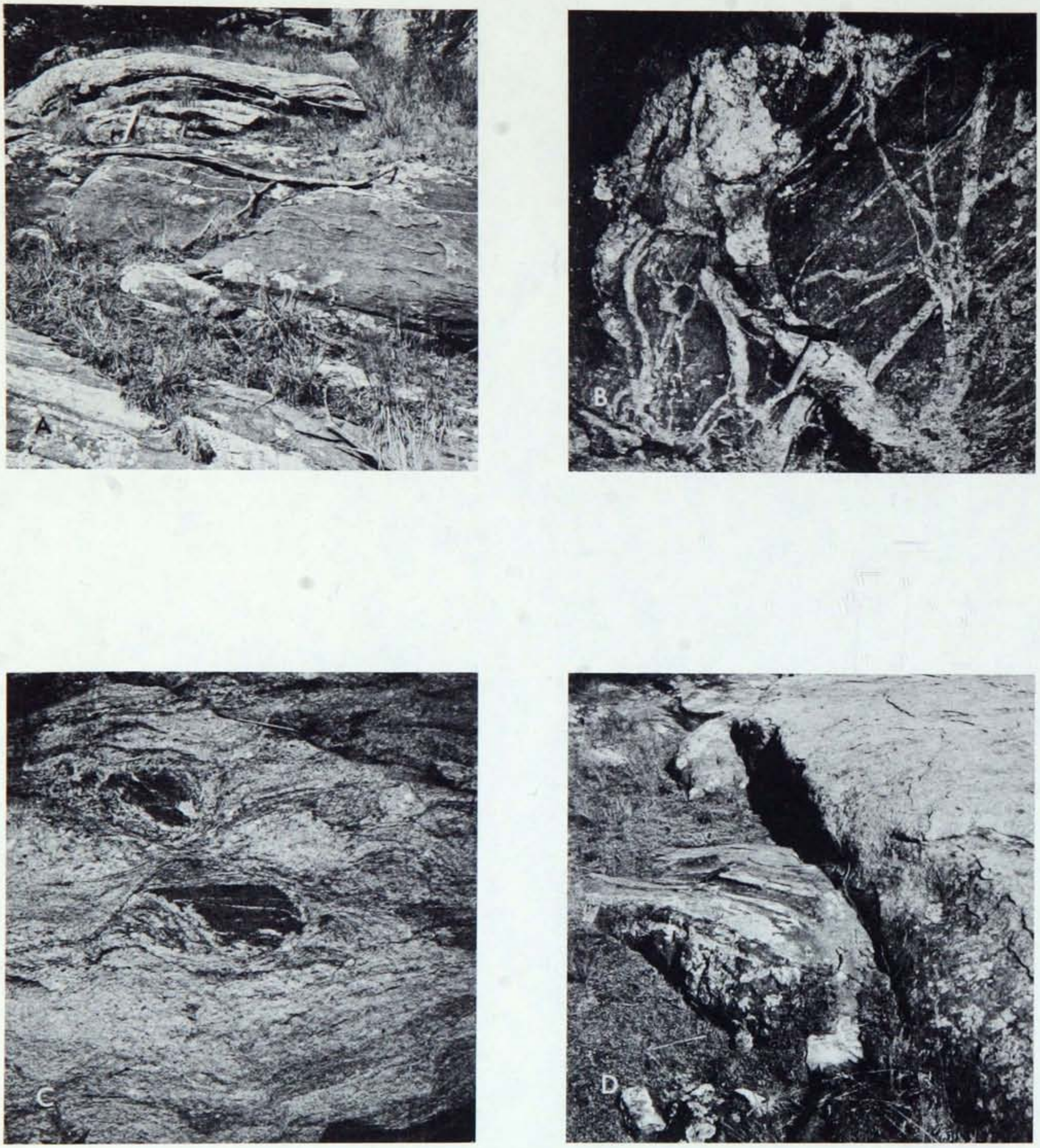


Figure III-83. Photographs of rocks in Minnesota River Valley. A, interlayered amphibolite and quartzofeldspathic gneiss. SE¼ sec. 23, T. 114 N., R. 37 W. B, network breccia of amphibolite with veinlets of gray quartzofeldspathic gneiss. SW¼ sec. 3, T. 113 N., R. 36 W. C, schlieric rafts of amphibolite and gray quartzofeldspathic gneiss in a matrix of pink and gray granitic gneiss, Morton. SE¼ sec. 31, T. 113 N., R. 34 W. D, discordant contact between hornblende-biotite-quartz plagioclase gneiss (on left) and quartz monzonite (on right). NE¼ sec. 12, T. 114 N., R. 38 W.

folds are common, again considered to be due to differences in the competences of the rocks involved. The amphibolite appears in discontinuous layers, lenses, and boudins, and is a black, medium-grained, granular, well foliated rock, consisting mainly of plagioclase and green hornblende, which is commonly associated, even in the same grain, with cummingtonite. On the margins of such bodies, rodded quartz-cummingtonite schist is common.

The biotite-rich gneisses are principally of two kinds. The dominant one is a gray-black, thinly banded gneiss, commonly containing the mineral association biotite-cordierite-garnet-anthophyllite in addition to quartz and plagioclase. The second is a heterogeneous, gray, banded gneiss, containing quartz, plagioclase, and biotite, with sillimanite knots and potassium feldspar-rich patches being common, and garnet or cordierite less common. Muscovite generally is developed around the sillimanite knots. Modes representative of the rocks of unit D are given in Table III-46.

A few small masses of coarse quartzofeldspathic gneiss are present in the area of unit D, and these may contain biotite, garnet, or anthophyllite. Also, a pink granitic dike, now albitized (?), traverses the extreme northeastern edge of the outcrop, and is at least spatially related to locally intense low-grade hydrothermal alteration of the gneisses.

The last major lithology to be described is quartz monzonite, which occurs principally as the Sacred Heart pluton, but also as major bodies northeast of Delhi and near North Redwood. Minor dikes of similar composition are found throughout, and the migmatites, as noted above, are locally permeated by granitic material. Representative modes are given in Table III-47.

The Sacred Heart pluton apparently intrudes the surrounding gneisses, the principal evidence for this being the dilatation implied by the correlation of the two amphibolite layers of unit A with amphibolite-rich zones near the northern and southern limits of exposure of the pluton, and the demonstrable discordance of several adjacent subconcordant dikes which closely resemble the dominant lithology in the Sacred Heart pluton, and are in part mapped as merging with it at its eastern end (fig. III-83D). The presence of rotated inclusions is not considered to be as good evidence as the above; it simply implies relative movement of the inclusions in a flowing matrix. The main body is typically a pink, medium-grained, homogeneous to faintly foliated and compositionally banded quartz monzonite, with as much as five percent biotite and two percent each of chlorite and muscovite, both of which may be secondary. On the north side of the pluton there is a distinctive salmon-pink, medium-grained granite with abundant rather nebulitic basic inclusions. A typical series of modal analyses from the granite into an inclusion is given in Table III-48; apparently the major reaction here depended on the incompatibility of quartz-biotite-clinopyroxene, yielding hornblende and potassium feldspar. Immediately north of this, except on the northeast side of the pluton, is a zone characterized by coarse, dark-green amphibolitic blocks separated by pink potassium feldspar-rich pegmatite and minor gray plagioclase-rich pegmatite, and there is gradation between this zone and that described immediately above. Similar rocks are exposed near the southern margin of the outcrop here, and it is these two zones that are tentatively correlated with the two amphibolite layers of unit A. To the north, three mappable subconcordant dikes of quartz monzonite are found, two of which were traced for more than 2 miles. These are slightly finer grained than the main body, and have foliated margins. Locally, even these dikes show faint compositional layering. In their vicinity, the gneisses tend to be migmatites having a wide range of structural forms, with the exception of the more nebulitic varieties; raft and veinitic types are very common.

To the southeast, in the valley east of Delhi, there is pink, medium-grained, massive to gneissic quartz monzonite. In detail, this rock crosscuts the gneisses, but the foliation is essentially concordant with that in the gneisses. There is, of course, no doubt as to the transgressive nature of many apophyses from this body (especially in the core of the Delhi anticlinorium). This mass is almost certainly continuous with the similar rock in the valley east of North Redwood, but here especially, the rock is more commonly inequigranular and very highly weathered. Locally, it is very difficult to separate from the associated gneisses, in part because of the intimate association and in part because of the considerable weathering that has affected both. (This is the area of the classic study on weathering by Goldich, 1938.)

It may be noted that pegmatitic dikes are especially common both on the north side of the Delhi synclinorium and near North Redwood; aplitic dikes are uncommon.

Three small diabase dikes have been noted to cut the gneisses in this region, but an isolated exposure in the core of the Delhi synclinorium is a granular rock having the assemblage colorless clinoamphibole-olivine-spinel, partly serpentinized. Whether this assemblage is related to the late mafic intrusions or is part of the metamorphic assemblage is unknown at present.

Structure

The major structures are defined not only in terms of the gross stratiformity described above, which depends strongly on lithologic correlations which are impressionistic over large distances in such a terrane, but also in terms of analysis of the foliations and lineations measured in the region.

Considering the foliations—schistosity, gneissosity, and compositional layering—in these gneisses, including the masses of quartz monzonite, no significant domains with different structures have been found. For example, dividing the region into three segments, at east Delhi and North Redwood, compilations of foliations yield girdles with B-axes N. 80° E., 18°, N. 77° E., 8°, and N. 80° E., 14°, respectively, from northwest to southeast. The overall compilation for the region (fig. III-84A) yields a broad girdle with two rather subdued maxima, and hence at least monoclinic and possibly orthorhombic symmetry. On the basis of this last assumption, an axial plane is defined, approximately N. 84° E., 82° N., and the B-axis is N. 83° E., 12°.

Elongate mineral grains and mineral aggregates yield a maximum about N. 84° E., 16° (fig. III-84B), interpreted as B-lineations congruent with the data from foliation planes. However, there is a spread in the data, and a very

Table III-46. Modes, in volume percent, from Unit D, Sacred Heart-Morton area.

	1a	1b	2	3	4	5	6	7
Quartz	72.0	7.9	12.9		19.9	1.9	44.4	90.0
K-feldspar						22.9	1.9	
Plagioclase	11.3	45.0	33.4	48.6	13.9	28.5	41.8	
Myrmekite						6.3		
Chlorite							Tr	3.0
Muscovite						3.0	Tr	7.0
Biotite	Tr				24.8	28.2	1.5	
Actinolite								
Hornblende		12.19	28.8	37.7				
Cummingtonite	16.3	30.2	17.1					
Anthophyllite					7.1			
Clinopyroxene				11.4				
Orthopyroxene								
Olivine								
Cordierite					28.6	5.4		
Garnet			0.5		4.3		10.5	
Sillimanite						2.1		
Epidote								
Sphene				2.3				Tr
Zircon							Tr	Tr
Apatite	Tr		Tr		Tr		Tr	
Opaques	0.4	3.9	7.1		1.3	0.5	Tr	Tr
Carbonate								
Andalusite						1.3		

1: Quartz-cummingtonite schist (1a) adjoining amphibolite (1b) (M14588-9)—NW¼sec. 33, 114N/36W
2: Garnet amphibolite (M14106)—NW¼sec. 33, 114N/36W
3: Pyroxene amphibolite (M14125)—SW¼sec. 33, 114N/36W
4: Biotite-cordierite-garnet-anthophyllite gneiss (M14173)—NE¼sec. 32, 114N/36W
5: Biotite-cordierite-sillimanite gneiss (M14557)—SE¼sec. 29, 114N/36W
6: Garnet-bearing quartzofeldspathic gneiss (M14680)—NW¼sec. 33, 114N/36W
7: Quartz-muscovite schist (M14113)—NE¼sec. 29, 114N/36W

subsidiary maximum at about N. 53° E., 20°. This subsidiary maximum would be overlooked were it not for the fact that the strong maximum in the plot of minor fold axes (fig. III-84C) is coincident, and a tail of fold axis data extends from here covering the region occupied by the mineral elongation data. This distribution is interpreted as resulting from the development of relatively common mineral elongation (L_1) and less common minor folds (F_1) congruent with the major fold system, and the development of relatively common minor folds (F_2) and rare mineral elongations (L_2) after this folding. That this interpretation is basically correct can be demonstrated in the field; in particular, warping of mineral elongations (L_1) in minor folds (F_2) is not uncommon.

Late deformation (L_2 and F_2) is exemplified not only by the lineations, but also, much more commonly, by warping or kinking of older structures along nearly vertical planes. The trends of such planes box the compass overall, but there is a very strong maximum at about N. 60° E. and a tendency for the apparent movement to be right-lateral (figs. III-82B and D). (The warping of foliations in the main quarry at Morton is of this nature.) Faulting is rare in this region as a whole, but one observed fault surface, and the only two mapped faults also trend northeast.

Two major shear zones trend northwest through the region, one from near North Redwood through Gold Mine Lake was traced for more than 10 miles, and the other crosses the axis of the Morton synclinorium. No major displacement has been found on either.

Post-L_1 deformation runs the gamut from ductile to brittle deformation. Such deformation not only partly controlled development of granitic material in the gneiss (fig. III-82D) but also gave rise to folding, shearing, fracturing, and faulting.

Metamorphism

The most instructive outcrops are those in unit D, in the core of the Delhi synclinorium. From these, the following equilibrium assemblages are inferred, omitting accessory phases:

1. quartz-plagioclase-hornblende-garnet
2. plagioclase-hornblende-clinopyroxene-sphene
3. quartz-plagioclase-biotite-cordierite-garnet-anthophyllite-ilmenite
4. quartz-potassium feldspar-plagioclase-biotite-cordierite-sillimanite
5. quartz-potassium feldspar-plagioclase-biotite-garnet

Cummingtonite is common in assemblage 1, either as discrete grains or homoaxially grown with hornblende, and is not considered part of the equilibrium assemblage. In the one thin section in which assemblage 4 was found, andalusite also is present (the similar assemblage lacking both andalusite and cordierite is common). Assemblage 5 with sillimanite as an additional phase occurs in hand specimen, but not in a single thin section.

Additional assemblages from elsewhere in the region are:

6. potassium feldspar-plagioclase-biotite-hornblende-clinopyroxene-sphene
7. plagioclase-hornblende-clinopyroxene-orthopyroxene
8. quartz-potassium feldspar-plagioclase-biotite-hornblende-sphene

In these assemblages, biotite and clinopyroxene do not appear together in the presence of more than one percent quartz, and the three minerals probably are not stable with respect to hornblende-potassium feldspar, involving a reaction such as:

$$4\,SiO_2 + KFm_3AlSi_3O_{10}(OH)_2 + 2\,CaFmSi_2O_6 \rightarrow KAlSi_3O_8 + Ca_2Fm_5Si_8O_{22}(OH)_2$$

Table III-47. Modes, in volume percent, of granitic rocks, Sacred Heart-Morton area.

	1	2	3a	3b	4	5	6
Quartz	19	23	25.1	20.2	27.3	25.2	29.0
K-feldspar	34	34	29.7	26.7	30.3	55.7	19.2
Plagioclase	39	35	38.8	42.6	36.1	14.2	40.2
Myrmekite	1	1	1.0	1.8	2.5	2.9	2.5
Chlorite	2	1	2.2	1.6		Tr	2.1
Muscovite	Tr	1	2.2	0.4		Tr	0.6
Biotite	4	3	0.2	5.3	2.7	Tr	5.0
Actinolite							
Hornblende						1.0	
Cummingtonite							
Anthophyllite							
Clinopyroxene							
Orthopyroxene							
Olivine							
Cordierite							
Garnet							
Sillimanite							
Epidote		1			0.2		
Sphene		Tr		0.2		Tr	Tr
Zircon		Tr	Tr				Tr
Apatite		Tr	0.2			Tr	Tr
Opaques	1	1	0.2	1.2	1.0	1.0	1.3
Carbonate		Tr	0.4				

1: Pink quartz monzonite (M14011)—Cen. of N½sec. 18, 114N/37W
2: Gray quartz monzonite (M14015)—SE¼sec. 13, 114N/38W
3: Banded quartz monzonite, 3a and 3b from adjacent bands (M14010)—Roadcut, SW¼sec. 8, 114N/37W
4: Quartz monzonite (M14509)—Cen. of sec. 28, 113N/35W
5: Granite (M15003)—NW¼sec. 11, 113N/36W
6: Granodiorite (M14072)—SW¼sec. 24, 114N/37W

Table III-48. Successive modal analyses, in volume percent, spanning a distance of about four inches from an inclusion into adjacent granite (M14215; roadcut, SW¼ sec. 8, 114N/37W).

	1	2	3	4	5	6
Quartz				3	12	14
K-feldspar	24	29	46	57	56	59
Plagioclase	13	17	16	12	15	19
Myrmekite					8	4
Chlorite						
Muscovite						
Biotite	6	9	Tr			
Actinolite						
Hornblende	52	31	4	Tr	1	2
Cummingtonite						
Anthophyllite						
Clinopyroxene	2	13	33	28	6	2
Orthopyroxene						
Olivine						
Cordierite						
Garnet						
Sillimanite						
Epidote						
Sphene	1	1	Tr	Tr	1	Tr
Zircon						
Apatite	1	Tr	1		1	Tr
Opaques	Tr	Tr	Tr	Tr	Tr	Tr
Carbonate	1	Tr	Tr		1	Tr

This reaction also has been invoked to explain the zoned inclusions in the granite of the Sacred Heart pluton.

Assemblage 7 has been found in only one thin section, from an amphibolitic inclusion in the gneiss at Morton. With this exception, the above assemblages (and of course the more common assemblages with lesser numbers of phases) are diagnostic of the upper amphibolite facies at pressure-temperature conditions where quartz-muscovite is unstable but quartz-biotite-sillimanite and quartz-cordierite-garnet are stable (Grant, 1968, p. 925).

Retrogressive metamorphism is in evidence here also. In particular, rocks bearing potassium feldspar and sillimanite show development of muscovite around the sillimanite knots, and in part intergrown with myrmekite, around and within the potassium feldspar, implying retrogression across the second sillimanite isograd, as discussed by Evans and Guidotti (1966). Zoning in garnet in assemblage 3, showing low Mn and high Mg/Mg+Fe in core relative to margin, suggests cryptic retrogression in this assemblage (Grant and Weiblen, 1971). As noted above, the

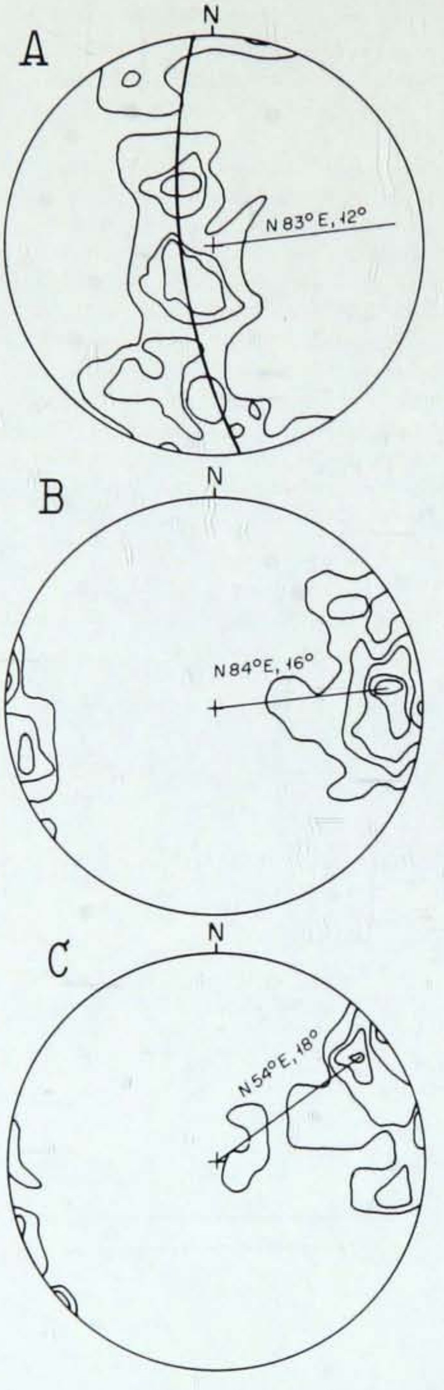


Figure III-84. Schmidt net diagrams. A, foliation in Sacred Heart-Morton area. Contours 2, 3, and 4 percent. B, lineation (elongate minerals and mineral aggregates) in Sacred Heart-Morton area. Contours 2, 4, 6, 8, and 10 percent. C, lineation (minor fold axes) in Sacred Heart-Morton area. Contours 3, 6, 9, and 12 percent.

development of cummingtonite may be caused by retrogression, as it is at Granite Falls. As would be expected, chloritization and sericitization are not uncommon, and extreme alteration to chloritic or albite- and epidote-bearing assemblages was found locally in unit D, and along the major shear zones.

The origins of these rocks present problems similar to those in the Granite Falls area, and they are discussed later. The gray quartzofeldspathic gneiss of units A, B, and C may represent an igneous protolith, which may have been more granitic than the quartz-plagioclase gneiss now found there. The amphibolites probably represent mafic igneous rocks, either extrusive or intrusive. The biotite-rich gneisses and quartzite of unit D are considered sedimentary in origin—graywacke and chert, respectively. The crux of the problem is that in such a high-grade plutonic environment, introduction of magma, partial melting, and metasomatism are all possible processes which could occur concurrently.

FRANKLIN-NEW ULM AREA

Outcrops in the remainder of the valley are sparse and highly weathered. In the valley south of Franklin, a pink to gray quartzofeldspathic gneiss with amphibolitic rafts, and a pink gneissic quartz monzonite with pyroxene-amphibolite rafts (mappable on the scale of 1:20,000) are closely associated and difficult to separate in many outcrops. In general, the migmatitic quartzofeldspathic gneiss lies north of the quartz monzonite and is similar to that found at Morton.

The most interesting facet of the Precambrian geology here concerns the late intrusions, principally the Cedar Mountain complex (Lund, 1956; Goldich and others, 1961) and a spectacular diabase dike system.

The Cedar Mountain complex is a body some 2,000 feet in diameter, composed of an outer shell of gabbro and an inner core of granodiorite (about 1,000 feet in diameter). The gabbro is a coarse-grained rock with steep swirling banding and extensively altered ferromagnesian minerals and plagioclase. Interstitial granophyre is common. The granodiorite is a pink, altered rock containing abundant chlorite and sericite. It is not found in contact with the gabbro. The body has a chilled margin against the gneisses, and there is splendid development of granophyre in the latter, suggesting partial melting resulting from the intrusion.

Several other small bodies having lithologies similar to the complex occur in the same area, and geophysical evidence suggests that still others are present in the surrounding region (Austin and others, 1970).

Among the more puzzling rocks here are a few scattered, isolated outcrops of serpentinized ultramafic rocks. One thin section shows colorless amphibole grains in a serpentine matrix, remarkably similar to the outcrop near Delhi, noted previously, but even less informative. As noted, a pair of diabase dikes trend southeastward, forming a marked ridge down the valley from near Cedar Mountain.

About 7 miles to the southeast near Fort Ridgely are outcrops of a porphyritic, well foliated granite that has schlieric, more mafic inclusions, and 6 miles farther on, a quartzofeldspathic gneiss that has amphibolitic rafts (table III-49). Finally, there is an outcrop of porphyritic foliated granite near New Ulm.

The Sioux Quartzite crops out in the vicinity of New Ulm. Here, it consists of a basal conglomerate with pebbles of quartz, jasper, and chert and, nearer Courtland, of quartzite with minor argillite. The conglomerate dips about 15-20° SE. and the quartzite dips as much as 30° S. (Goldich and others, 1961, p. 167-169).

Structure

In the Franklin area, foliation in the gneisses shows considerable variation, especially within about 2,000 feet of Cedar Mountain. The structure is best described as a monocline, striking eastward and dipping about 20° S., disrupted by the emplacement of the Cedar Mountain complex into a possible anticline. Although the monocline is compatible with the structures in the valley to the northwest, the few

Table III-49. Modes, in volume percent, of rocks in the Franklin-New Ulm area.

	1	2	3	4a	4b
Quartz	Tr		40.7	29.8	27.2
K-feldspar			32.8	18.7	60.2
Plagioclase	40.5	55.4	15.8	43.4	8.3
Myrmekite			6.1	3.3	3.3
Chlorite			Tr	0.5	Tr
Muscovite					
Biotite	Tr		3.9	1.5	1.0
Actinolite	19.4			0.7	
Hornblende	15.3	14.8			
Cummingtonite					
Anthophyllite					
Clinopyroxene	14.2	10.2			
Orthopyroxene	9.3	17.4			
Olivine					
Cordierite					
Garnet					
Sillimanite					
Epidote					
Sphene					
Zircon			Tr	Tr	
Apatite			Tr	0.2	
Opaques	1.3	2.2	0.7	0.8	Tr
Carbonate	Tr			1.1	
Rutile					Tr

1: Pyroxene amphibolite (M14703)—SE¼sec. 15, 112N/34W
2: Pyroxene amphibolite (M14752)—Cen. of sec. 22, 111N/32W
3: Fort Ridgely granite (M14758)—SW¼sec. 1, 112N/33W
4: Adjacent bands in granitic gneiss (M14751)—SW¼sec. 15, 111N/32W

measured lineations here average S. 17° E., 5°. This is approximately at right angles to the dominant lineations in the valley to the northwest, and may represent A-lineations within the same system, or a change in the relative magnitudes of the stresses involved, but not in their directions. The foliation in the Fort Ridgely granite is approximately S. 12° E. and vertical, whereas the quartzofeldspathic gneiss southeast of Fort Ridgely forms a northwestward-dipping monocline, with downdip lineation, and the granite at New Ulm again has a steep easterly foliation.

Metamorphism

In this part of the valley, assemblages that are sensitive to changes in metamorphic grade are virtually restricted to the amphibolite inclusions. At both Franklin and Fort Ridgely, these have the assemblage plagioclase-hornblende-clinopyroxene-orthopyroxene, with no more than a trace of quartz or biotite, and some secondary blue-green amphibole. This is compatible with the assemblage quartz-potassium feldspar-plagioclase-biotite in the granitic rocks, and further, is perhaps the assemblage typical of the granulite facies.

SYNOPSIS OF THE PRECAMBRIAN GEOLOGY

Lund (1956; see table III-50, this report) noted the mineralogic and chemical similarities in the rocks of his "granite series" (and to a lesser extent, in the rocks of his "basic complex"), and considered that the granitic rocks could have been cogenetic. It is this point, the coherence of the geology of the valley, that I wish to stress here, using, in particular, two different kinds of evidence—structure and metamorphism.

Structure

The major structure from Montevideo to Morton relates to folding about an axis approximately N. 85° E., 15°, derived from the gross stratiformity, foliations and lineations in quartz monzonitic, quartzofeldspathic, biotite-rich, and amphibolitic gneisses. No significant variation from this has been found. The major structure in the northwest part of the valley relates to an axis about N. 70° E., 32°. This is different in degree but not in kind from the fold system in the Montevideo-Morton segment; the change in trend is corroborated by the aeromagnetic data on the region (Zietz and Kirby, 1970). Certainly the major structure from Ortonville to at least Morton is related to folding on a shallow-plunging, east-northeasterly axis.

In the Franklin area (apart from the immediate vicinity of Cedar Mountain), the structure is compatible with this fold pattern, except that the lineations here average S. 17° E., 5°. One may compare with this structure the average foliation in the Fort Ridgely granite, which is S. 12° E. and nearly vertical, and the local anomalous fold near Ortonville, which has an axis S. 12° E., 16°. These last three attitudes are remarkably similar considering the paucity of data and the geographic distance involved. Moreover, they are essentially at right angles to the major axis for the Ortonville-Morton segment. This suggests that the stress system that produced the major structure from Ortonville to Fort Ridgely (and possibly to New Ulm) was everywhere similar. In the simplest terms, this similarity could be referred to a system with the least principal stress near vertical, and the intermediate and greatest principal stresses near horizontal, with the former being easterly, the latter northerly to produce the major structure, but with local reversal of the magnitudes of the stress along these two axes.

The only structure in the valley which is not compatible with this stress orientation is that southeast of Fort Ridgely —a northwestward-dipping monocline, comprising rocks that resemble the granitic rocks near Ortonville. The structure here is similar to that in two small outcrops of amphibolite outside the valley near Canby, but the relation of the structure in these outcrops to the general picture remains in doubt; possibly, the structure in these outcrops is related to the later deformation.

Later structures comprise in particular the dominantly northwestward-trending, steep shear planes having left-lateral apparent movement in the Granite Falls area and the dominantly northeastward-trending steep shear planes having right-lateral apparent movement in the Sacred Heart-Morton area. It will be noted that these are compatible with the alternate stress system conjectured above, but not with the major system. Perhaps a change in the magnitudes of the two greatest principal stresses did indeed occur, and perhaps this was a temporal rather than a spatial difference.

Table III-50. Comparison of some units used in this report with those of Lund (1956).

Lund	This paper
Morton quartz monzonite gneiss	Migmatitic gneiss (mu, mc, mb, ma); some granite (gr)
Montevideo granite gneiss	Migmatitic gneiss (mm, mi)
Gabbro gneiss	Hornblende-pyroxene gneiss (gnh); biotite gneiss and amphibolite of unit D (gnb); amphibolitic rafts
Garnetiferous quartz diorite gneiss	Garnet-biotite gneiss of Granite Falls area (gnb)
New Ulm, Fort Ridgely, Sacred Heart and Ortonville granites	Granite (gr)

Metamorphism

It has been said above that the inferred equilibrium assemblages from the metamorphic rocks (including the foliated granitic rocks) in the several areas indicate upper amphibolite facies in the Sacred Heart-Morton area, and either indicate or are compatible with granulite facies elsewhere. In order to portray the possible compatibilities or incompatibilities in these assemblages, one may try to show the chemographic relations between them; for such a range of compositions as is found here, two diagrams at least are required. First, let us consider those assemblages which may contain, in addition to quartz and plagioclase, the minerals potassium feldspar, biotite, cordierite, garnet, sillimanite, and either orthopyroxene or anthophyllite. These include the relevant assemblages of granitic rocks in the Ortonville-Odessa area, assemblages 6 and 7 of gneisses from the Montevideo-Granite Falls area, and assemblages 3, 4 and 5 of rocks from the Sacred Heart-Morton area. These may be compared with Figure III-85, a semiquantitative AKFM compositional tetrahedron in which the relative Fe/Mg ratios of the ferromagnesian phases are in accordance with available data (Himmelberg and Phinney, 1967; Grant and Weiblen, 1971), with the simplified portrayal of orthopyroxene and anthophyllite at the same point in the projection. All the assemblages that can be described in terms of this diagram may be compatible with one another, and the differences in the assemblages may result simply from differences in bulk composition, except for the presence of anthophyllite rather than orthopyroxene in the relevant assemblages from near Delhi. Moreover, the distribution of Fe and Mg in assemblages bearing garnet and biotite from Granite Falls (Himmelberg and Phinney, 1967, p. 333 and 339) and from near Delhi (Grant and Weiblen, 1971, p. 294) yield average values for the distribution coefficient, $Kd = (Mg/Fe)_{garnet}/(Mg/Fe)_{biotite}$, of 0.21 and 0.23, respectively.

Second, let us consider those assemblages which may contain the minerals garnet, hornblende, and clinopyroxene or orthopyroxene in addition to quartz, plagioclase, and a potassic phase (either feldspar or biotite). These include assemblages 2, 3, 4, and 5 of gneisses from the Montevideo-Granite Falls area, assemblages 1, 2, and 7 of rocks from the Sacred Heart-Morton area, and the relevant assemblage of granite inclusions from Franklin and Fort Ridgely. These may be compared with the semiquantitative compositional tetrahedron of Figure III-86 in which the only change from Figure III-85 is the replacement of K_2O by CaO, and in which the relative Fe/Mg ratios of the ferromagnesian minerals are in accordance with available data (Himmelberg and Phinney, 1967). There is only one four-

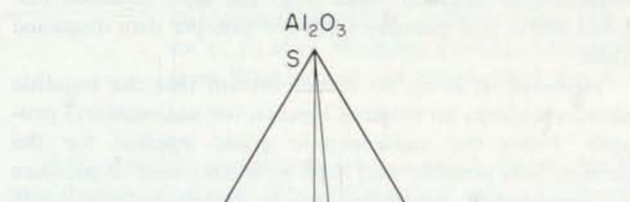


Figure III-85. Semiquantitative chemographic relationships between potassium feldspar (K), biotite (B), cordierite (C), garnet (G), sillimanite (S), and orthopyroxene or orthoamphibole (O), in an Al_2O_3-K_2O-FeO-MgO tetrahedron. Quartz, plagioclase, and water vapor are possible additional phases in all assemblages.

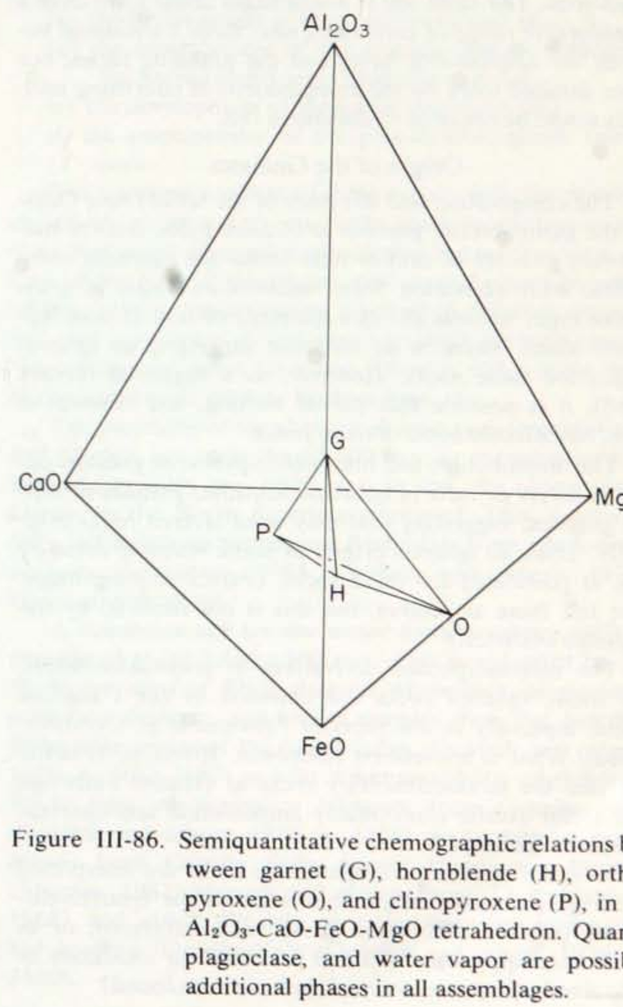


Figure III-86. Semiquantitative chemographic relations between garnet (G), hornblende (H), orthopyroxene (O), and clinopyroxene (P), in an Al_2O_3-CaO-FeO-MgO tetrahedron. Quartz, plagioclase, and water vapor are possible additional phases in all assemblages.

phase field involved, so that this simply illustrates the possible compatibility of all these assemblages, given differences in bulk composition.

The only assumed equilibrium assemblages that are not readily described in either of these two figures are those containing two potassic phases (potassium feldspar and biotite) and hornblende with or without clinopyroxene, including the assemblage from the hornblende monzonite in the Ortonville-Odessa area, assemblage 1 from gneisses in the Montevideo-Granite Falls area, and assemblages 6 and 8 from rocks in the Sacred Heart-Morton area. Comparing these assemblages with Figure III-86, one finds that two potassic phases are present only where there are less than the four phases determining the internal tetrahedron shown there.

The point to be made here is that, from the gross mineralogy of assumed equilibrium assemblages throughout the valley, the only major difference which could not be readily explained on the basis of differing bulk compositions is the appearance of anthophyllite rather than orthopyroxene in the rocks near Delhi. This, of course, does not mean that, with this one proviso, difference in bulk composition is the only reason for the differences in inferred equilibrium assemblages throughout the valley, nor that these rocks formed under the same pressure-temperature conditions. The same set of assemblages could form over a considerable range of conditions near those transitional between the amphibolite facies and the granulite facies, but more detailed work on the compositions of coexisting minerals would be required to document this.

Origin of the Gneisses

The composition and structure of the schists near Odessa, the garnet-biotite gneisses at Granite Falls, and the biotite-rich gneisses of unit D near Delhi are generally compatible with derivation from sedimentary rocks of graywacke type, whereas the rare quartzite of unit D may represent chert. There is no evidence suggesting an igneous origin for these rocks. However, as I suggested (Grant, 1968), it is possible that partial melting, and removal of melt, has affected some of these rocks.

The amphibolites and hornblende-pyroxene gneisses occur as layers or rafts in quartzofeldspathic gneisses or biotite gneisses, suggesting that they were layered rocks originally. Thus, an igneous origin, as mafic volcanic rocks or sills, is postulated for these rocks. (Partial melting might have left these as residua, but this is not required by the available evidence.)

The metamorphosed derivatives of graywacke, chert, and mafic igneous rocks are common in the Canadian Shield, especially in the Superior Province (*e.g.*, Goodwin, 1968a). What is uncommon (Goodwin, 1968a, p. 5) is the fact that the metasedimentary rocks at Granite Falls and near Delhi overlie conformably amphibolitic and quartzofeldspathic gneisses.

As noted above, the amphibolitic rocks are interpreted as being igneous in origin, but what of the quartzofeldspathic gneisses with which these are interlayered, or in which they appear as rafts and in which no inclusions of undoubted metasedimentary rocks have been found?

In the quartzofeldspathic gneisses, individual lithologies range from tonalitic to granitic. Let us first consider the gray tonalitic gneiss so common in the southeastern part of the valley. This consists of about 25-35 percent quartz, 50-60 percent plagioclase, and 5-10 percent biotite and hornblende. The composition is that of tonalite in general and trondhjemite in particular (see Goldich and others, 1970, table 3, sample 339 for a chemical analysis). The low FeO and MgO, high Na_2O, and low K_2O are atypical of graywackes and the high Al_2O_3 and Na_2O are atypical of arkoses. Thus, if this does represent an original protolith, it is probably of igneous origin and represents a tonalite or dacite. The main problem lies in the low FeO and MgO, high Na_2O and low K_2O relative to common tonalites and dacites (Nockolds, 1954, table 2). However, the analysis quoted above compares very closely with undoubted Lower Precambrian volcanic or hypabyssal rocks from northern Minnesota (Green, 1970a, table 3, "rhyodacites"). As Green noted (1970a, p. 89), high Na/K ratios are common in Lower Precambrian volcanic rocks.

If the gray tonalitic type of gneiss was the igneous protolith for all the quartzofeldspathic gneiss, then we must call for introduction of material by either magmatic or metasomatic processes, or both, to yield the granitic component so commonly found. There is evidence of such introduction of material, both from the field relations discussed above and possibly from the isotopic data discussed below.

However, it is by no means certain that the tonalitic gneiss represents an original igneous (or sedimentary) protolith. From the metamorphic grade inferred for the gneisses, it is possible that they were subjected to pressure and temperature conditions where partial melting could occur. If so, two possibilities exist: the tonalitic gneiss could represent a melt or a crystalline residuum.

The first possibility has been suggested by Goldich and others (1970, p. 3693) on the basis of experiments by Winkler and von Platen (1961). The second possibility is based on work in the Q-Or-Ab-An-H_2O system and related systems (*e.g.*, Tuttle and Bowen, 1958; Winkler, 1967; James and Hamilton, 1969) wherein quartz-plagioclase-melt-vapor can coexist over a large range in pressure, temperature and composition, the melt being more granitic than the crystalline assemblage. To illustrate the differences between the two possibilities let us use the data of Winkler and von Platen (1961). A prime example of the development of a "tonalitic" melt is that derived from the graywacke "Bicken" (Winkler and von Platen, 1961, p. 65, fig. 4; also see fig. III-87, this report). To produce such a melt, 94-96 percent of the rock had to be melted (Winkler and von Platen, 1961, table 5)—essentially all the salic constituents herein (James and Hamilton, 1969, p. 136-138)—and the melt contained 9.5 percent normative orthoclase and 4 percent normative biotite. However, the same system at 690° C yielded 24 percent melt of "granitic" composition and, at 710° C, 47 percent melt of "granodioritic" composition, in both cases with a quartz-plagioclase-rich crystalline assemblage having less than one percent normative orthoclase and less than three percent normative biotite (fig. III-87).

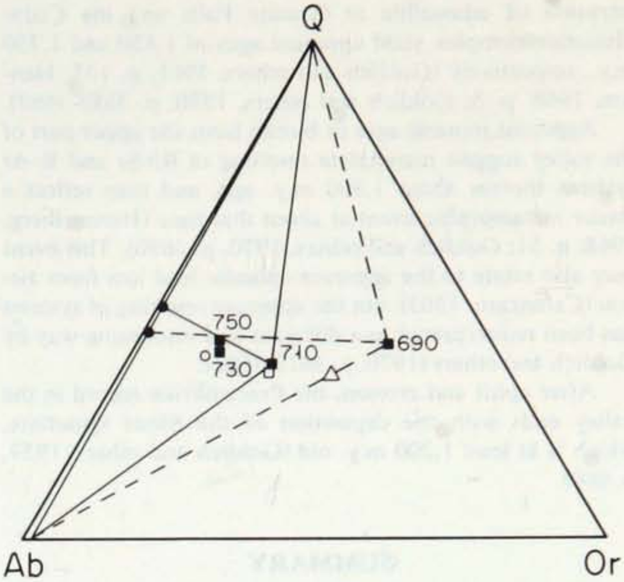


Figure III-87. Normative Q-Ab-Or diagram, showing the compositions of successive melts from graywacke "Bicken" (▪), and of residual crystalline material at 690°C and 710°C (•). Also shown are similar data for average tonalite (O), granodiorite (□), and calc-alkali granite (△); after Nockolds (1954). (Adapted from Winkler and von Platen, 1961, fig. 4.)

Both residual assemblages have a much higher Na/K ratio than the final melt. The separation of such a residuum from melt could yield both the tonalitic and granitic components of the migmatites. And such a coexistence of tonalitic crystalline material and granitic melt would be necessary en route to the production of the "tonalitic" melt at higher temperatures as required in the first possibility.

It must be noted that the above example is only an illustration of the envisaged relationship between the two possibilities. In Winkler and von Platen's work, the residual assemblages in particular, and the melts in part, contained cordierite and sillimanite, which are not present in the quartzofeldspathic gneisses. Further, the designation "tonalite," in particular (fig. III-87), is somewhat of an illusion; only two of Winkler and von Platen's melts (1961, table 5) contained less than 10 percent orthoclase in normative feldspar, and the anorthite component is omitted in the diagram.

If indeed the gray tonalitic gneiss represents a residuum from partial melting of a more potassic protolith, then the unanswered question is how much of the granitic component of the gneisses represents melt so produced? Apparently not all the more potassic material could be derived in this manner. The Rb-Sr data on the quartzofeldspathic gneisses (Goldich and others, 1970, Morton Gneiss and Montevideo gneiss) indicate a Sr^{87}/Sr^{86} ratio of 0.710 at an age of about 2,650 m.y., whereas the Sacred Heart and Ortonville granitic rocks, which might be considered possible partial melts related to the 2,650 m.y.-events, had at this time a Sr^{87}/Sr^{86} ratio of 0.702. Although there is considerable scatter in the data, this suggests that the Sacred Heart and Ortonville granitic rocks are not cogenetic with the granitic phases of the quartzofeldspathic gneisses, but are indeed allochthonous emplacements and, conversely, that at least some of the granitic phases of the quartzofeldspathic gneisses are cogenetic with the tonalitic gneiss.

Thus, it is considered that the quartzofeldspathic gneisses represent either an original suite of igneous rocks ranging from tonalitic to granitic in composition or a more homogeneous igneous protolith of approximately quartz monzonite composition which has been fractionated during partial melting into more and less potassic members. To these rocks more or less granitic material was added in conjunction with the emplacement of the Sacred Heart and Ortonville plutons. Obviously, more petrographic, analytical, and experimental work is required before these speculations can be refined satisfactorily.

Precambrian History of the Valley

In discussing the history of the valley, it is worthwhile to consider just what events we may hope to date. The most obvious of these are:

(a) the formation of the protoliths of the gneisses
(b) the metamorphism to upper amphibolite or granulite facies (M_1)
(c) the development of the major structures (F_1)
(d) the emplacement of the granitic bodies, especially the Sacred Heart and Ortonville plutons
(e) the development of the minor structures (F_2)
(f) the emplacement of the post-tectonic minor intrusions

The problems are that all these events, with the possible exceptions of (d) and (f), may span much more than an instant in geologic time, even considering the myopic view we have of the Early Precambrian, that several events have reset the useful isotopic systems involved to different extents, and that correlation of apparent ages with geologic events leaves plenty of room for error. There is still a great deal of geochronologic work to be done here.

The protoliths of the gneisses, at least near Granite Falls and Morton, are older than 3,000 m.y. (a conservative figure) and possibly are 3,600 m.y. in age, the oldest rocks known on the North American continent. This is derived from U-Pb data on zircons and Rb-Sr data from whole-rock analyses (Catanzaro, 1963; Goldich and others, 1970; Grant, unpub. data).

A minimum age for the major metamorphism (M_1) is considered to be $2{,}650 \pm 100$ m.y. This is indicated by (a) the linear array of Rb-Sr data for whole-rock, plagioclase, potassium feldspar, and biotite samples from the quartzofeldspathic gneiss of the lower valley (Goldich and others, 1970, p. 3683-3687) and (b) the compatibility of this with Rb-Sr data on potassium feldspars from Granite Falls (Goldich and others, 1970, p. 3688), K-Ar data on hornblende from Granite Falls, Sacred Heart, and Morton (Thomas, 1963; Hanson and Himmelberg, 1967; Hanson, 1968), and with U-Pb data on zircon from the garnet-biotite gneiss at Granite Falls (Goldich and others, 1970, p. 3680).

It should be emphasized that an apparent age of 2,650±100 m.y. is an approximate threshold for the last major setting of the several systems involved. It is not necessarily correlative with a "culmination" of metamorphism which may have preceded this time, but is unlikely to postdate it.

What of the development of the major structures (F_1)? On the geologic grounds that the high-grade metamorphic assemblages have crystallized in structural features, mineral lineations in particular, which are congruent with the major structure, the metamorphism (M_1) and deformation (F_1) are considered contemporaneous. One would wish to date lithologic bodies which may have originated during the deformation, and a prime target is the compositional heterogeneity in the migmatitic gneisses. Apparent ages may be calculated for each of the two pairs of adjacent but different lithologies reported by Goldich and others (1970, table 6). Of these, the older apparent age (samples 337D and 337L) is about 2,800 m.y., and is compatible with an array of Rb-Sr data from different adjacent lithologies (Grant, unpub. data) from Morton. (The problem is one of either incomplete homogenization of the strontium isotopes during the formation of the quartzofeldspathic migmatitic gneiss or later opening of the systems to diffusion or both.) The above evidence is compatible with the geologic conclusion that the metamorphism (M_1) and deformation (F_1) were synchronous. There is no evidence to support the hypothesis of earlier separate metamorphic or deformational events (see Goldich and others, 1970, p. 3693), although these may have occurred and been obliterated. There is, however, the probability that structural features, such as the gross stratiformity, deformed during F_1, are much older than 2,650 m.y. These are considered to be of igneous and sedimentary rather than tectonic origin.

The emplacement of dominantly quartz monzonite bodies, as near Sacred Heart and Ortonville, is considered on geologic grounds to be late syntectonic. Rb-Sr analyses of whole-rock samples from these localities yield apparent ages of 2,700 and 2,600 m.y., respectively (Goldich and others, 1970, p. 3687 and 3689), and the interpretation of U-Pb data on zircon from the Sacred Heart pluton is compatible with this apparent age (Goldich and others, 1970, p. 3679).

Development of late minor structures certainly postdated F_1, and at least shearing at Granite Falls postdated the emplacement of tholeiitic diabase which may be as young as 2,100 m.y. (Hanson and Himmelberg, 1967). Hanson and Himmelberg (1967) also found K-Ar apparent ages of 1,700-1,900 m.y. on hornblende and biotite from hornblende andesite dikes at Granite Falls, whereas the small intrusion of adamellite at Granite Falls and the Cedar Mountain complex yield apparent ages of 1,850 and 1,750 m.y., respectively (Goldich and others, 1961, p. 135; Hanson, 1968, p. 5; Goldich and others, 1970, p. 3688-3689).

Apparent mineral ages of biotite from the upper part of the valley suggest incomplete resetting of Rb-Sr and K-Ar systems therein about 1,800 m.y. ago, and may reflect a minor metamorphic event at about that time (Himmelberg, 1968, p. 31; Goldich and others, 1970, p. 3690). This event may also relate to the apparent episodic lead loss from zircon (Catanzaro, 1963), but the apparent resetting of systems has been reinterpreted in a different and interesting way by Goldich and others (1970, p. 3677-3682).

After uplift and erosion, the Precambrian record in the valley ends with the deposition of the Sioux Quartzite, which is at least 1,200 m.y. old (Goldich and others, 1959, p. 660).

SUMMARY

The Precambrian geology of the Minnesota River Valley can be summarized in the following working hypothesis.

Prior to 3,000 m.y., and probably as early as 3,600 m.y. ago, a layered sequence was formed composed of granitic rocks overlain by graywacke-like strata with basaltic rocks locally abundant in both. Whether the igneous rocks were extrusive, intrusive, or both is not known.

By about 2,650 m.y., these rocks had been deformed into a fold system on a shallow east-northeastward-plunging axis, and metamorphosed to the upper amphibolite or granulite facies. Possibly, partial melting of rocks of appropriate compositions, accompanying deformation, gave rise to the common migmatitic structures. In the later stages of these events, large bodies, dominantly of quartz monzonite, were emplaced near Ortonville and in the lower valley in particular.

Minor deformation continued after emplacement of these large bodies, overlapping the times of emplacement of mafic dikes prior to 1,800 m.y., when small intrusions such as the adamellite at Granite Falls and the Cedar Mountain complex were formed. Possibly, mild metamorphism occurred at this time, especially in the upper valley.

Deposition of the Sioux conglomerate and quartzite prior to 1,200 m.y. ends the Precambrian record in the valley.

We have a coherent geologic picture throughout this elongate window onto the oldest known rocks in North America. And these rocks differ little in kind from other rocks of Early Precambrian age known across the Canadian Shield.

Chapter IV

MIDDLE PRECAMBRIAN

General Geologic Setting, G. B. Morey
Mesabi Range, G. B. Morey
Gunflint Range, G. B. Morey
Cuyuna District, Ralph W. Marsden
East-central Minnesota, C. W. Keighin, G. B. Morey and S. S. Goldich
Diabase Dikes in Northern Minnesota, P. K. Sims and M. G. Mudrey, Jr.
Minnesota River Valley, G. B. Morey
Aitkin County Sulfide Deposits, G. B. Morey
Evidences of Precambrian Life in Minnesota, David G. Darby
Amino Acids in Some Middle Precambrian Rocks of Northern Minnesota and Southern Ontario, James R. Niehaus and Frederick M. Swain

GENERAL GEOLOGIC SETTING

G. B. Morey

In Minnesota, Middle Precambrian rocks are separated from older and younger rocks by major unconformities. As presently defined, these rocks have time boundaries of approximately 2,600 million years and 1,800 million years (Goldich and others, 1961; Goldich, 1968); as such, the Middle Precambrian corresponds to the early and middle parts of the Proterozoic in Canada (Stockwell, 1964).

A stratigraphic section recording the entire span of Middle Precambrian time is lacking in Minnesota, and subdivision of the 800 million years of time is not yet possible. However, several discrete geologic events can be delineated by isotopic data. For example, episodes of mafic dike intrusion occurred prior to 2,200 m.y. ago and at about 2,000 m.y. ago, in the early Middle Precambrian, and at 1,700-1,800 m.y. ago in the late Middle Precambrian (Hanson and Himmelberg, 1967; Hanson, 1968; Hanson and Malhotra, 1971). Likewise, there were at least two cycles of sedimentary deposition that were separated by a period of epeirogeny and erosion. An unknown thickness of slate and quartzite unconformably overlies Lower Precambrian rocks locally in east-central Minnesota and is overlain by dolomite of the Trout Lake formation of Marsden (this chapter). The dolomite in turn is unconformably overlain by sedimentary strata assigned to the Animikie Group. The Animikie rocks are widespread geographically, and it has been common practice to equate the terms "Middle Precambrian" and "Animikie"; however, radiometric dates have shown that Animikie deposition records only a small part of Middle Precambrian time. The Animikie strata represent a single depositional event that began with well-sorted clastic material characteristic of a stable shelf and ended with fine sand and mud characteristic of a deep basin with poor circulation. Deposition was terminated or closely followed by an orogeny that involved folding and metamorphism of the sedimentary strata and emplacement of granitic rocks in east-central Minnesota. Several individual plutons can be recognized within the batholith (Woyski, 1949), and Goldich and others (1961) have shown that the individual plutons were emplaced at slightly different times. In Minnesota, this orogeny has been termed the "Penokean." Elsewhere in Minnesota, emplacement of basalt and hornblende andesite dikes and small granitic plutons and a rather widespread re-equilibration of older isotopic ages through thermal metamorphism took place virtually contemporaneously with the Penokean orogeny.

GEOGRAPHIC DISTRIBUTION

The inferred distribution of Middle Precambrian rocks is shown in Plate 1 and is summarized in Figure IV-1. In general, Animikie strata unconformably overlie a variety of Lower Precambrian metasedimentary, metavolcanic, and plutonic rocks. Locally, they also overlie older Middle Precambrian sedimentary strata and diabasic gabbro or diorite dikes. In east-central Minnesota, Middle Precambrian metasedimentary and igneous rocks are overlain to the southeast by nearly flat-lying Upper Keweenawan and younger strata. To the east, they are overlain or intruded by Middle Keweenawan extrusive and intrusive rocks; they reappear again from beneath the Keweenawan rocks in northeastern Minnesota and extend into Canada for an additional 150 miles. Cretaceous rocks locally overlie the Middle Precambrian strata in the northwestern and western parts of the area.

Although Middle Precambrian rocks underlie a large part of northeastern and east-central Minnesota, exposures are restricted to a few small areas (fig. IV-1): (1) the Gunflint range in northern Cook County and the adjoining Thunder Bay district, Ontario; (2) the Mesabi range; (3) the Cuyuna district; and (4) east-central Minnesota. In addition, diabasic gabbro or diorite dikes of early Middle Precambrian age locally are exposed throughout northern Minnesota; and early and late Middle Precambrian basaltic and andesitic dikes, a small adamellite pluton, and several gabbro-diorite-granite bodies crop out locally in the Minnesota River Valley.

NOMENCLATURE AND CORRELATION

Problems of lithostratigraphic nomenclature in the Middle Precambrian have been discussed many times, and it is readily seen from the extensive literature that these problems have been the subject of much controversy. A brief summary follows; additional discussions are given by White (1954), Goldich and others (1961), Schmidt (1963), and Goldich (1968). The controversy in Minnesota has been at two levels: (1) the establishment and correlation of local stratigraphic successions; and (2) the correlation of strata in Minnesota with successions elsewhere in the Canadian Shield.

Today, there is a general consensus that all Middle Precambrian sedimentary rocks in Minnesota with the exception of the Trout Lake formation and associated rocks (see discussion of Cuyuna district, this chapter) are part of a single, more or less continuous depositional episode, and are assigned to the Animikie Group. The term "Animikie," an Indian word meaning "thunder," was first used by Hunt (1873, p. 339) for dark argillite and sandstone in the Thunder Bay area on the north shore of Lake Superior. Similar argillite exposed around Gunflint Lake in Minnesota was thought by Bell (1873, p. 93) to correlate with Animikie strata around Thunder Bay, and Irving (1883, p. 381-390) first recognized the lithologic similarity of the Animikie rocks to rocks exposed on the Mesabi range. However, the term "Animikie" first was used formally in Minnesota for a formation having three members (Grant, *in* Winchell and

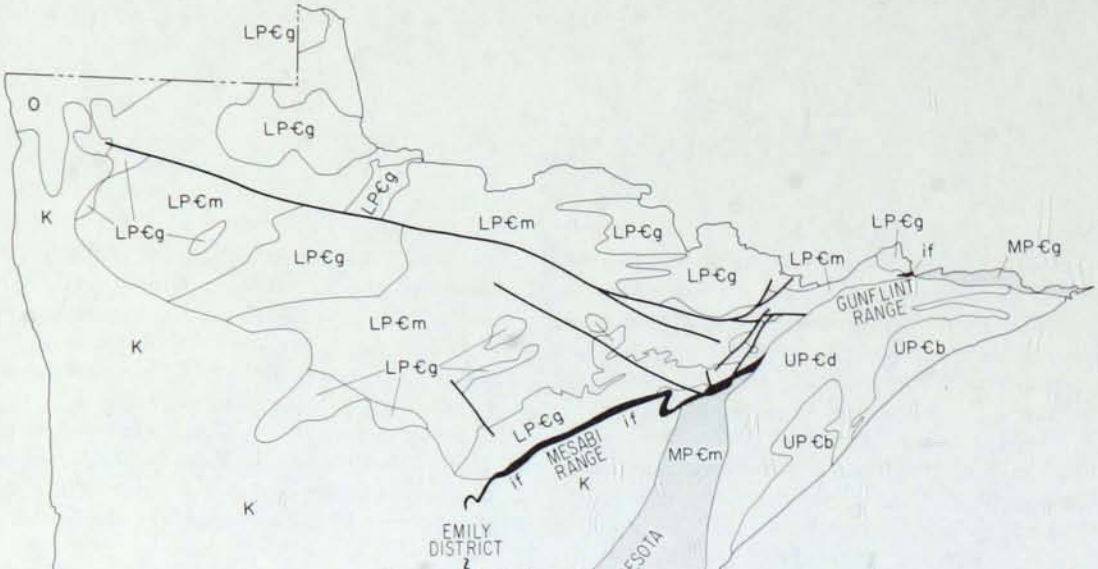


Figure IV-1. Generalized geologic map of Minnesota (modified from Sims, 1970) showing location of Middle Precambrian rocks referred to in text. The lower (southern) border of the state is approximately 265 miles long. $LP\epsilon_{m}$, Lower Precambrian metabasalt and graywacke, undivided; $LP\epsilon_{g}$, Lower Precambrian granitic rocks, undivided; $MP\epsilon_{m}$, Middle Precambrian graywacke and mudstone and their metamorphic equivalents (note that iron-formations [if] are shown in black); $MP\epsilon_{g}$, Middle Precambrian granitic rocks, undivided; $UP\epsilon_{q}$, Upper Precambrian quartzite; $UP\epsilon_{b}$, Upper Precambrian basalt, undivided; $UP\epsilon_{d}$, Upper Precambrian gabbroic rocks including the Duluth Complex; $UP\epsilon_{s}$, Upper Precambrian sandstones, undivided; ϵ, Cambrian sandstone and shale, undivided; O, Ordovician limestone and shale, undivided; D, Devonian limestone and dolomite, undivided; and K, Cretaceous shales, undivided.

others, 1899): (1) a "basal taconite member"; (2) an intermediate "black slate member"; and (3) an "upper graywacke-slate member." Workers of the U.S. Geological Survey (Van Hise and Clements, 1901; Clements, 1903; Van Hise and Leith, 1911) redefined the Animikie in Minnesota as a group having several formations, and renamed the "basal taconite member" the Gunflint Iron-formation. The "black slate member" and the "upper graywacke-slate member" were included in the Rove Slate. At about the same time, strata on the Mesabi range also were divided into three formations: (1) the basal Pokegama Quartzite (Winchell and Upham, 1888); (2) the intermediate Biwabik Iron-formation; and (3) the upper Virginia Slate, consisting of intercalated argillite and graywacke (Van Hise and Leith, 1901).

Although the stratigraphic continuity of rocks on the Mesabi and Gunflint ranges is well established, the stratigraphic position of rocks on the Cuyuna range has been the subject of much controversy (see Marsden, this chapter). Because the stratigraphic succession in the two areas is somewhat different, the main iron-formation of the Cuyuna range—the Trommald Formation of Schmidt (1963)—has been variously correlated with formations of the Animikie Group on the Mesabi range, either as the equivalent of the Biwabik Iron-formation (Grout and Wolff, 1955) or as an interbedded iron-rich member of the Virginia Formation (Zapffe, 1933). Most of this controversy resulted from the failure to recognize the possibility of facies changes between the underlying Mahnomen Formation on the Cuyuna range and the Pokegama Quartzite on the Mesabi range, and because the iron-formations have somewhat different tectonic settings. However, Schmidt (1963, p. 39) found ". . . no reason to consider a general correlation of the Trommald and Biwabik formations invalid. . . ."

The stratigraphic position of the Thomson Formation, as exposed in east-central Minnesota, also has been debated. Neither the top nor bottom is exposed; consequently, the formation has been variously correlated since Irving (1883) first suggested that it was equivalent to the Animikie strata at Thunder Bay. Subsequently, the formation has been assigned to either the Middle or Early Precambrian (see Grout and others, 1951, p. 1038). However, Goldich and others (1961, p. 5) suggested on the basis of radiometric dating that the formation is Middle Precambrian in age, and Morey and Ojakangas (1970) demonstrated that it is lithologically and sedimentologically similar to other Middle Precambrian graywacke-argillite successions in Minnesota. Thus, it is inferred that the Thomson Formation represents the down-basin extension of at least the upper part of the Animikie Group on the Mesabi range. The correlation of Middle Precambrian rocks in Minnesota is summarized in Table IV-1.

Because they are physically isolated from other Middle Precambrian rocks, regional correlation of the Animikie Group and associated igneous rocks in Minnesota has long been a major problem. Most of the controversy centers around the extent to which they can be correlated with similar-appearing rocks in northern Wisconsin and Michigan, and the use of the term "Huronian" as a time-stratigraphic term in Minnesota. The latter problem centers

Table IV-1. Stratigraphic nomenclature and inferred correlation of Middle Precambrian sedimentary rocks in Minnesota and adjoining Ontario.

GUNFLINT RANGE Minnesota and Ontario (Goldich and others, 1961; Tanton, 1931)	MESABI RANGE Minnesota (White, 1954)	CUYUNA RANGE Minnesota (Grout and Wolff, 1955; Schmidt, 1963)	EAST-CENTRAL MINNESOTA (Goldich and others, 1961; Morey and Ojakangas, 1970)
UPPER PRECAMBRIAN SEDIMENTARY AND IGNEOUS ROCKS (younger than 1.6 b. y.)			
unconformity			
Rove Formation	Virginia Formation	Rabbit Lake Formation	Thomson Formation
Gunflint Iron-formation	Biwabik Iron-formation	Trommald Formation	
"Kakabeka Quartzite"	Pokegama Quartzite	Mahnomen Formation	
unconformity			
		Trout Lake formation quartzite and slate	
unconformity			
LOWER PRECAMBRIAN IGNEOUS AND METAMORPHIC ROCKS (older than 2.6 b. y.)			

around the following question: did deposition of the Aphebian rocks on the north shore of Lake Huron, the Huronian Supergroup of Roscoe (1969, p. 4), take place approximately contemporaneously with deposition of the Animikie rocks, or do these rock successions represent separate and distinguishable events?

The use of the term "Huronian" in the Lake Superior region has been summarized by James (1958). Logan (1863) believed that rocks beneath the Animikie at the type locality were correlative with those of Murray's (1857) Huronian type locality. Irving (1883), however, directly equated the Animikie with the Huronian, and later Van Hise (1891) proposed a correlation between the rocks of the Marquette district of Michigan and the Huronian. The Huronian later was defined as consisting of three units, only two of which were represented in the type area; the Animikie Group was considered the third and upper unit because it was lithologically similar to the upper part of the Michigan sequence, which also was presumed to be Huronian. Because of the nomenclatural confusion, various workers from time to time have recommended that of the two terms, either "Animikie" or "Huronian" be abandoned in the Lake Superior region; however, the name "Animikie" has been retained in Minnesota because the rocks are clearly equivalent to those at the original Animikie type locality (Grout and others, 1951).

Because James (1958) concluded that the sedimentary strata in northern Michigan bore little resemblance to the Huronian rocks and some resemblance to the Animikie strata, he recommended that the term "Huronian" be abandoned in Michigan and replaced by the term "Animikie Series." However, the Animikie Series, as defined by James, contained rocks older and younger than those in the Animikie Group. Because of this, and because of the recognized dangers inherent in making correlations on the basis of homotaxial successions, Cannon and Gair (1970) recommended that the name "Animikie Series" be abandoned and replaced by the name "Marquette Range Supergroup." Nevertheless, they again emphasized that there was no evidence to support a correlation between that group and the Huronian Supergroup. However, because of similar iron-formations in both areas (James, 1954) and because of a remarkably similar stratigraphic succession in the Gogebic range of Wisconsin (Aldrich, 1929), the Animikie Group appears to be equivalent to at least the upper part of the Marquette Range Supergroup (table IV-2).

Table IV-2. Generalized correlation chart of Middle Precambrian rocks in the Lake Superior region (modified from Cannon and Gair, 1970; Schmidt, 1963; James, 1958; White, 1954; Aldrich, 1929).

MESABI RANGE	CUYUNA RANGE	GOGEBIC RANGE	DICKINSON COUNTY	MARQUETTE RANGE
UPPER PRECAMBRIAN SEDIMENTARY AND IGNEOUS ROCKS (younger than 1.6 b.y.)				
unconformity	unconformity	unconformity	unconformity	unconformity
Animikie Group: Virginia Formation	Rabbit Lake Formation	Tyler Slate	Badwater Greenstone Michigamme Slate Hemlock Formation	Baraga Group: Michigamme Slate Goodrich Quartzite
			disconformity	disconformity
Animikie Group: Biwabik Iron-formation	Trommald Formation	Ironwood Iron-formation	Vulcan Iron-formation Felch Formation	Menominee Group: Nagaunee Iron-formation Siamo Slate
Animikie Group: Pokegama Quartzite	Mahnomen Formation	Palms Quartzite		Menominee Group: Ajibik Quartzite
unconformity	unconformity	unconformity	unconformity	unconformity
	Trout Lake formation quartzite and slate?	Bad River Dolomite Sunday Quartzite	Randville Dolomite Sturgeon Quartzite Fern Creek Formation	Chocolay Group: Wewe Slate Kona Dolomite Mesnard Quartzite Enchantment Lake Formation
unconformity	unconformity	unconformity	unconformity	unconformity
LOWER PRECAMBRIAN IGNEOUS AND METAMORPHIC ROCKS (older than 2.6 b.y.)				

Available radiometric data from Minnesota and Ontario, although far from definitive, are consistent with the regional correlation outlined in Table IV-2. Although the Animikie rocks are believed to be younger than 2,000 m.y. (Hanson and Malhotra, 1971), uncertainties remain as to the actual time of deposition. Various attempts at determining that time have resulted in ages ranging from $1,635 \pm 24$ m.y. (Faure and Kovach, 1969) to $1,900 \pm 200$ m.y. (Hurley and others, 1962). Although the value observed by Hurley and others includes an arbitrary and perhaps unnecessary correction for 20 percent argon loss, it nevertheless has been widely quoted in the literature as the time of diagenesis. Faure and Kovach also interpreted the age they obtained as a time of diagenesis, and their data, if taken at face value, indicate that deposition and diagenesis on the Gunflint range took place after metamorphism in east-central Minnesota. Subsequent work, however, suggests several other possible interpretations. Misra and Faure (1970, p. 398) have shown that the apparent age of the Gunflint Iron-formation ". . . decreases from 1.7 b.y. at the eastern end to 1.2 b.y. . . . at the western end . . ." and that this systematic variation ". . . may be related to metamorphic effects caused by Keweenawan diabase sills. . . ." Although Misra and Faure concluded otherwise, their age may reflect partial re-equilibration resulting from Keweenawan igneous activity. A more interesting possibility has been suggested by Hanson and Malhotra (1971). Because the dike rocks they studied in northern Minnesota appear to have been metamorphosed about 1,600 m.y. ago, they suggested (p. 1110) that this event ". . . may not have been related to the Penokean orogeny . . . but may have been associated with burial by overlying sediments of the Animikie Group. . . ." However, the evidence for burial metamorphism is equivocal, and it may be that 1,600 m.y. was the time of a mild deformation younger than that of the Penokean orogeny. Thus, Faure and Kovach's age also may represent re-equilibration in response to this event; if so, the various mineralogic and textural features in the rocks of the Gunflint and Mesabi ranges, now ascribed to the Penokean orogeny, may be related to a post-Penokean event. Clearly, additional data are needed before this problem will be resolved, but it is significant that there are no data indicative of the Animikie Group being much older than 2,000 m.y.

The chronologic framework in Minnesota is consistent with that now evolving in northern Michigan. Aldrich and others (1965) showed that deposition of the Marquette Range Supergroup of Cannon and Gair (1970) predated a metamorphic event 1,900 m.y. ago, but they were unable to demonstrate precisely when the deposition started. More recently, Banks and Van Schmus (1971, p. 9) suggested from limited data that deposition of the Marquette Range Supergroup ". . . appears to be bracketed in the interval between 2,100 m.y. and 1,900 m.y. . . ." In contrast, radiometric age data suggest that the Huronian Supergroup is older than 2,000 m.y. Fairbairn and others (1969) have suggested that the Gowganda Formation in the upper part of the Huronian Supergroup was deposited $2,288 \pm 87$ million years ago. Furthermore, the Nipissing Diabase of Ontario has a Rb-Sr isochron age of 2,155 to 2,160 m.y. (Van Schmus, 1965; Fairbairn and others, 1969), and accordingly it is certain that some Huronian rocks were deposited before about 2,160 m.y. ago. In addition, there is abundant field evidence of an earlier, pre-diabase tectonic event that produced major folds in the Huronian rocks. Although direct radiometric age confirmation is lacking, this event must have occurred between 2,288 and 2,160 m.y. ago; Church (1968) has suggested approximately 2,200 m.y. as a reasonable time for this event. Elsewhere in the Huronian area, evidence also exists for post-diabase deformation. Fairbairn and others (1969) have suggested that this tectonic and metamorphic overprint occurred 1,750 m.y. ago, resulting in a composite "age" of $1,950 \pm 100$ m.y. in the rocks affected by the pre-diabase event.

For the reasons outlined above, Banks and Van Schmus (1971) concluded that the available radiometric data severely restrict possible correlation between the Marquette Range Supergroup in Michigan and the Huronian Supergroup of the north shore of Lake Huron. Thus, in Minnesota it seems appropriate to continue using local names for rock units inasmuch as they neither imply nor negate correlations between similar, but geographically isolated, sequences. For this reason, the term "Huronian" is not used in Minnesota.

MESABI RANGE

G. B. Morey

The Mesabi range is approximately 70 to 100 miles northwest of Duluth (fig. IV-1). The name designates the mostly buried, preglacial outcrop of the Biwabik Iron-formation, a narrow belt, one-fourth to three miles wide, of iron-rich strata that extend in an east-northeasterly direction for 120 miles along strike from eastern Cass County through Itasca County to Birch Lake in St. Louis County (fig. IV-2). The eastern end of the range is truncated by the Duluth Complex of Late Precambrian age, whereas the western end is covered by Cretaceous strata and thick glacial deposits; the extent and trend of the formation west of Grand Rapids are not as well known as in the remainder of the range. Throughout its exposed length, the Biwabik Iron-formation is underlain by the Pokegama Quartzite and overlain by the Virginia Formation. These rocks were deposited more or less continuously in the northern part of the Animikie basin, and subsequently were metamorphosed and deformed, perhaps during the Penokean orogeny and a later pre-Keweenawan event. The Animikie strata again were somewhat deformed and metamorphosed during Keweenawan time.

Since iron ore was first discovered in 1890, the Mesabi range has supplied approximately 2.8 billion tons of iron ore. Of that total, approximately 269 million tons, or 9.6 percent, represent ore concentrated from taconites since 1949. Today (1971) ore concentrated from taconite comprises 61 percent of the total ore produced.

DESCRIPTIVE STRATIGRAPHY

Pokegama Quartzite

General Statement

The Pokegama Quartzite crops out on the Mesabi range as a narrow belt between the northern edge of the overlying Biwabik Iron-formation and the underlying, truncated Lower Precambrian rocks. The maximum width of the outcrop belt is about half a mile on the western end of the range (fig. IV-2). On the East Mesabi, and locally elsewhere, the formation appears to be missing, because the iron-formation lies directly on older rocks. Scattered outliers occur as much as half a mile north of the main outcrop belt, indicating a much wider distribution previously. Natural exposures of the formation are rare because the outcrop belt is mostly buried by glacial drift. Several artificial exposures have been made in the iron ore mines along the Mesabi range, and the formation has been penetrated by diamond drilling at many places. However, at most of these localities data are available on only the upper few feet of the formation.

The name "Pokegama" was first used by H. V. Winchell (*in* Winchell, N. H., 1893a, p. 123) for exposures at Pokegama Falls on the Mississippi River north of Grand Rapids. Although this area is considered the type locality, neither the top nor the bottom of the formation can be seen, and

Figure IV-2. Generalized geologic map of the Mesabi range (compiled from White, 1954; Sims and others, 1970).

only a few of the rock types present elsewhere in the formation are exposed. The thickest and most complete section available for study is a diamond drill core from a hole located half a mile southeast of Eveleth. In this core, 117 feet of a total section of 167 feet consists of thin-bedded, fine-grained feldspathic quartzite and intercalated micaceous, quartzose argillite. A second relatively complete and readily available section, located north of Hibbing at the Pool Location Quarry, is 55 feet thick and contains most of the important rock types known to be present in the formation. Unfortunately, however, neither the upper nor lower contact is exposed. Thus, nowhere is the formation exposed in its entirety.

Lithology

Although quartzite is the most common lithology along the outcrop belt itself, feldspathic quartzite, feldspathic graywacke, and micaceous quartzose argillite are significant rock types in the few holes that have been drilled downdip from the outcrop area.

Commonly, beds of quartzite range in thickness from an inch to 20 feet and are white to greenish gray or pinkish gray. The framework grains, mostly quartz and minor microcline and plagioclase (fig. IV-3), range in size from fine

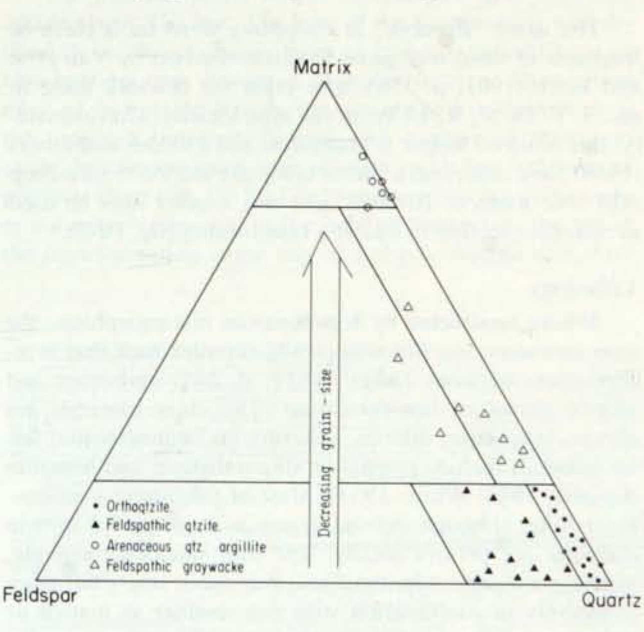


Figure IV-3. Summary of the mineralogic composition of rocks in the Pokegama Quartzite.

to coarse sand; some beds near the bottom of the formation are conglomeratic and have angular to subrounded pebble- and granule-size clasts of quartz, chert, granite, slate, and fine-grained igneous rocks that are difficult to identify.

The framework grains are cemented dominantly by silica, but carbonates, mainly calcite, and, near the top of the formation, siderite, locally are cementing agents. In addition, various proportions of a silt- and clay-size matrix material consisting mostly of chlorite and muscovite fill interstices. The presence of a matrix tends to give the rock a dark color, and there is a complete gradation from white quartzite to dark grayish-black or dark greenish-gray argillite. The gradation is expressed texturally by a decrease in the amount and size of the framework grains, and by an increase in the amount of matrix material.

Locally, thin, irregular beds of medium dark-gray, black, and bluish-gray cryptocrystalline chert, which may or may not contain granules, are interbedded near the top of the formation, as are beds of jasper, thinly laminated hematite, and various kinds of intraformational conglomerate.

Stratigraphy

The Pokegama Quartzite lies directly on an erosion surface cut into a variety of older rocks. This surface is inferred to have had a moderately irregular relief of about 100 feet. The contact between the Pokegama Quartzite and the overlying Biwabik Iron-formation, in contrast, is fairly regular and in most places appears to represent an abrupt change from dominantly clastic sedimentation to dominantly chemical precipitation. Where the contact is abrupt, it is marked by a thin, poorly developed intraformational conglomerate, but locally the contact is gradational and marked by intercalated beds of quartzite and iron-rich chert and shale. Ferruginous layers containing conglomeratic clasts and algal chert fragments are common in the transitional sequence, as are intercalated beds of siderite-cemented quartzite. Locally some of the siderite has been oxidized to hematite or goethite, which imparts a red color to the quartzite.

As defined, the Pokegama Quartzite is variable in thickness, in part as a consequence of local irregularities of the pre-Pokegama erosion surface. Available data suggest that the formation thickens to the west. In the East Mesabi district, the formation is thin and locally missing; it has an average thickness of about 25 feet and a known maximum of 55 feet. In the central part of the range, the formation averages about 60-70 feet thick, attains a known maximum of 167 feet and probably does not exceed 200 feet. Although data are sparse, White (1954, p. 20) concluded that the formation thickens westward to ". . . perhaps as much as 350 feet . . ." and Dolence (1961, unpub. M.S. thesis, Univ. Minn.) has suggested that it also thickens downdip from the outcrop.

Determination of vertical or lateral facies changes in the formation is handicapped by the sparse data; however, some general trends can be suggested. Massive quartzite dominates the upper part of the formation and that part of the formation exposed along the northern and western margins of the outcrop area. However, fine-grained quartzite, feldspathic graywacke, and micaceous quartzose argillite are more abundant to the south and west. Hence, fine-grained rock types may be the dominant lithology where the formation is thicker. Because the outcrop belt approximately parallels the probable strand line of the Animikie sea (White, 1954, p. 43), this facies pattern may indicate that the formation thickens and becomes finer grained southward.

Petrology

Pokegama sediments apparently were deposited on a slightly irregular erosional surface composed of a variety of igneous, metasedimentary, and metavolcanic rocks. It is inferred that the Animikie sea gradually advanced over this irregular terrane and deposited a conglomerate near the strand line, and at the same time deposited sand- and pebble-size material away from the strand line. Generally, the sand-size detritus is well sorted and rounded, indicating deposition in a fairly high-energy environment. Apparently the finer grained detritus was winnowed out and deposited still farther from the strand line in deeper and somewhat quieter water. Alternating beds of sand- or silt-size material possibly indicate minor advances and retreats of the strand line, but the existence of local, abrupt lateral changes in lithology argues that variable currents had a significant effect on sediment distribution. Thin laminations in the argillaceous rocks are indicative of quiet-water deposition, and the presence of disseminated iron sulfides implies a somewhat reducing environment.

The detrital material in the Pokegama Quartzite appears to have been derived from the Lower Precambrian rocks that crop out north of the Mesabi range, as indicated by the lithologic similarity of the basement rocks and the conglomeratic clasts, the variety and proportion of the heavy minerals, and the nature of the quartz and feldspar. In describing the basal conglomerate, Grout and others (1951, p. 1043) stated ". . . the conglomerate consists largely of fragments of the rock immediately below it. . . ." Although data are sparse, observations since that time have borne out this conclusion. For example, in a drill hole south of Biwabik (fig. IV-2), the lower part of the Pokegama consists of approximately 5 feet of material that resembles the underlying granite. In thin section, this unit is seen to consist of grus-like material that apparently formed by the mechanical disaggregation of the underlying granite. The grus passes transitionally upward into typical detrital grains that are well rounded and well sorted. Similarly, there is a close correlation between the heavy minerals of the Pokegama Quartzite and the accessory minerals of the Giants Range Granite (fig. IV-4). Tyler and others (1940, p. 1493) noted that the Pokegama sediment was in large part derived ". . . from the Giants Range batholith, although some was no doubt derived from the greenstone and slate and some from more distant formations. . . ." With two exceptions—(1) sphene in the granite is largely altered to anatase and/or leucoxene in the sediments, and (2) apatite and epidote are reduced in amount relative to zircon—the comparative data shown in Figure IV-4 indicate the similarity in composition of the mineral assemblages of the two rock units.

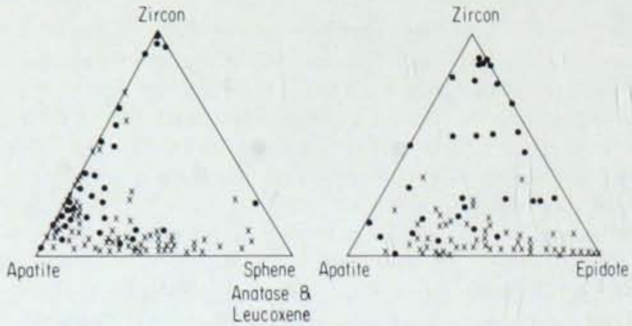


Figure IV-4. Relative amounts of selected heavy minerals in the Giants Range Granite and Pokegama Quartzite (modified from Tyler and others, 1941). A. Relative amounts of zircon, apatite, and sphene in Giants Range Granite (x) and zircon, apatite and sphene + anatase (+ leucoxene) in Pokegama Quartzite (•). Sphene is largely altered to anatase and/or leucoxene and apatite is reduced relative to zircon during sedimentation. B. Relative amounts of zircon, apatite, and epidote in Giants Range Granite (x) and in Pokegama Quartzite (•). Both apatite and epidote were reduced relative to zircon during sedimentation.

Biwabik Iron-formation

General Statement

The Biwabik Iron-formation, a ferruginous chert that contains 25-30 percent iron, has been studied more intensely than any other formation in Minnesota because of its value as a source of iron ore, which has been mined since 1892. The primary iron-formation was called "taconite" by Winchell (1893a) and this term is widely used interchangeably with "iron-formation."

The name "Biwabik" (a Chippewa word for a piece or fragment of iron) was given to this formation by Van Hise and Leith (1901, p. 356), who cited the Biwabik mine in sec. 3, T. 58 N., R. 16 W. as the type locality. Unfortunately, this mine no longer is accessible, but Pfleider and others (1968) have described a rather complete section from a deep drill hole south of Biwabik, and this locality may be used as reference section in lieu of a type locality (fig. IV-2).

Lithology

Where unaffected by Keweenawan metamorphism, the iron-formation is a mineralogically complex rock that is intermediate between James' (1954, p. 242) carbonate and silicate facies of iron-formation. The chief minerals are quartz, magnetite, siderite, ankerite, and minnesotaite; lesser minerals include greenalite, stilpnomelane, and hematite (Gruner, 1946; White, 1954). Most of the quartz is microcrystalline, although detrital grains as large as 0.2 mm in diameter are present locally. The iron silicates greenalite, minnesotaite, and stilpnomelane, may occur singly but more commonly in combination with one another as matted or radiating aggregates. Commonly, they are difficult to distinguish because of intimate intermixing and fine grain size; therefore, they are generally considered as a group. Magnetite forms tiny octahedra that are rarely more than 0.1 mm in diameter. Commonly, the individual octahedra are clustered together in regularly to irregularly banded, laminated, patchy, or mottled aggregates. The carbonates form small to large rhombs or irregularly rounded grains.

The iron-formation can be classified on the basis of textures observable in hand specimen or thin section as seemingly coarse-grained or "granular" types and fine-grained or "slaty" types. Actually, nearly all the mineral particles are very fine grained, but the aggregation of particles into spherical or ellipsoidal bodies, called "granules" by Spurr

(1894, p. 49), which average about 0.5 to 2.0 mm in diameter, gives some rocks a coarse-grained appearance. The granules are internally structureless and are composed of cherty quartz, iron silicates, iron carbonates, or iron oxides, and are enclosed in a fine-grained matrix of chert, iron silicates, iron carbonates, or iron oxides. In general, however, iron-formation composed dominantly of chert with iron silicates or magnetite is likely to be slaty. Thus, two fundamentally different kinds of iron-formation can be distinguished (Wolff, 1917; Gruner, 1946; White, 1954): (1) cherty iron-formation, which is characteristically massive, quartz-rich, and has a granular texture; and (2) slaty iron-formation, which is generally fine grained, finely laminated, and composed mostly, but not exclusively, of iron silicates and iron carbonates. The two types of iron-formation correspond approximately to James' (1954, p. 249, 267-270) "granular" and "nongranular" facies of iron-formation. The two types also differ somewhat in composition. Although the total iron content is about the same, there is more magnesium and aluminum in the slaty strata than in the cherty strata; *i.e.*, the slaty rocks are characterized by an alumina content of as much as 7 percent, whereas the cherty rocks contain only 1 percent (Gruner, 1946).

Stratigraphy

The iron-formation ranges in thickness from less than 200 to about 750 feet. The base of the formation is well defined by an abrupt change from an iron-poor quartzite to an iron-bearing rock having a granular texture. Throughout most of its outcrop length, the top of the iron-formation is the top of a limestone-bearing unit containing little iron oxide, but having some iron silicates and a few interbeds of granular chert (fig. IV-5). The limestone-bearing unit pinches out near Nashwauk; to the west of Nashwauk, the top of the iron-formation is the top of a cherty siderite unit. Still further west, the cherty siderite unit also fingers out, and on the westernmost Mesabi the top of the iron-formation is placed at the top of a graphitic argillite unit that commonly is iron-bearing. On the westernmost Mesabi, several iron-bearing members, one as much as 200 feet thick, are interlayered with argillite of the Virginia Formation. Clearly lithologies interfinger in this area.

Beds or groups of beds having cherty and slaty textures are interbedded on all scales in the iron-formation. Nevertheless, lithostratigraphic units that extend over long distances can be defined. Wolff (1917) was the first to divide the iron-formation into four major lithostratigraphic units, which, from bottom to top, are: (1) lower cherty; (2) lower slaty; (3) upper cherty; and (4) upper slaty. These units, which have been retained as members (Gruner, 1946; White, 1954), can be traced along much of the main Mesabi range and can be recognized even in the metamorphosed iron-formation adjacent to the Duluth Complex. Except for the lower cherty-lower slaty contact, which is rather sharp and can be defined accurately, the contacts between adjacent members are gradational. Because of vertical and lateral lithologic changes within the iron-formation, subdivision of the members at places is arbitrary, but nevertheless is useful from a genetic standpoint. In general, stratigraphic units included in the slaty members contain 40 percent or more slaty strata (White, 1954), although locally within a given member the proportion may lessen. The slaty strata characteristically contain sparse iron oxides, and the associated granular or cherty beds commonly are silicate-rich. The cherty members, on the other hand, contain between 10 and 30 percent slaty strata; the cherty beds are rich in magnetite (or locally hematite) although they also contain abundant silicates or carbonate.

Despite the variable lithology, the four members and the several subunits are recognizable throughout much of the range. Of the subunits, two algal-bearing beds that range

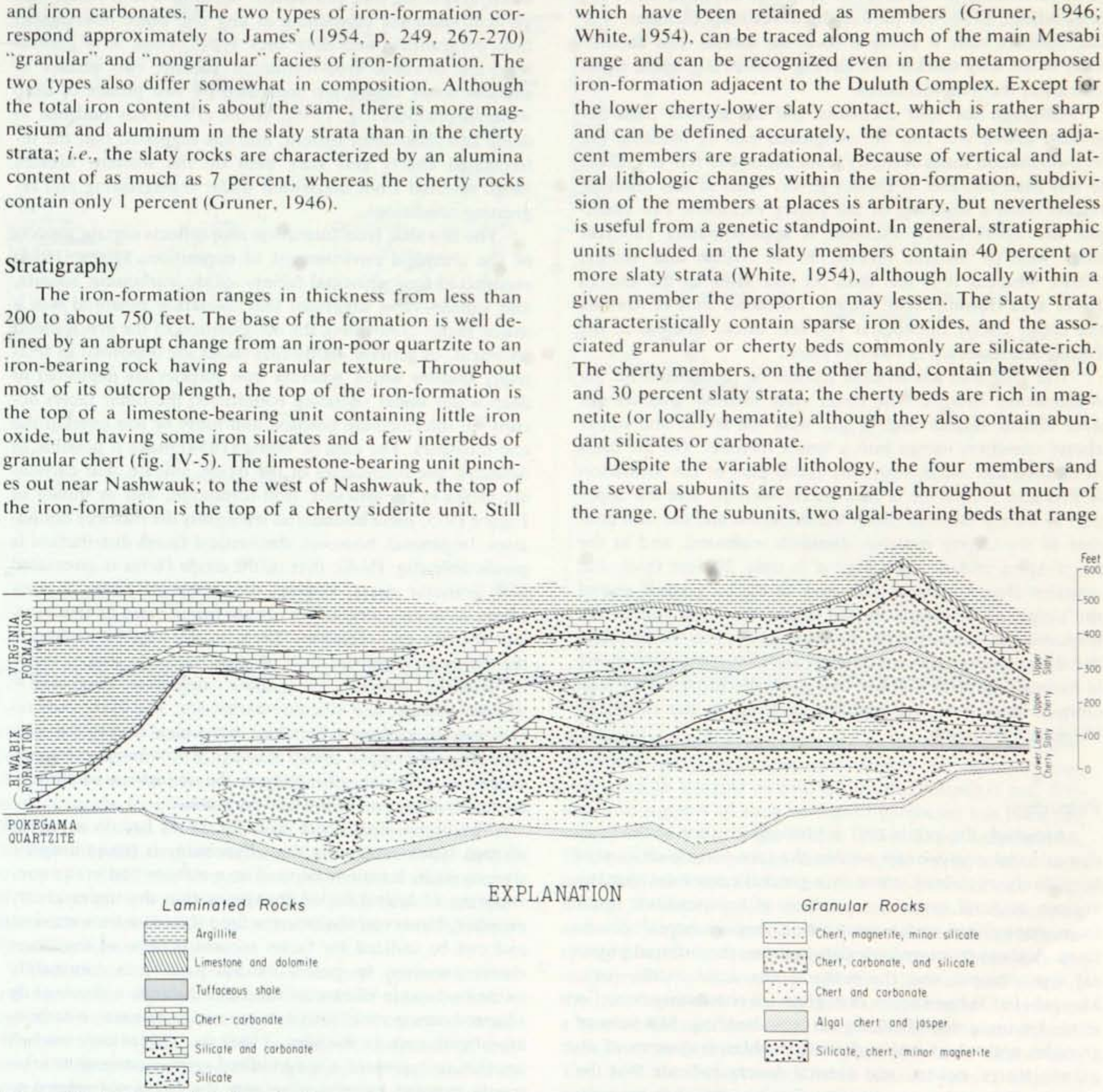


Figure IV-5. Generalized longitudinal stratigraphic section of the Biwabik Iron-formation showing relationship between member boundaries and various facies (modified from the data of White, 1954).

in thickness from several feet to several tens of feet constitute distinctive stratigraphic markers throughout most of the range (fig. IV-5). The thickest and most persistent algal unit is at the base of the iron-formation, whereas the other unit is in the middle of the upper cherty member. The algal structures occur either in their original growth position or as intraformational fragments. Most commonly, they are composed of laminated red and white chert, and are set in a jasper matrix that also contains granules and other kinds of intraformational conglomerate clasts.

A third marker bed, the so-called intermediate slate, is a dark gray or black, thinly laminated unit containing abundant carbonaceous material. It occurs at the base of, or locally within, the lower part of the lower slaty member. The intermediate slate is inferred to be an ashfall, and accordingly is the only marker bed having time-stratigraphic significance in the iron-formation.

Although the four members and the marker beds described above still can be recognized east of Mesaba, the iron-formation thins abruptly from about 400 feet at Gilbert to less than 200 feet at Dunka River. Most of this thinning results from a thinning of the cherty members. For example, the lower cherty member is approximately 150 feet thick west of Mesaba (White, 1954; Griffin and Morey, 1969), whereas it is less than 35 feet thick in the Dunka River area (Bonnichsen, 1969b). Likewise, west of Mesaba the upper cherty member is 180 feet thick, whereas in the Dunka River area it is 100 feet thick.

The four-fold subdivision cannot be recognized on the westernmost Mesabi. The lower slaty member pinches out near Grand Rapids and to the west the lower and upper cherty members merge into a single member. On the basis of texture and mineralogy, the lower part of the formation in this area is referred to as a cherty member and the upper part as a slaty member. Both the thickness and the iron content of the cherty member diminish westward, and at the far western end of the range it is only 20 feet thick and contains almost no iron. However, at the far western end of the range, as elsewhere on the range, the basal part of the formation is characterized by algal structures. In contrast, the slaty member consists of chert and abundant siderite and is intercalated with argillite that generally has a higher iron content than does the argillite assigned to the overlying Virginia Formation.

Petrology

Although the origin and subsequent history of the various mineral components within the iron-formation cannot be entirely explained, there is a general consensus that the various textural aspects, regardless of composition, result from deposition under contrasting environmental conditions. A close relationship exists between the inferred physical environment and the textural character of the rock. Mengel (1965) has shown that granules commonly occur in strata having graded bedding or cross-bedding. Mixtures of granules with chert and carbonate pebbles, fragments of algal structures, oolites, and detrital quartz indicate that the granules behaved as particulate detritus and that granule-bearing taconite may be regarded as a special kind of sandstone. LaBerge (1967a) suggested that the grain size and bedding aspects of the non-granular or slaty taconites are similar in many respects to siltstone or argillite, and further suggested that many of the granules in the cherty beds were derived from similar fine-grained material.

Thus, it seems reasonable to postulate that at any time during iron-formation deposition, cherty, granule-bearing sediments would be deposited relatively near the strand line in a shallow-water, agitated environment, whereas slaty or thin-bedded sediment would be deposited in deeper, less active water well away from the strand line. Fluctuations in the relative position of the strand line or periods of abrupt basin deepening and subsequent infilling would result in a vertical sequence containing at any one place intercalated beds of granular and slaty rock types. Thus, in a general way the Biwabik Iron-formation records two phases of shallow water deposition separated by two phases of deep water deposition (fig. IV-6). White (1954) has outlined in detail this lithofacies pattern, and has suggested that the intercalated cherty and slaty beds on the western Mesabi range resulted from deposition under transgressing and regressing conditions.

The Biwabik Iron-formation also reflects certain aspects of the chemical environment of deposition. James (1954) recognized four principal facies—oxide, carbonate, silicate, and sulfide—and suggested that oxidation potential was a major factor controlling the composition of the precipitated sediment. In general, oxide-rich facies are deposited in relatively shallow water, whereas iron sulfides are deposited in deep water under reducing conditions; the other facies occupy an intermediate position and more or less overlap the end members. The data of White (1954) make it possible to delineate the boundaries of the oxide, silicate, and carbonate facies in the Biwabik Iron-formation, and as shown in Figure IV-5, these boundaries transgress the member boundaries. In general, however, the vertical facies distribution is predictable (fig. IV-6); that is, the oxide facies is associated with granular strata, whereas the carbonate facies is associated with slaty strata; the silicate facies more or less overlaps the two end members. Thus, in any vertical sequence, the facies arrangement from shallow water to deeper water is: hematite; hematite+magnetite; magnetite; magnetite+silicate; silicate; and silicate+carbonate. The cycle then reverses itself, leading to hematite in the algal unit in the middle part of the upper cherty member. The water depth again deepened, resulting ultimately in the deposition of carbonate facies at the top of the iron-formation.

The upper algal chert unit (fig. IV-5) lies on a variety of rock types, indicating that it probably is time-transgressive; as such, it cannot be used as a marker bed in the construction of lateral facies changes within the upper cherty member. However, the intermediate slate is a time marker, and can be utilized for facies reconstructions of the lower cherty member. In general, facies pass from dominantly oxide and oxide-silicate in the east, through a dominantly silicate-bearing rock, into a dominant carbonate or carbonate-silicate rock in the west. Thus, the present outcrop belt appears to represent a longitudinal section somewhat arbitrarily exposed by tectonic events that were not related to the configuration of the original basin of deposition. How-

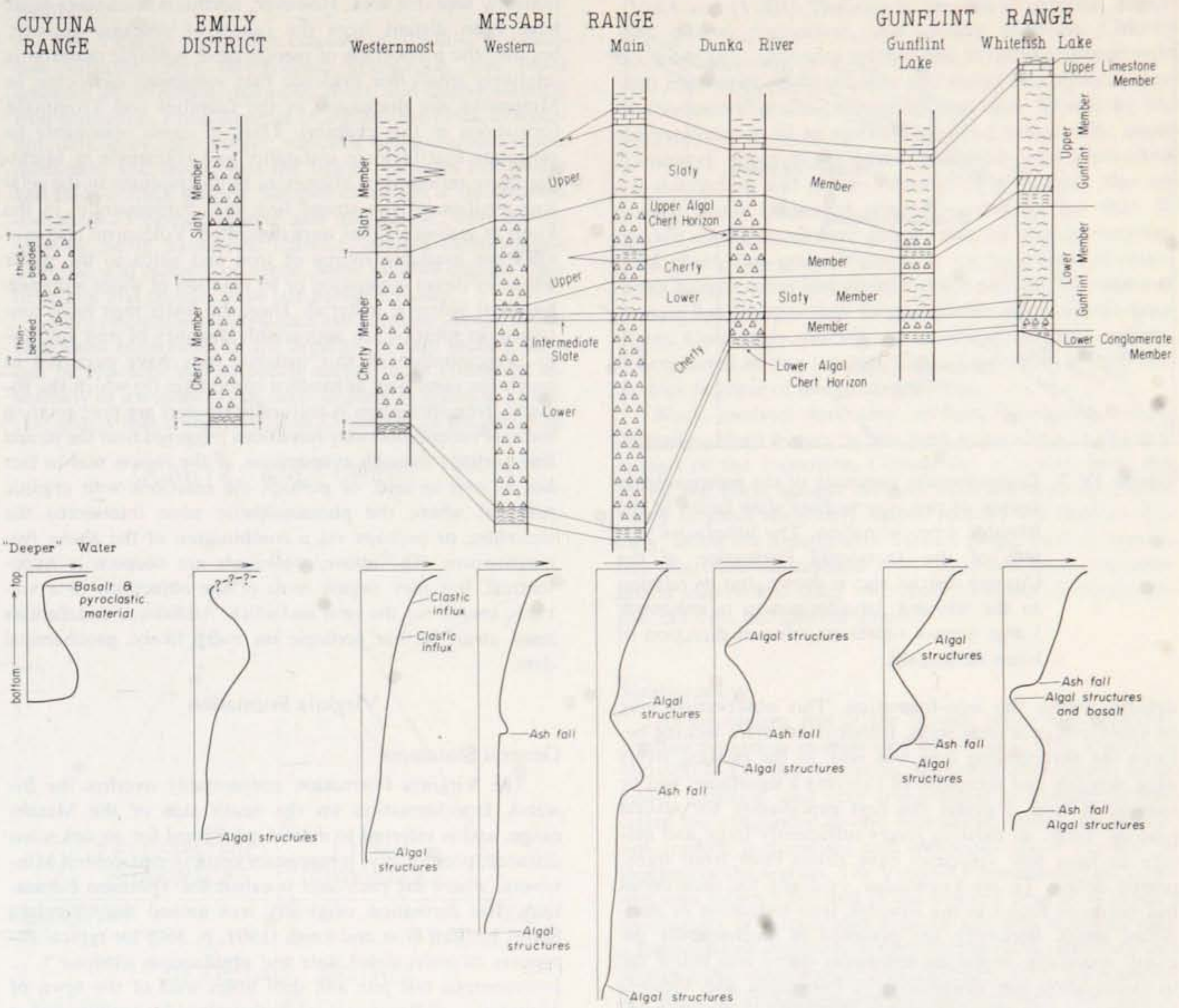


Figure IV-6. Summary of various lithotopes and their inferred sedimentologic setting in the Biwabik Iron-formation.

ever, an inferred strand line and facies boundaries can be delineated (fig. IV-7); the strand line is roughly parallel to the present outcrop pattern. Apparent local lithologic variations along the longitudinal section may be a result of facies change parallel to the strand line, but they are due also to the fact that the tectonic strike and the depositional strike are not parallel; iron-formation exposed at the eastern end of the section was deposited farther down basin from the inferred mean position of the strand line than was the iron-formation west of Virginia.

The source of the silica and iron now present in the iron-formation is unknown, and there is no evidence bearing on this problem inherent in the rocks themselves. Van Hise and Leith (1911, p. 516) concluded that much of the iron and silica was derived from magmatic springs and to a lesser extent from the reaction of sea water with hot submarine flows. Gruner (1922, p. 459) concluded, however, that weathering of a land mass of basaltic rocks under humid or semi-tropical conditions could supply the necessary iron and silica to streams emptying into the Animikie sea. Evidence in support of each of these hypotheses has been discussed many times (*e.g.*, White, 1954; Lepp and Goldich, 1965) and will not be reiterated here. However, several sedimentological features that bear on the problem are worth noting. Evidence exists that the detrital rocks underlying and overlying the iron-formation were derived from a dominantly granitic terrane located to the north. These data are incompatible with a model whereby the iron and silica in the iron-formation were derived through the weathering of a basaltic terrane. Secondly, the available evidence suggests that the iron-formation was deposited near a strand line. Yet, with the exception of some detrital quartz near the bottom of the formation, which may represent material reworked from underlying strata, there is virtually no clastic

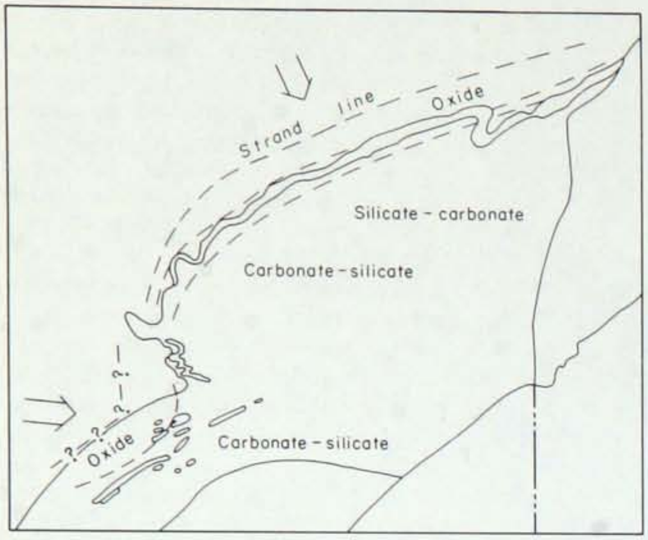


Figure IV-7. Diagrammatic summary of the paleogeologic setting of pre-intermediate slate facies in the Biwabik Iron-formation. The lithofacies pattern of the Trommald Formation of the Cuyuna district also is shown, but its relation to the Biwabik Iron-formation is unknown. Large arrows represent inferred direction of basin deepening.

detritus within the iron-formation. This observation may be explained in several ways. Either rivers were lacking because the surrounding area was arid, or the existing rivers were sluggish and incapable of carrying a significant bed or suspended load. I prefer the first explanation for several reasons. First, all existing rivers sufficiently large and mature to have low velocities have deltas built from transported debris. To my knowledge, evidence for such deltas has not been found in the Biwabik Iron-formation or associated strata. Secondly, the presence of mechanically derived, texturally immature sediments above and below the iron-formation (see discussions of Pokegama and Virginia mineralogy, this chapter) implies that little or no chemical weathering took place over a prolonged period of time in this region. Thus, it seems improbable that extensive weathering of the adjoining land and subsequent transport of the weathering products could provide the necessary iron and silica.

Arguments against a volcanic source have been enumerated by Gruner (1924), James (1954), and Tyler and Twenhofel (1952), and include: (1) lack of direct evidence of a volcanic association; (2) the large volume of volcanic material necessary to provide the needed amounts of silica and iron; and (3) the improbability that local magmatic sources could produce an iron-formation of such broad extent. A postulated local volcanic or magmatic source, however, may not be necessary for the Biwabik Iron-formation, inasmuch as there is ample evidence of extensive volcanic activity elsewhere in the basin. James (1958) noted that approximately equivalent rocks in northern Michigan contain much volcanic material and that individual formations are lenticular, indicating deposition in a series of small and relatively short-lived basins in a volcanically active and tectonically unstable area. However, northern Minnesota must have been distant from the center of volcanic activity, because the proportion of recognizable volcanic material is relatively small (for evidence that volcanism did occur in Minnesota, see discussion of the Gunflint and Trommald formations in this chapter). Thus, it seems reasonable to postulate that tectonic instability and volcanism in Michigan were reflected in Minnesota by fluctuations in the relative position of the strand line and, consequently, in the kinds of sediments that were deposited. Volcanism also provides an available source of iron and silica to the water either by direct emanation or by reaction of water with perhaps hot volcanic material. Thus, the water may have contained, in solution, an appreciable quantity of iron and silica. Precipitation of this material may have occurred in much the same way as modern carbonates (to which the Biwabik Iron-formation is texturally similar) are precipitating today. Precipitation may have been triggered near the strand line perhaps through evaporation, if the region was in fact hot as well as arid, or perhaps via reactions with organic material where the photosynthetic zone intersected the shoreline, or perhaps via a combination of the above two mechanisms. The above arguments are necessarily hypothetical, but they negate most of the objections to a volcanic source for the iron and silica. Additional conclusions must await further geologic or, more likely, geochemical data.

Virginia Formation

General Statement

The Virginia Formation conformably overlies the Biwabik Iron-formation on the south side of the Mesabi range, and is inferred to extend southward for an unknown distance; presumably, it reappears again in east-central Minnesota, where the rock unit is called the Thomson Formation. The formation originally was named the "Virginia Slate" by Van Hise and Leith (1901, p. 360) for typical exposures of intercalated slate and argillaceous siltstone ". . . in numerous test pits and drill holes west of the town of Virginia. . . ." Because it is characterized by argillite rather than true slate, however, the name was changed to "Virginia Formation" by White (1954, p. 18), and that terminology is used here. Aside from several localities in the East Mesabi district, there are no known natural exposures; the lower part of the formation is exposed, however, in several of the open pits along the range. Although the formation has been penetrated by diamond drilling at several places, no detailed stratigraphic or mineralogic studies have been completed. The most complete and continuous record of the formation is described by Pfleider and others (1968) in a report on the underground mining potential of taconite on the Mesabi range. Approximately 3,500 feet of Virginia core was obtained from four holes. Of this total, 1,485 feet was penetrated south of Biwabik and 1,129 feet south of Calumet (fig. IV-2). The cores obtained from these holes are considered representative of the Virginia Formation at the eastern and western ends of the Mesabi range, respectively. In lieu of accessible exposures at a well-defined type locality, these two diamond drill holes are considered alternative reference sections to the original type locality.

Lithology

The Virginia Formation consists of argillite, argillaceous siltstone, very fine-grained graywacke, and lesser amounts of limestone, dolomite, chert, and cherty sideritic iron-formation. Fresh samples of argillite are black to medium dark gray, the darker varieties being carbonaceous, mica-rich, and fissile, and the lighter varieties being silty and thin- to thick-bedded. Samples examined in thin section contain 5-15 percent quartz and plagioclase, with the remainder being muscovite, chlorite, opaque carbonaceous material, and euhedra of pyrite. Some samples contain minor calcite that appears to be late paragenetically.

Very fine-grained graywacke and siltstone are the coarser grained equivalents of the argillite. In thin section, they can be characterized as poorly sorted rocks composed essentially of a framework of sand- or silt-size grains in a clay-size matrix that comprises 15 to 75 percent of the rock (fig. IV-8A and IV-8B). The matrix consists of chlorite, muscovite, quartz, plagioclase, and opaque minerals. Chlorite occurs in lacy, radiating patches and as fibrous intergrowths with muscovite. Both chlorite and muscovite replace edges of framework grains, indicating that they formed by the recrystallization of an earlier clay-size fraction. The major framework minerals are quartz and sodic plagioclase (An_{2-15}); microcline and minor amounts of orthoclase also are present. Rock fragments generally comprise less than 10 percent of the framework grains, and are mostly recrystallized chert, fine-grained quartzite, or fragments of extensively altered schist and metabasalt. In general, the siltstones contain less feldspar than the graywacke and lack rock fragments. Commonly, the silt-size grains are angular, and are concentrated in thin laminae that are interlayered with thin to thick laminae of mica-rich argillite.

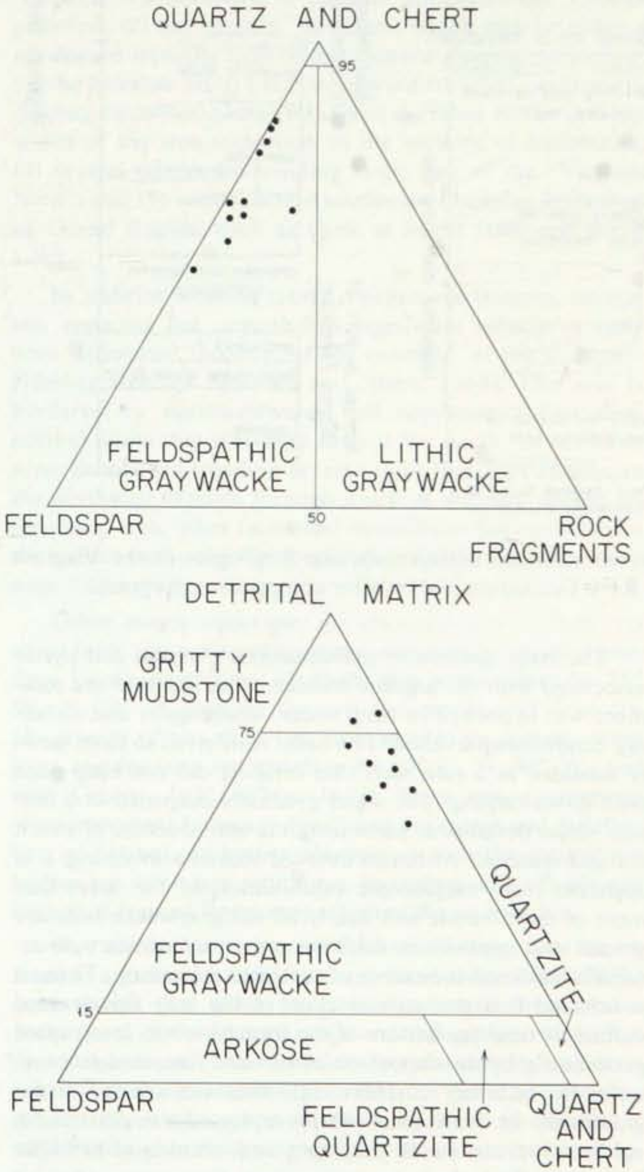


Figure IV-8. Summary of mineralogic composition of rocks in the Virginia Formation.

Black, medium dark gray, or light bluish-gray cryptocrystalline chert occurs as thin beds and nodules in the lower part of the formation. Commonly, it is very pure, but pyrite and black opaque carbonaceous material are present rarely as small irregularly shaped patches and laminae. Locally, as on the westernmost Mesabi range, beds of crystalline chert are intercalated with laminated layers of intermixed siderite and chert with lesser amounts of calcite, pyrite, chlorite, and detrital quartz.

Stratigraphy

At the eastern end of the Mesabi range near Biwabik, argillite comprises an estimated 80 percent of the lowermost 400-500 feet of the formation. Repeated alternations of argillite, siltstone, and very fine-grained graywacke characterize the remainder of the formation, where penetrated, but argillite is dominant and comprises about one-half of this part of the section. The Virginia Formation in the Calumet area differs from that in the Biwabik area in consisting dominantly of intercalated argillite and silty argillite, with a few scattered interbeds of very fine-grained sandstone and siltstone. An iron-bearing unit separated from the underlying Biwabik Iron-formation by approximately 200 feet of argillite and argillaceous siltstone also occurs in this area. This unit, at least 70 feet thick, consists of intercalated beds of argillite and cherty sideritic-iron-formation, and is similar lithologically to cherty sideritic rocks in the upper slaty member of the Biwabik Iron-formation.

The stratigraphic succession in the Biwabik area is much like that comprising the Rove Formation on the Gunflint range (Morey, 1969). The available data (fig. IV-9) suggest that the lower argillaceous succession thickens westward from Biwabik. This may be the result of real differences in the sedimentological regime of these sediments, or it may be the result of the fact that the east-west cross-section shown in Figure IV-9 does not parallel the inferred shoreline of the Animikie sea, and thus intersects different parts of the original depositional basin.

Petrology

The contact between the Biwabik Iron-formation and the Virginia Formation on the Mesabi range marks the gradual return of clastic deposition to the Animikie sea.

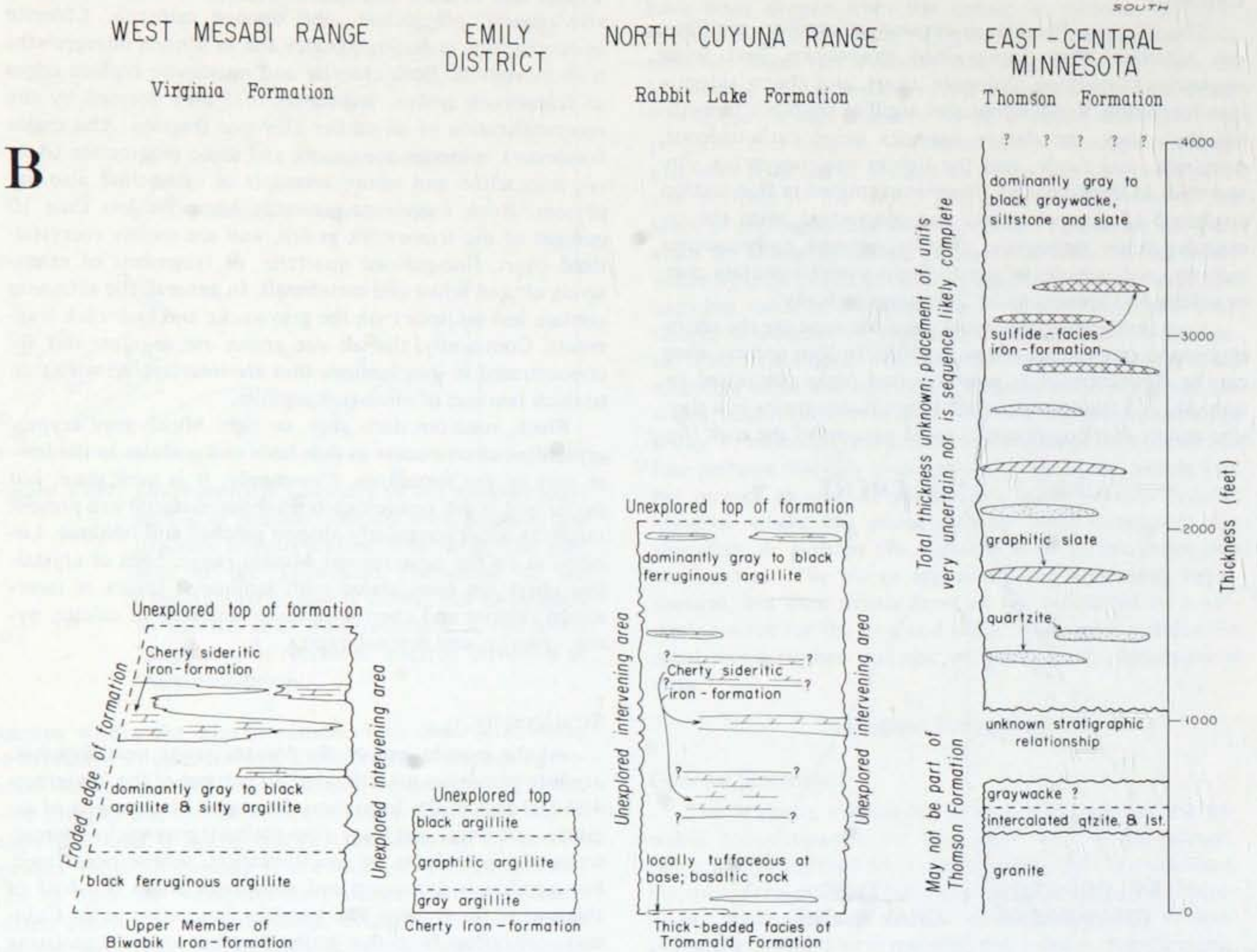


Figure IV-9. Diagrammatic east-west cross-section showing inferred relations between various lithologies in the Virginia Formation. Generalized lithologic succession in the Rove Formation is shown for comparative purposes.

Across much of the range, this transitional succession is characterized by interbedded limestone, dolomite, cryptocrystalline chert, and argillite. At some places the transition zone is as much as 30 feet thick; elsewhere, it is less than 10 feet thick. Commonly, the transitional succession contains conglomeratic layers consisting of chert, limestone, and argillite clasts as much as 20 inches in diameter, which appear to be of intraformational origin, and most likely were formed in a subaqueous environment. Thus, they do not represent a period of subaerial erosion as has been suggested by Wolff (1917) and by Leith and others (1935, p. 15). West of Hibbing, argillaceous rocks that directly overlie the iron-formation contain as much as 20 percent iron (Gruner, 1946). The presence of lean iron-formations well within the Virginia Formation indicates the gradational and transitional nature of this contact. Evidently there was only a slight change in environment from deposition of chert-siderite through simultaneous precipitation of iron carbonate and clastic deposits to deposition of clastic deposits in this part of the Animikie depositional basin.

The large amount of carbonaceous material and pyrite associated with the argillite indicates that much of the sediment was deposited in deep water, where quiet and reducing conditions prevailed. The basin is inferred to have slowly subsided at a rate such that infilling did not keep pace with downwarping. The water gradually deepened as a bottom slope developed, permitting the introduction of much silt-size material. Although detailed studies are lacking, it is apparent from megascopic examination of the cores that most of the siltstone and nearly all the graywacke beds are graded and contain, in addition, other sedimentary structures considered indicative of turbidite deposition. Thus, it is inferred that the accumulation of the very fine-grained sediments near the bottom of the formation was interrupted periodically by the deposition of silt- and fine-sand-size material by turbidity currents. The increase upward stratigraphically in the amount of silt- and sand-size detritus records an increase in the frequency and intensity of turbidite deposition.

The mineralogy of the Virginia Formation is much like that of the Rove and Thomson Formations. For these formations, Morey (1969) and Morey and Ojakangas (1970) concluded that a granitic plutonic terrane was the major source for the sediments, and there is no evidence indicating a different source for the detritus in the Virginia Formation.

STRUCTURE

The gross structure of the Mesabi range is a gently dipping homocline that strikes east-northeast and dips 5°-15° SE. This general trend is interrupted by several structural features that have caused noticeable bends in the Animikie strata or pronounced changes in the outcrop width of the Biwabik Iron-formation. Among the more prominent structures are (fig. IV-2): (1) the "Virginia horn," a broad, gentle fold consisting of a southwestward-plunging syncline—the Virginia syncline—and a parallel anticline—the Eveleth anticline; (2) the "Siphon" structure, which may be either a northward-trending fault or an eastward-dipping monocline; (3) the Biwabik fault, a northwestward-trending, southward-dipping fault that causes a marked decrease in the outcrop width of the iron-formation in the vicinity of Embarrass; (4) several northward-trending faults east of the "Virginia horn"; and (5) several broad southward-plunging folds west of Grand Rapids, such as those at Sugar Lake and Birch Lake.

In addition to these rather conspicuous features, several less apparent but nevertheless significant structures have been delineated recently, as for example, in the Calumet-Hibbing area (*cf.* Marsden and others, 1968). This area is bordered by northwestward- and northeastward-trending normal faults that converge toward the north. In addition, a northeastward-trending reverse fault that dips steeply to the northwest extends through much of the area. In the intervening area, other faults and monoclines fan out between the border faults and dip toward the middle, forming broad steps into a graben.

Other minor structures are common and include: (1) small folds and northeastward-trending monoclines that produce locally steep dips; (2) faults that strike about N. 75° W., N. 20° W., and, at the east end, north, and have displacements of less than 100 feet; and (3) prominent, steep joint and fracture sets trending N. 10° E., N. 40° W., and east (Gruner, 1922; White, 1954). These minor structures were important factors in localizing the extent and distribution of natural ore bodies. Associated with the natural ore bodies are secondary structures, including slump faults and folds that formed from compaction of the natural ore.

Not all the structures in the Mesabi range can be related to a particular tectonic event. Sims and others (1968b) have shown that some faults in the Lower Precambrian rocks north of the Mesabi range are contiguous with known faults that transect the Biwabik Iron-formation, and have inferred that many of the faults that cut the iron-formation resulted from renewed movements on the older structures, which formed initially during Early Precambrian time. Apparently, movement along some of these faults occurred during deposition of the iron-formation (*cf.* the thinning of the Biwabik Iron-formation across the Siphon structure; Gundersen and Schwartz, 1962, pl. 1), whereas movement on others clearly took place after consolidation of the Animikie strata. Other structures, however, appear to be unrelated to pre-existing Lower Precambrian structures. The "Virginia horn" for example, is folded on a northeastward-trending axis, which intersects the present outcrop pattern of the Biwabik Iron-formation at a low angle. Because the crest of the Eveleth anticline can be traced southwestward by geophysical methods for about 15 miles, the Virginia syncline is a large structure whose northwest limb is defined by what appears to be a southward-dipping homocline. The northeastward-trending fold axes on the Mesabi range are parallel to the fold axes in the Cuyuna district (see Marsden, this chapter) and possibly these folds formed contemporaneously during the Penokean orogeny. Thus, the present homoclinal dip of the Biwabik Iron-formation may be largely due to deformation during the Penokean orogeny.

Still other structures appear to be younger. Included in this category are the southward-plunging folds of the westernmost Mesabi range and faults that offset the northeastward-trending structures. At least some of the faults appear to cut and offset dikes that are presumed to be of Keweenawan age (White, 1954, p. 67), and accordingly may be Keweenawan or younger in age.

Clearly, the structure of the Mesabi range is complex and contains elements that are related temporally to several tectonic events. Any structural analysis is complicated also by the geographic distribution of the basement rocks, for as White (1954, p. 67) has pointed out, Animikie rocks are folded only where they are underlain by older metasedimentary and metavolcanic rocks. Where they are underlain by rocks of the Giants Range batholith, they are tilted and fractured and lack appreciable internal folding.

METAMORPHISM

The Animikie strata at the eastern extremity of the Mesabi range were thermally metamorphosed by the Duluth Complex, in Middle Keweenawan time. There is some evidence, also, particularly from the Biwabik Iron-formation, that the Animikie strata had been subjected previously to a widespread, mild pre-Keweenawan metamorphism; this event was related to the Penokean orogeny at around 1,850 m.y. or to a subsequent period of metamorphism around 1,650 m.y.

Pre-Keweenawan Metamorphism (?)

The Biwabik Iron-formation contains abundant minnesotaite and stilpnomelane, which are considered by some to be minerals indicative of the greenschist facies of metamorphic grade (James, 1955). White (1954) has shown that the distribution of these minerals is stratigraphically controlled, and he concluded, therefore, that they were primary in origin. However, textural data (French, 1968) indicate that much of the stilpnomelane and minnesotaite is secondary. Most of the minnesotaite is restricted to cherty beds, where it appears to have replaced greenalite, according to the reaction: greenalite + quartz = minnesotaite + water. On the other hand, stilpnomelane is almost entirely restricted to slaty beds which, on the average, contain a higher percentage of alumina than do the cherty beds (Gruner, 1946). The fine grain size and pervasive recrystallization of the slaty strata preclude definite recognition of the original material, but by analogy with experimental studies, several possibilities can be considered, including an iron chlorite-rich sediment (Yoder, 1957) and volcanically derived material (LaBerge, 1966) containing montmorillonite (Grubb, 1971). Of these possibilities, the iron-rich septechlorite, chamosite, has been identified by French (1968) in the Biwabik Iron-formation. Thus, selective transport during sedimentation may have concentrated clay-size material, such as chamosite and/or altered volcanic detritus, into particular stratigraphic zones. The unique bulk composition of these layers would strongly influence the mineral phases that subsequently formed, and reactions leading to the development of stilpnomelane would occur only in beds having a particular composition. Thus, the apparently stratigraphically controlled distribution of minnesotaite and stilpnomelane is not a limiting factor in considering the origin of these minerals.

There is neither textural evidence from the iron-formation nor diagnostic metamorphic mineral assemblages from the overlying Virginia Formation that are indicative of the physical conditions to which these rocks were subjected. The lack of any spatial relationship with Keweenawan rocks led French (1968, p. 74), however, to conclude that the Biwabik Iron-formation ". . . was subjected to a previous period of low grade metamorphism unrelated to the intrusion . . ." of Keweenawan rocks, and that ". . . such metamorphism may be related to an event at 1,600 to 1,700 m.y., which is recognized on the Cuyuna range and in east-central Minnesota. . . ." Although French undoubtedly was referring to the Penokean orogeny, Hanson and Malhotra (1971, p. 1110) have suggested that an event at this time ". . . may not have been related to the Penokean orogeny . . . but may have been associated with burial by overlying sediments . . .," which caused ". . . prehnite-pumpellyite-facies metamorphism. . . ." Alternatively, the metamorphism may have been related to a mild deformation that was younger than the Penokean orogeny. The latter suggestion is consistent with the data of Perry and Morse (1967), who, using oxygen-isotope ratios in quartz-magnetite pairs, demonstrated that temperatures on most of the Mesabi range never exceeded about 100° C. More recent work on fractionation factors by Becker (1971, unpub. Ph.D. dissertation, Univ. Chicago, p. 86-88) suggests that the temperature reported by Perry and Morse is approximately 50° C too low. If this temperature is geologically significant, it indicates that it will be difficult to make any distinction between high-grade diagenesis and low-grade metamorphism of the iron-formation.

Keweenawan Metamorphism

In the East Mesabi district, a second episode of metamorphism, related to emplacement of the Duluth Complex, is superposed on the broad metamorphic pattern outlined above. Broad features of the contact aureole in the Biwabik Iron-formation have been described by French (1968), who delineated four metamorphic zones on the basis of mineral assemblages. They are: (1) unmetamorphosed iron-formation, consisting of quartz, magnetite, hematite, ankerite, siderite, greenalite, minnesotaite, stilpnomelane, and chamosite; (2) a transitional zone, characterized by the appearance of secondary ankerite and quartz; (3) moderately metamorphosed iron-formation, marked by the disappearance of phyllosilicates and original carbonates and by the appearance of grunerite; and (4) highly metamorphosed iron-formation, consisting of quartz, amphibole, magnetite, pyroxene, fayalite, and secondary calcite. More recent studies have shown that French's metamorphic zone 4 can be subdivided further. In the Dunka River area, the iron-formation immediately adjacent to the contact with the Duluth Complex is characterized by inverted pigeonite and abundant orthopyroxene (Bonnichsen, 1969b), whereas that from near the zone 3-zone 4 transition lacks these minerals, and hedenbergite is the dominant pyroxene phase (Griffin and Morey, 1969; Morey and others, in press).

From O^{18}/O^{16} ratios of coexisting magnetite and quartz, Perry and Bonnichsen (1966) estimated that metamorphism in the Dunka River area attained a maximum temperature of 700°-750° C. They also estimated that temperatures ranged from 400° to 630° C (their fig. 2, open squares) across 402 feet of iron-formation that was studied from a drill core located approximately 16 miles southwest of the Dunka River area. At this locality, the iron-formation is separated from the Duluth Complex by 500 feet of Virginia Formation, and contains hedenbergite as the sole pyroxene phase. Recent work on fractionation factors by Becker (1971, *op. cit.*) suggests that the reported temperatures are approximately 100° too high (E. C. Perry, Jr., 1971, oral comm.); thus, the temperature range of metamorphism across zone 4 is from less than 400° C to about 650° C. Similarly, French (1968) estimated temperatures of 300°-400° C for moderately metamorphosed taconite two to three miles from the gabbro contact.

Morey and others (in press) have shown that most mineral assemblages in the thermally metamorphosed iron-formation can be attributed to original differences in bulk composition and behavior of oxygen and water. This conclusion agrees with those of French (1968) and Bonnichsen (1969b), who suggested that the metamorphism was largely isochemical and characterized chiefly by a progressive loss of H_2O and CO_2.

Thermal metamorphic effects on the pelitic rocks by the Duluth Complex are less well understood, for the metamorphism was complex and at many places it is difficult to define the contact precisely. Joel Renner (1969, unpub. M.S. thesis, Univ. Minn.) has shown that immediately adjacent to the contact a complex mixture of rock types exists that resulted from: (1) fractional crystallization and filter-pressing of the gabbro melt; (2) partial melting of country rock; and (3) variations in mineralogy and texture of the wall rocks. In general, hornfelses at the contact have a granoblastic or granular texture, and are vaguely layered; individual layers contain cordierite and hypersthene as well as minor biotite and ilmenite or hypersthene, plagioclase, K-feldspar, and biotite. Calcareous beds near the contact have a typical skarn mineralogy, consisting of wollastonite, diopside, tremolite, and grossularite garnet.

At distances of several tens to several hundreds of feet from the contact, most pelitic rocks are locally rich in biotite and some contain cordierite. Originally calcareous beds now contain various amounts and proportions of diopside, grossularite, plagioclase, calcite, and quartz.

ORE DEPOSITS

The iron ores of the Mesabi range are of two types, magnetite-taconite ore and natural ore. The magnetite-taconite ore, occurring in magnetite-rich zones in the Biwabik Iron-formation, can be mined and shipped after crushing and sizing or can be beneficiated using screening and washing or gravity methods to yield a high-iron product. The history, production, grade, and geology of each ore type has been described recently by Marsden and others (1968).

Magnetite-Taconite Ore

Magnetite occurs throughout the unoxidized iron-formation in widely varying amounts. It may be present as disseminations of individual octahedra, aggregates of individual octahedra, or layered clusters formed by interconnecting aggregates of grains. Very fine-grained magnetite, which occurs in both granules and matrix as disseminated and diffuse crystals 5 microns or less in size, is probably primary in origin. However, definite secondary magnetite euhedra, 0.05 to 0.1 mm in diameter, commonly replace earlier iron silicates in granules and also surround and embay fine-grained siderite in the thin slaty beds associated with cherty taconite (LaBerge, 1964; French, 1968). Replacement of granules by magnetite is most common at their margins, yielding an inner core of greenalite or minnesotaite surrounded by a rim of coarser magnetite that generally preserves the outline of the granule; more rarely, an entire granule is pseudomorphously replaced by magnetite. Thus, magnetite appears to have been one of the last minerals formed; in many cases there is little relationship between the primary sedimentary texture and the distribution of the magnetite.

Thick layers from which magnetite can be mined and concentrated using present technological methods occur principally in two areas, the main Mesabi range, between the towns of Nashwauk and Mesaba (now abandoned), and the East Mesabi district, from Mesaba to Dunka River. The distinction between these two areas has both a geological and a geographical basis, for east of Mesaba the Biwabik Iron-formation was modified through metamorphism by the Duluth Complex, which did not affect the remainder of the range. Nevertheless, the amount and size of individual grains of magnetite are everywhere nearly the same. Magnetite from both areas rarely forms aggregates larger than 0.5 mm in diameter, and within the aggregates, individual octahedra commonly range in diameter from 0.03 to 0.07 mm. However, gangue material, consisting mostly of quartz, is more abundant in the aggregates of the main Mesabi; the presence of gangue material results in aggregates that are less compact, and in individual magnetite grains that are slightly smaller. Nevertheless, the recoverable magnetite concentrate from any particular zone in the East Mesabi district is about the same as that from its non-metamorphic equivalent. This may result from the fact that the magnetite and gangue material are complexly intergrown in the more metamorphosed rocks, thus making grinding and liberation of magnetite more difficult.

In the area between Nashwauk and Mesaba on the main Mesabi range, ore bodies of mineable magnetite commonly occur within the cherty members; the slaty members are either too thin or too lean to be mineable. The ore zones are tabular bodies whose vertical extent is defined by a more or less arbitrary magnetite content, and whose lateral extent is limited by superposed oxidation patterns which were controlled largely by structure and to a lesser degree by stratigraphy and mineralogy. The middle part of the lower cherty member comprises the major ore zone, inasmuch as it has the greatest extent and is fairly uniform in magnetite content. In contrast, the upper cherty member is much more variable in lithology; consequently, it has a much less uniform iron content and the magnetite is distributed erratically.

West of Nashwauk, there is a profound lateral change from magnetite- and silicate-bearing to carbonate-bearing iron-formation, and as a consequence only relatively small horses or tongues of unoxidized magnetite-bearing strata exist here. In general, the magnetite ore occurs primarily in the lower cherty member in the southern part of the outcrop belt, where overlying strata protected the iron-formation from the pervasive, near-surface oxidation. In contrast, the lower cherty member is relatively thin in the East Mesabi district, and mineable magnetite deposits are restricted to the upper cherty member and the lower part of the upper slaty member.

Natural Ore

Natural ore bodies occur sporadically throughout all parts of the Mesabi range except the eastern part of the East Mesabi district (fig. IV-10). Approximately 27 percent of the outcrop area of the Biwabik Iron-formation contains natural ore mines, and although this figure does not take into account those areas of the iron-formation that were altered too little to form mineable material, it is a good indicator of the extent of alteration.

The natural ore bodies have a wide variety of shapes and sizes, but in general can be classed into one of three types: trough, fissure, or flat-lying. Trough ore bodies are as much as a mile long, 1,000 feet wide, and 200-400 feet deep. Fissure ore bodies are similar to but smaller than the trough-type. Both types are elongate ore bodies having steep and sharply defined walls. They may occur in parallel swarms. Individual ore bodies may converge, diverge, or intersect, so as to leave numerous rock islands or horses, which represent blocks of unaltered iron-formation within the natural ore bodies. Flat-lying ore bodies have a relatively great horizontal extent as compared with their thickness. In these, alteration appears to have been localized by bedding planes rather than by joints or fractures.

The major controls of ore localization are considered to be structural and lithologic on the main Mesabi and dominantly lithologic on the remainder of the range. Intense fracturing, including both joints and faults, localized in areas of abundant folding, may account for about 80 percent of the ore bodies on the main Mesabi. Most ore bodies east of Gilbert trend N. 40°-50° W., parallel to the strike

Figure IV-10. General map of the Mesabi range showing location, distribution, and shape of natural ore bodies in the Biwabik Iron-formation (modified from U.S. Steel map of 1956).

of a major joint set and perpendicular to the strike of the iron-formation. In this area, most bodies are of the fissure type, and appear to be distributed throughout the iron-formation without regard to the stratigraphy. Between Gilbert and Mountain Iron, as in the main Mesabi, the ore bodies are of the trough and fissure types, and appear to be localized by monoclinal flexures and by fractures of a major joint set that is perpendicular to the strike of the iron-formation. The ore bodies are either of the trough or fissure type and are deep. Some ore bodies extend vertically through the entire iron-formation, but most are confined to the more favorable parts of the iron-formation. Ore deposits between Mountain Iron and a hypothetical line about midway between Keewatin and Nashwauk trend either northwest or west, parallel to a major fracture set. Commonly, several parallel troughs coalesce to form fairly large bodies, particularly near the hanging wall surface. At depth, the ore bodies resemble those between Gilbert and Mountain Iron in containing large horses of unaltered iron-formation. Westward from east of Nashwauk, fissure- or trough-type ore bodies are inconspicuous, and most of the ore bodies are of the relatively shallow, flat-lying type. Here, the ore appears to be controlled by the lithologic character of the iron-formation. All the ore bodies reach or closely approach the present erosion surface, and none is completely covered by the Virginia Formation. Ore zones may extend beneath the Virginia Formation, and where this occurs the overlying rocks are kaolinized and bleached.

There is little doubt that the natural ores are the products of secondary oxidation and leaching of original iron-formation. The original minerals were oxidized to ferric oxides and, virtually at the same time, calcium, magnesium, and much of the silica were removed by leaching. Most commonly, hematite remained unchanged, whereas magnetite altered to martite, minnesotaite and greenalite altered to goethite, siderite altered to hematite or goethite, and stilpnomelane altered to goethite, hematite, and kaolinite.

Two principal hypotheses have been proposed for the source and nature of the oxidizing and leaching solutions: (1) oxidation and leaching resulting from weathering by downward-moving meteoric waters; and (2) oxidation and leaching resulting from upward-moving hydrothermal solutions. Neither field nor experimental and theoretical evidence can be considered as definitely indicating either mechanism. The general consensus, however, is that most of the natural ores on the Mesabi range were developed through normal weathering processes (*cf.* Marsden and others, 1968). All the natural ore bodies are related to an erosion surface, which, in general, is the present bedrock surface. They underlie either glacial drift or a thin veneer of Cretaceous strata. Although the ore extends to various depths, most of it is concentrated fairly near the surface. Thus, it is inferred that surface waters following permeable zones such as faults, joints, or other fractures, acting over a long period of time, could have produced the observed configuration, distribution, texture, and composition of the Mesabi natural ores.

The geologic restraints indicate that the ore formed subsequent to Middle Keweenawan time, when the iron-formation in the East Mesabi district was metamorphosed by the Duluth Complex, and prior to Turonian or Late Cretaceous time when the Coleraine Formation was deposited over a part of the iron-formation. The actual time of ore formation within that long time interval is not well documented, however. Because similar ore deposits in Wisconsin and Michigan are overlain by undisturbed Cambrian sandstones that contain clasts derived from the ore, oxidation and leaching there must have occurred prior to that time. There is no direct evidence on the Mesabi range, however, for a period of weathering at that time, but in east-central Minnesota, a thick regolith has been recognized beneath the Upper Cambrian strata (see Morey, this volume) and a similar regolith has been recognized throughout much of west-central Wisconsin (Ostrom, 1966). Thus, some of the natural ores in Minnesota also by analogy with the ore deposits in Wisconsin and Michigan may have been formed at that time.

However, it seems probable that most of the natural ores in Minnesota formed during Cretaceous time. Parham (1970) has recognized that a thick regolith that formed prior to the early Cenomanian in Late Cretaceous time (Austin, 1970a), extended originally from Manitoba, Canada to at least southern Minnesota. The eastern edge of that regolith in southern Minnesota at present has about the same altitude as the easternmost natural ore deposits on the Mesabi range. That the ore deposits on the Mesabi range formed during Cretaceous time is suggested by the paleomagnetic studies of Symons (1966, p. 1336), who suggested that ". . . the hypothesis of ore genesis which best fits . . ." his data ". . . is that meteoric solutions weathered the primary Animikie iron-formations during the Mesozoic-Cenozoic to form the soft, direct-shipping iron-ore deposits."

If the natural ores of the Mesabi range formed in a manner similar to those of the Cuyuna district, an additional possibility exists. Schmidt (1963, p. 62) has suggested that the Cuyuna ores may have formed in two different ways. The first process involved rising hydrothermal solutions that ". . . stimulated circulation of ordinary ground waters and accelerated to oxidizing and leaching capacity . . ." ultimately resulting in the formation of large, deep, tabular ore bodies, whereas the second stage apparently took place as a result of ordinary weathering and resulted in irregular, blanket-like ore bodies. Peterman (1966) showed that the Cuyuna rocks were metamorphosed about 1,850 m.y. ago and were later affected by ". . . hydrothermal leaching . . ." about 1,540 m.y. ago; the latter event may date the time the first-stage ores were formed. The blanket ores are at about the same elevation as the Cretaceous regolith, and Symons (1966) has shown that the paleomagnetic pole positions of some of these ores are similar to those of the Mesabi ores; thus, the blanket ores of the Cuyuna district may have formed in Cretaceous time. By analogy with the Cuyuna ores, the fissure- and trough-type deposits of the Mesabi range may have formed during one interval, whereas the blanket-type deposits formed at another time.

GUNFLINT RANGE

G. B. Morey

The Gunflint range is more or less continuously exposed from near Gunflint Lake on the International boundary to Thunder Bay on Lake Superior, a distance of approximately 100 miles (fig. IV-11). Isolated exposures as far east as Schreiber, Ontario, on the north shore of Lake Superior, indicate that the Middle Precambrian rocks once extended at least an additional 70 miles to the east. The range extends westward into Minnesota for nearly 12 miles, where it is truncated by the Duluth Complex. Diabase dikes and sills of Late Precambrian age are abundant in the Middle Precambrian rocks, particularly in the southwestern part of the range. The sills, as much as 1,100 feet thick (Morey, 1963), are generally conformable to the sedimentary bedding, but transect it in detail.

The rocks exposed on the Gunflint range are the northeastern extension of those on the Mesabi range. The two ranges are separated for a distance of approximately 40 miles by the Duluth Complex, which has cut out the intervening Middle Precambrian strata. The Gunflint Iron-formation is underlain by the Kakabeka Quartzite and overlain by the Rove Formation, an interbedded argillite and graywacke succession. The basal quartzite is thin and locally absent; where present, it is commonly included as a basal member of the iron-formation (Goodwin, 1956). The Gunflint Iron-formation and the Rove Formation are structurally conformable and commonly have a gradational contact; as on the Mesabi range, the top of a limestone-bearing member is considered the top of the iron-formation.

DESCRIPTIVE STRATIGRAPHY

Gunflint Iron-formation

Originally referred to in Minnesota as the "lower taconite member of the Animikie Formation" (Grant *in* Winchell and others, 1899), these iron-rich sediments were renamed by Clements (1903) the "Gunflint Iron-formation," for exposures along the north shore of Gunflint Lake. Broderick (1920) described the iron-formation in Minnesota, and applied Wolff's (1917) four-fold subdivision of the Biwabik Iron-formation to the Gunflint Iron-formation. Later, Gill (1927) and Tanton (1931) extended this work to the vicinity of Thunder Bay. More recently, the formation was remapped in Canada by Moorhouse (1960) and Goodwin (1960) and in Minnesota by Morey and others (1969, Minn. Geol. Survey, open file map, Long Island Lake 7.5-minute quadrangle).

The original character of the Gunflint Iron-formation in Minnesota is obscured by a thermal metamorphic overprint produced by Upper Precambrian igneous rocks. However, unmetamorphosed iron-formation can be observed nearby in Canada, and has been described by Goodwin (1956).

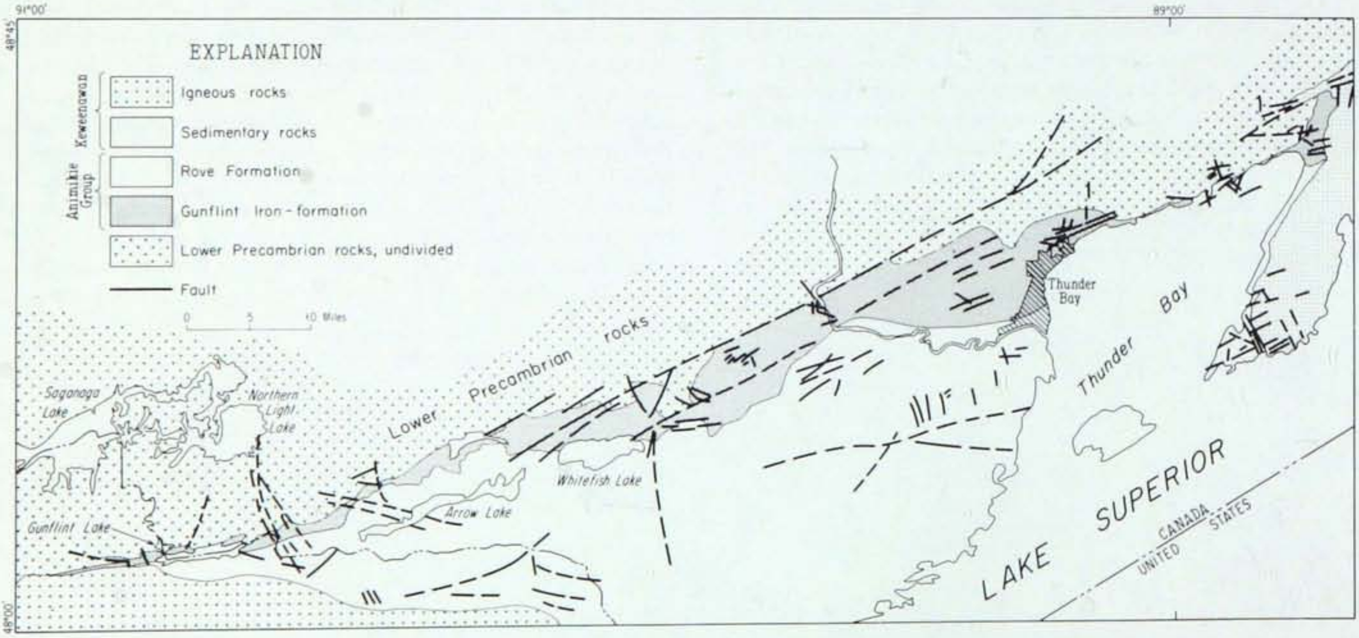


Figure IV-11. Generalized geologic map of the Gunflint range in the Thunder Bay district of Ontario and adjoining Cook County, Minnesota (modified from Goodwin, 1956).

Goodwin (1956, p. 568) divided the iron-formation into six sedimentary facies, each of which ". . . is an areally restricted unit with unique lithic characteristics. . . ." Two of the facies, a basal conglomerate and a topmost limestone, are lithologically distinct. Each of the other four facies occurs twice in cyclical repetition, and accordingly the iron-formation is divided into four members: basal conglomerate member; lower and upper Gunflint members; and upper limestone member (fig. IV-12). The basal conglomerate member, or the Kakabeka Quartzite of Tanton (1931), is equivalent to the Pokegama Quartzite on the Mesabi range. It is as much as 10 feet thick, and is a polymictic conglomerate containing clastic or Lower Precambrian metavolcanic rocks and granite in a matrix of quartz, feldspar, and lesser chlorite. In much of the western part of the Gunflint range, the unit is missing and the iron-formation directly overlies older rocks.

A lower algal chert facies consists of reef-like mounds of finely laminated black, red, and white chert. Individual mounds are either isolated or connected by thin layers of granular or oolitic chert. Many of the mounds are stromatolitic (Hofman, 1969), and the associated cherts contain an abundant microflora (Barghoorn and Tyler, 1965; Cloud, 1965). A lower tuffaceous shale facies overlies the lower algal chert facies, and is composed of fissile black shale containing volcanically derived material. Three facies comprise the upper part of the lower Gunflint member. The lower west taconite facies is composed of wavy-bedded granule-bearing chert, carbonate, and sparse iron oxides; greenalite occurs as granules; siderite forms thin beds, and the proportion of magnetite and hematite, both as disseminated grains and thin beds, increases upward. This facies grades laterally eastward into the lower bedded chert-carbonate facies, consisting of 4- to 6-inch-thick siderite-rich slaty beds intercalated with 2- to 6-inch-thick beds of gray chert. Carbonaceous material and pyrite are common in the shaly layers. This facies in turn grades northeastward into the lower east granular taconite facies. The lower 2 to 6 feet of this facies is composed of interbedded granular chert and ankerite. The upper 10 to 20 feet consists of interbedded red or green mottled chert and dolomitic limestone.

The base of the upper Gunflint member is marked by a granular cherty layer that is overlain by algal-bearing chert and, locally, by amygdaloidal basalt flows. The latter two units are overlain by granular chert and bedded jasper. The jasper beds grade upward into a tuffaceous shale facies consisting of ". . . black, tuffaceous shale and siltstone with considerable interbedded siderite and pyrite, together with extensive beds of volcanic ash" (Goodwin, 1956, p. 579). The ash contains ellipsoidal structures that resemble mudballs, and which are composed of concentric layers of small angular tuff fragments arranged about larger central clasts. The upper tuffaceous shale facies grades into an upper taconite facies and an upper bedded chert-carbonate facies. The upper taconite facies is composed of wavy beds of granular greenalite-bearing chert. This facies contains abundant hematite and magnetite in granules toward the top, and grades laterally eastward into the upper bedded chert-carbonate facies. The latter facies consists of intercalated gray chert and brown carbonate, consisting of siderite with lesser amounts of dolomite and/or ankerite. Brecciation and crumpling, apparently contemporaneous with deposition, are common in this facies.

The upper limestone member marks the top of the Gunflint Iron-formation. Minor chert beds, illite, and volcanic shards are present, and tuffaceous shale is common, especially in the eastern part of the outcrop area.

The Gunflint Iron-formation is approximately 300 feet thick in Minnesota, whereas it attains a maximum thickness of approximately 400 feet in Canada (Goodwin, 1956). The boundaries of Goodwin's members do not coincide with the boundaries of the four-fold classification scheme used in Minnesota; fortunately, however, the two schemes can be generally equated (fig. IV-12).

Because of the metamorphic overprint in Minnesota, the carbonates and greenalite are replaced by amphiboles, pyroxene, fayalite and, locally, by garnet and other silicates. In addition, many small-scale, pre-metamorphic, sedimentary textural features have been partially destroyed; however, the larger structures and complex bedding relationships are preserved.

The lower cherty member is thin in Minnesota, ranging

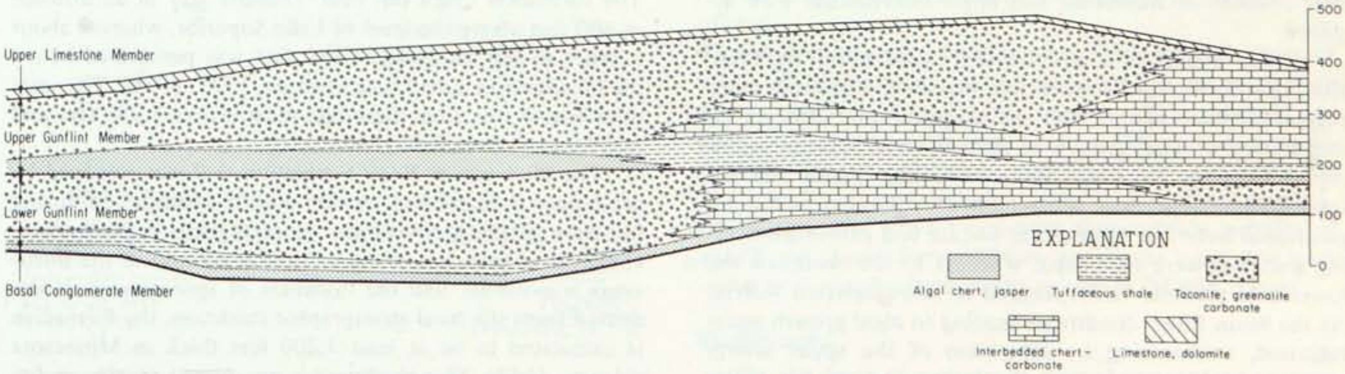


Figure IV-12. Generalized east-west cross-section showing inferred facies distribution within the Gunflint Iron-formation (modified from Goodwin, 1956).

in thickness from 15 to 45 feet. Feldspathic quartzite containing rounded pebbles and cobbles of granite is present locally at the base. A persistent magnetite-rich, silicate-bearing unit, 5 to 15 feet thick, occurs within this member and serves as an excellent marker bed. Most commonly, it lies on either the feldspathic quartzite or stromatolitic mounds separated by chert-cemented conglomerate containing fragments of algal structures. In turn, it is overlain by a massive, chert- and silicate-rich, magnetite-poor unit about 15 feet thick.

The lower slaty member generally is 80 to 95 feet thick. The lowermost 25 feet is a black, thin-bedded, volcanically derived argillite that generally lacks magnetite, and is equivalent to the intermediate slate on the Mesabi range and to the lower tuffaceous shale facies in Canada. The beds immediately above the argillite are massive and cherty, and resemble the upper part of the lower cherty member. This unit, however, passes abruptly upward into a unit consisting of cherty silicate-bearing beds, containing sparse magnetite, that are intercalated with a few thinly laminated beds composed almost entirely of silicates. The remaining 50 feet of the member is a thin-bedded or laminated unit containing various silicates and 20 to 35 percent magnetite. A few subordinate cherty silicate beds occur in this unit.

A complete gradation exists between the lower slaty, upper cherty, and upper slaty members. As presently defined, the upper cherty member is approximately 50 feet thick. The lower part of the member consists of irregularly bedded to lenticular chert layers intercalated with thinly laminated silicate- and magnetite-rich beds that decrease in abundance upward. Thin irregular layers of magnetite are common near the bottom of the member, but become less abundant upward. The middle of the member—equivalent to Goodwin's upper algal chert facies—is characterized by several granular chert beds containing algal structures, conglomerate fragments, and abundant magnetite. Thick lenticular chert beds with disseminated magnetite, separated by thin layers of a laminated silicate-rich rock, characterize the upper few tens of feet of the upper cherty member.

The upper slaty member is approximately 150 feet thick, and consists of a thin-bedded or laminated quartz-silicate rock intercalated with thinly laminated layers of graphitic iron-rich argillite, and 1- to 2-inch-thick beds of relatively pure chert. The upper 10-20 feet lacks magnetite, and consists of limestone and chert interbedded with argillite.

The arrangement of sedimentary facies within the Gunflint Iron-formation is much like that in the Biwabik Iron-formation. Goodwin (1956) has suggested that after an initial period of algal growth in shallow water, volcanism, as represented by the lower tuffaceous shale facies and the intermediate slate, was accompanied by sinking of the depositional basin. In Minnesota, silicate and carbonate-bearing material were deposited, whereas to the northeast the lower east taconite facies formed in more agitated waters. As the basin filled, conditions leading to algal growth again returned, culminating in deposition of the upper cherty member in Minnesota and the lower part of the upper Gunflint member in Canada.

Volcanic activity, marked by local basalt flows, terminated deposition of the upper algal chert facies and resulted in the widespread distribution of tuffaceous material. Accompanying downwarping resulted again in the deposition of granular iron silicate-bearing rocks in Minnesota, whereas in the shallow waters to the northeast, chert-carbonate rocks were deposited. As the basin filled, sporadic but violent volcanic activity, accompanied by downwarping, led to the deposition of the upper limestone member, and continued sinking set the stage for deposition of the Rove Formation.

Rove Formation

General Statement

The Rove Formation constitutes the upper part of the Animikie Group in northern Cook County, and is equivalent to the Virginia Formation of the Mesabi range. The name "Rove Slate" was first applied by Clements (1903, p. 390) to "slate" and quartzitic graywacke exposed on the south shore of Rove Lake in northern Cook County, Minnesota. Because much of the formation is argillite and graywacke rather than slate, the name was changed to "Rove Formation" by Grout and Schwartz (1933, p. 5). The formation crops out in an extensive area bordered on the north by the slightly older Gunflint Iron-formation and on the south by much younger rocks of Keweenawan age. To the east, the formation is covered by Lake Superior.

Geologic relationships are complicated by many tabular, concordant igneous bodies which commonly merge along strike to form several large bodies. Because such branching results in isolated "islands" of Rove between igneous masses, it is difficult to trace any stratigraphic marker bed in the Rove Formation over any great distance.

The thickness of the formation cannot be determined accurately because of poor, discontinuous exposures, local faults, and the presence of intrusive igneous rocks that have inflated the stratigraphic section. Eight hundred feet of Rove Formation was described by Tanton (1931) above Lake Superior on Sibley Peninsula, Ontario, but only 20 feet is present 24 miles to the north. The thinning toward the north is attributed to erosion prior to an overstep of Upper Precambrian sedimentary rocks (Moorhouse, 1960). The formation crops out near Thunder Bay at an altitude of 800 feet above the level of Lake Superior, whereas about 3 miles to the southeast, 1,280 feet was penetrated below the lake level in a drill hole (Tanton, 1931, p. 36). The total thickness in this area, therefore, must exceed 1,280 feet and may be more than 2,000 feet.

The thickness of Rove Formation that was removed by pre-Upper Precambrian erosion and by assimilation during intrusion of the Keweenawan igneous rocks is not known. However, if inflation of the section as a result of the intrusions is assumed, and the thickness of igneous rock is deducted from the total stratigraphic thickness, the formation is calculated to be at least 3,200 feet thick in Minnesota (Morey, 1969). This thickness is considered maximum because undetected igneous sills probably are present within the formation.

Lithology

Argillite comprises about 90 percent of the lower 500 feet of the formation and from 30 to 50 percent of the remainder. Repeated alternations of argillite, siltstone, and graywacke (40-80 percent) and minor quartzite (2-5 percent) characterize the upper part of the formation. Variegated argillite, limestone, dolomite, chert, conglomerate, and altered tuffaceous sediments make up a small percentage of the formation.

The mineralogic composition of representative samples is shown in Figure IV-13. The major framework minerals in the sandstones are quartz and sodic plagioclase. Rock fragments, which rarely exceed 1 mm in diameter, generally make up less than 5 percent but constitute as much as 10 percent of the framework grains. Regardless of composition, the larger grains typically are more rounded and fresher than the smaller ones; the shape of the smaller grains has been modified by recrystallization. The rock fragments are composed mostly of recrystallized chert, fine-grained quartzite, schist and, rarely, greenstone. Quartz rarely exceeds 2 mm in diameter and comprises 20 to 70 percent of the samples studied. Inclusions in individual grains are common and generally too small to be identified. Feldspar, ranging from 5 to 35 percent, is the second most abundant constituent of the framework grains. The grains are angular to subangular, and have an average diameter less than that of the accompanying quartz. Dusty and highly sericitized sodic plagioclase (An_2-An_{30}) is the most common feldspar. Both gridiron-twinned microcline and orthoclase are rare.

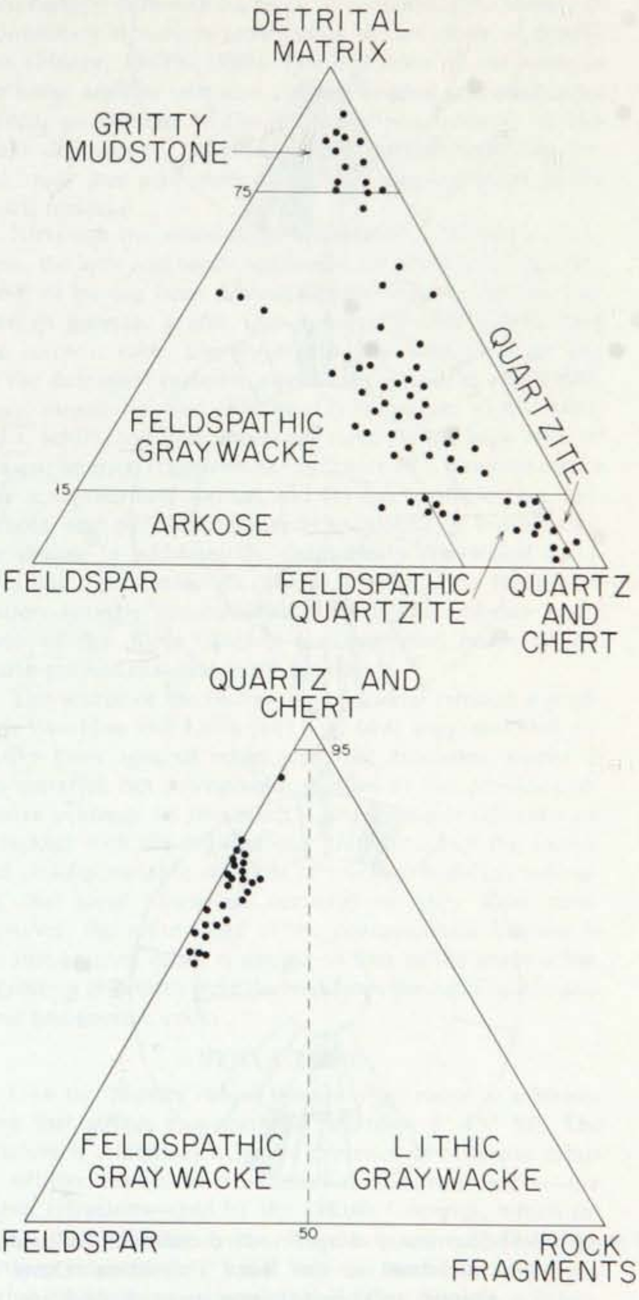


Figure IV-13. Summary of mineralogic composition of rocks in the Rove Formation (after Morey, 1969).

The matrix material consists of muscovite, chlorite (14 Å), septechlorite (7 Å), plagioclase, and quartz. Pyrite and carbonaceous material are present in all samples, and calcite or dolomite occurs in several. Although some samples contain $1M_d$ and 1M muscovite, the 2M polytype is most common. Both chlorite and muscovite occur as lacy, radiating patches and as fibrous intergrowths that replace edges of framework grains, implying that at least some of the layered silicates formed as the result of recrystallization.

The siltstones are fine-grained equivalents of the graywacke sandstones, and differ only in having more quartz and less feldspar and in lacking rock fragments. Silt-size grains are typically angular, and are concentrated in thin laminae that alternate with mica-rich laminae similar to those in most argillite. The samples that have been studied contain 5-50 percent quartz and plagioclase; the remainder is clay-size sericite, chlorite, quartz, opaque carbonaceous material, and euhedral pyrite. Most samples contain sparse calcite, but several contain more than 10 percent.

Several types of quartzitic sandstone also are present. Feldspathic quartzites are mineralogically identical to the graywackes but have less matrix material; they occur at the tops of some graywacke beds, and are interpreted as lag deposits from which the clay-size material has been winnowed out. Quartzites from the upper 700 feet of the formation contain small amounts of feldspar, muscovite, and calcite.

Intraformational conglomerates occur near the base of many sandstone beds; most of the clasts are not more than a few inches in diameter, but Tanton (1931) reported angular, disc-shaped clasts as much as 3 feet in maximum dimension that are only slightly removed from their original position. Grout and Schwartz (1933) described conglomerates from the Rove Formation that contain pebbles of quartz and granite derived from outside the basin, but such rocks were not observed during my study (Morey, 1969).

Stratigraphy

On the basis of a study of several incomplete sections (Morey, 1969), three lithostratigraphic units in the Rove Formation are tentatively distinguished. These are, from bottom to top, lower argillite, transition sequence, and thin-bedded graywacke.

The lower argillite unit is 400 to 500 feet thick and consists dominantly of three principal rock types, each of which has characteristic bedding features. The rock types are: (1) thin- to thick-bedded light-gray argillaceous siltstone, the upper parts of some of which are graded and have cross-laminae; (2) mostly very thin-bedded dark-gray, silty argillite; and (3) black, fissile, carbonaceous argillite having laminae generally less than half an inch thick. These three rock types generally occur in sequence from a basal siltstone bed grading upward to a silty argillite and finally to a carbonaceous argillite. The lower part of the lower argillite unit is composed almost completely of alternating beds of dark-gray silty argillite and black carbonaceous argillite. Silt-size material is, however, more abundant in stratigraphically higher beds of the unit. Lenses and irregular beds of limestone and dolomite occur near the base of the unit as do calcite and dolomite concretions of various sizes and shapes. These concretions have been described elsewhere by Tanton (1931) and Moorhouse (1963).

As the name implies, the transition sequence separates dominantly argillaceous rocks below from dominantly arenaceous rocks above. The bottom of the sequence is arbitrarily defined as that stratigraphic position where sandstone comprises a significant proportion of the section, and the top is marked by a relative change from argillite to argillaceous siltstone. Because of the arbitrary boundaries, the thickness differs from place to place but is as much as 100 feet.

Argillite layers ranging in thickness from less than 2 inches to 10 feet are intercalated with the sandstones. Although having a higher content of silt-size material and being lighter in color, these argillites are similar in appearance to those in the lower argillite. Intercalated sandstone or siltstone beds range in thickness from less than 6 inches to 1 foot. The texture of each sandstone bed is much coarser grained than that of the associated argillite. Most of the sandstone beds are dark gray or medium dark gray, poorly sorted, and composed of angular grains of quartz and feldspar in a matrix of muscovite and chlorite. Many of the beds are graded. Thin lag deposits of feldspathic quartzite gradationally overlie some of the graywacke beds. The quartzites are variable in thickness, and the top of each bed is irregular and may be ripple-marked; the change from quartzite to an overlying argillite bed always is abrupt.

The thin-bedded graywacke unit is as much as 2,700 feet thick, and consists of light- to dark-gray, fine-grained graywacke and intercalated argillaceous siltstone. Individual sandstone beds range in thickness from less than 1 to more than 3 feet; some graywacke beds, however, are more than 20 feet thick. In general, the proportion of graywacke, siltstone, and argillite varies without regard to stratigraphic position. Some parts of the member are dominated by argillite and argillaceous siltstone, whereas other parts consist almost entirely of interbedded graywacke and argillaceous siltstone. Thus, a definite stratigraphic succession within the member has not yet been determined. However, the upper 700 feet of the unit is distinguished by the presence of quartzite, which is feldspathic, generally gray, white, or pinkish-gray, and occurs in thin to thick beds having sharp soles and tops. A large number of the beds are structureless, but many are graded. Many of the original clastic grains in the quartzites were angular and others were rounded, but all have been modified by secondary overgrowths and by recrystallization and now form a sutured, interlocking mosaic texture.

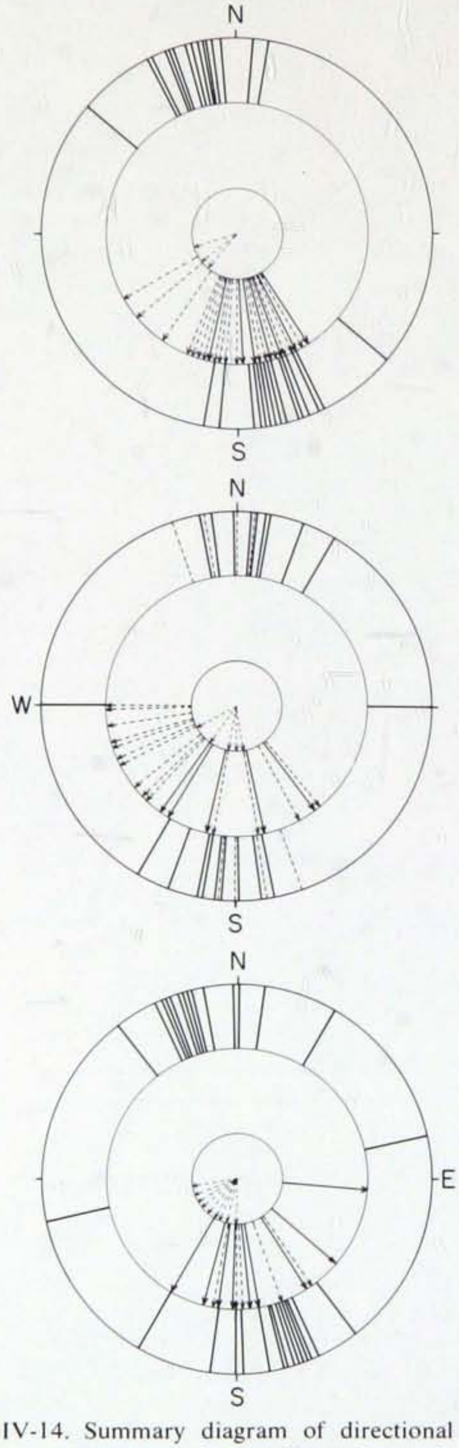


Figure IV-14. Summary diagram of directional features measured in the Rove Formation. Outer circle: solid line, groove casts; dashed line, flame casts; middle circle, solid line, flute casts; dashed line, cross-bedding; inner circle, ripple marks.

Petrology

Early Rove time was characterized by the deposition of black muds that accumulated in quiet water. The depth of water need not have been very great, as suggested by the black muds that are being deposited today in shallow-water basins such as the Baltic Sea. However, the Rove basin must have subsided slowly—at a rate such that filling failed to keep pace with downwarping. As the water gradually deepened and a gentle bottom slope was developed, much silt-size material was introduced into the basin. Accumulation of the very fine-grained sediments of the lower argillite unit was interrupted periodically by deposition of thin silt and sand beds by turbidity currents, as indicated by a variety of sedimentary structures attributable to this mode of deposition (Morey, 1967b, 1969). The transition of the rocks in the lower argillite unit into the thin-bedded graywacke unit records an increase in the frequency and intensity of turbidite deposition. Dispersal patterns of sediments (fig. IV-14) imply that a northern land mass supplied much of the clastic material.

Although the mineralogy of the Rove Formation is diverse, the silty and sandy sediments are interpreted (Morey, 1969) as having been derived mainly from a plutonic terrane of granite, gneiss, and metamorphosed sedimentary and volcanic rock. The reasons for this interpretation are: (1) the dominant material assemblage is quartz, alkali feldspars, muscovite, and chlorite; (2) fragments of quartzite, chert, schist, and greenstone are rare; (3) the high ratio of feldspar to rock fragments is indicative of a plutonic rather than a supracrustal source; and (4) calcic plagioclase, amphibole, and pyroxene, minerals indicative of basic rocks, are absent. In addition, the sedimentary directional structures and heavy minerals (Morey, 1965) in the Rove Formation strongly indicate that Lower Precambrian rocks north of the Rove outcrop area supplied much of the coarse-grained material to the formation.

The source of the fine-grained material remains a problem. Van Hise and Leith (1911, p. 614) suggested that extrusive basic igneous rocks were the dominant source of this material, but petrographic studies do not provide conclusive evidence on this point. Limestone and dolomite interbedded with the argillite near the bottom of the formation contain variable amounts of volcanic material, indicating that some volcanism occurred in early Rove time. However, the mineralogy of the coarse clastic fraction in the fine-grained rocks is similar to that in the graywackes, suggesting that both were derived from the same acidic plutonic and gneissic rocks.

STRUCTURE

Like the Mesabi range, the Gunflint range is a homocline that strikes east-northeast and dips 5°-15° SE. The structure is complicated by the presence of extensive tabular sill-like bodies of Keweenawan gabbroic rocks—the Logan intrusions—and by the Duluth Complex, which intrudes the Animikie rocks. This homocline defines the northern edge of the "Lake Superior basin," a structural feature of Keweenawan age. However, in Canada, a northward overstep of the Upper Precambrian Sibley Series over progressively older Middle Precambrian rocks indicates that some southward tilting occurred prior to deposition of the Upper Precambrian rocks (Moorhouse, 1960).

Except near the igneous bodies, folding is relatively inconsequential. Mapping of the Gunflint Iron-formation (see Sims and others, 1969) southwest of Gunflint Lake has disclosed several open asymmetric folds that plunge gently to the east-southeast. Inasmuch as the Logan intrusions also are deformed, this deformation must have taken place in Keweenawan time. Both structures, however, have axial planes that project into Lower Precambrian structures, suggesting that they formed in response to movements along older structures.

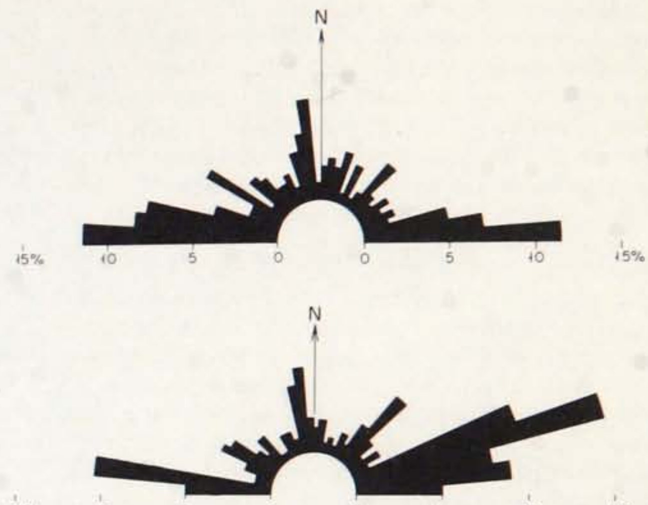


Figure IV-15. Rose diagrams showing strike of faults and joints in the Animikie rocks of the Gunflint range. A, faults, modified from data of Goodwin (1960) and Moorhouse (1960). B, joints in the Animikie rocks in northwestern Cook County.

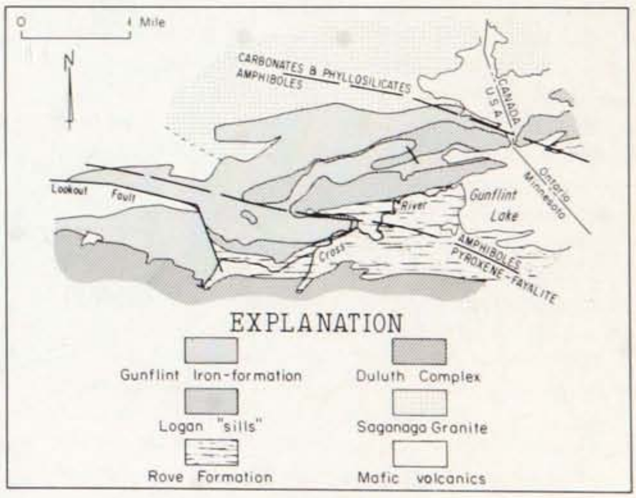


Figure IV-16. Generalized geologic map of the Gunflint Iron-formation in Minnesota showing the metamorphic mineral zones and their relation to the Duluth Complex (geology by Morey and Papike, 1967-68).

A well-developed fracture system consisting of both steeply-inclined faults and joints is prevalent throughout the area. Gravity faults are particularly common, and dominate the gross structure (fig. IV-11). Some of these faults have vertical movements of as much as about 300 feet (Tanton, 1931), but most displacements range from 20 to 100 feet. The faults generally trend N. 70° E. parallel to the strike of the Animikie strata, north-northwest, and N. 80°-90° E. (fig. IV-15). Joint sets generally have the same trend (fig. IV-16), but a subordinate set trends about N. 60° E., and sheeting joints in the igneous rocks strike more or less parallel to igneous rock contacts.

Inasmuch as the joints and faults affect or displace rocks of Keweenawan age they are inferred to have formed in response to stresses related to the development of the "Lake Superior basin." However, their geometry and distribution appear to have been controlled in substantial part by structures that originally formed in Early Precambrian time.

METAMORPHISM

Where unaffected by Keweenawan igneous activity, the Gunflint Iron-formation appears to be the least metamorphosed of the three major iron-formations in Minnesota. Much of the silica is chalcedonic, and greenalite, commonly believed to be a primary mineral, is the major silicate phase. Minnesotaite is rare and stilpnomelane occurs only where the iron-formation has been metamorphosed by Keweenawan intrusions (Goodwin, 1956). Mineral assemblages in the least altered facies of the overlying Rove Formation include: (1) in quartzitic and quartzofeldspathic rocks, (A) muscovite-chlorite-sodic plagioclase-quartz, and (B) muscovite-chlorite-potassium feldspar-quartz; (2) in calcareous rocks, (A) calcite-quartz-feldspar-muscovite-chlorite, and (B) dolomite-quartz-plagioclase-muscovite-chlorite, and (C) calcite-dolomite-quartz-plagioclase-muscovite-chlorite. The pelitic rocks are only partially reconstituted, inasmuch as much of the feldspar is still cloudy and turbid, and only rarely are the feldspar grains recrystallized. Thus, although the mineral assemblages are indicative of the greenschist facies grade of metamorphism, the lack of complete textural re-equilibrium suggests that they recrystallized at an even lower metamorphic grade. This is in agreement with the apparently unmetamorphosed Gunflint Iron-formation.

Unmetamorphosed iron-formation similar to that described by Goodwin (1956) is found in Minnesota only between Gunflint Lake and North Lake (plate 1). West of Gunflint Lake, three metamorphic zones have been distinguished by changes in mineralogy along the strike of the formation toward the cross-cutting Duluth Complex; the mineral assemblages are similar to those described from the Biwabik Iron-formation on the Mesabi range (French, 1968; Bonnichsen, 1969b; and Morey and others, in press). The three metamorphic zones that can be delineated are shown on Figure IV-16. Zone 1, or slightly metamorphosed iron-formation, occurs immediately west of Gunflint Lake. It contains finely divided chert, iron carbonates, greenalite, minnesotaite, and stilpnomelane. Finely divided hematite occurs at the east end of the zone, but disappears midway in it. Disseminated and interlocking grains of magnetite are abundant, especially as rims around granules. This part of the iron-formation is much like that described by French (1968) in "unmetamorphosed" Biwabik Iron-formation. Zone 2, or moderately metamorphosed iron-formation, is about 1.2 miles wide and extends to within 0.3 miles of the Duluth Complex. Grunerite-cummingtonite, hornblende, and actinolite, as well as quartz and magnetite, characterize this zone. As in zone 1, much of the magnetite is 0.002 to 0.02 mm in diameter, a size range similar to that observed in other Lake Superior iron-formations. Small-scale premetamorphic sedimentary structures such as granules and oolites are partly destroyed, but larger scale primary structures and bedding features are relatively unaffected. Zone 3, or highly metamorphosed iron-formation, occurs adjacent to the Duluth Complex and is characterized by a wholly metamorphic fabric. The rock is composed chiefly of quartz, magnetite, iron-rich pyroxenes, and fayalite. Very commonly, euhedral or subhedral grains of magnetite that are virtually the same size as the grains in lower grade rocks are poikilitically enclosed within large silicate grains. However, a significant part of the magnetite is extensively recrystallized and coarsened, and grains as much as a millimeter in diameter are concentrated along bedding planes. Actinolite is common in magnetite-rich layers, and both prograde and retrograde cummingtonite is abundant. In general, this zone is very similar to that described in the Dunka River area on the Mesabi range by Bonnichsen (1969b).

Dikes and sills assigned to the Logan intrusions also have metamorphosed the Gunflint and Rove Formations. The widths of the metamorphic aureoles are related directly to thicknesses of the sills, and range from less than a foot to more than 30 feet. In the Gunflint Iron-formation it is difficult to distinguish unique metamorphic assemblages adjacent to sills. In zone 2, for example, mineral assemblages characteristic of zone 3 occur in aureoles around the sills, and in zone 1, grunerite, which is characteristic of zone 2 metamorphism, is found adjacent to the sills.

In the Rove Formation, assemblages that can be assigned to the pyroxene-hornfels facies are found adjacent to thick sills and to the Duluth Complex, whereas assemblages characteristic of the hornblende-hornfels facies occur adjacent to thinner sills. Locally, andalusite (Grout and Schwartz, 1933; Morey, 1969) and chloritoid (Grant, 1970) have been identified.

ECONOMIC GEOLOGY

The Gunflint range in Minnesota resembles the East Mesabi district in having been intensely metamorphosed and in lacking appreciable secondary oxidation and leaching. Nevertheless, the range was prospected extensively at an early date for possible natural ore bodies. Numerous test pits were sunk unsuccessfully prior to 1892 (Grant *in* Winchell and others, 1899). About 1892, when a railroad was extended into the area from Thunder Bay, several shafts from 75 to 105 feet deep were sunk and some ore was removed. This ore consisted of layers of magnetite-rich iron-formation which had been hand-sorted and shipped. The mines were abandoned shortly thereafter at the time the rich Mesabi ores became available.

Interest in the area as a possible source of magnetite-taconite ore has occurred sporadically since about the turn of the century. For several reasons, however, no commercial venture has been undertaken. Among the major reasons are the relative difficulty of mining in an inaccessible area, the presence of large sill-like bodies of diabasic gabbro in the iron-formation, the fine grain size of the magnetite, and the apparently high concentration ratio of about 3.3:1 (Grout and others, 1959, p. 79). In addition, the amount of magnetite-rich ore that is readily available was believed to be limited because of the small dimensions of the outcropping magnetite-rich layers and their relatively steep dip.

Recent mapping of the Gunflint Iron-formation in the Long Island Lake 7.5-minute quadrangle (Morey and others, 1969, open-file map, Minn. Geol. Survey) suggests, however, that the structural configuration of the Gunflint range in Minnesota in many respects is analogous to that of the Dunka River area at the east end of the Mesabi range. Throughout the western part of the Gunflint range the iron-formation dips 20° to 60° S., and consequently forms a narrow outcrop belt. The relatively steep dips at the outcrop led Broderick (1920) and later workers to conclude that the iron-formation is conformable to and dips relatively steeply beneath the Duluth Complex, thus apparently limiting the recoverable magnetite-taconite ore to a small wedge-shaped body (Broderick, 1920, fig. 53, Section B-B'). Several lines of evidence indicated by the geologic mapping (see Sims and others, 1969) suggest, however, that the dip of the base of the Duluth Complex probably is less than 20° S. along much of the outcrop; thus, the amount of stripping necessary to expose additional iron-formation probably would not be commercially prohibitive. In addition, the Logan intrusions in this segment of the iron-formation are thin and probably would not constitute a significant mining problem.

The distribution and abundance of magnetite clearly is the most important factor in evaluating the economic potential of the Gunflint Iron-formation. Such an evaluation can be done properly only by metallurgical testing of carefully obtained diamond drill cores, which as yet has not been done. Some qualitative empirical conclusions based on outcrop observations can be made, however. It is worth noting that persistent magnetite-rich beds are abundant in certain parts of the iron-formation, but are rather thin and do not constitute ore zones in themselves. Additionally, magnetite disseminated in the form of individual euhedra, granules, and irregular beds is sparsely or abundantly present in the upper part of the lower slaty and lower part of the upper cherty members. Much of this magnetite is between 0.002 and 0.02 mm in diameter, a size range comparable to that noted by LaBerge (1964) in various Lake Superior iron-formations, by Gruner (1946) for unmetamorphosed parts of the Biwabik Iron-formation, and by Gundersen and Schwartz (1962) for the East Mesabi district. However, a significant portion of the magnetite in the iron-formation immediately adjacent to the Duluth Complex is extensively recrystallized, and interlocking grains as much as a millimeter in diameter are common along bedding planes. This coarsening of the grain size was noted previously by Broderick (1920), who suggested that the Duluth Complex may have contributed some material to the iron-formation during metamorphism. This possibility cannot be evaluated fully without a careful bed-by-bed comparison of metamorphosed and unmetamorphosed iron-formation, but it is interesting to note that the magnetite distribution is similar to that in the area now being mined at Dunka River (Bonnichsen, 1969b), at the eastern extremity of the Mesabi range. Accordingly, it is suggested that the Gunflint Iron-formation immediately adjacent to the Duluth Complex is worthy of careful consideration as a potential source of modest quantities of low-grade magnetite-taconite ore.