DULUTH COMPLEX, HISTORY AND NOMENCLATURE

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The Upper Precambrian rocks in northeastern Minnesota include about 2,500 square miles of anorthositic, troctolitic, gabbroic, granodioritic, and granitic intrusive rocks assigned to the Duluth Complex. These rocks crop out in an arcuate pattern extending from Duluth northeastward nearly to the northeastern tip of Minnesota, a distance of about 150 miles (fig. V-23). Except at Duluth, where it is underlain and overlain by flows of the Upper Precambrian North Shore Volcanic Group, the complex was intruded along an unconformity between the overlying volcanics and underlying older rocks of Early and Middle Precambrian ages. Radiometric dating of zircons from several rhyolitic flows of the adjacent North Shore Volcanic Group and from granitic to intermediate fractions of the Duluth Complex indicates that the intrusive rocks and flows are nearly contemporaneous and are $1{,}120 \pm 15$ m.y. old (Silver and Green, 1963; oral comm., 1970).

The earliest work on the Duluth Complex was done between 1880 and 1900 (see Taylor, 1964, p. 1-2 for complete list of references). During this period, the general rock types were described and their areal extents were established. Interpretations of the complex ranged from a reservoir for Keweenawan flows, through a great basal flow, to a laccolith. Later, on the basis of detailed studies around Duluth, Grout (1918a) concluded that the "Duluth gabbro" was a large differentiated intrusion occurring along the unconformity at the base of the Keweenawan. For this large basin-like structure he proposed the term "lopolith." In later work, Grout (1918b, 1918c, 1920) visualized the complex as having consisted of one large mass of magma (not necessarily homogeneous or resulting from one injection) which, through convective movements, developed into a layered and differentiated rock sequence. In describing the geology of Cook County, Grout and others (1959) included reconnaissance maps of several townships in the complex; but inasmuch as detailed studies were not attempted, the discussion was largely descriptive and did not attempt to explain the origin of the mass as a whole. Chemical analyses along two traverses across the complex caused Snyder (1959) to suggest the possibility of multiple intrusions. In the first integrated mapping, petrographic, and geochemical study of the complex (in the Duluth area), Taylor (1964) concluded that the complex consists of a series of multiple intrusions. In this area, the oldest unit, a coarse-grained anorthositic gabbro, comprises the upper part of the complex. The next younger unit, as indicated by crosscutting relations, comprises the lower two-thirds of the complex, and is a layered series of troctolite and gabbro, which itself is a multiple intrusion. Both these units are cut by intrusive ferrogranodiorite and granophyre as well as by late-stage basalt and aplite dikes. My work (Phinney, 1969) in Lake County has led me to the same conclusion and the same sequence determined by Taylor. Also, on the basis of new chemical data, I (Phinney, 1970) revived the 90-year-old idea that the complex is a series of chambers for Keweenawan flows.

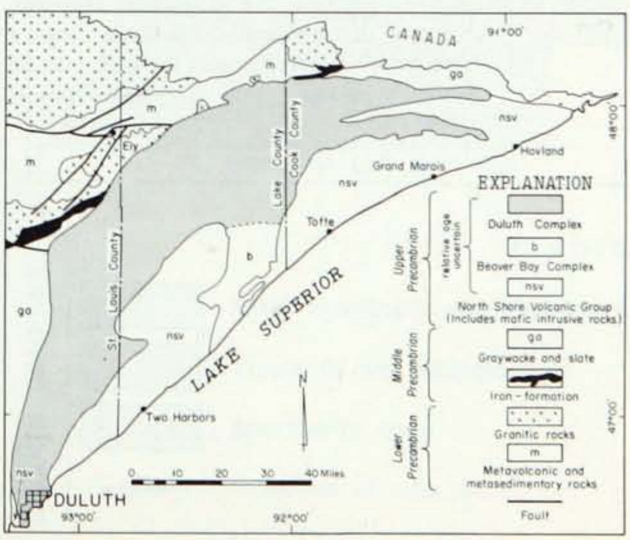

Figure V-23. Map showing Duluth Complex and adjacent rock types (after Sims, 1970).

Most of the recent work has been directed toward developing adequate geologic maps of the complex. Also, some work has been concerned with the occurrences of copper-nickel sulfide and titaniferous magnetite deposits and with the petrology and structure of various parts. Weiblen (1965, unpub. Ph.D. thesis, Univ. Minn.) investigated the funnel-shaped intrusion at Bald Eagle Lake in the Gabbro Lake quadrangle; Nathan (1969, unpub. Ph.D. thesis, Univ. Minn.) mapped part of the northern prong, emphasizing studies of the petrology of the oxide deposits; Davidson (1969a and b) mapped areas in the eastern part that previously were largely unknown; Hardyman (1969, unpub. M.S. thesis, Univ. Minn.) made a detailed study of both the silicate and sulfide mineralogy in a long drill core from near Babbitt; and Bonnichsen (1969b, 1971) has mapped the Babbitt-Hoyt Lakes region, which is the major area of known sulfide deposits, as well as the remaining poorly exposed parts of the southern half of the complex.

Since the early 1950's the Duluth Complex has been explored as a potential source of copper and nickel. Although most of the data obtained by drilling and test pitting are unavailable to the public, some have been placed in open files of the Minnesota Geological Survey.

The Duluth Complex is described on the following pages with respect to four areal subdivisions (fig. V-24).

These are: (1) the northwestern part, extending from the Kawishiwi River through Gabbro Lake and Lake Insula to the Ogishkemuncie Lake quadrangle; (2) the northern prong, extending eastward from the main body of the complex through the Gunflint Lake area; (3) the eastern part, extending eastward from Isabella Lake through the southern prong of the complex south of Brule Lake eastward to Lake Superior; and (4) the southern part, extending southward from a line through Babbitt and Greenwood Lake eastward to the western boundary of the Cramer quadrangle.

In reviewing the literature on the Duluth Complex as well as that on other mafic and ultramafic rocks, it is apparent that there is no complete systematic nomenclature. Accordingly, the classification scheme in Figure V-25 was adopted for this paper. The scheme allows the root name to be determined solely by the percentage of essential minerals without regard to texture, and thus conforms to general usage of root names in the past. If no textural terms are used, there are no textural implications. If further textural information is available, the terminology is expanded by use of modifiers, as discussed in the explanation for Figure V-25. This additional information allows one to determine within reasonable limits the proportions of various minerals and the nature of their occurrence (interstitial, early-formed liquidus phase, and others). Although some of these terms may result in rather lengthy rock names, the additional information is advantageous for both mapping and petrologic interpretations.

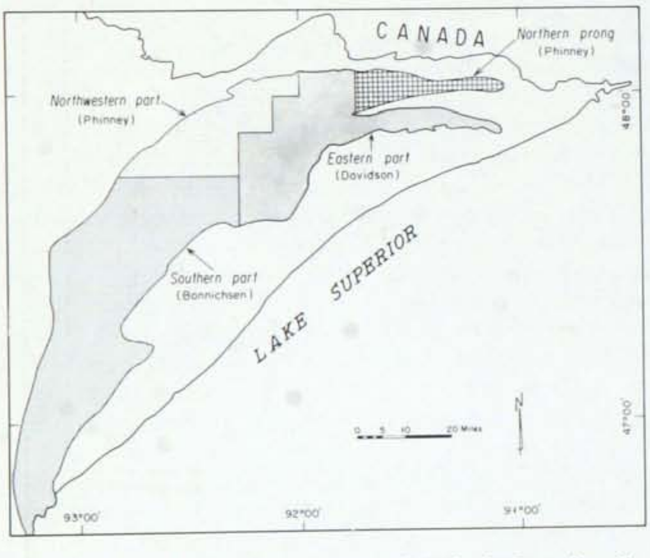

Figure V-24. Map showing areas of Duluth Complex discussed separately in text.

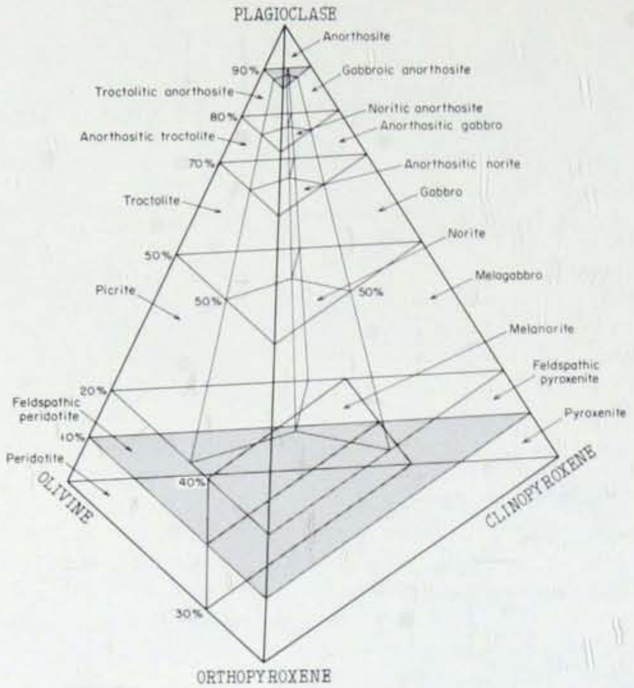

Figure V-25. Classification scheme for mafic rocks. The terms "clinopyroxene" and "orthopyroxene" may be replaced by the appropriate mineral names. If non-essential minerals constitute less than 10 percent by volume, prefix terms are used such as "olivine-bearing," "augite-bearing," "magnetite-bearing"; if non-essential minerals exceed 10 percent, the mineral name is used as a prefix, as "augite troctolite." Prefix textural terms are used where appropriate, for example, "poikilitic olivine-bearing anorthositic gabbro." Note: this classification breaks down when there are large amounts (>20%) of magnetite, ilmenite, chromite, or other spinels. In these cases, a more detailed description of the rock may be given. A general name for these types is "oxide rock."

NORTHWESTERN PART OF DULUTH COMPLEX

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Within the northwestern part of the Duluth Complex there are three major groups of rocks (fig. V-26)—anorthositic, troctolitic, and gabbroic types—and inclusions of various types. Of the approximately 95 square miles of Duluth Complex underlying the Gabbro Lake quadrangle (Green and others, 1966; Phinney, 1969), about 50 percent is made up of anorthositic rocks and 50 percent of troctolitic and gabbroic rocks. An estimated 5 to 10 percent of the area mapped as troctolitic rocks consists of inclusions. Of the approximately 170 square miles in the Forest Center quadrangle mapped as Duluth Complex, 77 percent is underlain by anorthositic rocks and 23 percent by troctolitic to gabbroic rocks. At least 15 percent of the latter is made up of inclusions.

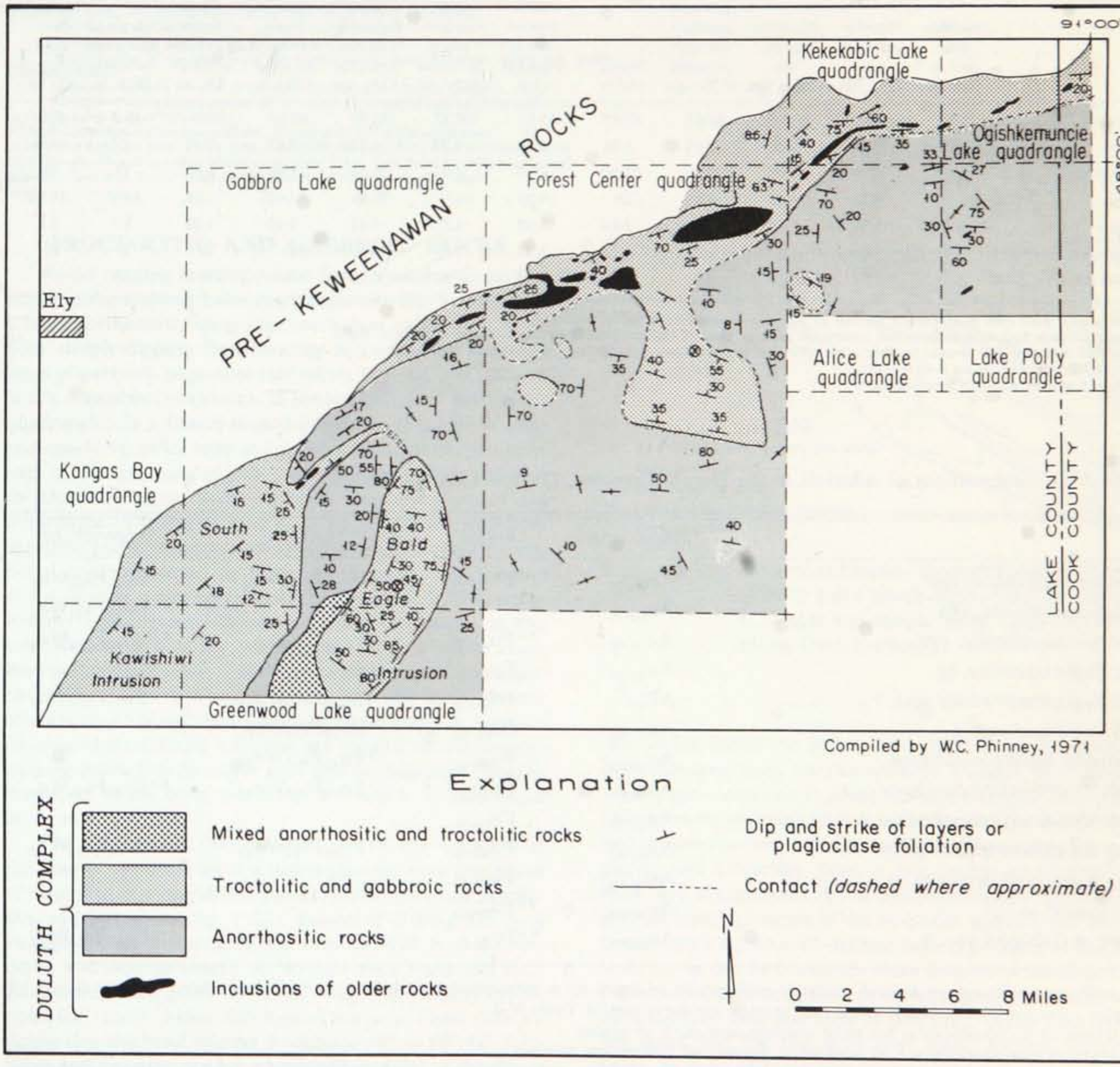

Figure V-26. Generalized geologic map of northwestern part of Duluth Complex.

ANORTHOSITIC ROCKS

In the Gabbro Lake and Forest Center quadrangles (Green and others, 1966; Phinney, 1969) anorthositic rocks are interpreted as the oldest of the intrusive units on the basis of inclusions of anorthosite in troctolite and gabbroic rocks and of crosscutting relationships of anorthositic rocks by troctolite and gabbro. A small noritic body also intrudes the anorthositic rocks in the Gabbro Lake quadrangle. Modes of the various anorthositic rocks are given in Tables V-8 and V-10, and mineral compositions are given in Table V-9. Within the anorthositic rocks, several stages of intrusion are recognized. Inclusions of one anorthositic unit in another, dikes of one in another, and crosscutting plagioclase lamination lead to the conclusion that there were several pulses of anorthositic melt. Within the anorthositic units the planar orientation of plagioclase laths generally is excellent, and indicates very irregular patterns interpreted as resulting from viscous flow in a crystal mush. The rock types and structures of the rocks within the Gabbro Lake quadrangle extend southward and southeastward into the Greenwood Lake and Kangas Bay quadrangles, respectively.

Table V-8. Average modes, in volume percent, of units in the Duluth Complex, Gabbro Lake quadrangle.

	Anorthositic gabbro (35) ago	Noritic anorthosite (37) agh	Gabbroic anorthosite (215) agu	Contact zone of gabbroic anorthosite with troctolite (12) agu	Norite (6) n	Basal contact zone of troctolite (24)	South Kawishiwi augite troctolite (13) sat	South Kawishiwi poikilitic troctolite (33) spt	Dike of South Kawishiwi intrusion (48) st	Anorthosite in South Kawishiwi intrusion (9) sa	Margin of Bald Eagle intrusion (12) bt	Central zone of Bald Eagle intrusion (7) bg
Plagioclase	78.87	87.44	87.56	85.67	67.55	65.51	68.57	71.16	63.65	95.2	57.3	42.2
Augite	8.90	2.63	3.10	5.00	3.91	6.31	9.73	4.52	8.63	0.85	2.1	46.0
Hypersthene	1.12	6.09	1.46	0.59	23.90	2.63	2.02	0.55	1.08	0.60	Tr	Tr
Olivine	3.24	0.24	4.45	6.17	Tr	14.23	14.82	19.34	22.42	1.26	34.0*	10.00**
Opaques	2.93	1.34	1.36	1.94	3.06	3.90	1.34	2.02	3.40	1.28	5.3	1.0
Biotite	2.50	1.07	1.33	.37	1.25	3.33	1.97	0.63	Tr	Tr	Tr	Tr
Symplectite	2.15	0.55	0.19	Tr	Tr	3.18	1.30	0.15	Tr	Tr		

Numbers after each rock type indicate number of thin sections used to calculate average mode
Letters after each rock type indicate unit symbol on geologic map of Gabbro Lake quadrangle (Green and others, 1966)
Modes total less than 100 percent because a few alteration and accessory minerals are excluded from the tabulation
* About 20 percent of olivine is serpentinized
** About 2 percent of olivine is serpentinized

Table V-9. Compositions of minerals in the Duluth Complex, Gabbro Lake quadrangle.

	*Plagioclase	Olivine	***Augite	Hypersthene
Anorthositic gabbro, ago	An_{54-70}	**Fo_{35-48}	**$En_{40}Fs_{21}Wo_{39}$	**$En_{45}Fs_{52}Wo_3$
Noritic anorthosite, agh	An_{58-70}			**$En_{51}Fs_{43}Wo_6$
Gabbroic anorthosite, agu	An_{55-85}	**Fo_{54-60}	**$En_{44}Fs_{16}Wo_{41}$	**$En_{46}Fs_{52}Wo_2$
Bald Eagle troctolite, bt	An_{62-81}	Fo_{71-74}	**$En_{39}Fs_{21}Wo_{41}$	
Bald Eagle intermediate rock	An_{67}	Fo_{69-70}	$En_{53}Fs_{12}Wo_{34}$	**$En_{74}Fs_{24}Wo_2$
Bald Eagle gabbro, bg	An_{57-63}	Fo_{61-62}	$En_{45}Fs_{14}Wo_{41}$	
Troctolite, basal contact zone	An_{53-67}	Fo_{59-65}	**$En_{45}Fs_{16}Wo_{39}$	
South Kawishiwi contact zone, scz	An_{55-65}	Fo_{50}	**$En_{38}Fs_{21}Wo_{42}$	**$En_{39}Fs_{59}Wo_2$
South Kawishiwi troctolite, sat	An_{57-70}	Fo_{50-55}		
South Kawishiwi troctolite, spt	An_{57-67}	Fo_{59-62}	**$En_{43}Fs_{16}Wo_{41}$	**$En_{64}Fs_{34}Wo_2$
South Kawishiwi troctolite, st	An_{58-72}	Fo_{55-63}		
Anorthosite, sa	An_{60-68}	**Fo_{60-62}		
Norite at Gabbro Lake, n	An_{35-46}	**Fo_{51}		$En_{60}Fs_{36}Wo_4$

All analyses are based on several replicate microprobe analyses on several points on each of several grains from between 2 and 10 samples from each unit; letters after rock names are explained in Table V-8
* Range of compositions is for most calcium-rich parts of grains
** Mineral compositions are of interstitial phases; all others are primary
*** Augite analyses show much variation within individual grains depending upon degree of exsolution and proximity to exsolved lamellae; only averages are listed

Table V-10. Modes, in volume percent, of rocks in Forest Center and adjacent quadrangles.

	FOREST CENTER QUADRANGLE											OTHER QUADRANGLES			
	Troctolitic rocks				Anorthositic rocks				Gabbroic rocks		Inclusion				
	M10404	M10460	M10400	M10441	M10463	M10632A	M10483	M10374	M10443	M10251	M10411	M10634	M10376	M10357	M10622
Plagioclase	79.73	76.93	70.82	23.30	98.02	85.98	90.33	77.69	49.44	57.21	62.12	100.00	96.82	90.87	56.02
Augite	3.39		4.07	2.09	.26	1.32	2.27	12.99	32.90	21.43	20.18			.13	1.43
Hypersthene	2.93	.40	.94	.13	.73	1.01	1.01	.99	.39					.19	.91
Olivine	10.11	22.05	19.04	71.01	.13	6.69	4.31	7.27	.59	16.87	17.50		.65	.44	39.82
Opaques	2.38	.19	2.50	2.68		1.07	.82	.44	9.81	3.61	.07		.71	4.76	1.43
Biotite	.78	.13	1.94	.72	.13	2.15	.69	.50	5.62	.81	.07		.58	1.76	.33
Matrix*	.69	.26	.69	.07	.73	1.70	.95	.18	1.24				1.36	1.84	.07

* Matrix material consists of symplectic intergrowths of hypersthene-plagioclase or biotite-ilmenite
Troctolitic rocks—Olivine and plagioclase of cumulate origin, other minerals interstitial
M10404 and M10400—Typical rocks of Lake 1 area
M10460—Southern margin of troctolite intrusion on Lake 2
M10441—From one of several thin (up to 1 foot) olivine-rich layers in troctolite
Anorthositic rocks—Plagioclase of cumulate origin, other minerals interstitial
M10463 and M10483—Typical rocks of Lakes 2, 3, and 4
M10632A—From southern part of quadrangle
M10374—One of the most mafic anorthositic rocks, much interstitial olivine and augite
Gabbroic rocks—
M10443—Within a few feet of the north contact of the Duluth Complex
M10251—Near southern contact of intrusion in east-central part of quadrangle
M10411—Large igneous inclusion
M10634—Anorthosite from Thomas Lake, Thomas Lake 7.5-minute quadrangle
M10376—Anorthosite from Adams Lake, Alice Lake and Lake Polly 7.5-minute quadrangles
M10357—Anorthosite typical of Malberg and Koma Lakes, Lake Polly 7.5-minute quadrangle
M10622—Troctolite from Fraser Lake, Kekekabic Lake 7.5 minute quadrangle

TROCTOLITIC AND GABBROIC ROCKS

Rocks ranging in composition from troctolite to gabbro occur in the Gabbro Lake quadrangle (see fig. V-26) as: 1) a large northeast-trending dike more than half a mile wide, with steeply-dipping flow-banding and late-stage segregations of gabbroic pegmatite (see tables V-8 and V-9, dike of South Kawishiwi intrusion), in the south-central part of the quadrangle; 2) a funnel-shaped intrusion, elliptical in plan and nearly 10 miles long and 2 miles wide, in the southern part of the quadrangle; 3) a shallow-dipping basin referred to informally as the South Kawishiwi intrusion, which has dimensions of about 6 miles by 7 miles and is defined by excellent planar orientation of plagioclase and cyclic gradational layering caused by higher concentrations of olivine at the base of many layers (tables V-8 and V-9, South Kawishiwi augite troctolite and poikilitic troctolite), in the southwestern part of the quadrangle; and 4) a wedge along the basal contact zone, in the northeastern part of the quadrangle (see tables V-8 and V-9, basal contact zone). Planar orientation of plagioclase and mineral layering in the funnel-shaped Bald Eagle intrusion are parallel to the steeply-dipping contact in the outer zone and become horizontal at the center of the body (see tables V-8 and V-9, Bald Eagle intrusion).

The lowermost 2,000 to 3,000 feet of the troctolite at the base of the complex is a heterogeneous zone composed of troctolite, olivine gabbro, picrite, norite, gabbroic pegmatite, and inclusions (fig. V-27). Extensive drilling has been carried out in this zone by the International Nickel Company, and descriptions of the various rock types and their relationships are given by R. E. Wager and others (1969, open-file report, Minn. Geological Survey). These data indicate that the basal contact is inclined 30° to 60° SE. Layering and foliation are not necessarily parallel to the dip of the basal contact, and may be somewhat shallower. Troctolite, which forms the bulk of the basal zone, is texturally heterogeneous even on the scale of a single thin section. Significant variations in grain size and changes from poikilitic to more granular texture may occur over distances of two to three inches. At some localities, alternating layers of picrite and troctolite, from a few inches to several feet thick, are conspicuous. A less conspicuous fine- and coarse-grained layering occurs in the troctolite, gabbro, and norite. Some norite shows well-defined layering. None of the separate layers can be traced for more than a few hundred feet. Lens-shaped concentrations of inclusions as much as several hundred feet long and tens of feet thick occur throughout the zone, but for the most part form distinct, traceable units. In most of these units the population of inclusions is greater than 30 percent.

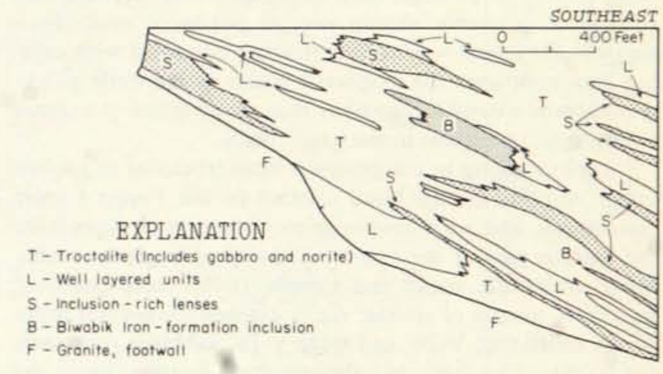

Figure V-27. Generalized section through the contact zone, Gabbro Lake quadrangle. Adapted from R. E. Wager and others, 1969 (open-file report, Minn. Geol. Survey).

Within the more mafic troctolite and picrite in the contact zone there are random lenses as much as several hundred feet long that contain abundant magnetite. The lenses range in thickness from a few inches to several tens of feet and are roughly parallel to the basal contact. The magnetite occurs with olivine, plagioclase, and some pyroxene and typically exists in modal concentrations of 30 to 50 percent; locally, magnetite constitutes nearly 90 percent of the rock over a vertical interval of several feet. Magnetic concentrates from these lenses and from inclusions of Biwabik Iron-formation were analyzed for titanium, vanadium, chromium, and manganese (Wager and others, 1969, *op. cit.*). From the results shown in Table V-11, it is clear that those magnetite grains interpreted as precipitates from melt are significantly enriched in titanium, vanadium, and chromium as compared to magnetite from inclusions of Biwabik Iron-formation. As would be expected, the igneous magnetite shows extensive exsolution of ulvospinel and pleonaste. From one 16-foot-thick magnetite-rich lens, analyses at two-foot intervals show a gradual increase from 3.0 percent titanium at the base to 7 percent at the top. The numerous inclusion-rich lenses, alternating troctolitic and picritic layers, and large plate-like inclusions of iron-formation—all parallel to the basal contact—indicate a significant amount of flowage in a melt laden with crystals and inclusions, probably during several pulses of melt. In a single melt, which was neither flowing nor filled with crystals and inclusions, the magnetite grains would settle quickly and form a basal layer rather than many lenses at various stratigraphic positions in the basal rocks.

Rocks ranging in composition from troctolite to gabbro occur adjacent to the basal contact in the Forest Center quadrangle, and are a continuation of these rock types from the eastern part of the Gabbro Lake quadrangle (see table V-10, troctolitic rocks and sample 10443 from gabbroic rocks). A tongue of similar rocks extends for several miles to the south (fig. V-26, and table V-10, gabbroic rock sample 10251). This body of gabbroic rocks contains many inclusions of anorthosite and has chilled borders against anorthosite. In general, a mineralogic layering and planar plagioclase orientation trend parallel to the contact, but the dips are quite variable, particularly near the southern contact where the lensoid and swirled appearance of the layers seems clearly to indicate flow structure. In the vicinity of Lake Insula, in the northeastern part of the quadrangle, there are several clear-cut dikes of troctolite in the anorthositic rocks, but in the vicinity of the major contact between the large areas of the two rock types the relationships are more obscure and the two rock types seem to be rather well mixed. The exposures along the shore in the northeastern part of the lake show nearly flat-lying interlayering of the two types, and in some outcrops there seem to be inclusions of each type in the other as though the anorthositic material was not solid at the time the troctolite was emplaced, allowing considerable intermingling of the two crystal mushes. There is a small positive gravity anomaly in this area (Ikola, 1968b, 1970) that may be related to a troctolitic mass that intrudes anorthositic rocks.

INCLUSIONS

Inclusions of variable shapes and sizes occur in the troctolitic rocks in both quadrangles. Although some of the inclusions are in the large dike within the Gabbro Lake quadrangle, mentioned previously, most occur in the lower part of the complex within horizontal distances of a mile or two from the basal contact. They are particularly abundant in the basal troctolite zone in the Forest Center quadrangle. From a series of traverses across the zone within two miles of the basal contact in the Forest Center quadrangle and other quadrangles to the east, Grout (1930) estimated rather conservatively that 15 percent by volume of this zone consists of inclusions. Some inclusions occur also in the anorthositic rocks, but they are much less numerous.

Some of the inclusions are clearly derived from iron-formation, for they contain interlayered recrystallized quartz, coarse magnetite, and iron silicate minerals. Granitic segregations, some of which are pegmatitic, are common near the contact in the southwestern part of the Gabbro Lake quadrangle, and quite likely represent partially to completely melted granitic inclusions. Other fine- to medium-grained granular inclusions have more obscure origins. Modal analyses and mineral compositions from some inclusions are given in Table V-12. Although these data (Renner, 1969, unpub. M.S. thesis, Univ. Minn.; Hardyman, 1969, unpub. M.S. thesis, Univ. Minn.) are from inclusions in the adjacent Babbitt area, the inclusions are similar to some of those in the northwestern part of the complex. Further discussion of the inclusions is given by Bonnichsen in the section on the southern part of the complex.

The presence of cordierite and/or biotite in some of the inclusions clearly indicates that the inclusions were derived from sedimentary rocks, but the mineral assemblages and compositions of other inclusions (for example D-3, D-17, or D-11-b) possibly are indicative of an igneous origin. For the latter type, there are two possible interpretations, either (1) that many were originally sedimentary rocks that

Table V-11. Mean Ti, V, Cr, and Mn content of magnetite of igneous and iron-formation derivation, Gabbro Lake quadrangle (data from Wager and others, 1969, open-file report, Minn. Geol. Survey).

	No. of samples	Ti (in percent)	V (in percent)	Cr (in percent)	Mn (in percent)
Igneous magnetite	95	4.26	0.24	0.53	0.20
Iron-formation magnetite	52	0.95	0.043	0.14	0.17

Table V-12. Modes and compositions of minerals in inclusions and associated gabbroic rocks near Babbitt (data for 61-1965 and 61-1968 from Hardyman, 1969, unpub. M.S. thesis, Univ. Minn.; remainder from Renner, 1969, unpub. M.S. thesis, Univ. Minn.).

	Plagioclase		Olivine		Hypersthene				Augite				Biotite			Cordierite			Opaques	Others
	Mode	An	Mode	Fo	Mode	FeO	MgO	CaO	Mode	FeO	MgO	CaO	Mode	FeO	MgO	Mode	FeO	MgO	Mode	Mode
																				kfsp
*D-2-Aa	51.2	37.3			33.4	30.0	16.4	0.19											7.8	7.6
*D-2-Ab					29.6								0.6	18.5	12.7	60.1	6.8	9.7	9.7	
*D-2-B	2.5	34.4			33.0	31.3	16.7	0.14					1.4	18.7	11.3	54.9	6.7	9.5	8.2	
†*D-3-a-1	55.5				40.5	27.6	18.8	0.60	Pr	10.7	11.4	18.1							4.0	
*D-3-a-2	49.2	46.5			37.4														8.3	
†*D-3-b-1	52.8	55.5			44.9	28.0	14.5	0.96	Pr	13.0	10.9	17.0	0.2	14.7	15.5				2.2	
*D-3-b-2	59.6				37.8														2.5	
*D-3-c	52.8				36.6														10.6	
*D-3-d	64.0				31.5														4.5	
D-4-B	58.0	45.2			36.0	27.6	18.2	0.86					0.8	19.2	11.6				5.2	
D-8	50.0	66.0			42.4	30.9	17.3	1.49	2.0	13.1	14.7	20.7	0.7	15.6	14.2				5.2	alt
D-9	52.4	60.6			22.2	28.3	19.2	1.41	4.9	12.6	12.6	20.6	5.3	19.3	11.9				3.9	11.3
D-7	65.8	62.8	4.4	44.8	2.0	25.3	18.7	1.44	9.8	10.4	12.7	20.7	7.4	16.1	15.3				2.8	7.7
D-11-a	[1]	62.3							81.5	8.8	14.7	22.1	1.1	18.9	12.3				1.6	15.8
*D-11-b	39.4	51.0	0.7						41.5	9.1	14.4	21.8	5.9	17.0	17.4				4.1	8.5
D-14	48.2	51.5			22.8	28.3	18.1	1.50	12.3	12.1	12.1	20.6	2.6	18.8	13.2				8.5	5.6
D-16	52.5	57.4			29.5	32.0	14.3	0.99					3.5	20.9	10.0				14.5	
D-15	55.5	60.1	3.1	34.4	16.6	27.1	19.3	1.10	13.2	11.2	13.0	20.9	3.1	15.5	16.7				3.7	4.8
*D-17	45.7	63.1	26.1	42.3	3.1	24.3	20.7	1.30	9.8	10.1	14.4	20.1	7.4	15.8	16.4				2.5	5.3
[2]D-18-a	26.2	63.0			16.7	29.8	13.5	1.20	8.6	10.1	12.1	21.4	1.2	12.8	18.8				4.3	5.9
*D-18-b	56.8	56.5			37.5	27.7	16.0	1.20	.4	13.5	10.9	21.2	0.1							5.2
*D-18-c	[1]	75.6							73.1	9.0	13.1	21.4	1.3						0.6	25.0
																				qtz
*F-4-c-A	64.7	50.2			30.0								0.5	17.4	14.5				1.5	3.2
*F-4-c-B	78.6				16.0														4.4	1.0
*F-4-c-C	27.4				72.0								Pr	17.7	11.8				0.6	
*F-4-c-D	67.4				30.1														2.5	
*F-4-c-E	65.5				21.8								7.1						5.5	
*F-4-c-F	59.7				37.8								Pr						2.5	
*F-4-c-G	56.7				28.1								10.5						4.6	
*F-3	1.1	35.1			12.4	33.7	14.9	0.27					28.7	19.8	9.6	56.2	7.8	9.6	1.7	kfsp
F-2A-a	50.3	60.6			42.3	33.3	17.2	0.59					1.3						6.2	
F-2A-b	55.6	37.2			19.1	33.3	17.2	0.59					22.1	18.9	12.2				1.0	2.0
F-2-B	39.0	35.0			14.9	30.1	18.1	0.49					28.2	18.4	12.4				0.5	17.4
																				Sym
F-1	47.3	59.4	16.0	37.1	5.6	25.8	17.9	1.20	9.1	13.0	12.2	19.8	12.4	17.2	14.7				5.2	4.4
F-0	53.7	60.4	13.1	37.4	6.6	26.4	20.4	1.40	10.7	12.0	13.6	21.7	7.4	16.3	15.2				4.5	4.0
61-1965	48.7	28			6.3								13.6			27.1			4.3	
61-1968	40.6	34			15.4								20.6						23.5	

[1] Included in alteration products
[2] Contains 37.0 percent symplectite
† Augite and hypersthene do not occur in contact with each other
* Inclusions
Sym = symplectite
61-1965, 61-1968 Metasediments from footwall
D-2, 3 Inclusion in contact with D-4

D-4, 8, 9, 7 Igneous rock at successively further distances from inclusion
D-4 is at contact, D-7 is about 2 feet away
D-11-a Dike through D-11-b
D-14, 15, 16, 18-a Igneous
D-17, 18-b, 18-c Inclusions associated with D-14, 15, 16, 18-a
F-3 and all F-4 Inclusions in contact with F-2
F-2, 1, 0 Igneous rocks at successively further distances from inclusion

were partially melted, leaving a more refractory residuum as the inclusion, or (2) that many were originally fine- to medium-grained gabbroic rocks. In the latter case the inclusions may be derived from dikes, sills, or chilled margins formed early in the magmatic history of the complex.

In the basal troctolite zone of the Gabbro Lake quadrangle, where many core holes have been drilled, a wide variety of rock types occur as inclusions, and these can be placed in three main categories: (1) light to medium-gray, fine-grained, sandy textured hornfelses of uncertain origin, which are most abundant; clinopyroxene is the dominant mafic mineral together with local olivine; (2) blocks of well banded Biwabik Iron-formation, typically 20 to 30 feet thick but as much as 200 feet thick and several hundred feet long, are quite common; (3) a variety of rocks that seem to be recrystallized olivine gabbro and troctolite; they vary considerably in mineralogy and texture and contain laths of zoned plagioclase; boundaries between all inclusions and surrounding troctolite are sharp both in thin section and in drill cores.

In a large inclusion east of Lake One, in the northwestern part of the Forest Center quadrangle (fig. V-26), plagioclase phenocrysts occur in a fine-grained, granular matrix of plagioclase, olivine, and pyroxene, indicating an igneous origin (see table V-13, inclusion). At the north end of Kiana Lake there are many fine-grained gabbroic inclusions in troctolite; the concentration of these inclusions increases northward until the rock type becomes entirely gabbro. Within this gabbroic rock are inclusions of granite, iron-formation, and anorthositic rocks. Thus, in this area, igneous rocks were incorporated into later intrusions of troctolitic melt. Other fine-grained inclusions of uncertain origin, some of which are associated with granitic rocks, also are present in the Forest Center quadrangle. One of these granitic associations occurs in the southeastern part of the troctolite intrusion. At this locality, an elongate mass of granite contains several large angular plagioclase grains or clumps of grains similar in size to the plagioclase in the surrounding troctolite. At the south end of the granitic outcrops there are several inclusions composed of a fine-grained, black, granular rock. The southernmost part of this outcrop is a large mass of this black, granular material which is an inclusion in the troctolite. Several similar relationships between granite and black, granular material occur in the same general area. In the Perent Lake and Kawishiwi Lake quadrangles (Davidson, 1969a and b), similar close associations exist between granitic rocks and granular gabbro or norite. It is clear from experimental data (Winkler and von Platen, 1958, 1960, 1961a and b) that inclusions of sedimentary rocks or intermediate igneous rocks in a basaltic melt will produce granitic melt plus a more gabbroic residual solid. Possibly, the associations described above represent the partial melting of large inclusions. Alternatively, the granitic rocks may represent inclusions from the Giants Range Granite, which can be inferred from gravity data (Ikola, 1970) to underlie the Duluth Complex in this area.

Table V-13. Chemical composition of clinopyroxene from troctolite, Gabbro Lake quadrangle (analysis obtained from Tadashi Konda, Univ. of Yamagata, Japan).

Sample	M10142
SiO_2	50.52
TiO_2	0.12
Al_2O_3	3.45
Fe_2O_3	1.47
FeO	14.43
MnO	0.32
MgO	16.02
CaO	12.77
Na_2O	0.24
K_2O	0.07
H_2O (+)	0.60

FURTHER OBSERVATIONS

Through reconnaissance mapping in adjacent areas—Ensign Lake, Alice Lake, Lake Polly, Ogishkemuncie Lake, and Kekekabic Lake 7.5-minute quadrangles—the general geology can be extended eastward from the Forest Center quadrangle. For the most part, the rock patterns are similar to those described above. The majority of the area is underlain by anorthositic rocks. Except for the zone along the base of the complex and extending southward for a distance of about 2 to 2½ miles, anorthositic rocks occur almost to the exclusion of other rock types. In Cook County, data from reconnaissance maps (Grout and others, 1959) of T. 63 N., R. 5 W., T. 64 N., R. 5 W., and T. 64 N., R. 4 W. indicate that anorthositic units extend a few miles into the complex. Several areas of these rocks have been examined, and they exhibit the same complex structural relationships—highly variable and crosscutting plagioclase lamination—between different anorthositic units as seen further westward. Rocks ranging in composition from troctolite through olivine gabbro occur within the lowermost 2 to 2½ miles from the basal contact and also as intrusions into the anorthositic rocks on the southwestern part of Alice Lake and the southern part of Thomas Lake. In the zone along the contact, the layers generally dip 30°-60° S. On Alice Lake, layering in the troctolite is rather shallow and dips less than 20° S. Orientation of the dips is somewhat variable, and may indicate a basin-shaped structure.

In the southeastern corner of the Ensign Lake quadrangle (fig. V-26), several troctolitic and gabbroic intrusive units form an irregular protrusion in the basal contact. These intrusions are stratigraphically below the major zone of inclusions, but contain numerous inclusions of anorthositic rocks, particularly in the vicinity of Ima Lake. The time relationships of these intrusions are uncertain, but at the northern end of Fraser Lake there appears to be a chilled margin on the south side of one intrusion against a large mass of troctolitic rocks, implying that the younger intrusions are nearer the base. Modal analyses of some of the anorthositic and troctolitic rocks in this reconnaissance area are given in Table V-10.

Several types of inclusions occur in the rocks of the Ensign Lake area, primarily in the 2-mile-wide zone of troctolitic rocks near the contact. Some of the traverses by Grout (1930), mentioned previously, were in this area. A few inclusions occur in the anorthositic rocks, but they are far less numerous than those in the troctolitic rocks. The inclusions in the troctolitic rocks consist of anorthosite, iron-formation, hornfels equivalents of the adjacent graywacke-slate and greenstone, and the fine-grained granular gabbroic rocks described previously. One of the most striking inclusions is a narrow band of iron-formation extending almost continuously from the north end of Kiana Lake, in the Forest Center quadrangle, northeastward through Thomas Lake, then eastward north of Fraser Lake almost to Little Saganaga Lake, in the east-central part of the Ogishkemuncie Lake quadrangle, a total distance of nearly 11 miles.

PETROLOGIC DATA

Textural relationships in the more common rock types in the Ensign Lake area are shown by the photomicrographs in Figures V-28, V-29, V-30 and V-31. In the anorthositic rocks, plagioclase was the only phase to precipitate during early crystallization, and it occurs as laths; all other minerals are interstitial (see figs. V-28B, C, and D; figs. V-30B, C, and D). The plagioclase laths generally have a well-defined orientation (fig. V-28B). Interstitial material may occur as: (1) overgrowths on plagioclase to form nearly pure anorthosite (fig. V-28B); (2) large poikilitic grains of augite or hypersthene (fig. V-28D); or (3) more complex intergrowths of hypersthene, plagioclase, biotite, and ilmenite (figs. V-30B, C, and D). The nature of the interstitial material probably depends upon the extent to which the interstitial fluid was trapped or prevented from maintaining diffusion equilibrium with the main reservoir of melt during consolidation of the crystalline mass. Nearly pure anorthosite would form with the least trapping of the interstitial fluid, and the symplectic intergrowths of hypersthene-plagioclase plus ilmenite and biotite would represent the most complete trapping. The change from hypersthene-plagioclase to biotite represents an increase in the H_2O fugacity as crystallization of the interstitial fluid progressed. In a few of the anorthositic rocks, small grains of olivine, some mantled by hypersthene, occur among the larger plagioclase laths (fig. V-28C). These may represent growth of olivine in the interstitial fluid.

In the troctolitic rocks, olivine and plagioclase are the early phases; other phases are interstitial (figs. V-29A, B, C, and D; fig. V-30A). Large poikilitic augite grains are a common interstitial component in the troctolitic rocks (figs. V-29B and C), although overgrowths of olivine and plagioclase occur at places (fig. V-29A). Rather complex exsolution lamellae of hypersthene, ilmenite, and magnetite occur in the poikilitic augite grains (figs. V-29D and V-30A). In the olivine gabbroic rocks, olivine and plagioclase are joined by augite in the early-formed phases. Clinopyroxene is most common within a few hundred feet of contacts. A chemical analysis of a clinopyroxene containing hypersthene exsolution lamellae from near the basal contact of the troctolite intrusion in the Gabbro Lake quadrangle is given in Table V-13. There are large exsolution lamellae of Ca-poor pyroxene in the augite host. The analysis should represent the bulk composition of the original pyroxene. The other pyroxene compositions given in the table were determined by the electron microprobe, and are indicative of the extent of exsolution consequent on slow cooling of the more pigeonitic original pyroxene.

The small norite body in the Gabbro Lake quadrangle contains hypersthene with well developed augite lamellae, which indicate that pigeonite was the original pyroxene (fig. V-31). One of the typical granoblastic textures of an inclusion presumed to be of igneous parentage is shown in Figure V-28A. Further details of the textural relations are given in my recent report (Phinney, 1969).

DISCUSSION

From the various crosscutting relations, inclusions, and chilled margins, described above, it is clear that the anorthositic rocks throughout the northwestern part of the Duluth Complex were intruded by rocks ranging in composition from troctolite to gabbro. The various structural, textural, and mineralogic relations that occur in the anorthositic rocks are indicative of several pulses of melt, most of which was in the form of a crystal mush.

The troctolitic rocks are clearly younger than the anorthositic rocks and quite probably resulted from several pulses of melt. Some troctolitic rocks show chilled margins, whereas others lack chilled margins and appear to have been intruded while the anorthositic material still was a mush. Although some of the units have a flow structure, gradational layering caused by the concentration of denser minerals at the bottom of units exists in several shallow basins. Settling is likely to have occurred in the troctolitic rocks because the olivine and pyroxene crystals have a density contrast to the melt that is an order of magnitude greater than that of plagioclase.

The settling velocities for crystals of various sizes can be calculated if density differences between crystals and melt as well as the viscosity of the melt are known. The data pertinent to the anorthositic rocks are given in Table V-14, and are based on melt compositions estimated by me (Phinney, 1970), the partial molar densities determined by Bottinga and Weill (1970), and the viscosity coefficients of Weill (1971, written comm.). These data clearly indicate that the density difference between plagioclase and melt is in the range ± 0.01, assuming a plagioclase composition in the range An_{60-70}. The presence of about 0.3 to 0.4 weight percent H_2O in the melts would decrease the density of the melts by about 0.02 to 0.03. Assuming a plagioclase composition of An_{60} at 1,250° C and a melt of density 2.62, one can calculate a terminal velocity for crystals of 0.5 cm diameter in a melt having a viscosity of 500 poises. Utilizing Stokes law, $V = \frac{2}{9} \frac{r^2 g}{h} (\rho - \rho o)$, the settling velocity would be 0.00054 cm/sec or 1.94 cm/hr. The viscosity increases rapidly with cooling below 1,200° C (over 2,000 poises at 1,150° C). There is an additional effect as crystals begin to form. Shaw (1969; Shaw and others, 1968) has shown that as crystals begin to grow and concentrate in a melt, the viscosity rapidly increases to 2,000 poises or

Figure V-28. Photomicrographs of anorthositic rocks and an inclusion, Duluth Complex. A, typical granoblastic texture in inclusion. Rock contains plagioclase with albite and carlsbad twinning, augite, olivine, and a few opaque grains (M6198); B, well oriented plagioclase grains in anorthosite (M5812); C, troctolitic anorthosite. Three olivine grains occur from top center to bottom center of photo. The middle grain is mantled by hypersthene (M5742); D, noritic anorthosite. Optically continuous hypersthene fills the interstices between plagioclase grains (M5700).

Figure V-29. Photomicrographs of troctolitic rocks, Duluth Complex. A, typical texture of troctolite. The troctolite consists almost entirely of plagioclase and olivine; interstices are filled by overgrowths on plagioclase and olivine (M5866); B, troctolite showing poikilitic augite grain enclosing plagioclase, olivine, and ilmenite (plates at lower left) (M5882); C, poikilitic augite in troctolite. Clusters of opaque exsolution lamellae occur in zones that are free of plagioclase and olivine (M5883); D, enlarged view of exsolution in interstitial augite of anorthositic troctolite. Plates and nearly vertical rods are ilmenite; nearly horizontal rods are magnetite. These are concentrated along hypersthene lamellae in augite host. Note biotite blades at contact with plagioclase grain on right. Biotites occur only where hypersthene lamellae are in contact with plagioclase (M10142).

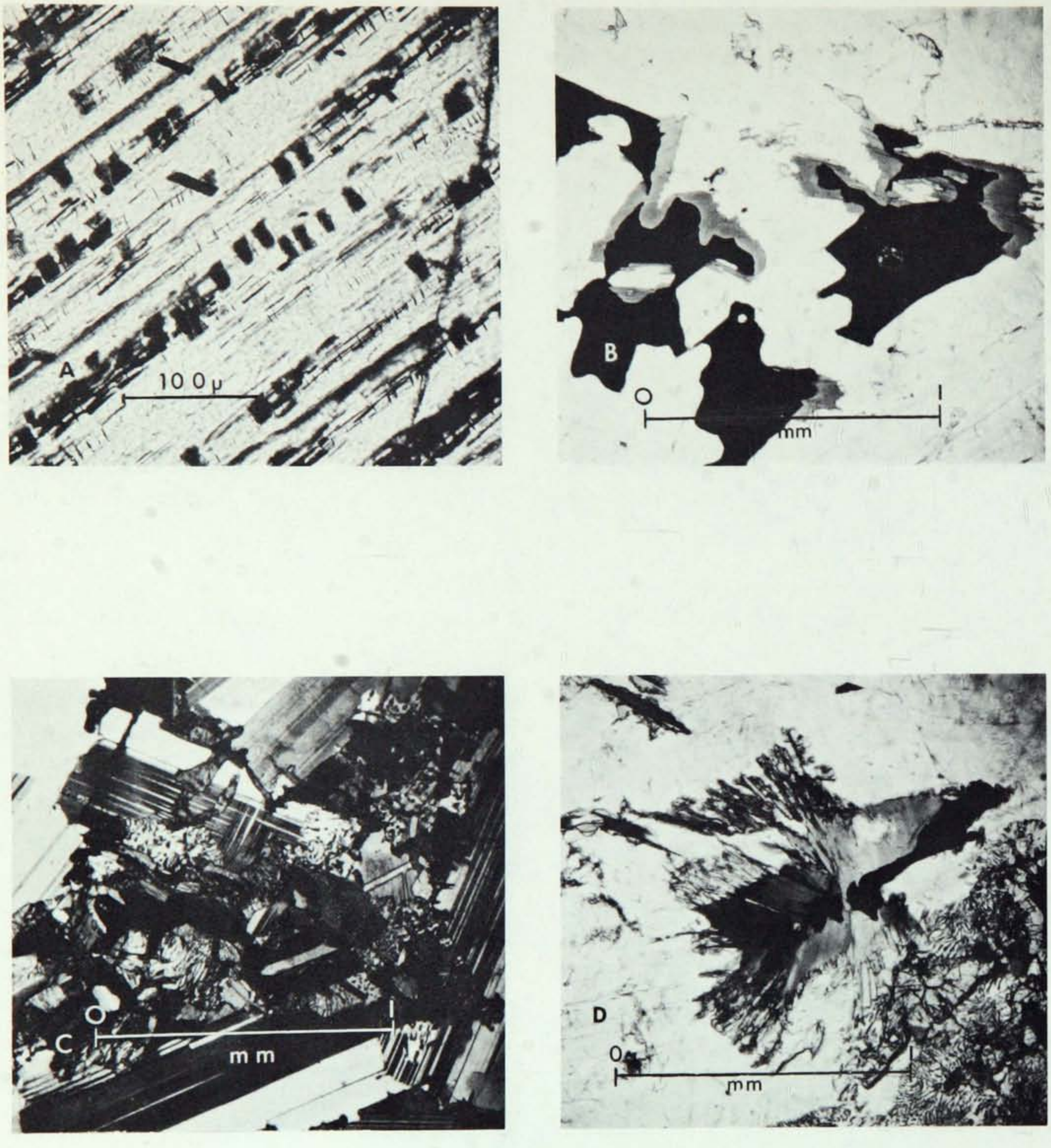

Figure V-30. Photomicrographs of anorthositic and troctolitic rocks, Duluth Complex. A, exsolution in augite of anorthositic troctolite. Plates are ilmenite; rods are magnetite. Plates and rods are concentrated along hypersthene lamellae. Two plates in top center are in augite host and show a different orientation than those in hypersthene lamellae (M10142); B, interstitial biotite and ilmenite (black) in anorthosite. Sequence of interstitial filling appears to have been biotite followed by ilmenite in the core of the filling (M5724); C, interstitial filling of symplectic intergrowth of hypersthene and plagioclase in anorthosite (M5700); D, interstitial filling in anorthosite. Sequence appears to be symplectic hypersthene and plagioclase followed by biotite and finally by ilmenite (M5724).

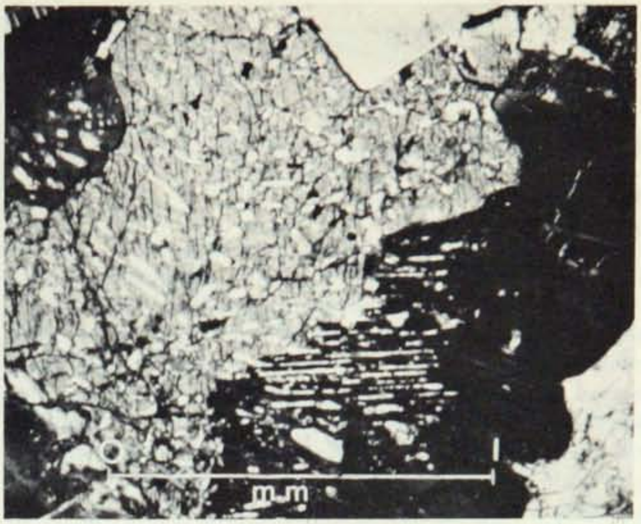

Figure V-31. Photomicrograph showing exsolution of augite in hypersthene host, from a small norite intrusion at Gabbro Lake. Orientation of lamellae indicates that inverted pigeonite was the original pyroxene (M5737).

more. Even without considering the interference between settling grains, the settling velocity decreases to only 0.38 cm/hr if the combined viscosity effects produce a value of 2,500 poises. Thus, inasmuch as this calculation was made on the basis of maximum density differences, large grain size, spherical grains, and no interference between settling grains, it seems reasonable to assume that the maximum settling velocity is on the order of a few tenths of a centimeter per hour. In many cases, the density difference may be near zero and the grains would remain in suspension. It is also possible that the crystals were less dense than the melt, in which case there might be a slight upward velocity for the crystals. Assuming maximum settling velocities of 0.1 to 0.2 cm/hr, it would require one year for crystals of plagioclase to settle only 8 to 15 meters. In such an environment, periodic currents with velocities of less than a centimeter per hour would keep the crystals from settling and the melt would move as a crystal mush. Such currents may result from convection due to cooling, influx of new melt, or extrusion of lava to the surface. The highly variable internal structure in the anorthositic rocks, indicated by the foliation given by plagioclase laths, is excellent evidence for this rather simple mechanical analysis.

Table V-14. Densities and viscosities of dry melts and plagioclase for anorthositic rocks of the Duluth Complex.

	Melts								
	T-56	T-45	F-96	KC-9	TH-2	LW-10	GFK-108	T-22	
ρ at 1250°C	2.65	2.68	2.64	2.68	2.64	2.66	2.65	2.67	gms/cc
η at 1200°C	1047	700	1250	761	1101	739	1183	750	poise
η at 1300°C	248	166	301	176	255	177	265	170	
	Plagioclase								
	An_{50}	An_{60}	An_{70}	An_{80}					
ρ at 1200°C	2.64	2.64	2.66	2.67					
ρ at 1000°C	2.64	2.65	2.67	2.68					

NORTHERN PRONG, DULUTH COMPLEX

William C. Phinney

The western half of the northern prong (fig. V-32) consists mainly of a series of thin intrusive sheets of gabbroic rocks. These intrusive sheets are transected along the western margin of the Gunflint Lake quadrangle by younger anorthositic rocks. The data that follow are taken largely from a report by H. L. Nathan (1969, unpub. Ph.D. thesis, Univ. Minn.), who mapped the Duluth Complex within the Gunflint Lake, South Lake, and Hungry Jack Lake 7.5-minute quadrangles. The remaining data are compiled from reports of Grout and others (1959) and R. C. Babcock (1959, unpub. Ph.D. dissert., Univ. Wisconsin).

Twenty-seven separate units were mapped by Nathan, and the major units are shown on the accompanying generalized map (fig. V-32) and on the sections in Figures V-33 and V-34. Several types of gabbroic, intermediate, and oxide-rich rocks are present; modes of the units and component mineral compositions, determined by microprobe, are given in Tables V-15 and V-16. From oldest to youngest, the sequence of intrusions, as defined by Nathan, begins with A, and extends through the alphabet to AA, the youngest. Evidence for his interpretation of the age relations of the units within the sequence are given by Nathan

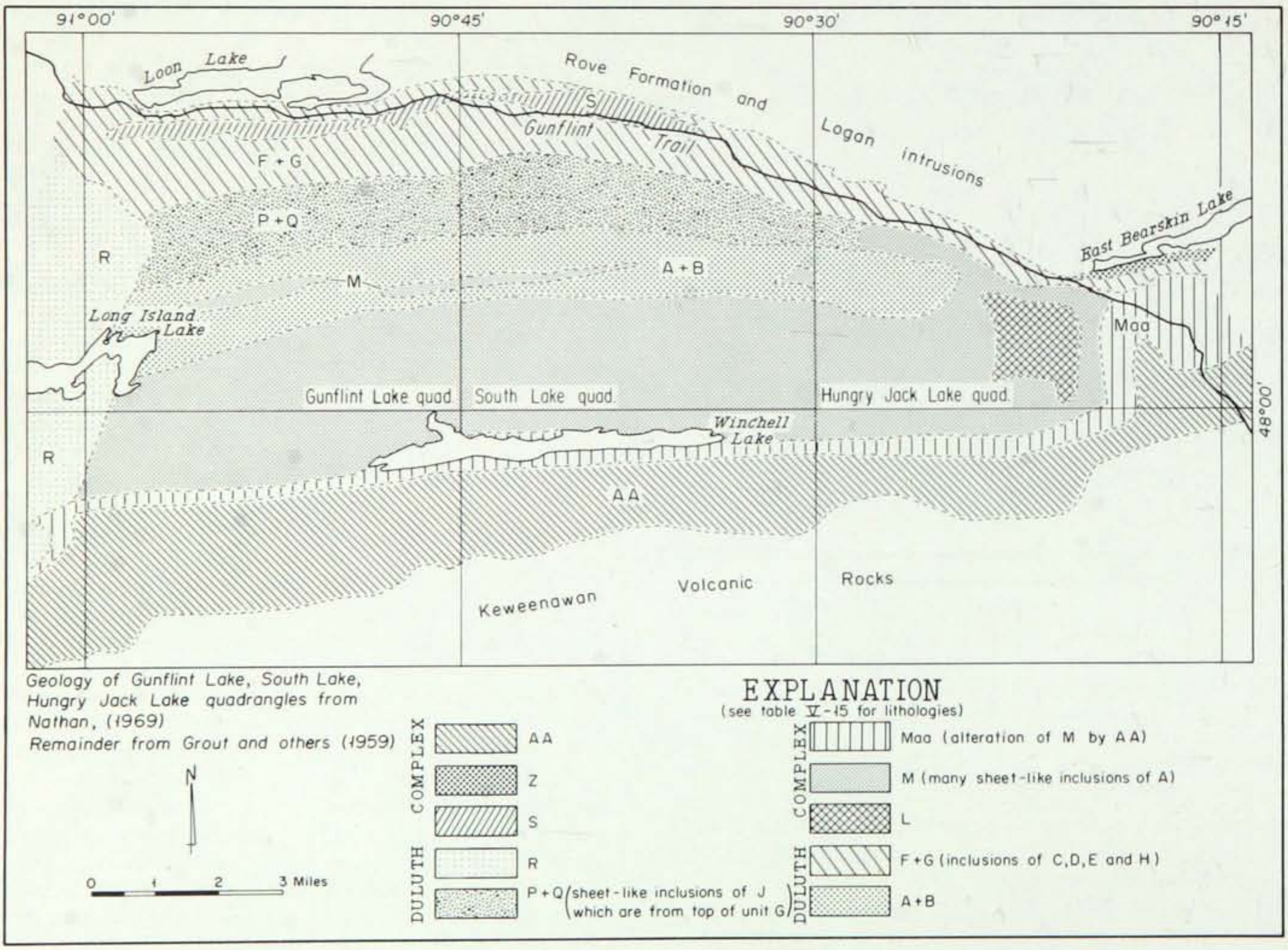

Figure V-32. Generalized geologic map of Duluth Complex in western half of northern prong (after H. L. Nathan, 1969, unpub. Ph.D. thesis, Univ. Minn.).

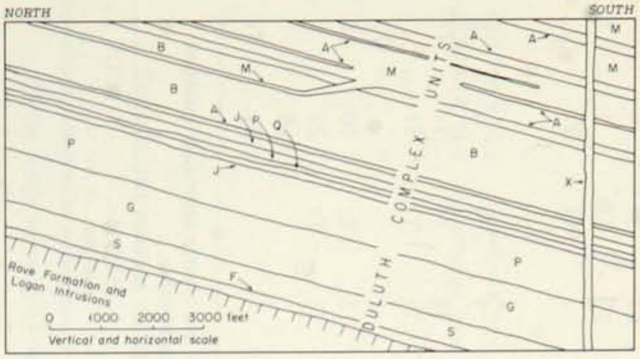

Figure V-33. Diagrammatic geologic section of south-central part of northern prong (after H. L. Nathan, 1969, unpub. Ph.D. thesis, Univ. Minn.).

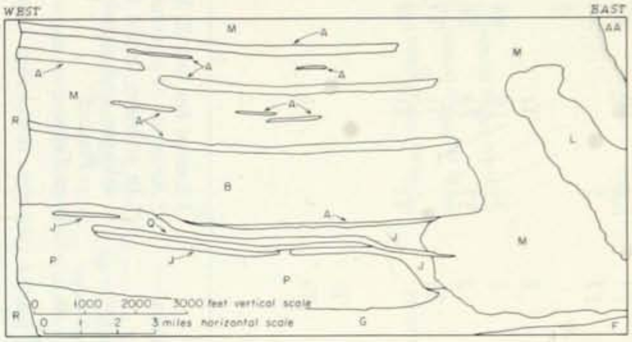

Figure V-34. Diagrammatic geologic section across southern margin of western half of northern prong.

(1969, *op. cit.*). Criteria that were used to define the sequence include inclusions, crosscutting relations, alteration, and fine-grained margins. The intrusive rocks form a more or less regular stack of tabular units, or sheets, that are inclined about 15° S. Some of the sheets of unit M are interpreted as apophyses of a large, generally discordant intrusive center in the Hungry Jack Lake quadrangle. The sheets are interrupted on the west by a younger anorthositic intrusion (unit R). Several minor stocks and dikes occur throughout the area. These include units described in Table V-15 (plus unit K) but not mentioned in the explanation of Figure V-32.

Several of the sheets have a mineralogic layering that is parallel to contacts of the individual sheets. The presence of density-sorted layers indicates that crystal settling was operative. Foliation of plagioclase tablets and oriented elongate olivine in the plane of foliation, which is parallel to the mineralogic layering, also suggest crystal settling. Plagioclase warping is common in some units. Although some of the slender laths are broken, most have a hinge zone in which curved, tapering twin lamellae interdigitate. Generally, the hinge zone is at a fulcrum formed by the corner of an underlying grain. Only the foliated rocks show this feature. Again, it is interpreted as a crystal settling phenomenon caused by the accumulating load on the grains in a mush on the floor of the magma. At the high temperatures involved, the slender plagioclase crystals could yield plastically around the fulcrum under the weight of the overlying crystals.

The interpretation from the field data that the rocks in this area are older than the main mass of the Duluth Complex to the west is supported by the paleomagnetic data of Beck (1970) and Beck and Lindsley (1969), who found that all samples from this area have reversed polarization whereas all samples from other parts of the Duluth Complex show normal polarization. By analogy with previous paleomagnetic data on some of the underlying sills (Logan intrusions) that show reversed polarization, Beck interpreted the reversely polarized rocks as being older than the main mass of the complex. This interpretation is discussed more fully elsewhere in this chapter by Green.

Along the southern margin and also in the eastern part of this area are several granitic intrusions (units Z and AA in table V-15), generally distinguished by their light-red to brick-red color. Although these may be late-stage differentiates of unit Y, they show intrusive relations with many older units through unit X. The intrusions were accompanied by H_2O-rich fluids, which severely modified the adjacent rocks (unit distinguished as Maa in figure V-32). Similar alteration occurs along fractures several hundred yards from the intrusions. Generally, the alteration consists of development of hornblende, commonly in the form of rosettes, sericitization or saussuritization of plagioclase, and introduction of quartz and alkali feldspar.

Units G, J, T, and V contain coexisting magnetite and ilmenite. Microprobe analyses of these phases allow estimating, by the method of Buddington and Lindsley (1964), the last temperature and oxygen fugacity conditions at which these rocks equilibrated. Table V-17 contains data for 16 samples analyzed by Nathan (1969, *op. cit.*). Except for unit T, the temperatures are significantly below the lowest temperatures normally assumed for basaltic melts. As shown in Figure V-35, there is a quite regular trend of oxygen fugacity with lowering of temperature. Because the temperatures of equilibration are spread over nearly 300° C in a rather regular way, there must be some variable controlling the minimum temperature to which this equilibration can continue. Figure V-36 indicates a very close relationship between the ulvospinel component of the magnetite and lowest temperature of equilibration; namely, the greater the iron content of the spinel, the lower the temperature to which equilibration will occur between ilmenite and titaniferous magnetite.

At the midpoint of the northern prong, east of Nathan's mapping, the width of the prong narrows rather abruptly from about 6 miles to slightly more than 2 miles. Also, the nature of the rocks changes from the series of sheets in the western half to a gradational sequence of gabbro through intermediate rock to granite in the eastern half. Although no detailed mapping of the units is available, petrographic data from four traverses across the intrusion are available (Babcock, 1959, *op. cit.*). The locations of the traverses are

Table V-15. Modes of units in Gunflint, South Lake, and Hungry Jack Lake quadrangles (after Nathan, 1969, unpub. Ph.D. thesis, Univ. Minn.).

	A (2 facies)	A	B*	C**	D***	E	F	G (upper)	G (lenses)	G (middle)	G (lower)	H	I	J (upper)	J (lower)	L†
Plagioclase	63	42	59	61	56	53	68	38	27	62	29	66	45	83	67	60
Augite	3	2	15	27	19	41	10	26	5	11	18	5	28	6	17	20
Hypersthene														Tr	Tr	Tr
Pigeonite	9	3	26	Tr		Tr	5					4		4	1	17
Olivine	25	53	Tr at base	10	14	Tr	10	15	4	5	24	20	2			
Opaques	Tr	Tr	Tr	1	11	6	7	16	64	14	15	4	23	4	15	1
Apatite	Tr	Tr	Tr	Tr		Tr	Tr	5	Tr	Tr	Tr	1	2	Tr	Tr	Tr
Biotite‡				Tr			Tr					Tr				Tr
Alkali feldspar							Tr							2		1
Quartz			Tr at top				Tr						Tr in 1 sec.	1	Tr	1
Hornblende‡				Tr										Tr	Tr	
Orthoclase																
Zircon																

	M	N††	O	P	Q	R	R†††	S	T	T	U	V‡‡	W	X	Y‡‡‡	Z	AA
Plagioclase	59	41	Pr	58	83	75	80	77	2	10	21	8	55	66	53	34	
Augite	24	40	Pr	32	Tr	15		17	5		4	44	35	22	31	3	3
Hypersthene	Tr			7							2	Tr			11	Tr	
Pigeonite	13			Tr	Tr			2		Tr		24					
Olivine	Pr in some	Pr in some		3	17	5	5	1	36	16	49	1		2			
Opaques	2	19	Tr	Tr	Tr	5		3	57	73	24	23	10	7	5	9	2
Apatite	Tr		Tr					Tr		Tr	Tr	Tr		Tr	Tr	Tr	Tr
Biotite‡									Tr	Tr	Tr	Tr		Tr	Tr	Tr	
Alkali feldspar	1		Pr											1			77
Quartz	1		Pr					Tr						2	Tr	10	18
Hornblende‡			Tr					Tr		1		Tr	Tr	Tr	Tr	29	
Orthoclase																15	
Zircon																Tr	Tr

* Augite > pigeonite in lower half of unit; pigeonite > augite in upper half

** Mode is of most common facies; other facies have various ferromagnesian values

*** Ferromagnesian minerals vary but remain between 10 and 20 percent

† Olivine + pigeonite in some, olivine but no pigeonite in some, pigeonite but no olivine in some

†† Modes variable in this unit

††† 15 percent very fine groundmass of plagioclase, augite, hypersthene, opaques

‡ Except for unit Z, biotite and hornblende are clearly alteration products

‡‡ Amounts of pigeonite, olivine, and opaques variable

‡‡‡ In addition to the listed minerals, one section contains zircon, alkali feldspar, and abundant quartz

Pr = Present

Tr = Trace

Table V-16. Compositions of mineral phases in units of Gunflint Lake, South Lake, and Hungry Jack Lake quadrangles (after Nathan, 1969, unpub. Ph.D. thesis, Univ. Minn.).

	Plagioclase	Olivine	Augite (Mg/Mg + Fe)	Pigeonite (Mg/Mg + Fe)	Titanomagnetite	Ilmenite $FeTiO_3$	Ilmenite $MgTiO_3$	Ilmenite Fe_2O_3
A	An_{61-64} upper An_{58-61} lower An_{59} mean	Fo_{63}						
B	An_{56-59} An_{57} mean		69					
C	An_{47-64} An_{56} mean	Fo_{50}						
D	An_{45-61} An_{51} mean							
E	An_{54} mean							
F	An_{52-57} fine fraction An_{57-60} coarse An_{56} mean	Fo_{29} fine Fo_{42} coarse	59 fine 66 coarse			83 99	10 1	7 0
G	An_{47-53} upper An_{50} mean	Fo_{36-50}, Fo_{43} mean(4)	62-64, mean 63(4)	53 and 61	usp_{22-44}, usp_{36} mean(4)	80	12	8
G	An_{49-56} middle An_{53} mean	Fo_{36-49}, Fo_{45} mean(13)	66-68, mean 67(4)	60	usp_{20-60}, usp_{42} mean(16)	79-86, 82 mean(9)	8-12, 10 mean(9)	5-10, 8 mean(9)
G	An_{50-55} lower An_{54} mean	Fo_{40-51}, Fo_{45} mean(5)	62 and 63		usp_{27-46}, usp_{33} mean(7)	75-96, 86 mean(9)	0-16, 8 mean(9)	4-10, 6 mean(9)
H	An_{51-54} An_{52} mean							
I	An_{52} mean		52					
J	An_{57} mean upper An_{50-59} An_{52} mean lower		61	52	usp_{28}	92	2	6
L	An_{57-59} An_{58} mean		59	51				
M	An_{55-63} An_{50} mean	Fo_{22} Fo_{55}	43 55 59	45 53				
N	An_{52} mean		71		usp_{86}			
P	An_{54-60} An_{68} mean	Fo_{60}	71	Hypersthene 66				
Q	An_{51-53} An_{52} mean	Fo_{63}						
R	An_{63}							
S	An_{52-59} An_{55} mean	Fo_{38}	58					
T	An_{44-63}	Fo_{52-62}, Fo_{57} ave	67		$usp_{58, 60, 64}$	71 76	21 16	8 8
U	An_{52}							
V	An_{61}	Fo_{48} Fo_{45}	69 66		usp_{33}	85	9	6
W	An_{56}							
X	An_{58-62} An_{60} mean							
Y	An_{47}							
Z	An_{60}							
AA	Intergrown alk fsp Or_{25} and Or_{47}							

Table V-17. Temperature-oxygen fugacity data from coexisting magnetite-ilmenite in units of northern prong of Duluth Complex (after Nathan, 1969, unpub. Ph.D. thesis, Univ. Minn.).

Unit	Sample	fo_2 as $-\log_{10}$	T°C	Mag % usp	Ilm FeTiO$_3$	Ilm MgTiO$_3$	Ilm Fe$_2$O$_3$
Gu	66HN263B	13	880	44	80	12	8
Gm	66HN293B	13	880	58	86	9	5
Gm	67HN829A	15	780	33	85	8	7
Gm	67HN830	13	870	43	81	12	7
Gm	67HN827	13	870	41	80	12	8
Gm	67HN826	13	830	39	83	9	8
Gm	67HN824	14	810	35	82	10	8
Gm	67HN824	16	760	32	85	9	6
Gl	67HN818	18	680	27	91	5	4
Gl	67HN818(incl)	18	680	27	94	2	4
Gl	67HN816	16	750	31	87	7	6
Gl	67HN814A	14	800	37	79	14	7
J	66HN575	17	720	28	92	2	6
T	67HN488A	11	1020	60	71	21	8
	67HN668C	11	980	58	76	16	8
V	67HN739	16	760	33	85	9	6

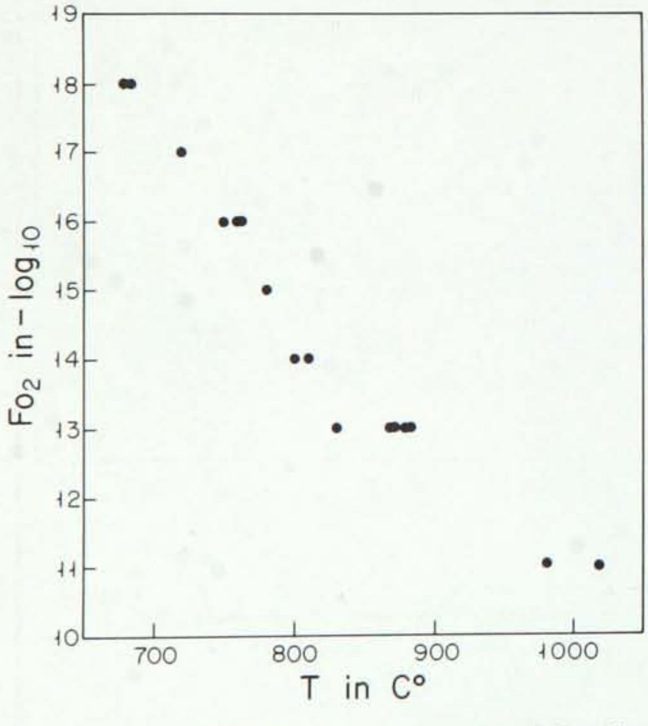

Figure V-35. Temperature-oxygen fugacity relationships from ilmenite-magnetite pairs, northern prong (data from H. L. Nathan, 1969, unpub. Ph.D. thesis, Univ. Minn.).

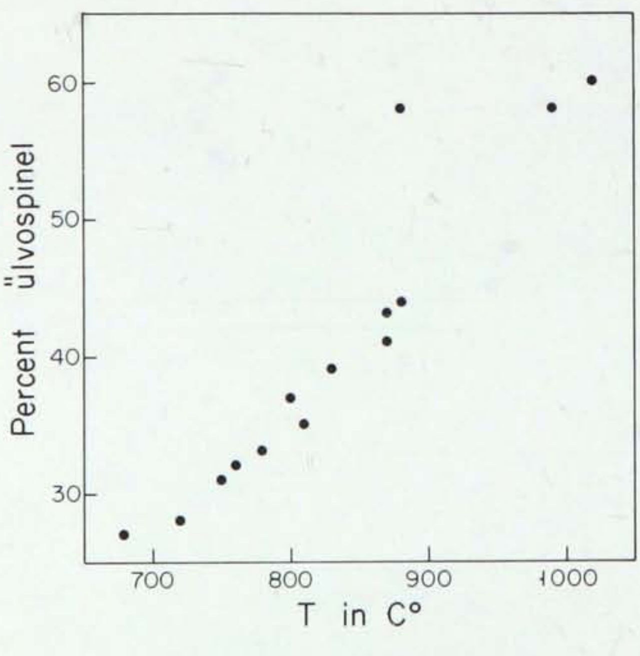

Figure V-36. Relationship of ilmenite-magnetite temperature and magnetite composition, northern prong (data from H. L. Nathan, 1969, unpub. Ph.D. thesis, Univ. Minn.).

shown on Figure V-37, and modal variations of the rocks are given in Tables V-18, V-19, V-20, and V-21. Babcock reported inclusions of gabbroic and intermediate rocks in the granite. Plagioclase compositions in the gabbroic rocks are reported as An_{45-55} in the centers of grains. "Fluxion structure" is mentioned as being parallel to the base of the complex, but specific measurements are not given. This feature plus the stratigraphic sequence of normally early-formed minerals at the base grading upward to the later-formed minerals seem to indicate crystal settling.

From these data Babcock (1959, *op. cit.*) concluded, "The presence of two distinct rock types, gabbro and granophyre, within the northeastern projection of the Duluth Complex is thought to be a result of fractional crystallization and differentiation through gravity settling and structural activity. Upon emplacement of magma the gabbroic minerals, plagioclase and pyroxene, crystallized; and, due to their greater density, accumulated in the lower portions of the magma chamber where they were knit together by continued crystallization. The liquid which remained in the interstices reacted slightly with the crystalline phase and then solidified in the form of intergrown quartz and potassium feldspar. The resulting rock is a gabbro with minor amounts of interstitial granophyre.

"The intermediate rock represents the gradational separation of gabbro and granophyre. Upward from the gabbro the amount of interstitial granophyre and alteration of mafic minerals increases, forming a rock which has a diabasic texture as does the gabbro but which contains abundant interstitial granophyre. As the granophyre becomes more abundant, the corresponding decrease in gabbroic minerals causes the diabasic texture to disappear and the rock appears as a mafic granophyre.

"The essentially complete crystallization and accumulation of the gabbroic constituents caused the residual liquid to become more acidic in composition and to crystallize as a granophyre. The texture of the granophyre is the result of crystallization of intergrown quartz and potassium feldspar from the liquid surrounding scattered euhedral plagioclase crystals. The mafic minerals which are disseminated throughout this rock are fine, scattered alteration products of pyroxene crystals which formed earlier."

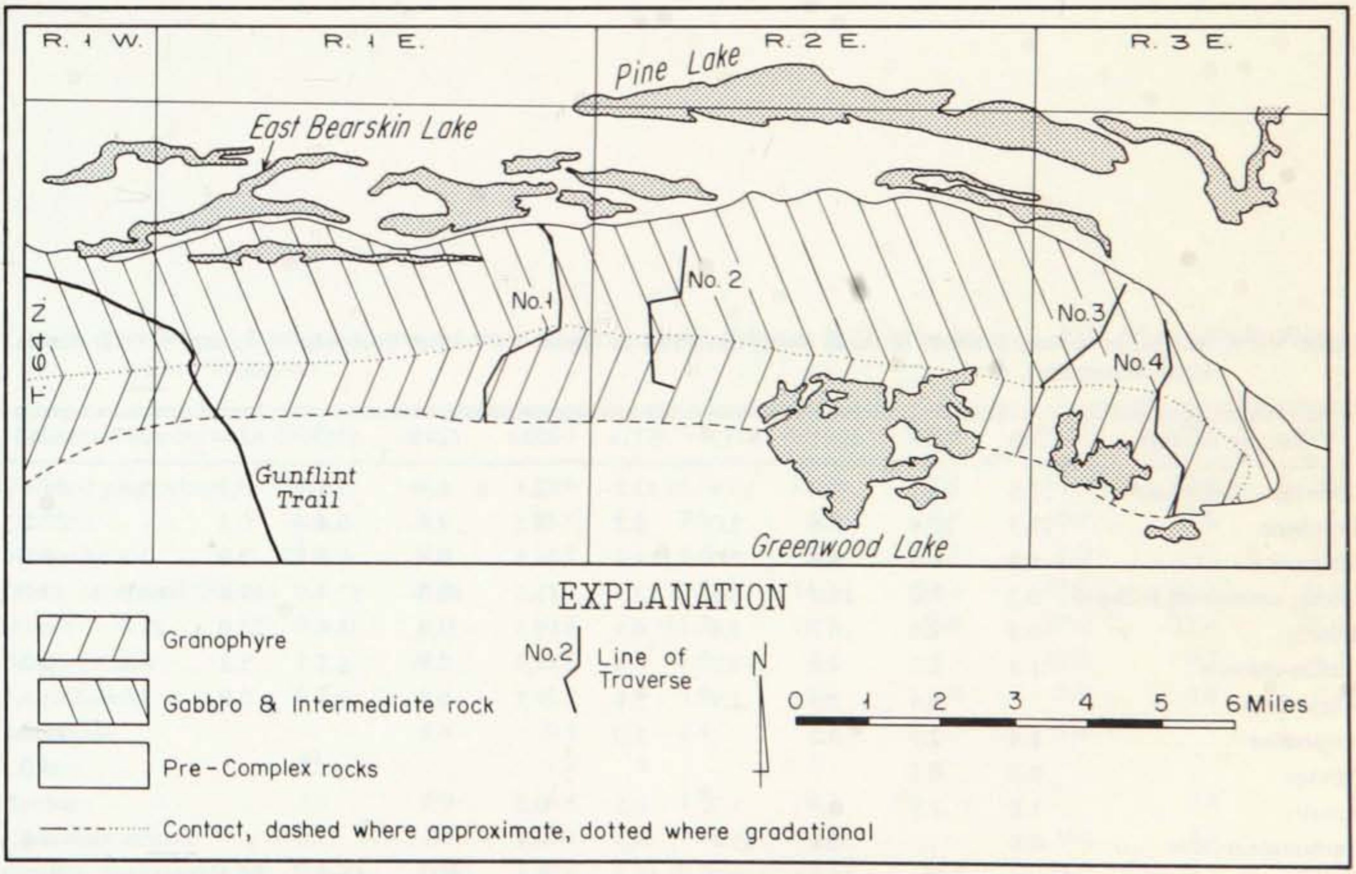

Figure V-37. Geologic map of eastern half of northern prong of Duluth Complex (after R. C. Babcock, 1959, unpub. Ph.D. dissert., Univ. Wisconsin).

Table V-18. Modes, in volume percent, of rocks in northern prong of Duluth Complex, traverse 1 of Figure V-37 (after Babcock, 1959, unpub. Ph.D. thesis, Univ. Wisconsin).

Sample number	7014	7012	5910	5909	5908	5907	5905	5904	5903	5276	5279	5162-a
Twinned plagioclase	32.3	42.8	50.1	45.0	36.2	35.5	20.9		1.0	1.3		0.2
Pyroxene	39.4	27.4	20.2	19.7	21.2	10.3	5.9	4.0	1.4	3.1	1.4	4.6
Magnetite	14.2	12.0	6.8	8.5	6.7	5.4	5.6	1.9	2.1	2.3	1.7	2.1
Dusty, untwinned feldspar	9.5		3.5	4.6	11.8	10.5	28.1	60.7	66.3	69.5	72.4	72.9
Quartz	1.4	0.3	1.9	4.3	3.9	6.7	11.8	15.5	15.7	17.4	23.5	18.5
Biotite-chlorite	0.6		0.8	1.7	1.1	8.6	19.5	16.8	10.6	6.3		0.5
Amphibole	0.6	0.8	3.7	3.4	5.0	2.4	1.4	0.5	0.5			
Serpentine	1.0	0.3	7.4	5.6	7.7	10.1	2.0		1.4			1.0
Olivine		12.2	0.7	0.5	1.4	3.0						
Apatite	1.1	3.6	0.5	3.2	3.4	1.6	0.2	0.4				
Carbonate-sericite		0.5	0.5	1.0	0.4	2.3	1.2	0.2	0.6			
Graphic intergrowth (quartz-feldspar)	0.6		8.2	3.4	3.4	13.6	21.5	7.1	42.4	54.9	59.6	62.3

Table V-19. Modes, in volume percent, of rocks along traverse 2 of Figure V-37 (after Babcock, 1959, unpub. Ph.D. thesis, Univ. Wisconsin).

Sample number	6269	6268	6267	5717	5716	6234	6242	6243	6244	6247	6246
Twinned plagioclase	57.6	34.1	51.8	53.4	42.8	32.4	4.9	5.8	0.2	0.4	51.6
Pyroxene	25.5	27.4	8.0	2.0	8.2	4.5	2.4	0.4	0.2		22.2
Magnetite	4.9	8.8	8.8	2.0	11.0	4.5	4.4	1.3	2.0	3.1	3.5
Dusty, untwinned feldspar	0.2	3.0	15.7	6.6	12.3	13.6	62.5	57.4	67.8	66.5	0.7
Quartz	0.5	2.5	6.7	3.4	6.3	11.3	23.4	28.0	23.0	21.4	
Biotite-chlorite	1.4	2.2	4.6	7.3	8.4	14.9	1.9	6.3	5.8	3.8	0.2
Amphibole		3.6	2.9	2.9	5.1	9.3	0.5	0.6	0.9	5.0	
Serpentine	1.4	2.7	0.2		1.0						15.8
Olivine	0.2	0.2									5.6
Apatite	1.8	2.9	0.5	1.2	0.5	0.2					
Carbonate-sericite	0.6		0.8	13.4	4.3	4.1					0.4
Graphic intergrowth (quartz-feldspar)			19.8	9.2	18.3	28.1	61.1	66.0	66.4		

Table V-20. Modes, in volume percent, of rocks along traverse 3 of Figure V-37 (after Babcock, 1959, unpub. Ph.D. thesis, Univ. Wisconsin).

Sample number	7511	7507	7514	7515	5711	5706
Twinned plagioclase	47.8	49.0	43.4	27.2	17.3	17.4
Pyroxene	24.8	19.6	6.0	14.6	3.4	2.3
Magnetite	4.2	6.2	9.6	3.1	4.8	4.1
Dusty, untwinned feldspar	4.4	5.5	5.2	24.3	38.3	47.4
Quartz	3.6	1.4	5.6	11.3	6.6	21.9
Biotite-chlorite	11.8	11.6	16.6	12.8	12.5	5.2
Amphibole	1.4	1.6	10.0	3.5	16.3	1.4
Serpentine	0.6			1.2		
Olivine		1.2		0.8		
Apatite	1.4	3.9	1.8		0.8	
Carbonate-sericite			1.8	1.4		
Graphic intergrowth (quartz-feldspar)	6.2	2.2	9.0	32.2	29.5	53.5

Table V-21. Modes, in volume percent, of rocks along traverse 4 of Figure V-37 (after Babcock, 1959, unpub. Ph.D. thesis, Univ. Wisconsin).

Sample number	6907	6197	6196	6195	6194	6910	7521	291	290
Twinned plagioclase	34.4	46.6	35.6	26.6	13.6	7.1	12.4	14.8	9.9
Pyroxene	36.6	15.4	15.6	7.9	2.3	0.2	0.6	0.2	
Magnetite	18.0	4.0	4.7	5.8	1.3	1.3	1.0	1.2	0.9
Dusty, untwinned feldspar	0.8	10.9	21.5	20.5	54.2	56.3	47.0	48.7	51.0
Quartz	0.3	4.0	10.7	11.5	21.0	25.8	22.0	23.4	28.4
Biotite-chlorite	5.3	10.9	2.8	13.9	4.4	5.4	12.2	9.7	4.2
Amphibole	0.2	3.2	1.2	7.9	2.6	3.0	2.9	0.8	0.4
Serpentine		0.8	0.5	2.4			1.8		
Olivine	0.5		1.9						
Apatite	3.8	3.0	5.1	1.0	0.2			0.2	
Carbonate-sericite	0.2	1.4	0.4	2.6	0.5		0.2	0.2	
Graphic intergrowth (quartz-feldspar)	0.3	9.9	29.4	30.4	27.4	56.1	54.2	54.1	69.2

EASTERN PART OF DULUTH COMPLEX

Donald M. Davidson, Jr.

The eastern part of the Duluth Complex, as defined here, includes the southern (or Brule River) prong (see Phinney, this chapter, fig. V-24) and areas immediately to the west. In this area, the major rocks of the complex are anorthositic rocks, troctolite and olivine gabbro, and felsic intrusive rocks. Inclusions of older rocks are common, especially in the western part of the area. Knowledge of the geology of the area is based largely on the early mapping by Grout and others (1959), my reconnaissance geologic mapping of several quadrangles from 1966 to 1970 (fig. V-38), and relatively detailed mapping of selected quadrangles (Perent Lake and Kawishiwi Lake, Davidson, 1969a and b; Long Island Lake, G. B. Morey and others, 1969, open-file map, Minn. Geol. Survey; and Gillis Lake, Weiblen and Beitsch, unpub. map). The rock terminology used in this section is the same as that described by Phinney (fig. V-25). The term "granophyric" is used to denote a rock texture; "granofels" is used rather than "hornfels" for the metamorphosed volcanic rocks included within the complex because of their medium-grain size.

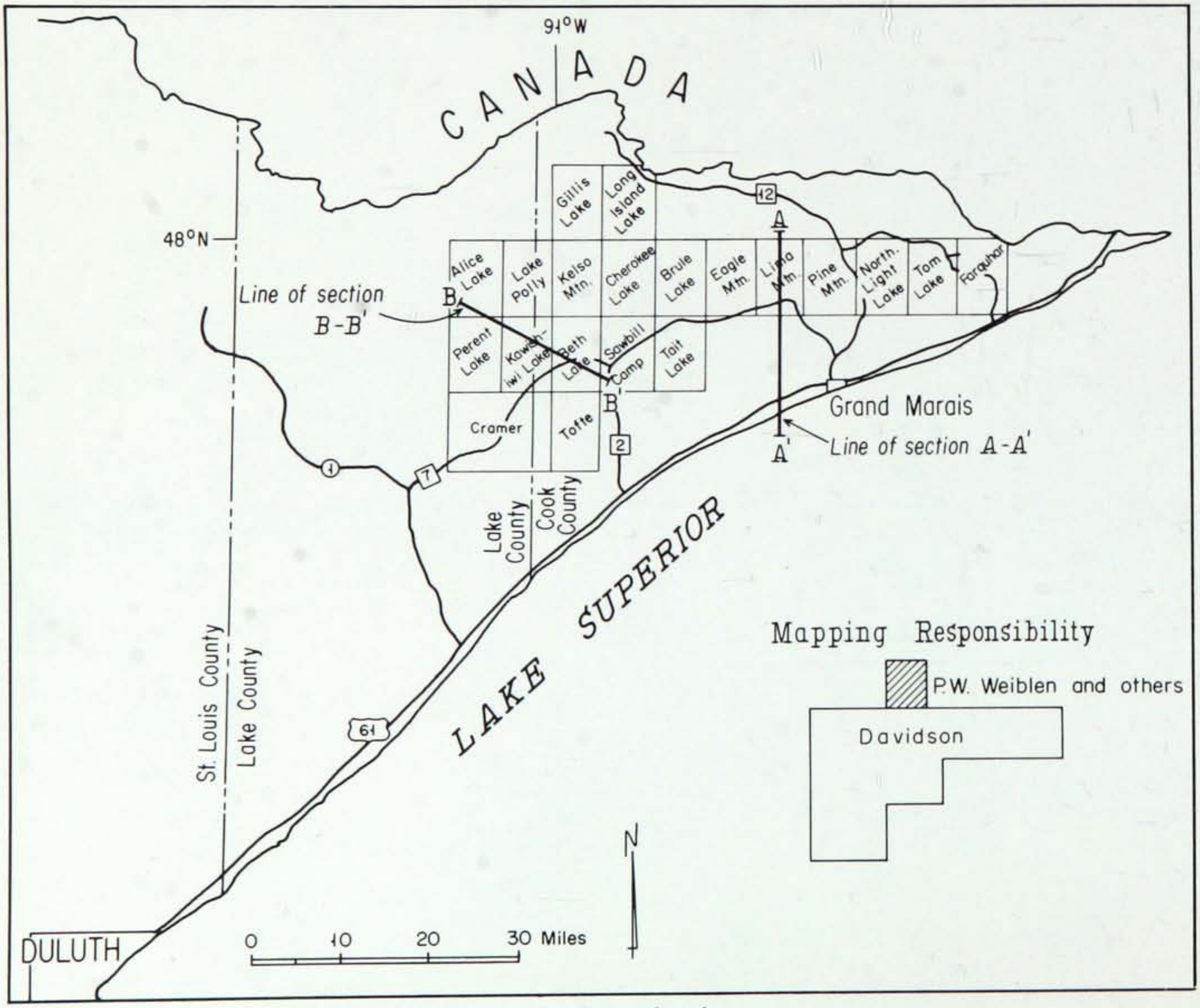

Figure V-38. Map of eastern part of Duluth Complex, showing quadrangles.

ROCK UNITS

Except for the older, complex series of thin intrusive sheets along the northern margin of the area, described in the preceding section by Phinney (this chapter), the oldest rocks of the complex in this area are anorthositic rocks. These are intruded by troctolite and gabbro, which in turn are cut by the felsic rocks. For convenience of discussion, these three rock types are referred to in this paper as anorthositic series, troctolite-olivine gabbro series, and felsic series.

ANORTHOSITIC SERIES

Nearly half of the bedrock in the eastern part of the Duluth Complex consists of anorthositic rocks. They comprise an area of about 300 square miles along the western margin, particularly west of longitude 91° W., and form a thin strip along the southern margin of the northern prong (fig. V-39). In addition, anorthositic rocks probably occur in the eastern extremity of the southern prong, south and east of the center of the Pine Mountain quadrangle (fig. V-38). This body may be continuous with the Reservation River diabase unit of Grout and others (1959), which has a similar lithology. Areas containing anorthositic rocks generally are characterized by low relief between altitudes of 1,500 and 1,800 feet, poor bedrock exposures, and highly weathered rocks.

The rocks are coarse grained, locally foliated, and contain 85-90 percent plagioclase (An_{48-70}; typically An_{55-60}). They are predominantly gabbroic anorthosite, but include troctolitic and oxide-rich units (tables V-22 and V-23). Augite and oxide minerals generally occur interstitially in poikilitic textures, whereas olivine generally is a cumulus mineral. Orthopyroxene occurs in symplectic intergrowths with late-stage plagioclase; biotite is associated with the oxide minerals.

As noted by Bonnichsen and Phinney elsewhere in this chapter, inclusions of relatively pure anorthosite (>95 percent An_{50-60}) are present in the units of the anorthositic

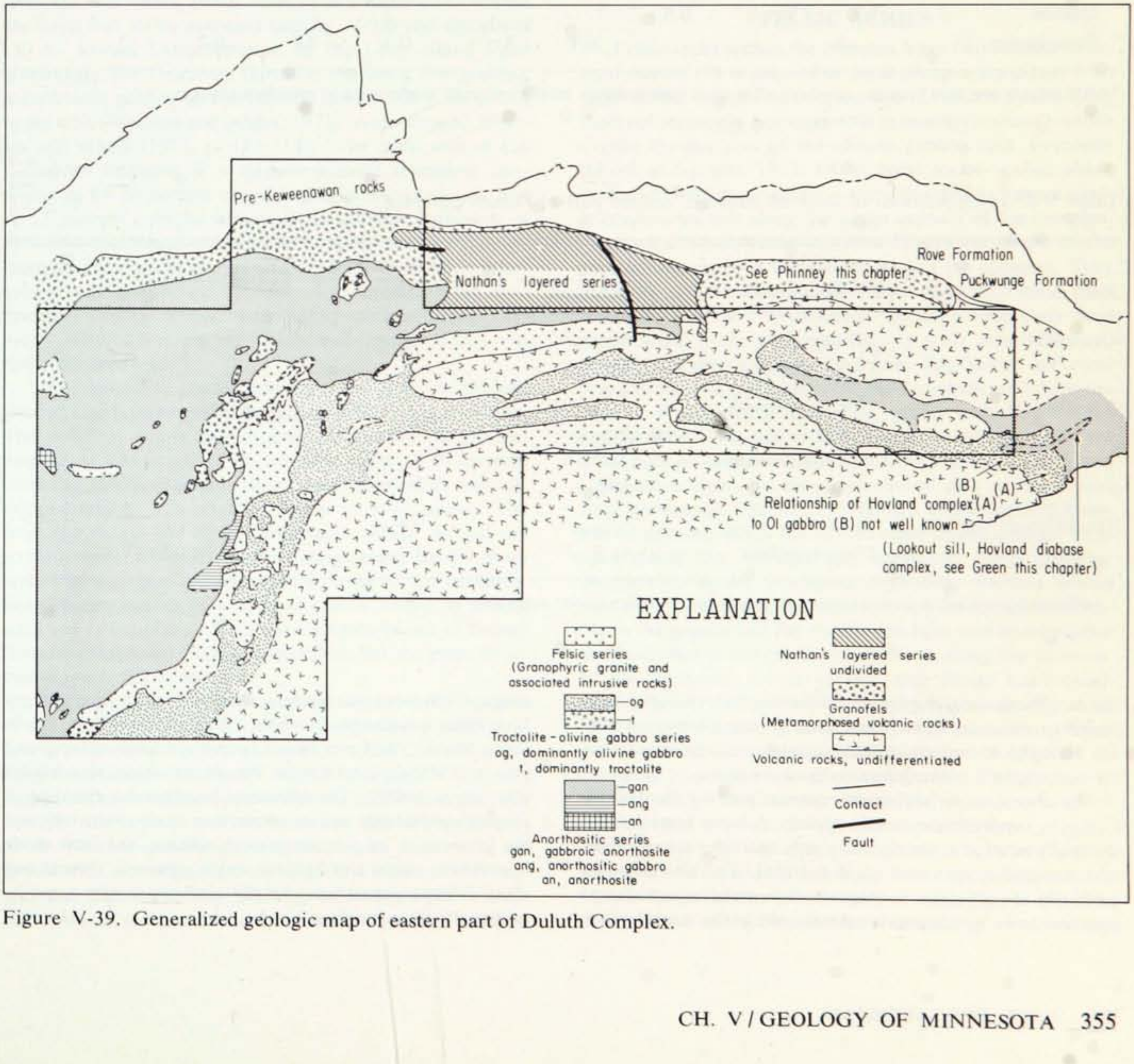

Figure V-39. Generalized geologic map of eastern part of Duluth Complex.

Table V-22. Modes, in volume percent, of major rock units in eastern part of Duluth Complex.[1]

Rock series	Granofels		Anorthositic			Troctolite-olivine gabbro				Felsic
Rock type	Micro-grabbo (4)	Quartz diorite (2)	Anortho-site (4)	Gabbroic anortho-site (27)	Tracto-litic anortho-site (5)	Tracto-lite	Olivine gabbro (31)	Oxide rocks (2)		Grano-phyric granite (10)
Plagioclase	58.7	39.5	93.7	86.2	87.7	77.2	64.4	58.3	23.3	18.3
K-spar		2.5								44.7
Clinopyroxene										
Augite	12.3		4.2	6.8	1.0	2.0	11.7	26.1	4.3	
Pigeonite	6.9									
Orthopyroxene	0.7			1.1	1.0	1.5	0.3			
Olivine			Tr	3.1	7.7	16.1	15.0	8.4		
Opaque oxides	8.3	8.6	1.0	1.9	1.2	1.8	2.6	6.8	72.4	5.4
Hornblende		24.6								
Quartz		15.3	0.5							31.6
Biotite	5.2	9.3	Tr	0.6	1.3	1.1	1.4	0.2		
Uralite	1.9	Tr	0.6							
Symplectite				0.3	Tr	0.3	4.1	0.2		

[1] 1500 counts per thin section; numbers after each unit indicate total number of modes used to calculate average

Table V-23. Compositions of selected minerals, eastern part of Duluth Complex.

Rock series	Rock type	Compositions: Plagioclase	Olivine	Augite
Granofels	Microgabbro	An_{53-58}[1]		
	Quartz diorite	An_{36-48}[1]		
Anorthositic	Gabbroic anorthosite	An_{52-65}	Fo_{55-57}	
	Troctolitic anorthosite	An_{48-69}	Fo_{56-57}	$En_{38}Fs_{22}Wo_{40}$
Troctolite-olivine gabbro	Troctolite	An_{55-60}	Fo_{50}	
	Olivine gabbro	An_{55-70}[1]	Fo_{28-55}	$En_{38}Fs_{22}Wo_{40}$

[1] Analyses by optical determinations; all others are by microprobe analysis

series. These coarse-grained inclusions are foliated and range in diameter from one meter to one kilometer; they are thought to represent early crystal accumulations that formed within a convecting anorthosite magma.

In several areas within the eastern part of the Duluth Complex, anorthositic rocks appear to have been hydrothermally altered contemporaneously with the introduction of interstitial quartz and alkali feldspar. The altered areas generally are adjacent to felsic intrusive bodies, which appear to have gradational contacts with the anorthositic rocks. Such occurrences have been noted in the Hungry Jack Lake quadrangle (Nathan, 1969, unpub. Ph.D. thesis, Univ. Minn.), the Long Island Lake quadrangle (Morey and others, 1969, *op. cit.*), and the Perent Lake quadrangle (Davidson, 1969b). The alteration involves the clouding of plagioclase feldspar and its conversion to clay minerals, and the conversion of ferromagnesian silicate and iron oxide minerals to micas and hydrous oxide minerals. Quartz and alkali feldspar occur interstitially, and apparently were introduced during the alteration.

In the Perent Lake quadrangle (Davidson, 1969b), an area of pegmatitic gabbroic anorthosite with local concentrations of micronorite and microgabbro occurs adjacent to a large felsic intrusive body. The pegmatitic material forms dike-like, irregular patches 10 to 30 meters across. Augite crystals more than 4 centimeters in length have been noted, and plagioclase laths 3 to 4 centimeters long are common.

TROCTOLITE-OLIVINE GABBRO SERIES

Olivine-bearing rock units occur at both the lower and upper margins of the Duluth Complex in this area (fig. V-39). As yet, the genetic relationships between these units have not been determined, but both appear to be younger than the rocks of the anorthositic series.

As noted by Phinney (this chapter), the troctolitic rocks along the base of the complex form a zone 2.5-3.5 km wide that extends nearly continuously from Duluth northeastward to the eastern margin of the Long Island Lake quadrangle. In the Gillis Lake and Long Island Lake quadrangles, this zone, referred to informally as the Tuscarora intrusion (Weiblen and others, 1971), consists of a sequence of troctolite units that strike eastward (see fig. V-39) and dip about 15° S., toward Lake Superior. In the Long Island Lake quadrangle, the Tuscarora intrusion overlies a fine-grained, granoblastic gabbro (granofels) and consists of a lower and upper unit (Weiblen and others, 1971). According to Weiblen and others (1971, p. 113-114), "The main unit of the Tuscarora Intrusion is a medium-grained troctolite, consisting of 65-70 percent cumulus plagioclase (An_{55-60}), and 10-15 percent cumulus olivine (Fo_{50}). Relative amounts of poikilitic augite and iron-titanium oxides vary locally. Orthopyroxene mantles olivine and occurs in symplectic intergrowth with plagioclase. Biotite is associated with the iron-titanium oxides. Planar orientation of plagioclase and modal-mineral layering are locally well-developed and mutually concordant.

"The troctolite grades into an upper unit which consists of interlayered poikilitic augite gabbro and troctolite. The poikilitic augite gabbro consists of about 70 percent plagioclase (An_{50-60}), 15-20 percent augite, 5-10 percent ilmenite, and is medium- to coarse-grained with well developed augite orthocrysts as much as $1\frac{1}{2}''$ across. The troctolite within the layered interval is similar to that described above. Contacts between layers are generally sharp and in general conformable with layering in the troctolite. Interlayering occurs on a scale of several inches to several feet, and is undulatory with wave lengths of ten to twenty feet and amplitudes of two to three feet, but the gross structure is nearly flat-lying."

The upper unit, along the eastern and southern margins of the Duluth Complex, is dominantly olivine gabbro; it forms an arcuate outcrop pattern from the Cramer quadrangle northeastward nearly to Hovland (fig. V-39), and constitutes most of the southern prong of the complex. Although little is known of the extent of this unit southwest of the Cramer quadrangle, reconnaissance field work together with aeromagnetic data (Zietz and Kirby, 1970) suggest that rocks of similar lithology (olivine gabbro or troctolite) extend southwestward from the Cramer quadrangle as far as the Beaver Bay Complex, described earlier by Gehman (1957, unpub. Ph.D. thesis, Univ. Minn.).

Although olivine gabbro is the dominant unit, variations in the amount of augite or oxide minerals locally give rise to other units—ophitic gabbro, troctolite, or olivine-oxide rock. Troctolite layers about one meter thick occur locally within the olivine gabbro, and large lens-like concentrations of iron-titanium oxide as much as two meters thick and several meters long have been observed (Grout and others, 1959; Grout, 1950). The abundant oxide minerals distinguish this unit from the northern (lower) troctolitic unit.

Mineralogically, the gabbroic rocks consist, in order of decreasing abundance, of cumulus euhedral plagioclase, poikilitic subhedral augite, cumulus anhedral olivine, and cumulus subhedral oxides (table V-22). Metamorphosed inclusions of various rock types occur in the gabbroic rocks, and include Keweenawan volcanic rocks, anorthositic rocks and, at one locality (sec. 25, T. 63 N., R. 4 W., Cherokee Lake quadrangle), possible Virginia Formation.

FELSIC SERIES

Felsic rocks within the complex have two distinct structural habits: (1) small, rather local plutons associated with rocks of the anorthositic series; and (2) relatively flat-lying sheets of seemingly homogeneous granophyric granite which overlie the margins of the olivine gabbro unit. Formerly (Grout and others, 1932, 1959), these rocks—called either "red rock" or "granophyre"—were thought to form a nearly continuous belt along the upper contact of the complex.

Several small, felsic intrusive bodies cut rocks of the anorthositic series in the eastern part of the complex. They generally lack small-scale structures other than local flow lineation and have gradational contacts with their host rocks. Small dikes, a few centimeters wide, however, commonly have distinct chilled borders. The bodies are composed of several rock types, the most common of which are granite, syenogranite, adamellite, granodiorite, and ferrogranodiorite. Two bodies that appear to be typical of the small plutons occur in the Perent Lake and the Long Island Lake quadrangles. In the north-central part of the Perent Lake quadrangle (Davidson, 1969b), a coarse-grained hornblende granite, which has approximate dimensions of 4 km by one-half km, intrudes and apparently alters gabbroic anorthosite, locally producing pegmatitic textures in the host rock. Generally, the contacts appear gradational between the granite and the anorthosite host, and microgabbro and micronorite are developed at places along the contacts. In the southwest corner of the Long Island Lake quadrangle (Morey and others, 1969, *op. cit.*), medium-grained ferrogranodiorite has gradational contacts over tens of feet with both the underlying anorthositic host rocks and an overlying granophyric to granitoid granite. The granite, in turn, intrudes a black fine-grained metavolcanic rock, which is interpreted as a relict Middle Keweenawan flow.

Subhorizontal sheets of granophyric granite as much as 70 meters thick and 10 km long, occur sporadically along both contacts of the olivine gabbro unit (fig. V-39). The sheets occur along the southern margin of the southern prong and also adjacent to the western and northern margins of the olivine gabbro unit in the Cramer, Beth Lake, and Kelso Mountain quadrangles. The sheets either underlie (as in the Beth Lake quadrangle) or overlie (as in the Kelso Mountain quadrangle) metamorphosed Keweenawan volcanic rocks. Rocks of similar lithology have been found also in the subsurface at depths of at least 60 meters by drilling in the olivine gabbro unit.

The predominant lithology of the sheets is a fine- to medium-grained granophyric granite that is remarkably uniform in mineralogy. Except for a few contained fractures and dikes, the granite is mesoscopically structureless; accordingly, at places it is virtually impossible to distinguish from rhyolite. The rock consists of hematite-stained, subhedral alkali feldspar and anhedral quartz, generally in granophyric intergrowths, together with minor amounts of euhedral plagioclase and subhedral iron-titanium oxides (table V-22). Several rock types, including syenogranite, adamellite, granodiorite, and ferrogranodiorite, are associated with the granophyric granite, as are substantial quantities of granofels, which overlies the granite, and apparently are derived from pre-existing Keweenawan flows. The intermediate rock types, cited above, appear to be restricted to the basal contact of the sheets, and commonly have exceedingly complex intrusive relationships, as noted previously in the Kelso Mountain quadrangle (Grout and others, 1959), where granophyric granite occurs both topographically above and marginal to several intermediate rock units.

Further studies are needed to clarify the petrogenetic relationships of the felsic rocks, but certain observations appear to bear on the problem. Judged from anomalies on the gravity maps of Craddock and others (1970) and Ikola (1970), it appears that several of the smaller felsic intrusive bodies, such as the one in the Perent Lake quadrangle, are situated above the buried projection of the Giants Range Granite beneath the Duluth Complex. Such a relationship suggests that these intrusive rocks may have been derived by partial melting and mobilization of the Giants Range Granite. Moreover, the sporadic occurrence of the larger granophyric granite sheets along both margins of the olivine gabbro unit is consistent with a hypothesis that these rocks may be either differentiates squeezed out of the gabbro unit during postintrusive subsidence of the Keweenawan basin or differentiates that migrated updip after basinal tilting. Those parts of the magma that reached the surface would have tended to be fine grained, whereas the lower parts would have tended to produce multiple, coarser grained bodies.

GRANOFELS

Medium- to fine-grained granofelses occur extensively in the topographically higher parts of the western third of the area, from the Cramer quadrangle northeastward through the Long Island Lake quadrangle (figs. V-38 and V-39). The granofels occurs as subhorizontal bodies ranging in thickness from less than one to more than 30 meters. Except in the Beth Lake and Kelso Mountain quadrangles, where these rocks are locally overlain by subhorizontal, granophyric granite and associated felsic rocks, the rocks appear to comprise a unit that structurally overlies the Duluth Complex rocks. Mineralogically, the rocks can be classified as diorite or gabbro (table V-22). They contain granoblastic plagioclase (An_{36-58}), subhedral, subpoikilitic hornblende or rounded, subpoikilitic augite, anhedral blebs of oxides, biotite, and minor amounts of K-feldspar, quartz, and apatite. The rocks have granoblastic textures and rarely are porphyritic. In the southern part of the Beth Lake quadrangle, preserved remnant snowflake textures have been noted, although elsewhere metamorphism and alteration appear to have been sufficiently intense to obliterate such textures. The degree to which a metamorphic texture is developed varies within and between outcrops, and although a granoblastic texture is ubiquitous, it is best developed adjacent to the contacts of adjacent intrusive rocks.

Although previously interpreted as a younger intrusive unit of microgabbro and quartz diorite (Davidson, 1969a), it is now believed that the granofels series represents metamorphosed Keweenawan volcanic flows, which locally have undergone hydrothermal alteration. Evidence for this conclusion is the ubiquitous granoblastic texture and the spatial proximity of these rocks to relatively unmetamorphosed volcanic rocks of comparable thickness.

STRUCTURAL GEOLOGY

The rocks of both the anorthositic series and the troctolite-olivine gabbro series have a foliation given by the alignment of plagioclase grains. The degree to which the foliation is developed depends to some extent upon the mineralogy of the rocks. In general, rocks of the troctolite-olivine gabbro series tend to have a well developed consistent foliation, which is most conspicuous in layers that are relatively rich in olivine. Rocks of the anorthositic series, which contain plagioclase laths 2 to 4 cm long, on the other hand, tend to have a randomly oriented foliation. Exceptions have been noted, however, in both the Kawishiwi Lake and Long Island Lake quadrangles, where the foliations are moderately consistent and nearly horizontal.

Three conjugate, nearly vertical fracture sets occur in the eastern part of the Duluth Complex (fig. V-40): 1) N. 40° E.—N. 20°—30° W.; 2) N.—N. 60° E. or N. 60° W.; and 3) N. 20° E.—N. 80° E. As noted in Table V-24, the predominant N. 40° E.—N. 20°—30° W. conjugate set is common in all units of the complex.

Conjugate sets 1 and 2 appear to be unrelated to the host rock structure, whereas set 3 appears to be subparallel to the predominant foliation in the olivine gabbro unit, and forms the predominant "grain" in sheeted granophyric granite units. Commonly this "grain" fracture is accompanied by a northward-dipping, low-angle "rift" fracture that is parallel to "grain" strike, which gives rise to the predominant east-west-trending valley and ridge topography and elongate drainage pattern typical of areas underlain by such rocks within the eastern part of the Boundary Waters Canoe Area.

GEOPHYSICAL INTERPRETATION

A large, elliptical, positive Bouguer anomaly (+70 mgals) dominates the gravity pattern over the eastern part of the Duluth Complex (Craddock and others, 1970; Ikola, 1970). The high is elongate in an easterly direction and asymmetrical, and is centered over Pipe Lake in the Brule Lake quadrangle. Its axis is coincident with a septum of granophyric granite along the southern contact of the southern tongue of the complex. The high is believed to result from a large volume of comparatively dense intrusive rock at depth, although a thick pile of flows cannot be ruled out.

The western margin of the anomaly has a very steep gradient (20 mgals in three kilometers) that trends northerly and approximately coincides with the Sawbill Trail. The steep gradient may be attributable to: (1) the onlap of the olivine gabbro unit over anorthositic series rocks; (2) thinning of the olivine gabbro between the volcanics and the anorthositic series rocks; or (3) an absence of felsic rocks along this axis. Interestingly, foliations in the olivine gabbro unit strike parallel to the trend of the gradient in this area.

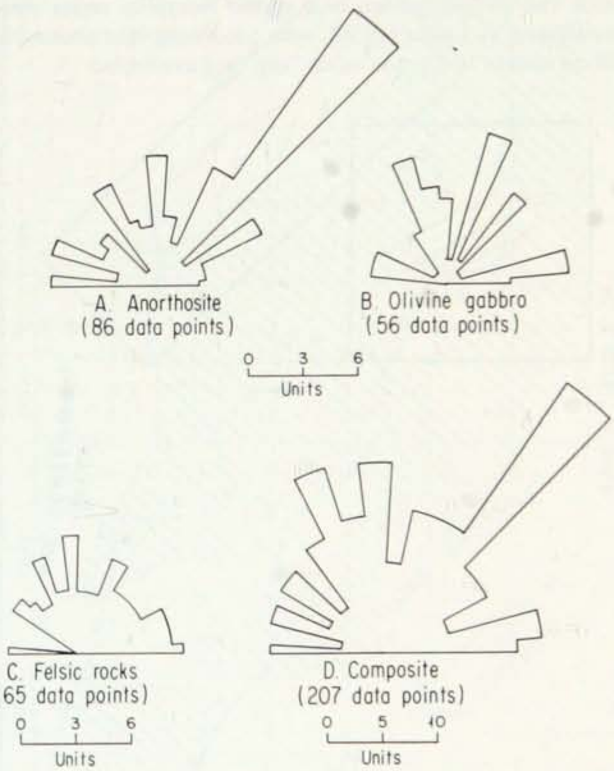

Figure V-40. Vertical fracture patterns, eastern part of Duluth Complex.

STRUCTURAL INTERPRETATION

Two inferred geologic sections across the southern prong of the Duluth Complex are shown in Figure V-41. Judged from structural attitudes, the rocks of the Duluth Complex as well as the intercalated Keweenawan flows dip about 15°-20° S.

ECONOMIC GEOLOGY

Two types of mineral deposits occur in the eastern part of the Duluth Complex as defined herein: 1) low-grade copper-nickel concentrations and 2) titanomagnetite-rich rocks. Both types are in the layered units of the early mafic series and units of the troctolite-olivine gabbro series.

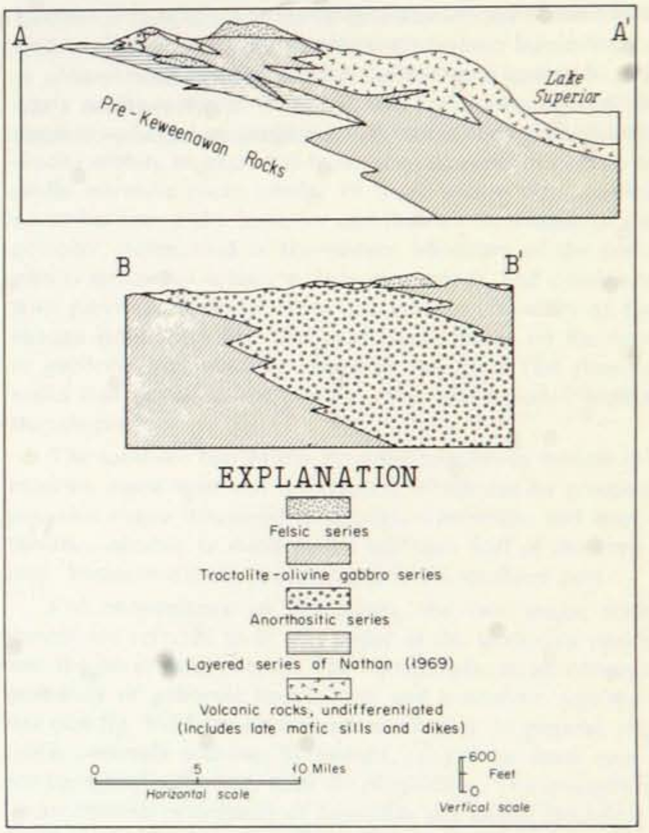

Figure V-41. Inferred geologic sections across southern prong of Duluth Complex. Lines of sections shown on Figure V-38.

Table V-24. Conjugate fracture sets in eastern part of Duluth Complex.

Conjugate fracture set	Anorthositic series	Olivine gabbro series	Felsic series
N. 40° E. - N. 20°-30° W.	Strong	Weak	Common
N. - N. 60° E. or N. 60° W.	Common	Common	Strong
N. 20° E. - N. 80° E.	Weak	Strong	Common

Minor amounts of copper-nickel sulfide minerals have been noted in the eastern part of the complex. Of particular interest is the Long Island Lake quadrangle, where discontinuous zones of gossans and visible sulfide minerals occur near the base of the Tuscarora intrusion (Johnson, 1969, unpub. M.S. thesis, Univ. Iowa; 1970, unpub. Ph.D. thesis, Univ. Iowa; Weiblen and others, 1971). Similar isolated exposures have been found at the top of the intrusion, at the contact with anorthositic gabbro, in the Long Island Lake quadrangle. The sulfide minerals, consisting of chalcopyrite, pyrrhotite, and minor pentlandite, occur interstitially to plagioclase and olivine. Drilling (Johnson, 1969, *op. cit.*) has indicated a tabular, possibly continuous zone of low-grade material (0.3 percent combined copper-nickel) about 50 feet thick at the base of the Tuscarora intrusion. A thinner zone, 10 to 20 feet thick, which is 50 to 100 feet above the lower mineralized zone, has a higher combined copper-nickel content (near one percent, according to Johnson). Another occurrence of potential economic interest is in secs. 21 and 22, T. 62 N., R. 6 W. (Kawishiwi Lake quadrangle); at this locality, gabbroic anorthosite containing visible chalcopyrite and pyrrhotite intrudes granofels (Davidson, 1969a). Exploratory drilling and geophysical work appear warranted at this locality. Also, surface outcrops containing sulfide mineralization occur in the upper troctolite unit south of Brule Lake, near the contact with the Keweenawan volcanic rocks.

Locally significant amounts of titaniferous oxides occur in the eastern part of the complex as: 1) banded segregations; 2) irregular bodies; or 3) dike-like intrusive units (Grout, 1950; Nathan, 1969, *op. cit.*), within both the layered units mapped by Nathan and in the olivine gabbro unit of the troctolite-olivine gabbro series. The primary titanium oxide phases are ilmenite solid solution (Fe_2O_3-$MgTiO_3$-$FeTiO_3$) and titanomagnetite (Fe_3O_4-Fe_2TiO_4). Subsolidus exsolution has resulted in complex intergrowths (Nathan, 1969, *op. cit.*).

The economic potential of selected oxide-rich deposits within the olivine gabbro unit of the troctolite series was investigated by Grout (1950), who concluded that about 80 million tons of low-grade oxide "ore" are available.

SOUTHERN PART OF DULUTH COMPLEX*

Bill Bonnichsen

The southern part of the Duluth Complex includes the area extending from Duluth northward to the vicinity of Babbitt and Birch Lake. Thick glacial drift covers much of this area so that reliable geologic knowledge is available for only certain parts (fig. V-42). Knowledge of the drift-covered areas is limited to what can be inferred from published geophysical surveys (Ikola, 1968b; Bath and others, 1964, 1965; U.S. Geological Survey Map GP-639, 1969), examination of cores from scattered drill holes, and geophysical surveys conducted by private groups.

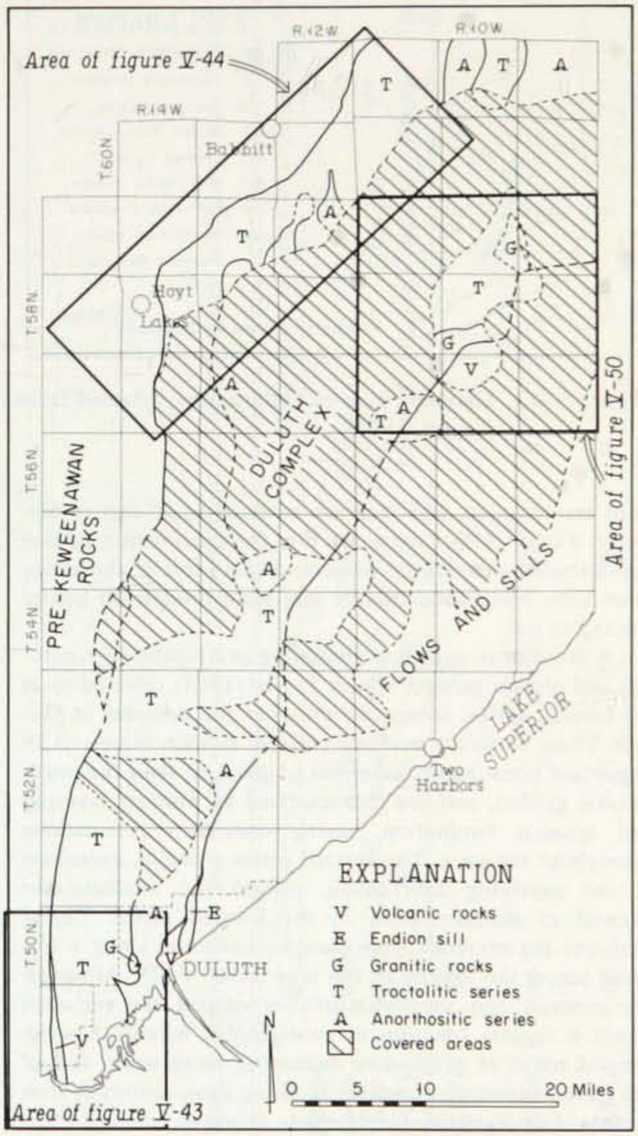

Figure V-42. Geologic map of southern part of Duluth Complex.

* Contribution number 534, Department of Geological Sciences, Cornell University, Ithaca, New York 14850.

The western, or footwall, margin of the complex is accurately located only in areas having good outcrop; its position in covered areas is known within approximately one-tenth to half a mile from geophysical surveys. The western contact dips eastward at variable angles.

The eastern margin of the complex is not located as precisely as the western margin because of the paucity of exposures and the low geophysical contrast between rocks included within, or excluded from, the complex. Inasmuch as mafic intrusive rocks similar to those within the complex occur between Lake Superior and the eastern margin of the complex, delineation of the eastern boundary of the complex is somewhat arbitrary. It is convenient, and consistent with previous work, to define the eastern boundary as the change from troctolitic and anorthositic rocks on the west to gabbroic and volcanic rocks on the east. The granitic rocks that occur in the contact zone are included within the complex.

The southern half of the complex consists of several intrusions, some large and some small, which can be grouped into two major lithologic categories—troctolitic and anorthositic—similar to those in the northern half of the complex. Troctolitic rocks predominate in the southern part.

For convenience of discussion, the two major rock groups are referred to in this paper as the troctolitic series and the anorthositic series. The anorthositic series consists primarily of gabbroic anorthosite and troctolitic anorthosite (see fig. V-25 for classification scheme). In general, the mafic minerals (olivine, pyroxenes, oxides) in these rocks are paragenetically later than the plagioclase. The troctolitic series consists principally of troctolite and augite troctolite. In these rocks, plagioclase and olivine commonly are contemporaneous; however, it is not uncommon for plagioclase to be earlier than olivine and, locally, olivine is earlier than plagioclase. Generally, both plagioclase and olivine are earlier than associated pyroxenes and oxides. In general, rocks assigned to the troctolitic series are characterized by a higher content of mafic minerals than are the rocks assigned to the anorthositic series. In various areas, gabbro, ferrogabbro, norite, picrite, dunite, peridotite, and intermediate or granitic intrusive rocks are closely associated with troctolite and seem to be genetically related to it. Such occurrences are considered to constitute part of the troctolitic series.

The troctolitic series occurs adjacent to the western margin in the southern part of the complex. The anorthositic series generally is located east and southeast of the troctolitic series. At several places, troctolite intrusions cut the anorthositic series and anorthositic rocks occur as inclusions in troctolite. These consistent age relationships between the two major rock types have led me to consider them as two fundamentally distinct groups which, together, make up the southern half of the complex. It remains to be seen, however, whether the distinction between the two groups will remain as clear-cut as it is presented here.

Granitic rocks are exposed at several localities along the eastern margin of the complex in the Greenwood Lake area, and tentatively are considered to be related to troctolite and ferrogabbro of the troctolitic series. Similar granitic rocks, and closely associated rocks having intermediate compositions, occur in the Duluth area. Most of the granitic rocks are characterized by graphic quartz-alkali feldspar intergrowths. Many earlier investigators referred to these rocks as "granophyre" because of this texture; others have used the term "red rock." The feldspars characteristically are clouded with abundant fine-grained material, presumably hematite, which gives the characteristic red color to these rocks. Many contain plagioclase phenocrysts, and the mafics commonly are hydrous minerals such as biotite, chlorite, and amphibole.

Numerous inclusions of fine-grained, generally granular hornfelses, ranging in size from a few inches to thousands of feet, are present in the southern part of the complex, and are especially abundant within the troctolitic series adjacent to the footwall. Many are metamorphosed Virginia Formation, some are fine-grained troctolites or other mafic intrusive rock types, and a few are Biwabik Iron-formation. Hornfelses derived from mafic volcanic rocks probably are most abundant, however.

On the following pages, three relatively well exposed regions (Duluth, Babbitt-Hoyt Lakes and Greenwood Lake; see fig. V-42) are described. These descriptions are followed by an account of some late-stage troctolitic and ultramafic rocks. The later sections are concerned with the mineralogic, chemical, and petrologic aspects of the rocks that comprise the troctolitic series and the various types of hornfelses.

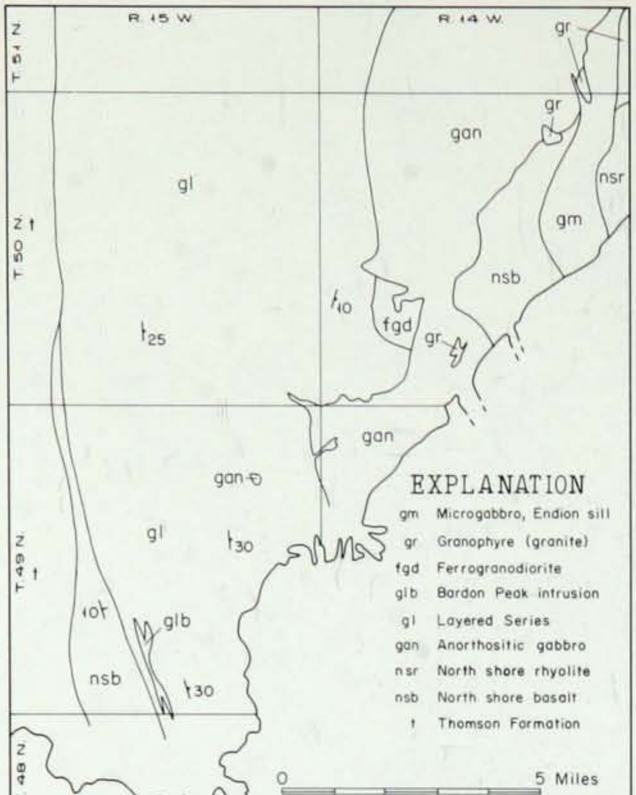

Figure V-43. Geologic map of Duluth area (adapted from Taylor, 1964).

DULUTH AREA

The southern end of the Duluth Complex is well exposed in the vicinity of Duluth, and has been mapped (scale 1:24,000) by Taylor (1964). Data from Taylor's comprehensive report and many of his conclusions are summarized below.

At Duluth (fig. V-43), the Duluth Complex overlies a Keweenawan basalt footwall; the basalt pinches out northward, and the Middle Precambrian Thomson Formation constitutes the footwall. The basalt has a maximum thickness of 2,500 feet, and is grossly conformable with the basal contact and internal structures of the overlying intrusive mass. Anorthositic gabbro, which generally is coarse grained and contains 75-90 percent calcic labradorite, is the oldest intrusive unit at Duluth. The unit contains abundant inclusions, ranging in composition from gabbro to anorthosite, and many outcrops consist of a jumble of large blocks in a coarse feldspathic matrix. Generally, the inclusions are rounded, and they are as much as 25 feet in diameter. Taylor (1964) suggested that the anorthositic gabbro was intruded as a crystal mush and that most of the inclusions were early solid phases that were broken up before coming to rest.

A stratiform sequence consisting principally of troctolite and olivine gabbro, which Taylor (1964) referred to as the layered series, forms two-thirds of the complex at Duluth. These rocks are medium grained, contain olivine as an important constituent, have less plagioclase than the anorthositic gabbro, and are characterized by primary layering and igneous lamination having consistent orientations throughout the area. The layered series contains inclusions of the overlying anorthositic gabbro. To evaluate the amount of differentiation in the layered series, Taylor analyzed the minerals from samples collected along a traverse across the middle of the area (table V-25). Although the minerals vary somewhat in composition, the variation is not a regular function of stratigraphic height. A weak general trend of plagioclase becoming more sodic and of the ferromagnesian minerals becoming more iron-rich with increased stratigraphic height exists, however.

Rhythmic layering, with units from a few inches to a few feet thick, is well developed in many parts of the area. Locally the layering was disturbed by the intrusion of younger gabbros. Large-scale layering, with lenticular units on the order of tens of feet thick, also is present.

Table V-25. Mineral compositions and modal analyses of samples from the layered series at Duluth (after Taylor, 1964); P = less than 1 percent.

		Compositions			Abundance of minerals (volume percent)									
Sample number	Height above base (feet)	Plagioclase (An)	Olivine (Fo)	Orthopyroxene (Fs)	Plagioclase	Clinopyroxene	Orthopyroxene	Olivine	Magnetite	Apatite	Biotite	Orthoclase	Quartz	Rock name
G580	16,000	53			44	27	1		6	2	1	5	1	Microsyenogabbro
M3728	14,300	54		43	45	26	8		12	3	2	4	2	Magnetite syenogabbro
M3725	11,300	60		37	45	37	P	7	9	1	1			Olivine melagabbro
M3720	8,500	62	58	37	78	P	9	8	2	1	2			Feldspathic olivine norite
M3716-3	6,000	60	59	37	81	P	4	12	1	1	1			Feldspathic hypersthene troctolite
M3716-1	6,000	59	57		50	25		20	5	P	P			Olivine gabbro
M3711	1,800	65	59		81	4	P	12	2	P	P			Feldspathic augite troctolite
G1006	0	61	61		59	32		4	2	P	P			Olivine gabbro

Taylor (1964) suggested that the layered series resulted from the bottom accumulation of crystals forming in an actively circulating magma. He attributed the lack of well developed cryptic layering to either periodic magma renewal or to multiple small intrusions, and suggested that crystallization occurred in an environment of tectonic instability characterized by multiple magma injections.

Adjacent to the base of the complex, near its southern termination, the main troctolite of the layered series is intruded by a semi-concordant lens-like body of coarse-grained olivine gabbro and local peridotite (Bardon Peak intrusion). Taylor suggested that this body was emplaced shortly after the lower part of the layered series had crystallized, but while it was still hot.

Several bodies of rocks of intermediate to granitic composition occur in the Duluth area, and seem to be closely related to the layered series. The largest such body, along the contact between the layered series and the overlying anorthositic gabbro (fig. V-43), is ferrogranodiorite, which Taylor (1964) considered to be chemically and mineralogically gradational with the layered series. He suggested that it may have formed from the same magma that previously had given rise to the layered series, but was uncertain whether it is a separate intrusion or a late segregation of differentiated magma from the adjacent layered series. Taylor considered that most of the granitic bodies at Duluth are magmatic in origin, but that a few probably had resulted from the replacement of previous mafic rocks by material deposited from hydrothermal fluids.

BABBITT-HOYT LAKES REGION

I have mapped the region along the northwestern side of the complex between Babbitt and Hoyt Lakes, which is relatively well exposed. It includes the area between the Gabbro Lake (Green and others, 1966) and Kangas Bay (Phinney, open-file map, Minn. Geol. Survey) quadrangles to the northeast and the large poorly exposed region extending southward. Exposures are sufficiently good for a distance of 2 to 5 miles outward from the footwall of the complex to define and trace geologic units (fig. V-44). Additional data have been obtained from studies of drill core made available to me and others.

Birch Lake-Dunka River Area

The troctolitic series occupies the greater part of the Babbitt-Hoyt Lakes region, and occurs mainly between pre-Keweenawan footwall rocks to the northwest and older Keweenawan rocks to the southeast (fig. V-44). East of Babbitt, the series has been divided into several distinct units that are separated by gradational contacts. The most extensive unit, a medium-grained poikilitic augite-bearing troctolite, is characterized by augite oikocrysts that are slightly separated from one another. The augite content generally decreases from northwest to southeast—stratigraphically upward—across this unit. This troctolite is very similar to, and is considered the equivalent of, the poikilitic augite troctolite in the Gabbro Lake quadrangle, which forms a large part of the South Kawishiwi intrusion (Green and others, 1966).

Another troctolite (unit t, fig. V-44), distinguished by a sparseness of minerals other than plagioclase and olivine, lies southeast of the poikilitic augite troctolite. The contact between the two units has been traced for nearly 15 miles subparallel to the base of the complex. The contact between this unit and the underlying poikilitic augite troctolite is gradational across a distance of a few tens or hundreds of feet. Consequently, the stratigraphically higher augite-poor troctolite is considered to be part of the same intrusion as the augite troctolite.

The troctolite to the northwest of the poikilitic augite troctolite unit is characterized by having a higher augite content than the poikilitic augite troctolite and by having abundant interstitial oxides. The two units are gradational. The rock is medium to coarse grained, is quite heterogeneous, and contains numerous incipient pyroxene-oxide pegmatites. This augite troctolite may be equivalent to the augite troctolite (sat) of the South Kawishiwi intrusion, inasmuch as both are northwest of the poikilitic augite troctolite.

Several medium- to coarse-grained anorthosite, troctolitic anorthosite, and gabbroic anorthosite bodies that range in size from inches to hundreds of feet across occur within the various troctolite units south and southeast of Birch Lake. Clearly, most are inclusions within the troctolites; a few, however, might be cogenetic plagioclase-rich segregations, as has been suggested for similar segregations in the

Figure V-44. Geologic map of Babbitt-Hoyt Lakes region.

Gabbro Lake quadrangle (Phinney, 1969, 1970, and this chapter). These bodies are more common in the two troctolites that flank the poikilitic augite unit than within it.

The intrusive rocks immediately above the footwall contact in the area east of Babbitt (Dunka River area) are an assortment of fine- to medium-grained troctolites, gabbros, and norites (unit g_2 on fig. V-44, unit g on fig. V-45). Included in the unit are medium- and coarse-grained troctolites and poikilitic augite troctolites similar to those farther to the east in the complex, medium- to coarse-grained anorthositic rocks, and several types of hornfels.

The footwall of the Duluth Complex in the Dunka River area (Bonnichsen, 1968, unpub. Ph.D. thesis, Univ. Minn., 1969a and b) consists of as much as 100 feet of hornfelsed Virginia Formation, which overlies the Biwabik Iron-formation (fig. V-44). Sills of diabase, believed to be related to the complex, cut the footwall rocks. Irregular granitic segregations as much as a few inches across, believed to be the result of partial melting, are common in the Virginia Formation adjacent to the basal intrusive unit of the complex. As can be seen from Figure V-45, the basal contact is irregular and several faults occur within the footwall and locally along the contact. Inasmuch as this is the only area in the southern part of the complex where the basal contact is well exposed, the relationships here may be indicative of the contact as a whole.

Drilling in the Dunka River area by mining companies indicates that the geology of the basal part of the complex is complicated. Descriptions of four deep drill holes from sec. 2, T. 60 N., R. 12 W. (see fig. V-45), which penetrate the footwall of the Duluth Complex, are summarized in Figure V-46. In all four holes, the general vertical sequence of rock types encountered above the footwall is (1) a basal intrusive unit, overlain by (2) an augite-poor troctolite, overlain by (3) a poikilitic augite-troctolite. Inclusions of rock types such as anorthositic gabbro and hornfels occur in each of these units.

The basal intrusive unit (unit 7 of fig. V-46) which correlates with unit "g" on Figure V-45, is a heterogeneous mixture of mostly fine-grained rock types, including gabbro, troctolite, and norite. The unit contains several types of inclusions and local segregations of hypersthenite, picrite, oxide-rich rocks, and massive sulfides. Except in these segregations, the textural relations suggest that most rocks in this unit probably crystallized in place from a liquid. Probably, the unit owes much of its variability to contamination from granitic rocks of the underlying Giants Range batholith and the Virginia Formation.

The augite-poor troctolite (unit 6 of fig. V-46) generally is medium grained and contains approximately 60 percent plagioclase and 35 to 40 percent olivine, with only minor augite or oxides, but locally it is more anorthositic, and in several places is interlayered with picrite, dunite, and minor peridotite (unit 5). The picrite and dunite units are gradational with the troctolite, and are considered part of the same intrusive body. The sparseness of augite and oxides,

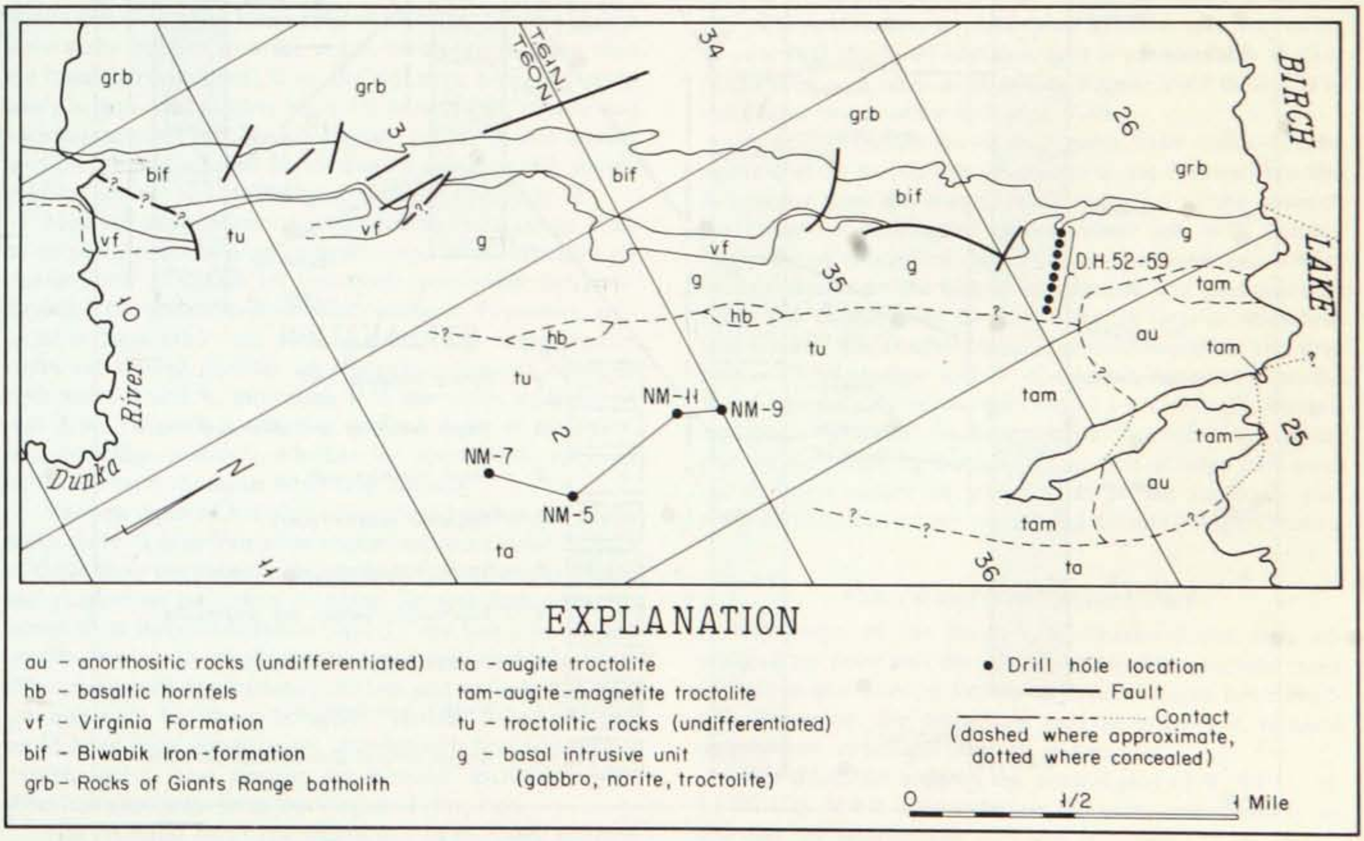

Figure V-45. Geologic map of Birch Lake-Dunka River area.

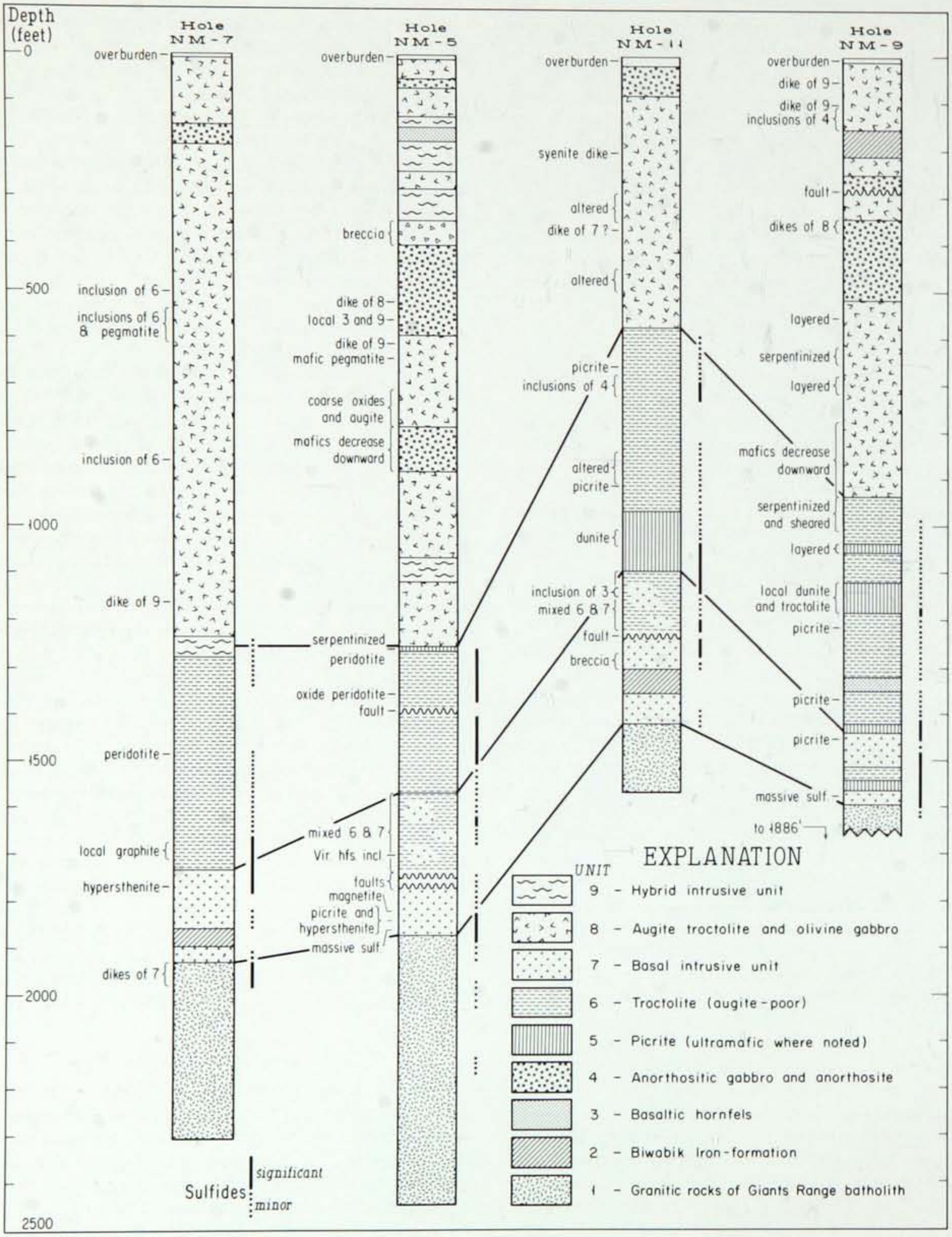

Figure V-46. Geologic logs of four long drill cores from the Dunka River area.

and the presence of picrite and dunite layers, suggest that the augite-poor troctolite originated, at least in part, by crystal settling. Included within this intrusion are inclusions of anorthositic (unit 4) and hornfelsic rocks (unit 3). Locally, the picrite and dunite are moderately to strongly serpentinized. Much of this intrusive body, particularly the upper part, has been recrystallized.

An augite troctolite (unit 8), which overlies the augite-poor troctolite, is somewhat variable in composition, and locally grades into olivine gabbro, troctolitic anorthosite, or gabbroic anorthosite. It contains inclusions of augite-poor troctolite (unit 6) and rocks of unit 4 (anorthositic gabbro, anorthosite, and troctolitic anorthosite), as well as inclusions of basaltic hornfels (unit 3). This troctolite is characterized by augite oikocrysts 1 to 4 inches across. Olivine generally is more abundant than augite, although it is not uncommon for augite to be more abundant than olivine so that the rock is olivine gabbro; locally olivine is sparse or absent. The quantity of oxides varies in this unit and locally is sufficiently abundant for the rock to be called an augite-magnetite troctolite. The unit may have been intruded as a crystal-rich mush.

The relative ages of the augite troctolite (unit 8), the augite-poor troctolite (unit 6), and the basal intrusive unit (unit 7) are not known definitely. Probably, the augite-poor troctolite is older than the basal intrusive unit and occurs as inclusions within it. It is suggested, because of apparent inclusion relationships and recrystallization effects, that the augite-poor troctolite is older than the overlying augite troctolite. If so, the augite-poor troctolite is the oldest of these three troctolitic units. Observations made outside the Dunka River area suggest that the augite troctolite is older than the basal intrusive unit. If so, the intrusive sequence tentatively is suggested to have been: (1) anorthositic gabbro and anorthosite (unit 4); (2) augite-poor troctolite and associated picrite (units 5 and 6); (3) augite troctolite and olivine gabbro (unit 8); and (4) the basal intrusive unit (unit 7).

Medium-grained anorthositic gabbro, containing little or no olivine, is the most common type of anorthositic inclusion (unit 4). Others are troctolitic anorthosite and anorthosite. The inclusions have sharp contacts. Pegmatitic material is associated with some contacts and some contact zones are altered. Similar anorthositic inclusions occur in both units 6 and 8, indicating that the rocks included in unit 4 are older than either of the two types of troctolite. It is not clear, however, whether the anorthositic rocks of unit 4 are from the same, or diverse, sources.

Various types of hornfels are enclosed within the mafic rocks. Several iron-formation inclusions occur (unit 2); part of these have been metasomatized, so that quartz is absent and plagioclase has taken its place. Several inclusions, referred to as basaltic hornfels (unit 3), are fine grained and consist dominantly of plagioclase and augite and have variable amounts of hypersthene, olivine, and oxides. The parent materials for these hornfelses remain uncertain, but could have been either basalt or relatively fine-grained intrusive rocks. Also present are hornfels inclusions with abundant cordierite from the Virginia Formation.

The youngest rock type shown in Figure V-46, referred to as the hybrid intrusive unit (unit 9), is heterogeneous, but seems to consist mainly of calcium pyroxene, biotite, and amphiboles with variable amounts of plagioclase and local quartz. The rocks vary in grain size from fine grained to pegmatitic and modally from pyroxene- or hornblende-rich to plagioclase-rich. The amphiboles commonly replace the calcium pyroxene. This unit is closely associated with hornfels inclusions, and may have originated by contamination or some sort of partial melting process. It is interesting to note that in drill core NM-7, this unit was intruded along the contact between the augite troctolite and the underlying augite-poor troctolite.

Several thin (one-half inch to two feet), nearly vertical granitic dikes cut the various mafic units shown in Figure V-46, and evidently formed along fractures that developed during the cooling of the mafic rocks. Except for local hydrothermal alteration they had little effect on the enclosing mafic rocks. Some of the dikes were sheared after emplacement, for they contain broken minerals and thin serpentinized veins. A few evidently followed fractures and faults that previously had been serpentinized.

The basal intrusive unit is shown in more detail in Figure V-47, which is a summary of the geology of 8 drill holes spaced 200 feet apart in the southeastern part of sec. 26, T. 61 N., R. 12 W. (fig. V-45). As shown in Figure V-47, a wide variety of inclusions occur within the basal intrusive unit; locally these are so abundant that the rock has been referred to as an intrusive breccia. Thin sections of the intrusive rocks from these cores indicate that the rocks vary gradationally from troctolite to gabbro to norite. Generally, they are fine grained and have decussate textures, suggestive of crystallization in place from a liquid. It is interesting to note that the basal intrusive unit is considerably thicker in the drill core sections shown in Figure V-47 than it is in the deeper holes shown in Figure V-46.

In the drill holes shown in Figures V-46 and V-47, the uppermost 50 to 100 feet of granite in the footwall has less K-feldspar than the deeper parts. Adjacent to the contact, the rocks are thoroughly recrystallized and most contain hypersthene. Many are dioritic in composition, but similar in appearance to the quartz monzonitic and granodioritic rocks that characterize the Giants Range Granite elsewhere (see Green, this chapter) and at greater depths in the drill holes. The apparent loss of K-feldspar suggests that the rocks immediately below the contact were partially melted, and that a "granitic" fraction was lost. Probably, the noritic part of the overlying basal intrusive unit in large part owes its chemical nature to assimilation of the materials that were distilled out of the underlying Giants Range Granite.

Central and Southwestern Part

Southwest of the Birch Lake-Dunka River area exposures are poor and the relationships of the various units in the complex are not known in detail. Judged from available exposures, the troctolites vary in grain size, mineral proportions, structural attitude, and texture.

The troctolite body in the central part of T. 59 N., R. 12 W. (fig. V-44) is bounded on the west, and probably on the east, by anorthositic rocks; it is covered to the south. The layering and igneous lamination in it are highly vari-

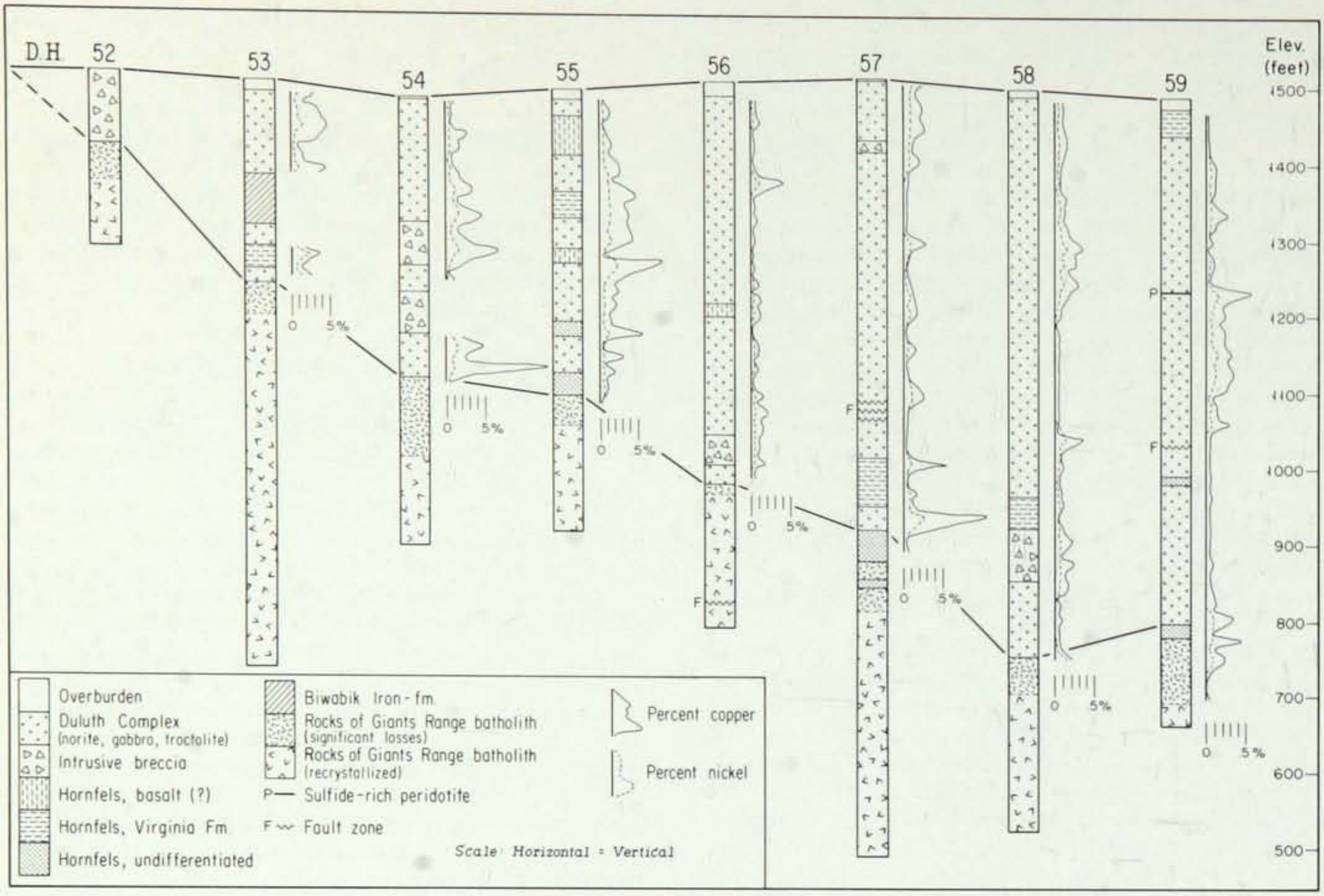

Figure V-47. Geologic logs and Cu-Ni assays for eight drill cores from Dunka River area.

able in attitude. Troctolite and gabbro are interlayered locally, particularly in the eastern exposures, in a manner similar to that found north of Greenwood Lake, 8 to 10 miles to the east. Inclusions are more abundant than in most other troctolite bodies, and include anorthositic and troctolitic rocks and various hornfelses, part of which are metasediments from an unknown source.

A generally coarse-grained gabbro in the western part of T. 59 N., R. 12 W. evidently overlies the extension of the troctolite exposed to the north, except along the northeastern part of their mutual contact where a thin zone of anorthositic rocks separates the two (fig. V-44). Apparently, the southwestern part of this contact is sharp and probably intrusive. The gabbro body is about 3 miles long, locally strongly layered, and dips shallowly to the southeast; it contains anorthositic and troctolitic inclusions. The gabbro is somewhat heterogeneous but typically contains about 50 percent early crystallized plagioclase and variable amounts of texturally late augite, olivine, and oxides, generally in that order of decreasing abundance. A relatively high apatite content and abundant exsolved opaque inclusions within the augite characterize the intrusion. The sinuous southeastern margin of the body probably results from erosion of a slightly irregular, gently-dipping contact. On this side of the body are scattered outcrops of hornfelsed igneous rocks, mainly fine-grained porphyritic types; a few contain abundant brown hornblende.

The anorthositic series is exposed several miles south of Babbitt and east of Hoyt Lakes (fig. V-44); outcrops of it are abundant in the northern part of T. 59 N., R. 12 W., and in the southern part of T. 60 N., R. 12 W., but become sparse to the south and east. Marked lithologic variation characterizes the series, with troctolitic and gabbroic anorthosites containing distinctive olivine, augite, and oxide oikocrysts as the principal rock types.

The footwall rocks throughout the Babbitt-Hoyt Lakes region—Lower Precambrian granitic rocks of the Giants Range batholith overlain by the Middle Precambrian Biwabik Iron-formation and Virginia Formation—are much the same as in the Dunka River area. Inclusions of metamorphosed Virginia Formation are common in the lower few hundred feet of the complex. Most of these inclusions are small, generally being less than 100 feet thick.

Several large, lithologically diverse, concordant bodies of hornfels occur in the interval between 1 and 6 miles away from the footwall contact (fig. V-44). Probably most represent metamorphosed mafic volcanic rocks, others represent earlier fine-grained intrusive rocks, and a few were sedimentary rocks. The largest body of hornfels (Colvin Creek body in fig. V-44) has an arcuate shape and is 4 to 5 miles long, as inferred from aeromagnetic maps. Another body in sec. 33, T. 60 N., R. 12 W. is exposed in a rock cut along Erie Mining Company's railroad. Apparently it is located along the boundary between the troctolitic series to the northwest and older anorthositic and other rocks lying to the southeast, in a structural position similar to that of the Colvin Creek body. An interesting occurrence of distinctly layered hornfelsed basalt, well exposed in sec. 18, T. 59 N., R. 13 W. in a rock cut along the same railway tracks, locally forms the footwall for the troctolitic series. The layering dips 53° SE. and strikes northeastward, approximately parallel to the adjacent basal contact. Extensive hornfels outcrops also occur on and northeast of Moose Mountain in the northeastern part of T. 58 N., R. 14 W., within troctolite.

The presence of the Colvin Creek body and the hornfels in sec. 33, T. 60 N., R. 12 W. along the hanging wall of the troctolitic series, and of similar volcanic hornfelses within and at the base of the series, suggests that prior to intrusion of the troctolites an extensive volcanic unit was present along the footwall of the complex. If so, the volcanics were broken up in such a manner that the Colvin Creek body and the hornfels in sec. 33, T. 60 N., R. 12 W. remained attached to the eastern side of the detachment zone along which the troctolitic series was emplaced.

Other large hornfels bodies that differ from the volcanic types and the Virginia Formation occur in the Babbitt-Hoyt Lakes region. The largest of several bodies in the central part of T. 59 N., R. 12 W. is a plagioclase-Ca pyroxene-magnetite hornfels, with minor sphene and local biotite, apatite, and calcite; probably it is of sedimentary origin. A small outcrop of plagioclase-augite-magnetite hornfels along Erie Mining Company's railroad in the northern part of sec. 21, T. 59 N., R. 12 W. has well preserved sedimentary cross-bedding and cut-and-fill structures. The compositions and structures of these hornfelses are similar to those in an isolated group of unmetamorphosed sedimentary outcrops of unknown age located in the Greenwood Lake area, in the S½ sec. 4, T. 57 N., R. 10 W., between the Duluth Complex and Lake Superior.

Structural Attitude of Base of Duluth Complex

Exploratory drilling for copper-nickel sulfides in the Babbitt-Hoyt Lakes region and adjacent Kangas Bay and Gabbro Lake quadrangles to the northeast has provided information on the structural configuration of the base of the complex. Information from a part of this area is given in Figure V-48, which shows structural contours on the base area of the complex and the downdip limit of the Virginia and Biwabik formations.

Structural contours to depths of as much as 4,200 feet indicate that the dip of the basal contact is variable. In two areas the dips are shallow. In an area of about one square mile within about 1,000 feet of the surface in the northern part of T. 60 N., R. 12 W. (Dunka River area), the contact dips approximately 10° SE. A larger terrace, about 4 square miles in area—mainly at depths between 2,000 and 3,000 feet, in the southwestern part of T. 60 N., R. 12 W. and the northwestern part of T. 59 N., R. 12 W—dips less than 20° SE. In adjoining areas, the dip generally is 35-40° SE. On the other hand, along the north-south segment of contact in T. 58 N., R. 14 W. near Hoyt Lakes, the basal contact dips as much as 60° SE. Northeast of the Babbitt-Hoyt Lakes region, in the southwestern part of the Gabbro Lake quadrangle, the basal contact is known to be flatter at a depth of a few thousand feet than it is near the surface.

Throughout this region, the iron-formation generally dips 5 to 15° SE. at the surface (Grout and Broderick, 1919a), approximately the same as its dip on the structural terraces beneath the complex. Accordingly, there appears to have been little rotation of the basement in this area. In the Dunka River area, on the other hand, the iron-formation at the surface dips 15° to 35° SE. (Bonnichsen, 1968, *op. cit.*), indicating that moderate deformation and rotation did occur locally there.

Mancuso and Dolence (1970) have suggested that the older rocks were not materially deformed by intrusion of the lower components of the complex; instead, the attitude of the basal part of the complex was determined by the preexisting structural configuration of the country rocks. I generally concur with this view, particularly when considering the region as a whole; but in some areas, such as at the north end of the taconite mine in the Dunka River area, local faulting and tight folding in the iron-formation apparently are the result of forceful emplacement of the complex rocks.

Drill Core from Southwestern Part of T. 60 N., R. 12 W.

A drill core in the southwestern part of T. 60 N., R. 12 W., near Babbitt, which intersects the contact between the Duluth Complex and underlying Virginia Formation at a depth of approximately 1,900 feet, was studied by Hardyman (1969, unpub. M.S. thesis, Univ. Minn.). Medium-grained troctolite is the main rock type intersected in the core; it grades locally into olivine gabbro and at places is pegmatitic. Two picritic zones, each approximately 10 feet thick, occur at depths of 930 and 1,880 feet in the drill hole. The interval between samples available for study was fairly large in the upper part of the core, and it is not known whether one or more distinct intrusive units is present.

Much of the analytic data obtained on the core by Hardyman is incorporated in Figure V-49, which shows the variation with depth in modal abundance of olivine and plagioclase and the compositional variations of plagioclase, olivine, and pyroxene. Other data from the study are included in the section by Phinney in this chapter. Plagioclase compositions are fairly uniform throughout the hole but are slightly more sodic in the lower 200 feet. The olivines and pyroxenes, however, vary together in composition and show strong iron enrichment in the lower 200 feet. Augite generally is somewhat more abundant in the lower 200 feet of the core. Sulfides also are considerably more common in that section. The sample nearest to the basal contact in this core (depth 1,894 feet) is an augite norite lacking olivine, but rich in sulfides. Both the augite and hypersthene

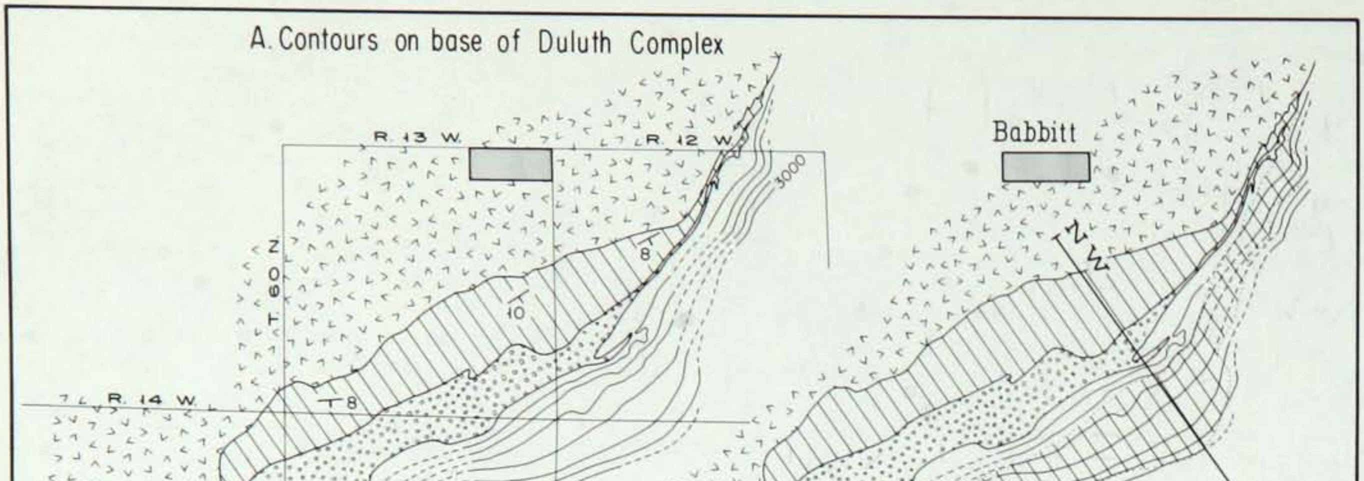

Figure V-48. Structural contours and cross-section of Babbitt-Hoyt Lakes region.

grains of this sample have shapes indicating that they are paragenetically early. The average modes for the principal rock types are:

	plag.	olv.	cpx.	opx.	bio.	opq.	total
troctolite	65.5	26.3	3.8	1.3	1.5	2.2	98
olivine gabbro	53.8	18.7	15.1	1.9	3.3	5.8	99
picrite	32.1	61.6	1.3	—	0.9	3.5	99

Note that biotite, orthopyroxene, and the opaque minerals are more abundant in those rocks that are enriched in augite.

The upper picrite contains fresh olivine, whereas the lower one is highly serpentinized. Plagioclase within the upper picrite zone averages An_{62}, whereas just above and below the picrite the average is about An_{58}. Olivine and augite are slightly Mg-enriched above the upper picrite. Both the percentage and Mg content of olivine decrease upward above this zone. Hardyman suggested that olivine is richer in Mg in those rocks where it is more abundant.

The olivines in the picrite zones characteristically are euhedral to subhedral and well sorted according to size. Hardyman (1969, *op. cit.*) noted that olivine in the picrites is Mg-rich and plagioclase is Ca-rich in comparison to adjacent troctolite, and he suggested that these zones represent marginal phases of the magma in which the olivines became concentrated by settling. In view of the known discontinuous nature of the picritic zones, he suggested that they were torn apart and redistributed by subsequent movements in the semi-consolidated magma.

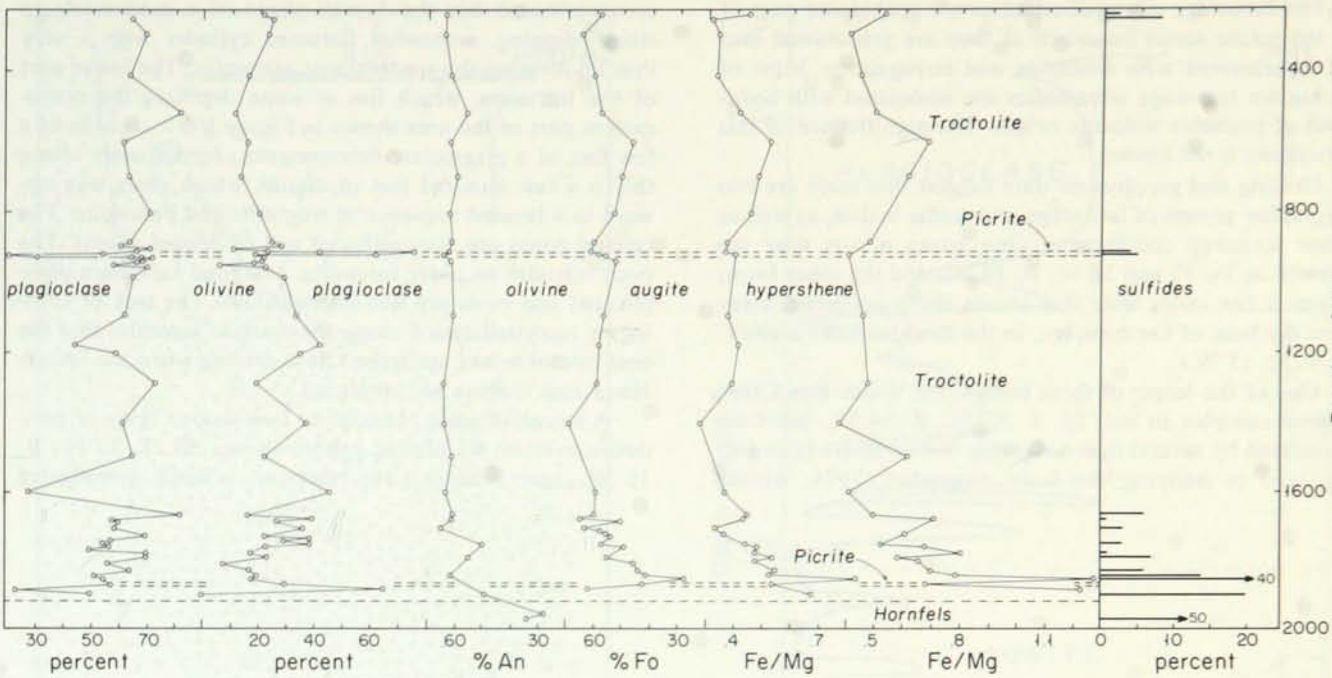

Figure V-49. Mineral abundances and compositions in a drill core from southwestern part of T. 60 N., R. 12 W. (after Hardyman, 1969, unpub. M.S. thesis, Univ. Minn.).

GREENWOOD LAKE REGION

In the Greenwood Lake region two granitic bodies are present along the eastern margin of the complex (fig. V-50), between an area of anorthositic, troctolitic, and gabbroic rocks (mainly ferrogabbro) to the northwest and an area of volcanic rocks to the southeast. Apparently, the troctolite and ferrogabbro are interlayered and intergradational, and typically contain more augite and magnetite than the average troctolite in the Babbitt-Hoyt Lakes region. In the ferrogabbro, augite and magnetite commonly are paragenetically contemporaneous with the plagioclase and olivine. A chemical and modal analysis of typical ferrogabbro is included in Table V-27 (no. 12). The volcanic rocks are mainly felsites, but include non-porphyritic basaltic rocks characterized by a high content of magnetite. An analysis of a magnetite-rich basalt is included in Table V-30 (no. 5).

The granitic rocks are medium grained and uniform, and consist of a granophyric intergrowth of quartz and alkali feldspar, accompanied by quartz and plagioclase grains. The rocks contain small amounts of biotite or hornblende and traces of pyrite. Tentatively, these rocks are considered to be closely related in origin to the troctolites and ferrogabbros of the area because of their close, apparently conformable spatial relationship. The felsites and magnetite-rich basalts that lie to the southeast may be the extrusive equivalents of the ferrogabbros and granitic rocks.

The isolated sedimentary rocks that occur in the south half of sec. 4, T. 57 N., R. 10 W. (fig. V-50) are unmetamorphosed but highly indurated, sand-size clastic rocks with well developed cross-bedding and local cut-and-fill structures. Bedding is variable in attitude but generally inclined gently to the southeast. In thin section, the rock is seen to consist of subangular to subrounded plagioclase grains in a chloritic matrix. Probably, these rocks are Keweenawan in age, the plagioclase having been derived from older Keweenawan mafic or anorthositic intrusive rocks.

The gravity high (Ikola, 1968b) in the southeastern corner of Figure V-50 coincides with the crest of the Midcontinent Gravity High. The rock types underlying that area are not known. The slight deviation from the regional trend of the gravity contours over the areas underlain by the granitic rocks is interpreted as indicating that the granitic rocks are sheetlike in form. The cause of most of the aeromagnetic anomalies (U.S. Geological Survey Map GP-639, 1969) is not known, although it can be seen that magnetite-rich basalts coincide with and probably account for some of them. The most intense magnetic anomaly—the sinuous trend that skirts the northwestern margin of the granitic bodies—evidently is caused by a rather iron-rich intrusive unit underlying the granitic bodies.

LATE-STAGE ULTRAMAFICS

Several late-stage ultramafic bodies occur within the southern part of the complex, but are poorly exposed. Peridotite and dunite are most common; pyroxenite and rocks consisting largely or wholly of Fe-Ti oxides also are present. Plagioclase, apatite, and sulfides are present in some. The bodies range in shape from thin sills or dikes to large lenses or irregular bodies having maximum dimensions of more than 1,000 feet. Parts of some bodies have well developed layering.

The late-stage ultramafic bodies are considered part of the troctolitic series inasmuch as they are gradational into and interlayered with troctolite and ferrogabbro. Most of the known late-stage ultramafics are associated with hornfelses of probable volcanic origin. The significance of this association is not known.

Drilling and geophysical data suggest that there are two geographic groups of late-stage ultramafic bodies, as well as a few scattered occurrences. One group occurs near the footwall in Ts. 57 and 58 N., R. 14 W., and the other forms a zone a few miles wide that trends north-northeast, away from the base of the complex, in the Boulder Lake area (T. 53 N., R. 15 W.).

One of the larger of these bodies, the Water-Hen Creek layered complex in sec. 28, T. 57 N., R. 14 W., has been penetrated by several drill holes (fig. V-51). P. R. Mainwaring, who is studying the body, suggested (1971, written comm.) that it has the overall shape of a moderately to steeply-dipping, somewhat flattened cylinder with a very thin lip forming the westernmost extremity. The lower part of the intrusion, which lies at some depth in the northeastern part of the area shown in Figure V-51, consists of a few feet of a plagioclase-ilmenite rock. Immediately above this is a few hundred feet of dunite, which gives way upward to a layered sequence of troctolite and peridotite. The layered zones are discontinuous and of limited extent. The body intrudes an older troctolite. Marginal facies are complicated and evidently rich in inclusions. The lack of chilling or recrystallization along the margins indicates that the host troctolite had undergone little cooling when the Water-Hen Creek magma was emplaced.

A steeply-dipping, tabular or lens-shaped body of peridotite overlain by olivine gabbro in sec. 36, T. 53 N., R. 15 W., near Boulder Lake reservoir, is being investigated

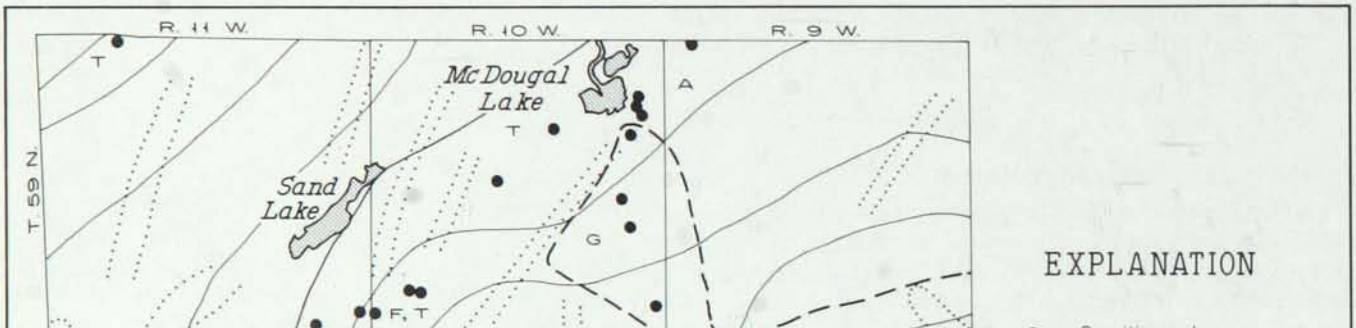

Figure V-50. Geologic map of Greenwood Lake area.

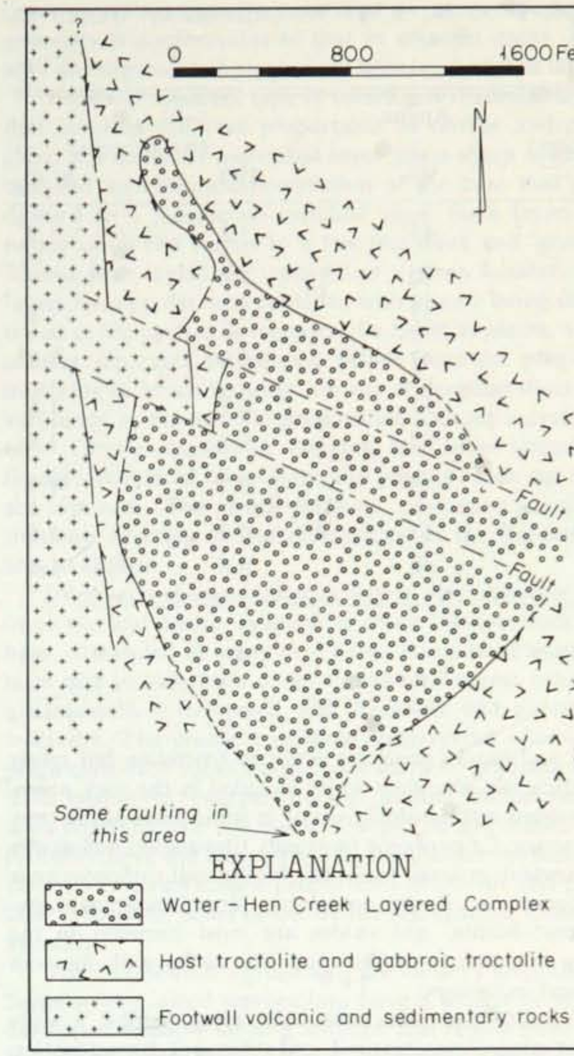

Figure V-51. Geologic map of the Water-Hen Creek layered complex (geology by P. W. Mainwaring).

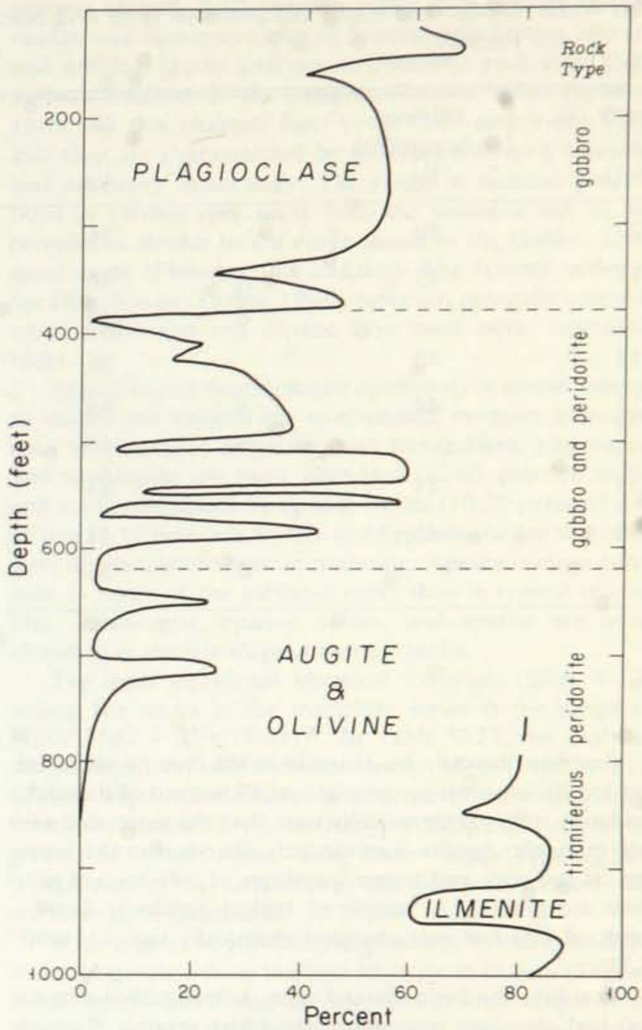

Figure V-52. Mineral abundances in drill core 5-3, sec. 36, T. 53 N., R. 15 W. (drill hole is inclined about 45° W.).

by J. W. Delano, Warren C. Forbes, and myself (fig. V-52). The body is approximately a quarter of a mile long and is elongated in a north-south direction; almost its entire thickness was penetrated by a drill hole. The upper part of the core is gabbro; it grades downward through a zone of interlayered gabbro and peridotite, to peridotite and local dunite at the base of the intrusion. The mafic minerals are altered (deuteric?) in the upper part of the hole, but are fresh in the lower part. The footwall rock is a volcanic hornfels (see analysis 3, table V-30, and later discussion). The footwall contact dips approximately 60° E., as indicated by its position in another hole 700 feet to the west.

Plagioclase (An_{50-65}) is abundant in the upper part of the core. Two distinct generations—earlier deformed xenocrysts and crystal fragments as large as 1 cm long, and later, smaller, undeformed subhedral laths—occur throughout much of the upper gabbro zone. In the upper part of the peridotite zone plagioclase is interstitial to the other silicate minerals.

Distinctly purple Ca pyroxene is the most abundant mineral in the peridotite, and is second in abundance to plagioclase in the upper gabbro. Exsolved ilmenite blebs occur in the Ca pyroxene throughout most of the core. Olivine occurs throughout the core. In the upper part, it is paragenetically later than the plagioclase, but in the peridotite it crystallized earlier than much of the accompanying pyroxene and ilmenite. Only locally, in the olivine-rich lower few feet of the peridotite zone, is it well size-sorted, probably as a consequence of accumulation by crystal settling. Throughout the core, the olivine has been recrystallized so that it lacks zoning, and many initial single crystals are now polygranular.

Electron microprobe analyses of pyroxenes and olivines (table V-26) indicate that the mineral compositions are fairly uniform throughout most of the core. Only in the uppermost sample, where the minerals are richer in Mg, does much deviation occur. A decrease in the TiO_2 and Al_2O_3 content of the augites with depth also is indicated.

Table V-26. Olivine and augite compositions from drill hole 5-3, sec. 36, T. 53 N., R. 15 W. (analyses by Warren C. Forbes).

Depth (feet)	Olivine (mole percent) Fo		Augite (mole percent) En	Fs	Wo	(wt. percent) TiO_2	Al_2O_3
120	50		36	19	45	1.1	1.9
375	43	A	36	24	40	1.0	
		B	40	23	37	2.1	1.8
515	40		35	22	43	0.9	1.7
713	44	A	36	21	43	0.7	1.8
		B	33	20	47	0.9	1.8
815	43		33	18	49	0.7	1.7
990	41	A	34	23	43	0.9	1.5
		B	36	23	41	0.6	1.4

Ilmenite, the only Fe-Ti oxide in the core, is abundant and locally accounts for as much as 40 percent of the rock. Evidently, it is paragenetically later than the associated silicate minerals. Apatite is moderately abundant in the lower part of the core and minor quantities of sulfides and graphite are present. A sample of typical peridotite from a depth of 880 feet was analyzed chemically (no. 13, table V-27).

Possibly, the body formed from a ferrogabbro magma that had abundant suspended plagioclase crystals. Because of the liquid fraction's high specific gravity, the initial load of suspended plagioclase crystals tended to rise to the upper side of the steeply-dipping body. This left a liquid that had the composition of an iron-rich titaniferous peridotite probably similar to that of no. 13, Table V-27. Inasmuch as ilmenite and apatite are paragenetically the latest minerals, the last liquid to crystallize evidently was even more enriched in Fe, Ti, and P than the peridotite magma. Judged from the presence of graphite, the abundance of ilmenite, and the total lack of magnetite-ulvospinel, the peridotite magma evidently was in a highly reduced condition. Thin dikes of fine-grained but similar peridotite and dunite cut the footwall hornfelses.

CHARACTERISTICS OF THE TROCTOLITIC SERIES

Troctolite and associated rock types of the troctolitic series comprise most of the exposed part of the southern half of the Duluth Complex. Augite-bearing troctolite is the most abundant type and augite troctolite is next most common. These types are gradational with olivine gabbro, augite-free troctolite, and anorthositic troctolite; local facies include picrite, dunite, troctolitic anorthosite, and norite. Oxides (ilmenite and magnetite-ulvospinel), hypersthene, and biotite generally occur in troctolite, but rarely are sufficiently abundant to be included in the rock name. Ferrogabbro and peridotite occur in some parts of the troctolitic series; Ca pyroxene (normally titanaugite) and oxides are abundant in these rocks but biotite and orthopyroxene are uncommon. Rocks containing large amounts of hypersthene, biotite, and oxides are most common in the vicinity of contacts of the troctolitic series with footwall rocks and inclusions.

Troctolite and associated rock types exhibit a wide range of mineralogic, textural, and structural characteristics. Many subtle variations of grain size, mineral proportions, and texture occur within short distances; these can be observed in some outcrops, but are most readily seen in drill core. In general, nearly as much variation can be seen in some long drill cores as can be seen throughout the entire southern half of the complex; other drill holes are monotonously uniform. Contacts between various textural and mineralogic varieties of troctolite or other rock types may be either sharp or gradational. Sharp contacts between various textural and mineralogic modifications of essentially the same rock type are sufficiently common that they are not considered as evidence for intrusion.

Structures

A wide variety of structures, including xenoliths, mineral segregations, layering, mafic pegmatites, igneous laminations, local dikes, and joints along which the minerals have been altered, characterize the troctolitic series. In some outcrops, the rocks contain structures indicative of a complicated geologic history; taking the southern half of the complex as a whole, however, the rocks in most outcrops are essentially homogeneous or have only a weakly developed igneous lamination. The most abundant structures are those involving layering and inclusions. The attitude of igneous lamination is internally consistent in most outcrops and generally is conformable to that in adjacent rocks. Invariably the alignment of plagioclase laths is parallel to layering.

The most apparent type of layering is rhythmic layering that involves different proportions of olivine and plagioclase. An idealized individual layer has a sharp lower contact and an olivine concentration at the base that grades upward to a plagioclase-enriched zone. Such layers commonly are a few inches to a few feet thick and, generally, several such layers are repeated at a given locality. Many layers have gradational contacts, with olivine being concentrated in the central or top part of a layer; at places, a sharp contact separates the olivine below from the plagioclase overlying it. Much layering consists of irregular short-range variations in the mineral proportions without a cyclic pattern. Olivine-plagioclase layering has little lateral continuity; individual layers normally pinch or fade out within an exposure. The most perfectly developed small-scale rhythmic layering in troctolite occurs in the Bardon Peak area at Duluth.

Irregularly shaped bodies as well as discontinuous lenses of picrite and dunite, as much as about 100 feet thick, have been intersected in many drill cores. Some have sharp contacts and include little or no troctolite, whereas others are gradationally interlayered with troctolite and anorthositic troctolite. The contacts between interlayered olivine- and plagioclase-rich rocks range from sharp to gradational and from planar to interpenetrating; mutual inclusions have been observed in a few cores. The picrite and dunite layers probably have the same origin as the thinner rhythmic layers that involve variable proportions of olivine and plagioclase. However, some of the dunite segregations clearly are xenoliths.

The olivine-rich segregations are variably serpentinized. Some serpentinized segregations have a weakly or strongly developed, shallow-dipping foliation that results from closely spaced, very thin, parallel serpentine veinlets which formed when olivine was converted to serpentine. As a consequence of the veinlets, the rocks appear strongly sheared; however, thin sections show that the initial igneous fabrics are well preserved.

Small- and large-scale layering involving changes in grain size or proportions of augite and Fe-Ti oxides occurs locally in troctolitic rocks, but is most common in ferrogabbro and late-stage ultramafic bodies. Layering involving pyroxenes and oxides is excellently developed in the previously described Greenwood Lake area, in the Water-Hen Creek layered complex, and in the core from drill hole 5-3 (fig. V-52).

Crosscutting relationships among different varieties of troctolite are common, as is irregular or highly contorted layering. Most of the complicated structures that have been observed are interpreted as having resulted from deformation that occurred when the troctolite was solid, or when it contained only a small amount of residual melt.

Composition

Although variable in texture and structure, the troctolitic rocks are fairly uniform mineralogically. A typical troctolite might contain 65-75 percent plagioclase, 15-30 percent olivine, 0-10 percent augite, 0-5 percent opaque oxides, and minor amounts of hypersthene, biotite, apatite, and sulfides. Modal analyses of troctolitic rock units from various localities in the complex (Taylor, 1964; Phinney, 1969 and this chapter; figs. V-44, V-45 and V-46) show that they are characterized by differences in both essential and accessory mineralogy. The range in mineral proportions in various rock units from the southern half of the complex is similar to the range found in the Gabbro Lake quadrangle (Phinney, this chapter). The layered series in the Duluth area (Taylor, 1964), however, generally contains more augite and less olivine than most other troctolitic rocks.

In picrite and dunite, augite commonly is sparse, whereas oxides and sulfides are as abundant or more abundant than in troctolite. In gabbro and ferrogabbro, plagioclase and titanaugite are most abundant (35-45 percent each), and are accompanied by opaque oxides (10-20 percent) and olivine (0-10 percent). Biotite and hypersthene are less common in ferrogabbro than in troctolite. Apatite is more common in many of the gabbroic rocks than in typical troctolites. Titanaugite, opaque oxides, and apatite are most abundant in the late-stage ultramafic rocks.

The most significant chemical variation (table V-27) among the rocks in the troctolitic series is the range in $MgO/[MgO + \Sigma Fe\ (FeO)]$[1]. In Table V-27, the analyses are arranged in order of decreasing $MgO/[MgO + \Sigma Fe\ (FeO)]$ which, presumably, is in order of increasing fractionation. Except for the tendency of TiO_2 and P_2O_5 to be greater in those rocks with lower $MgO/[MgO + \Sigma Fe\ (FeO)]$ ratios, the variations in the other components do not seem to be systematic.

With respect to mineral compositions in the troctolitic rocks, plagioclase from the layered series at Duluth (Taylor, 1964) ranges from An_{53} to An_{65} and olivine ranges from Fo_{53} to Fo_{61} (table V-25). In a drill core from near Babbitt (Hardyman, 1969, *op. cit.*), olivine generally ranges from Fo_{44} to Fo_{67}, but is as iron-rich as Fo_{27} in the lower 200 feet (fig. V-49). The plagioclase in this core ranges from An_{50} to An_{64}; most is more calcic than An_{60}.

A comparison of olivine and plagioclase compositions from various troctolitic units (fig. V-53) shows that the outer troctolite of the Bald Eagle intrusion has a more calcic plagioclase and a more magnesian olivine than the other bodies; presumably, the outer troctolite is less differentiated than the other troctolitic bodies. The compositions of olivine and plagioclase from other units that have been analyzed overlap considerably, but no strong compositional trends are evident. The most pronounced trend is an iron enrichment in olivine and a sodium enrichment in plagioclase from adjacent to the footwall in the core studied by Hardyman (1969, *op. cit.*).

The olivine in the peridotite of drill hole 5-3 (table V-26) is richer in iron than most of that in troctolitic rocks, but is not as iron-rich as some in the lower 200 feet of the core studied by Hardyman (1969, *op. cit.*).

The ranges in pyroxene compositions determined by Taylor (1964) from the layered series at Duluth (table V-25), by optical means, and by Hardyman (1969, *op. cit.*) by the electron microprobe (fig. V-49) are:

Table V-27. Chemical analyses of troctolitic series mafic and ultramafic rocks from the southern half of the Duluth Complex (1-10 troctolites, 11-13 ferrogabbro and ultramafic rocks).

	1	2	3	4	5	6	7	8	9	10	11	12	13
SiO_2	45.45	47.90	47.70	48.20	46.45	47.36	46.90	46.90	48.30	45.95	32.90	38.80	25.15
Al_2O_3	17.86	20.73	19.04	19.53	21.30	18.81	15.68	19.08	17.14	16.02	1.59	11.06	1.32
Fe_2O_3	0.09	0.22	0.87	Tr	0.81	1.30	1.23	1.52	4.01	0.10	13.25	4.79	2.40
FeO	9.48	7.52	8.84	10.60	9.57	10.65	10.58	8.72	8.64	11.26	21.06	13.44	31.96
MgO	11.20	7.73	8.65	9.28	7.90	8.59	8.61	7.20	7.95	7.30	20.14	7.65	13.70
CaO	10.90	10.42	8.96	8.51	9.83	8.33	10.11	8.02	8.18	12.56	0.50	17.10	6.20
Na_2O	2.58	3.01	2.53	2.52	2.14	2.94	2.32	2.58	2.80	3.41	Tr	1.99	0.081
K_2O	0.21	0.19	0.53	0.32	0.34	0.42	0.57	0.83	0.21	0.54	Tr	0.44	<0.01
TiO_2	0.87	0.79	1.80	0.65	1.19	1.10	2.59	1.03	1.09	1.65	5.36	3.73	16.03
P_2O_5	0.08	0.09		0.19	0.02	0.11	0.10	0.17	0.18	0.24	Tr	0.13	0.98
MnO	0.13	0.12	Tr	0.14	Tr	0.14	0.16	0.10	0.11	0.15	0.40	0.18	0.44
S	0.017	0.016		0.03				1.102	0.071	0.071	0.05	0.054	0.60
H_2O	0.68	0.62	1.38	0.73	1.16	0.29	1.00	1.30	0.71	0.81	5.11	0.60	0.50
CO_2	0.03	0.09		0.02		0.01	0.03		0.15	0.11	0.10	0.08	0.08
Total	99.577	99.446	100.30	100.72	100.75	100.06	99.80	98.922	99.541	100.171	100.71	100.044	99.441
$\frac{MgO}{MgO+Fe(FeO)}$	.540	.500	.473	.469	.434	.423	.419	.417	.394	.391	.379	.301	.287

Description and location of samples:

1. Med.-gr. foliated troctolite, 68% plagioclase, 26½% olivine, 4% pyroxene and 1½% oxides; olivine and plagioclase are well sorted by size; SE¼ sec. 32, T. 60 N., R. 12 W.; exposure at Y in railroad; sample A-208
2. Med.-gr. poikilitic augite-bearing troctolite, decussate texture, poor size sorting; cen. N½ sec. 11, T. 59 N., R. 13 W.; 300 feet E. of railroad jct.; sample A-69
3. Olivine gabbro; Birch L., sec. 35 (?), T. 61 N., R. 12 W. (Ruotsala and Tufford, 1965, no. 17, p. 29)
4. Olivine gabbro, layered series; West Duluth, sec. 23, T. 49 N., R. 15 W. (Taylor, 1964, no. 8, p. 29)
5. Olivine gabbro, 0.04% NiO (Ruotsala and Tufford, 1965, no. 8, p. 27); near S¼ post, sec. 35, T. 61 N., R. 12 W.
6. Banded troctolite, layered series, 57% plagioclase, 40% olivine, 2% magnetite and 1% clinopyroxene; includes 0.01% SrO; Bardon Peak, upper railroad tracks, sec. 34, T. 49 N., R. 15 W. (Taylor, 1964, no. 7, p. 29); sample M4634
7. Olivine gabbro, Bardon Peak intrusion; lower railroad tracks, sec. 33, T. 49 N., R. 15 W. (Taylor, 1964, no. 11, p. 29); sample M4633
8. Coarse-gr. hypersthene-bearing anorthositic troctolite, 0.37% C; drill hole B1-134, depth 1,199 ft.; lower contact of 32-foot hornfels inclusion
9. Med.-gr. poikilitic augite-bearing troctolite from drill hole USS26029, depth 593 ft.; 4 ft. above top of 53-foot hornfels inclusion
10. Med.-gr. poikilitic augite-bearing troctolite; drill hole USS17700, depth 222½ ft.; basal contact of complex
11. Peridotite, Bardon Peak intrusion, 0.04% Cr_2O_3, 0.15% Cu_2O and 0.06% rarer elements; Short Line Park, sec. 34, T. 49 N., R. 15 W. (Taylor, 1964, no. 12, p. 29)
12. Med.-gr. foliated ferrogabbro, 41% plagioclase, 41% titanaugite, 15% magnetite and 3% olivine; SE¼ sec. 35, T. 59 N., R. 11 W., along railroad, sample T-22
13. Coarse-gr. peridotite, mainly titanaugite with ilmenite and olivine, minor apatite; sec. 36, T. 53 N., R. 15 W.; drill hole 5-3, depth 880 ft.

The locations of drill holes USS17700, USS26029 and B1-134 and the analytical reference for samples 1, 2, 8, 9, 10, 12, and 13 are given in Table V-29.

	Augite			Hypersthene		
	En	Fs	Wo	En	Fs	Wo
Layered series	36-45	18-26	35-39	57-66		
Core studied by Hardyman	38-47	12-21	39-48	56-73	27-41	1-2
Lower 200 feet studied by Hardyman	33-42	18-28	42-46	41-62	36-56	1-2

[1] ΣFe (FeO) = total iron expressed as FeO.

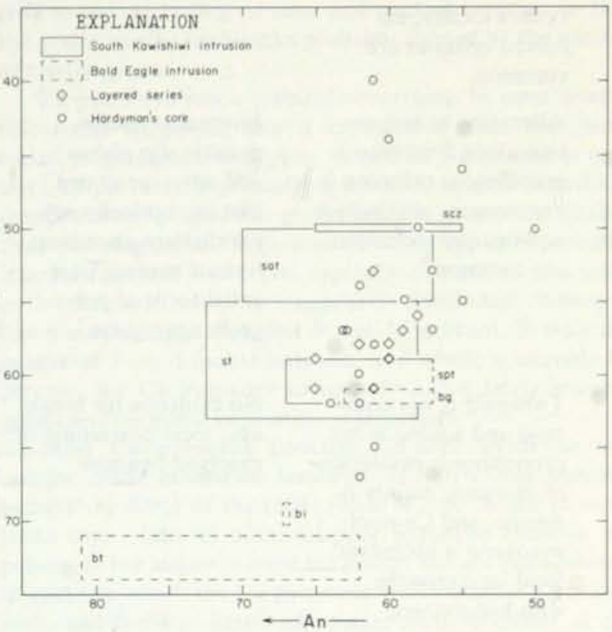

Figure V-53. Plot of olivine vs. plagioclase compositions for troctolitic series rocks.

Comparison of augite compositions from the late-stage peridotite in drill hole 5-3 (table V-26) with those listed above and in the Gabbro Lake quadrangle (see Phinney, this chapter), reveals that the augite in the peridotite is only slightly richer in iron than that in troctolitic rocks, and that augites in some troctolitic rocks, such as in the lower 200 feet of the core studied by Hardyman (1969, *op. cit.*) are richer in iron than those in the peridotite.

Most of the Ca pyroxene from the layered series is titanaugite, which Taylor (1964) believed is chemically and optically similar to that reported by Hess (1949). Three samples from the layered series contain 3.48, 2.83, and 2.70 percent Al_2O_3, and 1.03, 0.99, and 1.05 percent TiO_2, respectively. Their Al_2O_3 content is very similar to that (3.45 percent) of the augite from the Gabbro Lake quadrangle, reported in the accompanying paper by Phinney; however, the TiO_2 content (0.12 percent) of the Gabbro Lake augite is quite low. The available data on the TiO_2 content of augites from rocks of the troctolitic series (table V-26) suggest that a trend exists involving Ti-enrichment with a decrease in the Mg/Fe ratio of the augites.

In the layered series, Taylor (1964) reported that the orthopyroxene mainly is inverted pigeonite. This contrasts with orthopyroxenes from other parts of the troctolitic series; neither Phinney (1969 and this chapter) nor Hardyman (1969, *op. cit.*) mentioned evidence of inverted pigeonite in troctolitic rocks. Inverted pigeonite is uncommon among troctolitic and gabbroic samples that I collected from the southern half of the complex.

Textures

The textures of troctolites and other rocks reported from various parts of the complex (Taylor, 1964; Hardyman, 1969, *op. cit.*; Phinney, 1969) suggest that the physical processes which formed the rocks were variable. The common textural characteristics of plagioclase, olivine, pyroxene, and opaque oxides are summarized in Table V-28.

A fairly well-defined paragenetic sequence, deduced from grain shapes and distribution, can be established in most samples. In most troctolites, augite troctolites, anorthositic troctolites, and olivine gabbros, olivine and plagioclase are in large part texturally earlier than the other minerals. In many rocks they appear essentially contemporaneous, but in some the plagioclase appears to have started crystallizing prior to the olivine. Evidently, nearly all olivine in some troctolites crystallized later than the associated plagioclase. In picrites, and especially dunites, the olivine is partly or entirely earlier than the plagioclase. In nearly all troctolites the pyroxenes and opaque oxides appear to have crystallized later than olivine and plagioclase. In the ferrogabbros, however, Ca pyroxene and the oxides commonly appear to be approximately contemporaneous with olivine, and in some, contemporaneous with plagioclase as well. In norites, plagioclase, olivine, and the pyroxenes commonly are approximately contemporaneous; in some, however, plagioclase is earlier than the mafic minerals and hypersthene locally is the earliest phase.

Inasmuch as plagioclase is the principal constituent of the troctolitic rocks, its form and distribution determine the main elements of the rock fabrics. Generally it occurs as anhedral or subhedral prismatic tablets 1 to 10 mm long that have large weakly zoned core areas enclosed in relatively thin sodic rims. The boundaries between most zones are gradational. It is uncommon for plagioclase to lack internal zoning, but it is even less common for it to be strongly zoned. Most plagioclase contains both carlsbad and albite twins and some also contains other types. The albite twins commonly form rather complicated patterns and are interrelated with the fracturing, bending, and other internal features of the grains.

Table V-28. Textural characteristics of minerals in rocks from the troctolitic series.

Mineral	Occurrence	Size	Internal features	Evidence for breakage and recrystallization
Plagioclase	Anhedral and subhedral prismatic tablets, and irregular grains; rarely euhedral; interstitial in ultramafic rocks	Laths typically 1-10 mm long; length to width ratio generally 2:1 to 5:1; sorting according to size generally is poor; some troctolites are porphyritic	Twinning is ubiquitous; zoning generally is present but weak, both normal and oscillatory are common, patchy zoning occurs locally; exsolved opaques are common	Broken external forms, bent and offset twin lamellae; zoning weak or absent and twinning less complex in recrystallized rocks
Olivine	Commonly equidimensional to irregular grains in troctolite, but locally poikilitic; subhedral to euhedral in picrite, dunite and some troctolites; interstitial in anorthositic rocks and where low in abundance	Grains 0.1-5 mm in diameter; poor size sorting except where abundant; excellent sorting in most dunites; phenocrysts are uncommon	Alteration to serpentine along fractures is common; twinning is uncommon; plagioclase and opaque inclusions are common	Internal crystallographic slip planes and adjacent grains that are optically subparallel are abundant; lack of zoning; interstitial form of polygrain aggregates
Augite	Interstitial or poikilitic grains in most rocks; irregular grains and subhedral prisms in some gabbros, ferrogabbros and norites; locally replaces olivine margins	Oikocrysts are as much as 3-4 inches in diameter in many rocks; oikocryst size is quite variable but commonly is uniform in a given rock	Twinning is not common and zoning is not pronounced; exsolution of opaques, mainly ilmenite, and Ca-poor pyroxene is abundant; local intergrowths with hypersthene	No evidence for breakage; local coarsening of exsolved opaques
Hypersthene	Interstitial grains, rims on olivine, and symplectic intergrowths with plagioclase and ilmenite; locally as irregular grains and rarely as subhedral grains in norite	Oikocrysts are large only where hypersthene is abundant; grains as much as a few mm in diameter occur locally in norite and hypersthenite	Twinning and zoning are not common; locally intergrown with clinopyroxene; exsolved augite along (100) is common but evidence for inverted pigeonite occurs only locally in troctolites	No evidence for breakage; symplectic intergrowths with plagioclase are very sensitive to recrystallization so that vermicular hypersthene is changed to equidimensional inclusions in plagioclase
Opaque oxides (magnetite and ilmenite)	Common as irregular, poikilitic and interstitial grains but not abundant as euhedral or subhedral grains; both are common as inclusions in plagioclase; ilmenite commonly occurs in intergrowths with hypersthene and exsolved in augite	Oikocrysts seldom over 0.5 inch across; euhedral and subhedral grains normally less than 0.1 mm in diameter	Ilmenite locally is twinned; magnetite commonly contains exsolved ilmenite, ulvospinel and spinel	Ilmenite locally has exsolved to magnetite grain margins; some poikilitic opaque areas are polygranular

In troctolitic rocks, olivine occurs mainly as grains 0.1 to 5 mm in diameter; the grains are either equidimensional and evenly dispersed or occur as aggregates containing a few equidimensional to irregular grains. The aggregates commonly have overall shapes suggesting that they originated by recrystallization of a larger single grain. Many olivines contain internal dislocations across which slight optical disorientations are seen. This evidently is an intermediate step in the recrystallization of larger initial grains to form polygrain aggregates. In picrites and dunites, olivine commonly occurs as equidimensional grains that are well sorted according to size, and packed together so that the grains touch. Such rocks probably formed by the settling of olivine grains.

Ca pyroxene has a textural occurrence in most troctolitic rocks suggesting that it crystallized later than associated plagioclase and olivine. Where its abundance is very low (<1 percent), it generally is confined to local interstitial areas between plagioclase grains and the grains are smaller than the plagioclase. Where more abundant (1-5 percent), it occurs as small oikocrysts, typically one-half to two inches in diameter, or it is irregularly distributed. Where it has a modal range between 5 and 15 percent, it typically occurs as 1- to 4-inch oikocrysts, and where it exceeds 15 percent, the Ca pyroxene commonly forms fairly massive grains and incipient pegmatitic segregations.

Most Ca pyroxene contains Ca-poor pyroxene and opaque oxide exsolution lamellae. The pyroxene lamellae occupy the (001) or the (100) plane of their hosts; in some rocks they occur in both. Ca-poor pyroxene lamellae are present in the augite in most troctolites but are not common in rocks in which the Ca pyroxene is titanaugite. Lamellae, rods, and blebs of exsolved opaque material occur in the Ca pyroxene in most troctolites, and commonly are more conspicuous than Ca-poor pyroxene lamellae; they are even more abundant in the pyroxene in some gabbros and ferrogabbros. The opaque material that I checked is ilmenite; possibly it is magnetite in other rocks, and pseudobrookite was reported by Phinney (1969). Apparently, the inclusions are quite susceptible to grain-size coarsening; in many pyroxene grains the opaque particles are relatively large, and it seems clear that they were formed by the merging of initially smaller bodies.

Orthopyroxene is interstitial to the plagioclase and olivine in troctolitic rocks, but seems to be approximately contemporaneous with Ca pyroxene and opaque oxides. It rarely occurs as discrete grains approaching equidimensional or prismatic shapes. Where abundant, it occurs as oikocrysts or as irregular to interstitial grains. Thin discontinuous orthopyroxene rims mantle olivine grains in many rocks; very likely, these have formed from the olivine by reaction with late-stage SiO_2-enriched interstitial melt.

Vermicular or symplectic intergrowths of orthopyroxene in a crystallographically continuous plagioclase host are common in troctolitic rocks, especially those from the basal zone of the complex and adjacent to Virginia Formation inclusions. The orthopyroxene generally forms one-fourth to one-third of the volume of such intergrowths. These intergrowths occur mainly along grain boundaries between olivine and plagioclase, along boundaries between plagioclase grains, and in mesostasis areas where oxides, biotite, sulfides, and apatite are localized. Most occurrences of this intergrowth probably formed after the enclosing rocks had crystallized. This is suggested by the manner in which the material embays the edges of plagioclase laths and by its distinctly intergranular distribution. The plagioclase-orthopyroxene intergrowths, as well as local orthopyroxene-oxide and orthopyroxene-sulfide intergrowths, which occur in mesostasis areas, may have formed from the last stage of melt crystallization, however.

The opaque oxides ilmenite and magnetite-ulvospinel, are very common in the troctolitic rocks; they are mainly interstitial to plagioclase and olivine, but appear to be contemporaneous with the pyroxene. Oxides form distinct oikocrysts where relatively abundant, but more commonly occupy small interstitial corners between plagioclase laths. A few troctolitic rocks contain subhedral to euhedral magnetite or ilmenite, which probably was deposited synchronously with associated plagioclase and olivine.

Reddish-brown biotite is a minor constituent in almost all troctolitic rocks, and is especially abundant in rocks from the basal zone of the complex; it is uncommon in ferrogabbro. Biotite shows a strong textural association with opaque oxides, and to a lesser extent with olivine and pyroxene. Its most typical occurrence, and its only occurrence in many rocks, is as a reaction corona partially rimming or replacing the outer part of opaque oxide grains. In most rocks, biotite is interpreted as having formed after the rock had crystallized.

Apatite is widely present and most abundant in some gabbro and late-stage peridotite bodies, but its concentration seldom exceeds one percent. It occurs in a variety of morphologies ranging from long slender needles to short stubby prisms and anhedral masses.

The degree of sorting according to grain size varies considerably. In some rocks, all plagioclase grains are nearly equal in size; in others, there is a large range of sizes or the range is distinctly bimodal. For bimodal distributions, the small size fraction generally occupies the largest volume (porphyritic texture). The large size fraction occupies the greater volume in some rocks, however. In many such rocks, this texture may have been formed by granulation, and subsequent annealing, of previously larger grains. If the plagioclase grains in a rock are uniform in size, olivine grains also tend to be uniform. Troctolites are generally non-porphyritic and poorly sorted. Rocks with porphyritic textures seem to be more common than rocks with well developed sorting. In picrites and dunites, the olivine grains tend to be very nearly the same size. In some ferrogabbros all the minerals are well sorted; such rocks also commonly have moderately to strongly developed igneous lamination.

The grain-to-grain packing varies considerably among troctolitic rocks from various areas. In most, the arrangement suggests crystallization from a mush comprised of a liquid containing a high percentage of crystals. Paragenetically early grains generally are packed sufficiently closely to have formed a self-supporting pile, or to have substantially interfered with one another in flow. Plagioclase is dispersed in some rocks so that most grain margins are in contact with other minerals. Local parts of many bodies, however, consist only of plagioclase, and in these, plagioclase grains abut other plagioclase grains. Such plagioclase domains range from 2 or 3 grains across to areas a few inches across that involve hundreds of grains. Commonly, the interstitial minerals are abundant at the periphery of such plagioclase domains. Such an arrangement may be the result of slow cooling, during which various elements in the melt migrated to established crystal growth surfaces.

A common textural feature of troctolite is the presence of one-half to six-inch oikocrysts of pyroxenes, oxides and, locally, olivine. These also may be the result of very slow cooling which resulted in widely spaced crystal nucleii. Generally, plagioclase grains enclosed within the oikocrysts are isolated from one another and are subhedral or slightly rounded. Plagioclases between oikocrysts, however, commonly share mutual boundaries with little or no interstitial material.

Bending and internal slip surfaces within plagioclase grains and other evidence of grain deformation and breakage are common in troctolite and associated rocks (see table V-28). Deformed plagioclase laths are present in the majority of troctolites from the southern part of the complex, but are uncommon in rocks with well sorted plagioclase and well developed layering and igneous lamination.

Evidence of post-consolidation recrystallization (table V-28) is common in the troctolitic rocks. Complete recrystallization, involving virtual destruction of zoning and simplification of the twinning in plagioclase, removal of small irregularities along grain boundaries, development of generally anhedral equidimensional grains, and development of mutual inclusion relations, has resulted in textures resembling those in hornfels inclusions. In most rocks, the finer grained portions, which are more sensitive to recrystallization, provide evidence for incipient or partial recrystallization. The symplectic plagioclase-hypersthene intergrowths are one of the most sensitive indicators of partial textural re-equilibration. In some rocks, the fine-grained areas are hornfelsed, whereas coarser grained parts of the same rock have retained an igneous texture. Apparently, olivine is more easily recrystallized than the coarse-grained plagioclase, for it is common for olivine to occur in polygranular aggregates that have formed from initially large single grains, whereas adjacent plagioclase laths have remained essentially undisturbed.

Origin

The wide range of structural and textural fabrics in rocks of the troctolitic series indicates considerable variation in the physical processes of magmatic evolution. Clearly, a few rocks, particularly some of the dunite and picrite layers within troctolite, were formed by settling of crystals from a melt. Others have textures and structures suggesting that they may be marginal or internal flow segregations, or float accumulations from exceptionally dense melts. Still other rocks evidently crystallized in place, and have the same composition as the melt they formed from. Internal deformation and crystal granulation and related recrystallization, which are strongly developed in some rocks, suggest that such rocks were emplaced as crystal mushes, in which the crystals interfered with one another as the mass moved.

The late-stage peridotite encountered in drill hole 5-3 (fig. V-52) is particularly instructive. The abundant broken plagioclase laths that evidently floated upward in this body suggest that its emplacement was accompanied by rather intense deformation of the host rocks. This, and the body's shape, suggest that it was localized in a dilatational shear that resulted from tectonic movements late in the history of the complex. The mineralogic similarity of the peridotite to the interstitial components in troctolitic rocks suggests that the peridotite magma was local in origin. Most likely it formed from interstitial liquids that were present in nearby partly consolidated troctolitic rocks, or possibly in anorthositic rocks. Such interstitial liquids would have migrated into a dilatation zone as it developed.

It is my opinion that many of the troctolitic magmas were intruded as crystal mushes, in which crystallization continued for a considerable time as the magmas moved into their positions of final consolidation. Where there was a high proportion of melt, the crystals easily segregated to form such rocks as picrite, dunite, and augite-poor troctolite, but where the melt fraction was low, the liquid parts crystallized in contact with the earlier suspended grains. Where the amount of melt was exceptionally low, extensive grain granulation occurred. The high proportion of crystals in the magmas imparted considerable mechanical strength to them; this is manifested by plucking and local deformation of the footwall and the general comminution of inclusions, wedges of footwall material, and previously crystallized intrusive rocks. Apparently, the ferrogabbro and late-stage ultramafic rocks crystallized from liquids left over after the separation of troctolite. On the other hand, norite and other rocks with considerable hypersthene most likely owe their origin in part to contamination processes, which are discussed in more detail below.

Many lines of evidence support Taylor's (1964) suggestion that tectonic instability characterized the environment of crystallization. The Duluth Complex is on the flank of the Midcontinent Gravity High, a feature I interpret as a rift zone filled with mafic material upwelled from the earth's mantle. Probably considerable tectonic activity occurred while this zone was being formed and filled.

HORNFELSES

Hornfels bodies as much as thousands of feet in maximum dimension are abundant in many parts of the Duluth Complex. The smaller bodies definitely are inclusions within the intrusive rocks, whereas some of the larger ones are situated between adjacent intrusive bodies and may not be true inclusions. The hornfelsed inclusions represent a variety of rock types, many of which constitute a part of the footwall. Their distinguishing characteristics are their fine grain size (generally less than one mm average diameter), granoblastic texture, and their fabrics or structures, which indicate that they are older than the igneous rocks in which they occur.

Footwall Rocks

The principal rocks constituting the footwall of the southern part of the Duluth Complex are granitic rocks of the Lower Precambrian Giants Range batholith, argillite and siltstone of the Middle Precambrian Virginia Formation, the Middle Precambrian Biwabik Iron-formation, and Keweenawan basaltic lavas. The granitic rocks form the footwall of the complex at the surface northeast of Birch Lake and in the subsurface southwest of Birch Lake at least to Hoyt Lakes. The Virginia Formation forms the footwall throughout most of the Babbitt-Hoyt Lakes region (see fig. V-44), and, together with the Thomson Formation, its metamorphosed equivalent to the south, probably comprises the footwall in most of the region between Hoyt Lakes and Duluth (Sims, 1970). Also, the Virginia Formation occurs abundantly as inclusions within the lower part of the complex. The Biwabik Iron-formation is in contact with the complex at the surface in parts of the Dunka River area and in the subsurface throughout the Babbitt-Hoyt Lakes region, and also occurs locally as inclusions in the Duluth Complex.

Laminations, graded bedding, local intraformational pebbles, slump structures, and other primary structures are well preserved in the metamorphosed footwall rocks and generally are evident in all but the smallest inclusions.

In the Duluth area, Keweenawan basalt, which has been hornfelsed adjacent to the contact, forms the footwall of the complex (Taylor, 1964, pl. 1). Hornfelsed basalt also occurs immediately below the complex in sec. 18, T. 59 N., R. 13 W. (fig. V-44), and the presence of this occurrence together with the substantial volume of hornfelsed volcanic rocks within the complex suggest that volcanic rocks may form the footwall at places between Hoyt Lakes and Duluth.

Virginia Formation Hornfels

Adjacent to the Duluth Complex, the argillaceous rocks of the Virginia Formation have been converted to feldspar-bearing cordierite-biotite hornfelses that vary from fine to very fine grained, becoming progressively finer grained away from the contact. Pyrrhotite, graphite, and orthopyroxene are common minor constituents, and ilmenite, chalcopyrite, pyrite, quartz, vesuvianite, garnet, andalusite, staurolite, chlorite, muscovite, anthophyllite, wollastonite, and Ca pyroxene occur locally. The inclusions of Virginia Formation in the complex are similar to the footwall rocks, but differ in being coarser grained and in having slightly different proportions of the constituent minerals. In particular, plagioclase and orthopyroxene are more abundant and biotite is less abundant. Plagioclase, orthopyroxene, and opaque minerals occur mainly as anhedral, equidimensional grains, but some of the biotite and orthopyroxene and much of the cordierite is poikiloblastic. Cordierite also occurs as small grains; biotite and graphite generally occur as flakes. Potassium feldspar occurs as grains and as exsolved blebs in plagioclase. The equilibrium silicate mineral assemblage of the hornfelses derived from the predominant argillaceous facies of the Virginia is considered to be cordierite, orthopyroxene, biotite, plagioclase, and locally potassium feldspar.

In some samples from the footwall and from small inclusions, part of the cordierite occurs as irregularly shaped grains varying from intergranular to poikilitic. These occur between and enclose rounded grains of feldspars and additional cordierite. In some of these samples the feldspars also are partly poikilitic, enclosing one another and cordierite. These textures are interpreted as having originated by the recrystallization of a partial melt.

To determine the chemical changes resulting from metamorphism of the Virginia Formation and to aid in distinguishing it from hornfelsed volcanic rocks, several samples of both unmetamorphosed and metamorphosed Virginia as well as volcanics were analyzed (tables V-29 and V-30). Oxide ratios for the rocks are given in Table V-31. Additional analyses of similar materials were published previously (Grout, 1930, 1933a; Morey, 1969). An $Al_2O_3-SiO_2-(FeO+MgO)$ plot of various hornfelses (fig. V-54) shows that the Virginia is more aluminous than the other hornfelses, although both groups have a considerable range in proportions of SiO_2. In a $CaO/(Al_2O_3+CaO)$ vs. K_2O plot (fig. V-55A), the sedimentary materials clearly have lower ratios. The exception, analysis 3, is an unmetamorphosed calcareous siltstone. In the CaO vs. $MgO/[MgO+\Sigma Fe(FeO)]$ plot (fig. V-55B), it can be seen that the samples of Virginia Formation are low in CaO and have a relatively narrow range of $MgO/[MgO + \Sigma Fe(FeO)]$ values as compared to the volcanic rocks. The exceptions are the calcareous siltstone (no. 3, table V-29), with a high CaO content, and a contact rock (no. 18, table V-29), that is enriched in Mg relative to Fe.

The analyzed samples of the Virginia Formation show systematic variations in proportions of SiO_2, Al_2O_3 and $(FeO+MgO)$, as indicated by the three groups (A, B and C) on Figure V-54. The most siliceous group (A) includes all the unmetamorphosed and all but two of the metamorphosed footwall samples, but none of the inclusions. The exceptions are sample 10, which has a texture indicating recrystallization from a partial melt, and sample 17, which was taken within inches of the basal contact. Group B, with an intermediate SiO_2 content, includes all samples of inclusions except 18, which is from the edge of a large inclusion. The two samples (17 and 18) forming the least siliceous group (C) were taken within a few inches of contacts. The extensive SiO_2 loss inferred for samples 17 and 18 is accompanied by an apparent decrease in K_2O (fig. V-55). The other samples (11, 12 and 19) that may have lost K_2O are the remaining three from large inclusions. The small inclusions show little, if any, deviation in K_2O content from their initial values. Evidently, Na_2O has not been depleted in as many samples as has K_2O; it has been lost from 17 and 18, the two contact samples, however.

The $CaO/(CaO+Na_2O)$ ratio (table V-31), which should indicate approximate plagioclase compositions, generally shows higher values for the samples having relatively low SiO_2. For the intermediate and low-silica groups (B and C) of Figure V-54, the $CaO/(CaO+Na_2O)$ ratios are .452 or lower for those with partial melt textures, and .493 or higher for those that lack the texture. Similarly, the $K_2O/(K_2O+Na_2O)$ ratio of the four samples with partial melt textures is .531 or higher, whereas the other rocks have ratios of .372 or lower.

Except for sample 18, the $MgO/[MgO + \Sigma Fe(FeO)]$ ratios for all the analyses in Table V-29 fall within a relatively narrow range. Sample 18 is from the margin of a large inclusion; its Fe impoverishment with respect to Mg,

Table V-29. Chemical analyses of unmetamorphosed and thermally metamorphosed Virginia Formation and of hornfelsed Virginia inclusions from the Duluth Complex.

	Unmetamorphosed				Thermally metamorphosed (from footwall of Duluth Complex)						From interiors of large inclusions		From small inclusions				From contacts		
	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19
SiO_2	59.90	65.10	45.15	57.10	57.65	62.80	58.85	60.40	62.40	53.40	45.15	46.15	58.15	47.55	54.65	54.15	44.95	45.85	49.75
Al_2O_3	15.16	12.09	9.86	17.04	18.74	15.66	17.18	17.53	16.65	19.56	18.73	16.84	18.13	23.82	17.42	19.09	18.56	19.88	18.10
Fe_2O_3	0.98	0.38	2.11	0.93	0.29	0.42	0.72	0.38	0.40	0.53	0.28	0.83	0.35	0.62	1.17	0.61	0.70	1.49	1.36
FeO	5.36	4.47	3.04	6.64	7.82	5.68	7.54	7.12	5.92	10.60	12.20	13.48	8.58	11.04	9.20	9.90	17.80	13.02	9.88
MgO	3.35	2.53	2.02	3.96	4.80	3.20	4.15	3.97	3.55	6.30	5.65	5.20	5.10	6.68	4.85	6.03	9.55	14.10	5.60
CaO	2.60	5.10	16.05	0.98	1.02	2.62	2.33	1.21	1.18	0.98	6.37	6.56	1.76	1.18	4.04	1.66	1.46	0.78	6.92
Na_2O	2.79	3.80	2.56	1.99	2.44	3.88	2.26	2.36	2.59	1.72	2.88	3.07	2.13	1.71	4.16	2.11	1.03	0.312	3.63
K_2O	2.36	1.04	1.31	3.27	3.32	2.10	3.20	3.34	3.19	3.34	0.68	0.25	2.41	3.03	1.91	2.45	0.61	0.17	0.41
TiO_2	0.80	0.62	0.48	0.80	0.89	0.86	0.89	0.87	0.78	1.17	0.88	1.02	0.96	1.36	0.94	1.10	0.91	1.37	1.03
P_2O_5	0.26	0.23	0.18	0.21	0.21	0.24	0.21	0.22	0.19	0.20	0.14	0.14	0.17	0.07	0.17	0.26	0.15	0.12	0.18
MnO	0.10	0.07	0.34	0.07	0.07	0.07	0.08	0.06	0.06	0.10	0.14	0.18	0.06	0.08	0.08	0.09	0.16	0.16	0.17
S	0.455	0.180	1.73	0.874	0.22	0.09	0.09	0.089	0.48	0.290	3.44	3.48	0.178	1.18	0.795	0.308	3.86	1.32	1.60
H_2O	3.57	2.04	1.88	4.40	1.62	1.46	1.43	1.56	1.59	1.12	1.16	0.64	1.55	1.32	0.97	1.85	1.04	0.70	0.74
SO_2	1.44	1.94	13.17		0.18	0.29	0.39	0.21			1.68		0.15	0.07			0.22		
S	0.39		0.54	1.64					0.86	0.16	2.01	3.67			0.17	0.18	0.28	0.51	0.53
Total	99.515	99.59	100.42	99.904	99.27	99.37	99.32	99.319	99.84	99.47	101.39	101.51	99.678	99.71	100.525	99.788	101.28	99.782	99.90

Description and location of samples (only the most abundant minerals are listed):

1. Distinctly laminated black and gray silty argillite; drill hole MDD2, depth 231 ft.; hole 2, Mesabi deep drilling project (Biwabik hole) SW¼SE¼ sec. 22, T. 58 N., R. 16 W.; details of MDD2 given in Pfleider and others, 1968
2. Indistinctly laminated gray argillaceous siltstone; drill hole MDD2, depth 511 ft.
3. Indistinctly layered gray calcareous siltstone, locally pyritic; drill hole MDD2, depth 1,243 ft.
4. Thinly laminated black argillite; drill hole MDD2, depth 1,532 ft.
5. Cordierite-biotite-plagioclase-bearing hornfels, 7 ft. below base of Duluth Complex; drill hole USS17700, depth 230 ft.; NE¼NW¼ sec. 34, T. 59 N., R. 14 W.
6. Cordierite-biotite-plagioclase-bearing hornfels, 17 ft. below base of Duluth Complex; drill hole USS17700, depth 240 ft.
7. Cordierite-biotite-plagioclase-bearing hornfels, 227 ft. below base of Duluth Complex; drill hole USS17700, depth 450 ft.
8. Cordierite-biotite-plagioclase-bearing hornfels, 326 ft. below base of Duluth Complex; drill hole USS17700, depth 549 ft.
9. Cordierite-biotite-plagioclase-bearing hornfels, 345 ft. below base of Duluth Complex; drill hole USS17700, depth 568 ft.
10. Cordierite-potassium feldspar-orthopyroxene-biotite-plagioclase-bearing hornfels, sample M-12233; near cen. NE¼ sec. 3, T. 60 N., R. 12 W.; 10,300′ N., 2370′ E., Erie Mining Co. grid (plate 1, Bonnichsen, 1968, unpub. Ph.D. thesis, Univ. Minn.)
11. Plagioclase-orthopyroxene-pyrrhotite-bearing hornfels, 18 ft. above base of 32-foot-thick hornfels inclusion; drill hole B1-134, depth 1,181 ft.; SW¼NW¼ sec. 33, T. 60 N., R. 12 W.
12. Plagioclase-orthopyroxene-pyrrhotite- and graphite-bearing hornfels, 4 ft. below top of 53-foot-thick hornfels inclusion; drill hole USS26029, depth 619 ft.; NW¼SE¼ sec. 9, T. 59 N., R. 13 W.
13. Cordierite-biotite-orthopyroxene-plagioclase-bearing hornfels, cen. of one-foot-thick inclusion 157 ft. above base of Duluth Complex; drill hole USS17700, depth 66 ft.
14. Cordierite-potassium feldspar-orthopyroxene-biotite-plagioclase-bearing hornfels; from an indistinct inclusion a few inches thick, 19 ft. above base of Duluth Complex; drill hole USS26029, depth 1,080 ft.
15. Plagioclase-orthopyroxene-biotite hornfels inclusion (2-3 in. dia.) in biotite "norite" host of sample 16; taken from loose material blasted from basal zone of Duluth Complex in Dunka River area; sample M-12275, inclusion; E. side, SE¼SW¼ sec. 3, T. 60 N., R. 12 W.
16. Cordierite-biotite-plagioclase-orthopyroxene hornfels or norite enclosing sample 15; a sharp contact occurs between 15 and 16; sample M-12275, host
17. Cordierite-orthopyroxene-pyrrhotite-bearing hornfels, from less than 6 in. below base of Duluth Complex; drill hole USS17700, depth 223 ft.
18. Cordierite-orthopyroxene-pyrrhotite-bearing hornfels, from lower contact of 32-foot-thick inclusion; drill hole B1-134, depth 1,199 ft.
19. Plagioclase-orthopyroxene hornfels, from approx. 4 ft. below top of 53-foot-thick inclusion; drill hole USS26029, depth 601 ft.

The above samples, as well as numbers 1, 2, 8, 9, 10, 12 and 13 of Table V-27 and numbers 1, 2, 3, 4 and 5 of Table V-30 were analyzed by K. Ramlal, Univ. of Manitoba. Details of this method are reported in Wilson and others, 1969.

Table V-30. Chemical analyses of various volcanic rocks and hornfelses from the southern part of the Duluth Complex (analyses 1, 2, 6, and 7 are basalt hornfels).

	1	2	3	4	5	6	7
SiO_2	47.70	49.55	47.25	40.20	57.55	41.31	42.98
Al_2O_3	15.61	14.07	12.84	10.98	10.84	12.12	13.09
Fe_2O_3	1.16	1.44	3.63	4.29	6.56	3.52	3.16
FeO	6.56	9.52	7.72	12.42	6.04	14.57	13.38
MgO	10.92	8.60	7.75	8.35	4.60	6.58	6.44
CaO	12.04	10.86	15.90	17.09	7.21	11.07	11.33
Na_2O	2.09	2.56	2.19	1.72	3.02	2.06	2.22
K_2O	0.21	0.33	0.09	<0.01	0.14	0.16	0.23
TiO_2	0.31	0.76	1.23	3.47	2.13	7.04	6.52
P_2O_5	0.05	0.07	0.27	0.36	0.35	0.63	
MnO	0.13	0.23	0.19	0.26	0.21	0.21	0.21
S	0.013	0.010	0.016	0.020	0.013	0.10	
H_2O	2.75	2.12	0.64	0.27	0.94	0.50	0.55
CO_2	0.10	0.06	0.21	0.05	0.05	0.05	
Total	99.613	100.18	99.926	99.48	99.653	99.92	100.11

Description and location of samples:

1. Sample D-257; NW¼SE¼ sec. 33, T. 49 N., R. 15 W., from along D. W. & P. railroad, 2 or 3 ft. below basal contact of Duluth Complex
2. Sample D-259; same location as D-257, 100-200 ft. below basal contact of complex
3. Plagioclase-augite-orthopyroxene hornfels; drill hole 5-3, depth 1,200 ft.; sec. 36, T. 53 N., R. 15 W.; 200 ft. beneath the titaniferous peridotite body
4. Plagioclase-augite-magnetite-orthopyroxene hornfels; drill hole NM4, depth 1,248 ft.; at corner of secs. 3, 4, 9 and 10, T. 61 N., R. 11 W.; 163 ft. below top of 580-foot-thick hornfels inclusion
5. Sample T-111; vesicular, nonporphyritic, magnetite-rich unmetamorphosed basalt; sec. 26, T. 59 N., R. 11 W.
6. Sample M3763; 42% plagioclase (An_{60}), 38% augite, 14% magnetite-ilmenite, and 6% olivine; NW of 57th Ave. W. quarry, Duluth (Taylor, 1964, no. 1, p. 13)
7. Sample from 57th Ave. W. quarry, Duluth (Taylor, 1964, no. 2, p. 13)

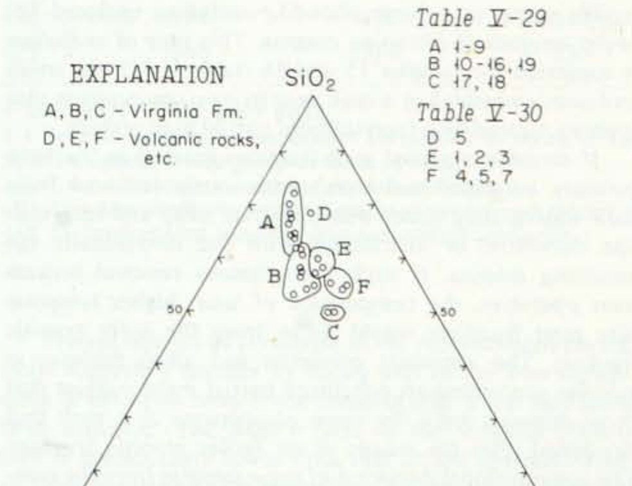

Figure V-54. Plot of the relative abundances of SiO_2, Al_2O_3, and FeO+MgO for various hornfelses.

and its impoverishment in SiO_2, K_2O, and Na_2O suggest that iron, as well as the granitic components, was lost.

Notable in all the analyzed Virginia hornfels samples is the absence of olivine and Ca pyroxene, even though the enclosing troctolites contain abundant quantities of both. The lack of Ca pyroxene suggests that cordierite and Ca pyroxene are antipathetic in this metamorphic environment, an interpretation consistent with ACF compatibility diagrams such as those given by Winkler (1967). This antipathy is further brought out by the hornfels mineral assemblages tabulated by Renner (1969, unpub. M.S. thesis, Univ. Minn.). Renner's data indicate not only that cordierite and Ca pyroxene are mutually exclusive, but also that the plagioclase occurring with cordierite is substantially less calcic (maximum An of 35.1) than the plagioclase occurring with Ca pyroxene (minimum An of 51.0).

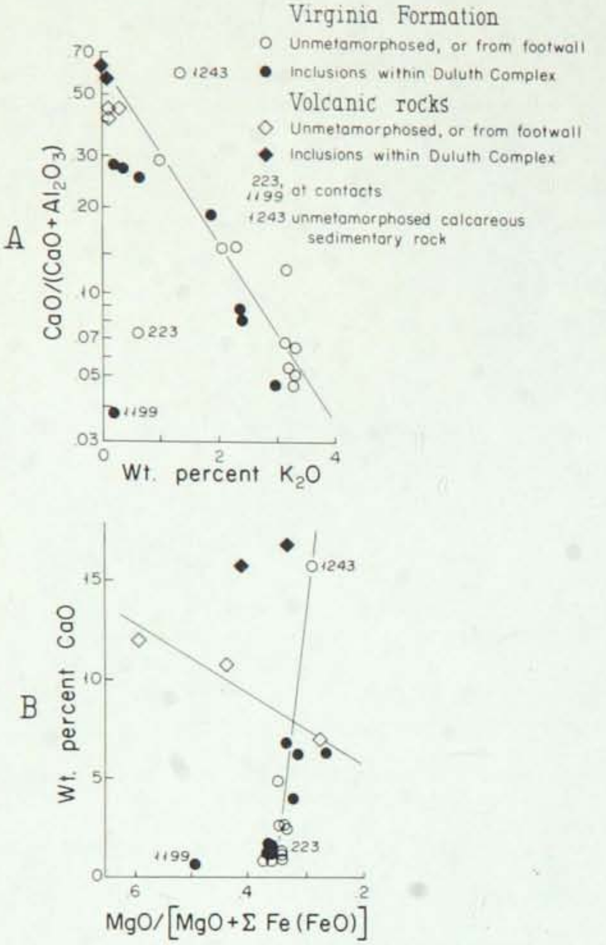

Figure V-55. Plots of K_2O vs. $CaO/(CaO+Al_2O_3)$ and CaO vs. $MgO/[MgO+\Sigma Fe(FeO)]$ for hornfelses.

Five samples of troctolite—three taken adjacent to Virginia hornfels bodies (8, 9 and 10, table V-27) and two taken from probably normal troctolite (1 and 2) without associated Virginia hornfels—were analyzed to determine possible effects of contamination by wall rocks. The most significant compositional deviation shown by the troctolite adjacent to the hornfelses is a low $MgO/[MgO + \Sigma Fe(FeO)]$ ratio, suggesting contamination by material having an even lower ratio, probably the Virginia Formation. Other chemical differences between the troctolite at contacts and that elsewhere are not pronounced, although the contact rocks appear to have been slightly enriched in K_2O, TiO_2, P_2O_5, MnO, CO_2, and Fe_2O_3. Also, it is quite probable that the troctolite in contact zones was enriched in SiO_2. This is indicated petrographically by the partial replacement of olivine by orthopyroxene and by the development of comparatively large quantities of intergranular plagioclase-orthopyroxene symplectite. The intergranular symplectic distribution suggests that SiO_2 migrated along grain boundaries into the intrusive rocks after they had solidified.

Partial Melting

The petrographic features and chemical variations discussed above strongly support the concept that partial melts, generally granitic in composition, formed at high temperatures and escaped from the Virginia inclusions and from the footwall rocks. The hornfelses from which the melt escaped became enriched in the remaining constituents, notably CaO, Al_2O_3, MgO, and FeO, and thus remained as inert inclusions within the igneous rocks. The following general evolutionary scheme regarding partial melting of the Virginia Formation, and concomitant contamination of adjacent igneous rocks, is suggested.

As the argillite was heated, volatiles accompanied by a small fraction of dissolved solids were driven out. This was accompanied by appropriate dehydration and decarbonation mineralogic reactions. As the temperature increased further, the minimum melting temperature in the granite system was reached, so that a small amount of granite melt formed. Depending on the dynamic structural situation and the quantity of melt produced, it either was lost from the inclusion or remained to eventually recrystallize. If the enclosing magma contained a high proportion of melt, the contaminating materials were dispersed and a hybrid igneous rock resulted. Where the enclosing intrusive rock was solid and undergoing fracturing, as it was stressed by cooling or by movement in adjacent areas, much of the granitic liquid became localized in the fractures to form dikes after the enclosing mass had cooled.

If the temperature of an inclusion was raised considerably above the minimum temperature at which a granitic liquid would form, a higher proportion of the inclusion would melt. Depending on the structural situation, two general evolutionary courses that are end members of a continuum of possibilities can be traced.

If the partial melt did not escape from the site in which it was formed, the rocks in that part would attain mobility at some temperature, and eventually could move independently as an intrusive breccia. Such a circumstance could readily occur in a large plate-like inclusion enclosed between tongues of intruding magma. This type of evolution is suggested for samples 15 and 16 (table V-29), the small inclusion embedded in a rock near its own composition that is characterized by a recrystallized partial melt texture.

If successive partial melt fractions formed as the temperature increased and were continuously removed from their source, they either would migrate away and recrystallize elsewhere, or intermingle with and contaminate the enclosing magma. If such a continuous removal process were operative, the composition of later, higher temperature melt fractions would differ from the early granitic fraction. The abundant cordierite and alkali feldspar in samples containing recrystallized partial melts suggest that these minerals were the main components of a melt that developed after the escape of an earlier granitic fraction. The compositional deviation of these samples from the composition of unmetamorphosed Virginia (fig. V-54) suggests that an earlier granitic melt had in fact escaped. The refractory residual type of hornfels, exemplified by samples 17 and 18 (table V-29) taken from contact zones, probably was formed by the complete removal of the interstitial melt.

Table V-31. Oxide ratios for the hornfelses listed in Tables V-29 and V-30.

	$\frac{SiO_2}{SiO_2+Al_2O_3+FeO+MgO}$	$\frac{CaO}{CaO+Al_2O_3}$	$\frac{CaO}{CaO+Na_2O}$	$\frac{K_2O}{K_2O+Na_2O}$	$\frac{MgO}{MgO+\Sigma Fe(FeO)^*}$	
1	.715	.146	.482	.458	.349	A
2	.773	.297	.573	.215	.345	A
3	.752	.620	.862	.339	.290	A
4	.674	.054	.330	.622	.346	A
5	.648	.052	.295	.576	.373	A (b)
6	.719	.143	.403	.351	.346	A (b)
7	.671	.119	.508	.586	.336	A (b)
8	.678	.065	.339	.586	.347	A (b)
9	.705	.066	.313	.552	.361	A (b)
10	.594	.048	.363	.660	.363	B (a)
11	.553	.254	.689	.191	.312	B (c)
12	.565	.280	.681	.075	.268	B (c)
13	.646	.089	.452	.531	.364	B (a)
14	.534	.047	.408	.639	.366	B (a)
15	.635	.188	.493	.315	.321	B (c)
16	.607	.080	.440	.537	.366	B (a)
17	.495	.073	.586	.372	.341	C (b)
18	.494	.038	.714	.353	.496	C (b)
19	.597	.277	.656	.102	.335	B (c)
(1)	.590	.435	.852	.091	.590	E
(2)	.606	.436	.809	.114	.443	E
(3)	.625	.553	.879	.040	.414	E
(4)	.559	.568	.909	.00X	.339	F
(5)	.728	.399	.705	.044	.278	D
(6)	.554	.477	.843	.072	.271	F
(7)	.566	.464	.836	.094	.284	F

* $\Sigma Fe(FeO) = FeO + 0.9\ Fe_2O_3$

( ) Sample numbers in parentheses correspond to rocks in Table V-30; others are rocks in Table V-29

A-F Designates which group of those in Figure V-54

(a) Contains cordierite and has partial melt texture

(b) Contains cordierite but does not have partial melt texture

(c) Metamorphosed Virginia, does not contain cordierite

The abundance of cordierite in the recrystallized partial melts suggests a manner by which part of the iron enrichment in the basal zone of the complex (fig. V-49) may have been achieved. The Mg/Fe ratio of the Virginia Formation is considerably lower than that of the igneous rocks, and the ferromagnesian component of any partial melt developed from it most likely would have had an even lower ratio.

The addition of contaminating material enriched in iron would cause the resulting igneous rocks to have lower Mg/Fe ratios. If we consider the ferromagnesian component of the contaminating material to have the composition of cordierite, then the addition of a melt consisting essentially of cordierite and alkali feldspar to a basaltic melt would produce a norite, or some rock intermediate between norite and gabbro or troctolite. Consider the following general reaction (ignoring the Fe-Mg substitutions for the sake of simplicity): $Mg_2Al_4Si_5O_{18}$ (cordierite from inclusions) + $CaSiO_3$ (available in magma)$\rightarrow$$2MgSiO_3 + 2CaAl_2Si_2O_8$ (combines with albite) + SiO_2 (reacts with olivine). The consumption of $CaSiO_3$ from the magma would increase the quantity of orthopyroxene present in the final rock at the expense of Ca pyroxene, inasmuch as: $CaMgSi_2O_6 = MgSiO_3 + CaSiO_3$. The anorthite produced would combine with any albite in the partial melt or magma to form plagioclase. The composition of plagioclase produced by such contamination would vary, depending on the relative amounts and compositions of the partial melt and magma, but most likely would be more sodic than plagioclase crystallized from uncontaminated magma. In the core studied by Hardyman (1969, *op. cit.*; fig. V-49), for example, the plagioclase is An_{50} in the hypersthene gabbro immediately above the Virginia footwall, in contrast to An_{60-65} throughout most of that core. Any SiO_2 that was released would react with olivine to form additional orthopyroxene: $Mg_2SiO_4 + SiO_2 \rightarrow 2MgSiO_3$. The potassium feldspar component could produce biotite by combining with the ferromagnesian constituents, or it might migrate further away, to constitute part of a granitic body.

The results predicted above are consistent with observed rock types in the basal zone of the complex. Orthopyroxene and biotite are much more common there than elsewhere, and rocks such as hypersthene gabbro, hypersthene troctolite, norite, and even hypersthenite commonly are present in that zone. For example, in the Dunka River area (figs. V-45, V-46, and V-47), rocks having a moderate or high content of orthopyroxene, and moderate amounts of biotite, are as much as several hundred feet thick at the base of the complex.

As noted earlier, the granitic rocks of the Giants Range batholith in the contact zone were partially melted and inasmuch as these rocks form the footwall over a large area, and contain an even greater potential quantity of low temperature-melting material than the Virginia Formation, it is likely that they contributed substantially to the contamination of the mafic rocks. The interaction between the mafic rocks and the underlying granitic rocks probably was the most important source of contaminating materials in the complex. The Biwabik Iron-formation and hornfelsed volcanic rocks are other potential sources of contamination, and may have contributed iron, silica, and other elements to the complex.

The chemical exchanges between inclusions and magma discussed above are directly contradictory to Grout's (1933a) earlier views. He believed that argillaceous materials such as the Virginia were so thoroughly reconstituted, including the addition of a large quantity of CaO, that they became identical in composition to the enclosing intrusive rocks. He suggested that these changes were accomplished by materials migrating into the inclusions; evidently he did not believe that there was much change in the enclosing intrusive rocks.

Volcanic Hornfels

A large proportion of the hornfels bodies in the complex differ in mineralogic, chemical, or structural characteristics from the inclusions of Virginia Formation. Many of the rocks were volcanic in origin and a few of these are discussed below.

The Colvin Creek body (fig. V-44) contains 40-60 percent plagioclase, 30-40 percent pyroxene (mainly augite), 10-20 percent magnetite, and minor biotite and apatite. The texture is granoblastic, and a few recrystallized plagioclase phenocrysts are present. Small, round, elliptical or irregular plagioclase segregations, one-fourth to one inch across, locally accompanied by coarser grained Ca pyroxene or oxides, probably represent metamorphosed amygdules. Thin dikes and irregular veins of medium- or coarse-grained rock containing Ca pyroxene, oxides, and minor apatite cut the hornfels. The configuration and location of these segregations indicate that they probably originated prior to or during metamorphism, rather than having been introduced from outside the hornfels mass.

A hornfels body exposed along the Erie Mining Company railroad in sec. 33, T. 60 N., R. 12 W. (fig. V-44) contains vaguely defined thin lenses and discontinuous layers of small elliptical or irregular Ca pyroxene concentrations. This hornfels contains 50-55 percent plagioclase, about 15 percent augite, 0-5 percent opaque oxides, about 35 percent orthopyroxene that has inverted from pigeonite, and sparse olivine. These rocks have granoblastic textures, although locally the plagioclase occurs as laths, giving the rock an igneous-appearing texture. Thin granite dikes and larger gabbro dikes cut the body, and several vertical peridotite dikes several feet in thickness are present. The peridotite dikes consist mainly of medium-grained titanaugite, which locally is partially altered to biotite and amphiboles, and variable amounts of olivine, ilmenite, sulfides and, locally, plagioclase. These dikes are nearly identical to the titaniferous peridotite that occurs in drill hole 5-3 (figure V-52).

The large rock cut along the Erie Mining Company railroad at the base of the complex in sec. 18, T. 59 N., R. 13 W. (fig. V-44), contains hornfelses of varying texture and mineral proportions. Most samples have about 50 percent plagioclase, abundant olivine, augite, and a small percentage of oxides; biotite is locally abundant and apatite and brown hornblende are minor constituents. Round to elliptical plagioclase-rich "spots," which probably were amygdules, locally are conspicuous. These rocks vary in texture from granoblastic to diabasic, and definitely are volcanic. Local, widely spaced, thin layers containing abundant biotite evidently represent interflow zones.

About 580 feet of fine-grained plagioclase-Ca pyroxene-orthopyroxene-magnetite hornfels, very similar to the Colvin Creek body, was intersected in a drill core at the intersection of secs. 3, 4, 9 and 10, T. 61 N., R. 11 W., in the southwestern part of the Gabbro Lake quadrangle. Thinner intercepts of the same rock occur at similar depths in adjacent holes and crop out in SE¼ sec. 34, T. 62 N., R. 11 W., making it probable that hornfels forms a sheet at least 2 miles long enclosed within the intrusive rocks in the area. The hornfels lacks olivine, the orthopyroxene is inverted pigeonite, and apatite and biotite occur in trace amounts. The core contains small plagioclase segregations, probably metamorphosed amygdules, at certain horizons. An analyzed sample from the middle of the body (no. 4, table V-30) has a very low SiO_2 content (40.20 percent) and a high CaO content (17.09 percent), which are anomalous for a volcanic rock. Cutting this rock are a few thin granite dikes and local veinlets, up to a few inches thick, composed of medium- to coarse-grained Ca pyroxene and oxides that are similar to the irregular pyroxene-oxide veinlets in the Colvin Creek body.

The hornfels beneath the peridotite in drill hole 5-3 (fig. V-52) is mainly plagioclase-Ca pyroxene-olivine hornfels with traces of opaque oxides, biotite, and apatite. Locally the place of olivine is taken by orthopyroxene (inverted pigeonite). This hornfels contains a few small plagioclase segregations which may either be metamorphosed amygdules or recrystallized phenocrysts. Also cutting the hornfels are a few thin, medium-grained augite-olivine-ilmenite-apatite-peridotite dikes very similar to the overlying peridotite. The hornfels sample (no. 3, table V-30) from 200 feet below the peridotite contains inverted pigeonite, rather than olivine. It is similar to the hornfels represented by analysis no. 4 in having high CaO and low K_2O contents, but appears to be more closely related to the two samples (1 and 2) of metamorphosed footwall basalt from the Duluth area.

Analyses of seven rocks, which represent the known compositional range of these hornfelses, are given in Table V-30. They are listed in order of decreasing $MgO/[MgO + \Sigma Fe(FeO)]$ ratios, the values for which are in Table V-31. Samples 1 and 2 are hornfelsed basalts from the footwall of the complex west of Duluth, and samples 3 and 4 are typical materials from the two cores discussed above. Sample 5 is an unmetamorphosed, vesicular, non-porphyritic basaltic rock, having a high proportion of magnetite, from sec. 26, T. 59 N., R. 11 W., between the complex and Lake Superior in the Greenwood Lake region (fig. V-50). This flow, which may have been derived from magma left over after the separation of abundant troctolite and ferrogabbro, is interlayered with felsites. Its iron/magnesium ratio is similar to that of samples 4, 6 and 7, but it is notably richer in silica, and poorer in Fe, Mg and Ca. Samples 6 and 7 are basaltic hornfelses from Duluth, which Taylor (1964) interpreted as representing a broken, recrystallized chill zone of a late-stage apophysis from the layered series. If this interpretation is correct, their composition may be that of a late-stage liquid developed during the evolution of the troctolitic series. The analyzed samples are notable in having a low SiO_2 content and substantial amounts of TiO_2 and FeO, and are of interest because of their general chemical similarity to some of the other hornfelses.

The veins and dikes (consisting mainly of Ca pyroxene and oxide minerals) that traverse several of the hornfels bodies are distinctive, and some at least, particularly those in the Colvin Creek body and the drill hole from the southwestern part of the Gabbro Lake quadrangle (no. 4), probably formed within, and from, the enclosing hornfels. Many of these veins are zoned, and have a peripheral oxide concentration enclosing a core that consists mainly of Ca pyroxene. The proportions of pyroxene to oxides vary from one vein to another, and plagioclase is abundant locally. The veins seem restricted to the hornfelses with high oxide and apatite concentrations. It is not yet clear if these veins formed prior to metamorphism or are the result of some high-temperature process.

SULFIDE MINERALS IN THE DULUTH COMPLEX*

Bill Bonnichsen

Extensive low-grade sulfide deposits, primarily consisting of pyrrhotite, chalcopyrite, cubanite, and pentlandite, occur in the Duluth Complex, and are of interest as potential ores of copper and nickel. The deposits are estimated to contain hundreds of millions or even billions of tons of mineralized rock having a combined nickel and copper content between one-half and one percent. The possibility of commercial deposits has prompted considerable activity by several mining companies, including exploratory drilling, feasibility studies, and examination of mining and metallurgical problems. Although no ores have been mined, it is probable that significant quantities will be mined eventually.

Knowledge of the deposits is meager at the present time and the following is based mainly on the examination of sulfide minerals in drill cores and in polished sections from the Dunka River area and, to a lesser extent, from other parts of the complex. Information available from mining companies or the literature also has been drawn upon.

DISTRIBUTION OF SULFIDES

The region that has been most intensively explored extends for a distance of approximately 40 miles along the northwestern side of the complex from the edge of the Boundary Waters Canoe Area, east of Ely, to a few miles south of Hoyt Lakes (see my accompanying paper and that of Phinney for a discussion of the geology). The majority of the sulfides occur within a few hundred feet of the basal contact of the complex. Only scattered occurrences that are economically interesting have been encountered farther outward in the complex, although traces of sulfides have been found at many localities.

Structural and Lithologic Controls

Sulfide-bearing rock is fairly continuous within the basal zone of the complex between Hoyt Lakes and the Boundary Waters Canoe Area, and certain concentrations are of sufficient size and grade to be of economic interest. Individual deposits within this zone vary in shape, from lenses that are parallel to the base, to lenses or very elongate bodies that are inclined to the base, to irregular lenses. So far as known, the lateral boundaries of the concentrations are gradational, except where they abut against large, nonmineralized inclusions. On the other hand, the upper limits of some are abrupt, and may coincide with the contacts between intrusive units.

Sulfide minerals occur in a wide variety of rock types in the basal zone of the complex as well as in some of the footwall rocks and in inclusions. Table V-32 summarizes the types of sulfides that occur in several lithologic units in and adjacent to the complex.

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The estimated abundance of sulfide minerals in four selected cores from the Dunka River area is shown in Figure V-46. Significant, as well as minor, sulfide occurrences are restricted to the lower few hundred feet of the complex, and their upper limit generally coincides with the contact between the augite-poor troctolite and the overlying augite troctolite. Although the augite troctolite contains scattered traces of sulfides, their abundance is markedly less than in the lower units. This apparent segregation into specific intrusive units suggests that the sulfide minerals are syngenetic.

In Figure V-47, showing copper and nickel assays for several holes drilled 200 feet apart through the basal intrusive unit, it can be seen that copper and nickel sulfides are erratically distributed. Although details are lacking, the abundance of total sulfide minerals has approximately the same distribution in these cores.

In the drill core studied by Hardyman (1969, unpub. M.S. thesis, Univ. Minn.), most of the sulfides are concentrated in the lower few hundred feet (fig. V-49). Inasmuch as the sulfide abundances were determined by modal analyses of individual polished sections, the average quantity for 10-foot core intercepts would be much less than the values shown.

The structural controls for the concentrations of copper and nickel sulfides in the complex are poorly known at present. Basin-like depressions or plunging synformal structures at the base of the complex appear to have localized some concentrations, although there is no evidence that these are the major structural controls. Other types of inflections in the base of the complex, the juxtaposition of sulfides with the various lithologic units that occur in the footwall, and abundant inclusions of Virginia hornfels also have been considered as possible ore controls.

In the Dunka River area and at other localities, economically interesting concentrations of sulfides also occur locally in the older granitic rocks as much as a few hundred feet beneath the complex. Such occurrences clearly indicate that sulfide minerals were introduced into the host rocks of the complex and that, in places at least, they were independently mobile with respect to their mafic source rocks. It is not known presently whether these concentrations resulted from the infiltration of molten sulfides or from hydrothermal or diffusion processes. Similar epigenetic sulfides occur in some inclusions in the complex, as for example, in the quartzite inclusion from the Dunka River area listed in Table V-32.

The Virginia Formation contains substantial quantities of indigenous sulfides (for typical analyses see table V-29). In hornfelsed Virginia Formation, the principal sulfide, pyrrhotite, is distributed along relict sedimentary structures. Prior to metamorphism, the sulfur in the Virginia probably was primarily in the form of FeS_2. Concurrent with the development of pyrrhotite from FeS_2 in the inclusions and footwall rocks, the sulfur liberated by this reaction probably migrated into the intrusive body to form local sulfide concentrations. However, this type of sulfide occurrence is not considered to be as important as those developed by magmatic processes, inasmuch as many of the sulfide occurrences are not associated closely with Virginia Formation hornfelses.

The hornfelsed volcanic rocks that are included in the complex and which are easily mistaken for Virginia Formation hornfelses have extremely low abundances of sulfide minerals.

Table V-32. General characteristics of sulfide concentrations in rock units within and adjacent to the Duluth Complex.

Unit	Principal rock types	Sulfide mineralogy	Sulfide assemblages	Oxides
Basal intrusive unit, Dunka River area*	augite troctolite, norite, gabbro	major: po, cp, pn minor: cb, bn, sph	mainly Mss, local Mss + indep cp, local cb repl po	ilm > mgt
Augite-poor troctolite, Dunka River area*	troctolite, picrite, dunite	major: po, cp, cb, pn minor: sph, bn, "X"	mainly cb repl po and Mss, local Mss + indep cp	ilm > mgt
Basal zone rocks, INCO shaft and drift, Lake Co.	troctolite, augite troctolite, picrite, troctolitic anorthosite	major: cb, cp, po, pn minor: sph, bn	mainly cb repl po, local Mss + indep cp	ilm > mgt
Core studied by Hardyman (1969, unpub. M.S. thesis, Univ. Minn.), Babbitt area*	troctolite, augite troctolite, biotite-bearing troctolite	major: po, cp, cb, pn minor: bn, "X"	Mss, Mss + indep cp, and cb repl po	ilm > mgt
Late-stage ultramafic body, PD 5-3*	olivine gabbro, ilmenite peridotite	major: po, cp, pn minor: cb, sph	Mss and Mss + indep cp	ilm
Virginia Fm. in footwall and inclusions*	plagioclase-hypersthene-biotite-cordierite hornfels	major: po minor: cp, pn, sph	mainly po	ilm
Volcanic hornfels in footwall and inclusions*	plagioclase, augite hornfels with variable magnetite	Sulfides are very sparse minor: cp, bn	cp + bn	mgt > ilm
Giants Range granitic rocks in footwall of complex*	plagioclase-rich hypersthene, biotite-bearing hornfels	major: cp minor: bn, sph, po, pn	indep cp	mgt > ilm
Quartzite inclusion in basal intrusive unit, Dunka River area	quartzite with minor biotite and chlorite	major: po, cp, pn	Mss + indep cp	ilm

* See accompanying paper by Bonnichsen for geologic information on these units

Abbreviations: indep, independent; Mss, relict monosulfide solid solution grains; repl, replacing; sph, sphalerite; "X", talnahkite phase "X", etc.; for other abbreviations see footnotes of Table V-33

Textural and Structural Form

The sulfide minerals occur in a variety of textural and structural forms, ranging from rather uniform or quite irregular disseminations to veins and massive deposits. In the intrusive rocks, the sulfides generally are texturally later than associated silicate minerals, although a small proportion are intergrown with the silicate minerals and small sulfide grains commonly are enclosed within them.

Irregularly distributed sulfides are more abundant than uniformly disseminated sulfides. Among the irregularly distributed types, a distinctive spotted variety in which the sulfides occur as scattered one-eighth to three-fourths-inch composite grains is common. Another common type occurs in rocks characterized by inhomogeneities in grain size and texture. In these rocks, the sulfides tend to be more abundant and to occur in larger aggregates in the coarser parts of the rocks. Such aggregates in the more pegmatitic areas are as much as a few inches across; coarse-grained biotite, clinopyroxene, and oxides commonly are closely associated with this type of sulfide. The various textures are gradational from one place to another within individual sulfide occurrences.

Net textures, in which silicate grains, mainly olivine, are enclosed in a continuous network of sulfide minerals, are comparatively rare. Such occurrences have been noted principally at the margins of some of the late-stage ultramafic bodies, such as at the lower edge of the titaniferous peridotite body cut in drill hole 5-3 (see my accompanying paper).

Massive or nearly massive sulfide deposits are uncommon. Inasmuch as the shapes of many such segregations have to be assumed from drill intersections, many of the smaller, apparently massive deposits may instead be coarse-grained sulfide aggregates of the pegmatitic type, mentioned above. Others may be thin, vertical veins. Some that occur in the basal contact zone have the form of dikes or are highly irregular lenslike bodies, and are as much as a few feet across.

Some massive sulfide deposits, that are thought to be subhorizontal, are considered to represent immiscible sulfide liquid segregations. Such layers as much as a few feet thick have been encountered near the base of the complex; their lateral dimensions are unknown. Two such bodies were intersected in the cores illustrated in Figure V-46. The sulfide zone in NM-5 is several feet thick, and contains rounded fragments of silicate rocks. In both cores, the massive sulfide concentrations occur within the same stratigraphic interval as other mineral segregations. The silicate rock inclusions in the NM-5 sulfide concentration suggest that local mobilization of the liquid sulfide mass occurred after it had segregated.

The sulfides in the granitic footwall rocks and in some inclusions occur in thin veinlets and especially as interstitial grains. Some of the sulfides in the granitic rocks are localized in zones that appear to have been sheared and subsequently recrystallized. Adjacent zones that appear unsheared are nearly barren.

MINERALOGY

Pyrrhotite is the most abundant and chalcopyrite the second most abundant sulfide mineral within and adjacent to the Duluth Complex. Cubanite and pentlandite also are common. The proportions of the major minerals differ considerably from place to place. Other minerals, including bornite, sphalerite, native copper, digenite-chalcocite, pyrite, mackinawite, several secondary nickel sulfides associated with pentlandite, talnahkite, and phase "X" ($Cu_3Fe_4S_6$) have been reported (Anderson, 1956, unpub. M.S. thesis, Univ. Minn.; Hall and Weiblen, 1967; Hardyman, 1969, *op. cit.*) or observed by me. Data on the sulfides and accompanying opaque oxides that occur in several rock units are summarized in Table V-32. Much of Hardyman's modal and analytical sulfide data are summarized in Table V-33.

Pyrrhotite

Although pyrrhotite is the most abundant sulfide, its concentration relative to the other sulfides varies widely, depending on the specific sample. In the intrusive rocks, it occurs mainly as anhedral or irregular grains that are paragenetically later than associated silicates. Locally, it occurs in vermicular intergrowths with hypersthene and ilmenite, in thin veins cutting the associated silicates, or as small inclusions within the silicates. Both hexagonal and monoclinic pyrrhotite are present in many samples, as was reported by Hall and Weiblen (1967) and by Hardyman (1969, *op. cit.*). Typically, pyrrhotite contains blebs and lamellae of chalcopyrite and pentlandite, and locally sphalerite, that have unmixed from the pyrrhotite. These textural relationships indicate the breakdown during cooling of a monosulfide solid solution (Mss) that had formed at high temperatures (Kullerud and others, 1969; Craig and Kullerud, 1969).

Pyrrhotite is by far the most abundant sulfide in hornfelsed Virginia Formation. It occurs as homogeneous grains or as grains that contain minor quantities of chalcopyrite, pentlandite, and sphalerite. It is commonly associated with graphite. In the granitic rocks, pyrrhotite occurs only in the first few feet beneath the basal contact of the complex.

Chalcopyrite

Chalcopyrite is the second most abundant sulfide and is more widespread than pyrrhotite. It occurs in virtually every polished section I examined that contains sulfide minerals. In the intrusive rocks, chalcopyrite occurs principally as irregular blebs, thick lamellae, and edge concentrations in the relict Mss grains, and as independent grains having various textural forms. In the latter, chalcopyrite occurs as veinlets cutting the silicates, as vermicular intergrowths with the silicates, and as small inclusions within the silicate grains. Irregular to very irregular grains, which usually are quite small, are the most common form of independent chalcopyrite.

The quantity of chalcopyrite is small in hornfelsed Virginia inclusions, but where pyrrhotite is present it generally is also. Chalcopyrite is the most abundant sulfide in the granitic rocks beneath the complex and in many of the hornfelsed volcanic inclusions.

Cubanite

Cubanite is relatively abundant in some of the mineralized intrusive rocks, but is uncommon in inclusions and in the footwall rocks. The variation in the abundance of cubanite, relative to the abundances of the other sulfides, is illustrated in Table V-33 for the core examined by Hardyman (1969, *op. cit.*). Cubanite occurs principally as exsolution lamellae intergrown with chalcopyrite, and in replacement intergrowths with pyrrhotite. In the intergrowths with chalcopyrite, the quantity of chalcopyrite generally is greater than that of cubanite. Where associated with pyrrhotite, the proportions of the two minerals vary rather widely from grain to grain. The pyrrhotite in these intergrowths commonly occurs as crystallographically parallel islands in the cubanite. If the amount of cubanite is low, it tends to be confined to the marginal parts of the pyrrhotite masses. Similar relationships between cubanite and pyrrhotite were interpreted by Hardyman (1969, *op. cit.*) as resulting from exsolution; I, however, believe that the cubanite replaces pyrrhotite, for in some rocks having these textures the cubanite occurs between pyrrhotite and chalcopyrite grains.

Table V-33. Modes and compositions of sulfide minerals from the drill core studied by Hardyman (1969, unpub. M.S. thesis, Univ. Minn.).

		Modal proportions									
Depth		oxides			sulfides			Weight percent			
(feet)	Rock type	mgt	ilm	po	cp	cb	pn	Cu	Fe	S	Mineral
249	aug troc	14	86	6	85	18	6				
255	troc	45	55	5	74	16	5	23	42	34	cb
933.7	aug troc	18	82	3	76	19	2	87		19	cc
								37	32	34	cp
								38	31	33	ta?
								30	37	34	X?
949	aug and bio-brng troc	39	61	24	76	Tr	Tr				
1671	trc anth	18	82	55	45	Tr	Tr				
1691	pyx-brng troc	15	85	57	32	11	Tr				
1724	troc	46	54	10	90		Tr	34	32	35	cp
								63	12	28	bn
1729	troc							22	41	35	cb
								32	33	34	cp
1746	bio-brng troc							38	28	32	tg?
								35	35	33	X?
								22	45	37	bn
								36	31	34	cp
1764	aug oxide troc	37	63	5	33	46	15				
1779	bio-brng troc	9	91	60	40	Tr	Tr				
1796	troc	37	63	20	34	31	15				
1831	aug troc	Tr	100	76	24	Tr	Tr				
1843	bio-brng olv gb	4	96	56	8	28	9	19	45	35	cb?
									58	38	po
1854	aug and bio-brng troc	24	76	65	5	15	15	23	43	36	cb
								33	31	35	cp
									64	38	po
1894	aug norite	8	92	54	10	17	19	34	32	36	cp
								23	45	37	cb
									62	39	po
1991	skarn			60	40	Tr		33	33	35	cp?
									60	39	po

Abbreviations: aug, augite; bio, biotite; bn, bornite; brng, bearing; cb, cubanite; cc, chalcocite; cp, chalcopyrite; gb, gabbro; ilm, ilmenite; mgt, magnetite; olv, olivine; pn, pentlandite; po, pyrrhotite; pyx, pyroxene; ta, talnahkite; tg, tarnished grain; Tr, trace; trc anth, troctolitic anorthosite; troc, troctolite; X, phase "X"

Pentlandite

Pentlandite is part of the sulfide mineral assemblage in most of the intrusive rocks, but is uncommon in the hornfels inclusions and in the granitic rocks beneath the complex. It rarely occurs as grains that are isolated from other sulfides, but instead generally is associated with pyrrhotite and chalcopyrite in the relict Mss grains. Pentlandite occurs as distinct, approximately equidimensional grains, as exsolution flames, and as local veins within pyrrhotite. Equidimensional grains are most common.

Several other nickel-bearing minerals, including violarite (Anderson, 1956, *op. cit.*), millerite, and bravoite (tentative identifications) have been observed in association with pentlandite. These occur commonly as thin veinlets or patches within the pentlandite and most likely are the result of late-stage replacement reactions.

Bornite

Bornite is relatively uncommon in the intrusive rocks. It occurs locally as exsolution lamellae within chalcopyrite, and is associated with independent chalcopyrite grains rather than with chalcopyrite enclosed in relict Mss grains. Bornite is more common relative to chalcopyrite in the granitic footwall rocks than in the mafic intrusive rocks. In the granitic rocks its abundance, relative to that of chalcopyrite, increases with increasing distance away from the complex. It has not been observed in Virginia hornfels, but it is the most abundant sulfide in some of the volcanic hornfelses.

Sphalerite

In many of the intrusive rocks, sparse, small sphalerite grains are enclosed in the pyrrhotite of the relict Mss grains. Locally, it occurs as edge concentrations and as exsolution "stars" within chalcopyrite in the massive sulfide occurrences and in the granitic rocks. In turn, the sphalerite may contain minute exsolved particles of chalcopyrite. The mineral is easily overlooked in polished sections because of its low reflectivity.

Other Minerals

Other minerals in the deposits include pyrite, mackinawite, native copper, chalcocite-digenite, talnahkite, phase "X" ($Cu_3Fe_4S_6$), as well as the secondary nickel-bearing sulfides associated with pentlandite. Both pyrite and mackinawite are closely associated with, and probably replace, pyrrhotite. Pyrite is rather uncommon. Anderson (1956, *op. cit.*) found small quantities of it replacing pyrrhotite in several massive sulfide samples, and I have observed it in a few polished sections. Euhedral pyrite crystals occur in some of the cavities in mafic pegmatites and in a few of the miarolitic cavities in the intrusive rocks in the basal zone of the complex. Such crystals are more abundant in cavities in the granitic rocks (granophyres) associated with the eastern margin of the complex.

Judged from available samples, traces of probable mackinawite are widely distributed in the Duluth Complex, although some grains could be millerite or valleriite. It occurs as feathery replacements in and at the edges of pyrrhotite grains. Analysis of several mackinawite grains with the electron microprobe (Hardyman, 1969, *op. cit.*) gave the following average composition:

	Weight Percent
Fe	55.0
S	34.6
Ni	2.8
Cu	3.3
	95.7

According to Hardyman (1969, *op. cit.*), the upper limit for zinc and arsenic in mackinawite is three percent.

In many sections containing sulfides from the intrusive rocks, there are tiny, distinctly orange, readily tarnished grains that are similar in appearance to chalcopyrite. They generally are isolated from the other sulfides, and commonly occur as inclusions within the silicate minerals. Hall and Weiblen (1967) and later Hardyman (1969, *op. cit.*) attempted to identify these minerals by use of the electron microprobe. Hardyman reported the presence of the recently reported mineral talnahkite, $Cu_{18}(Fe,Ni)_{18}S_{32}$, (Cabri, 1967) and the presence of an unnamed similar-appearing phase having the approximate composition of $Cu_3Fe_4S_6$, which has been referred to as phase "X." Hardyman was unable to find optical criteria by which to characterize phase "X" or talnahkite. Analysis of several grains of talnahkite (Hardyman, 1969, *op. cit.*) gave the following average composition:

	Weight Percent
Cu	37.56
Fe	31.11
S	30.65
	99.32

Traces of material similar to talnahkite and phase "X" appear to be relatively widespread in samples from within the complex but are rare in the footwall rocks and in inclusions within the complex.

Traces of native copper in tiny veinlets cutting the sulfide minerals and of chalcocite-digenite replacing chalcopyrite and bornite have been noted by Hardyman (1969, *op. cit.*), Hall and Weiblen (1967), and me.

CLASSIFICATION OF SULFIDES

I have used the following classification to characterize and compare the sulfide occurrences in individual samples and in the different rock units, and the same classification is used in Table V-32:

(1) relict Mss grains

(2) relict Mss grains and independent chalcopyrite

(3) independent chalcopyrite (±bornite)

(4) extensive cubanite replacement of pyrrhotite

Category (1), relict Mss grains, is considered to represent conditions in which a single sulfide phase crystallized at high temperatures. Category (3) is for rocks that contain little or no pyrrhotite and in which independent chalcopyrite is the principal sulfide. Bornite or cubanite lamellae occur locally in the chalcopyrite. Category (2) refers to rocks that have a mixture of relict Mss grains and independent chalcopyrite. Category (4), although not mutually exclusive with categories (1-3), is a distinct and important group in the Duluth Complex sulfides. The occurrence of replacement cubanite is thought to represent cooling events that were not strictly isochemical, but may have involved sulfur loss, superimposed on the pre-existing patterns of relict Mss grains and independent chalcopyrite. This classification is based largely on textural interpretations; the greatest source of uncertainty is the assignment of specific polymineralic masses containing both chalcopyrite and pyrrhotite to either category (1) or (2).

The sulfide assemblages in the various rock units listed in Table V-32 are classified by the above scheme. For each of the intrusive units, more than one assemblage commonly is present; also, the abundance of the various assemblages differs from one unit to another.

CONCLUSIONS

(1) A fairly continuous zone of sulfide-enriched rock occurs adjacent to the footwall of the complex throughout the Hoyt Lakes-Ely region, along the northwestern margin of the complex. Relatively enriched concentrations within this zone probably will be mined in the near future.

(2) The sulfide concentrations appear to have certain structural and lithologic controls and, in part, the occurrences are related to specific intrusions.

(3) The principal sulfides in the intrusive rocks are pyrrhotite, chalcopyrite, cubanite, and pentlandite; they vary considerably in abundance and in textural relationships.

(4) The majority of the sulfides in the Duluth Complex appear to be syngenetic magmatic concentrations. However, evidence for some mobility of the sulfides is indicated by the occurrence of local disseminated sulfides in the granitic wall rocks as much as a few hundred feet beneath the complex. These deposits are relatively enriched in copper with respect to those in the complex itself.

(5) Hornfelsed Virginia Formation, both in the footwall and in inclusions, contains abundant pyrrhotite and minor chalcopyrite but only traces of other sulfides. Volcanic hornfels, on the other hand, contains only sparse sulfides, mainly chalcopyrite and bornite. The sulfides are mainly indigenous in both types of hornfels.

ACKNOWLEDGMENTS

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LOGAN INTRUSIONS

P. W. Weiblen, E. A. Mathez, and G. B. Morey

The geologic terrane of northern Cook County is characterized by southward-dipping sill-like bodies and crosscutting dikes of diabase and basalt which intrude the Middle Precambrian Gunflint and Rove formations. This terrane extends more or less continuously from west of Gunflint Lake to the northwest shore of Lake Superior in the Thunder Bay district of Ontario (fig. V-56). Within the northern and central parts of the area, differential erosion has resulted in a series of easterly-trending ridges which, in cross-section, are characterized by steep, north-facing cliffs and gently-dipping, south-facing dip slopes. Intervening low areas, characterized by numerous lakes and swamps, are underlain by strata of the Rove Formation. The sills become thinner and less abundant toward the east and northeast, and in the Thunder Bay district only a few more or less flat-lying sills have been recognized (see Geul, 1970, p. 9). There, the resulting topography is more irregularly dissected and characterized by large mesas of Rove Formation capped by diabasic gabbro. In the southern part of the Thunder Bay district along the northwest shore of Lake Superior and in the Grand Portage-Pigeon River area of Cook County (fig. V-56), the sill-like intrusions are fairly thin and more irregular in shape. In addition, vertical dikes of variable thickness are prominent. The thicker dikes form

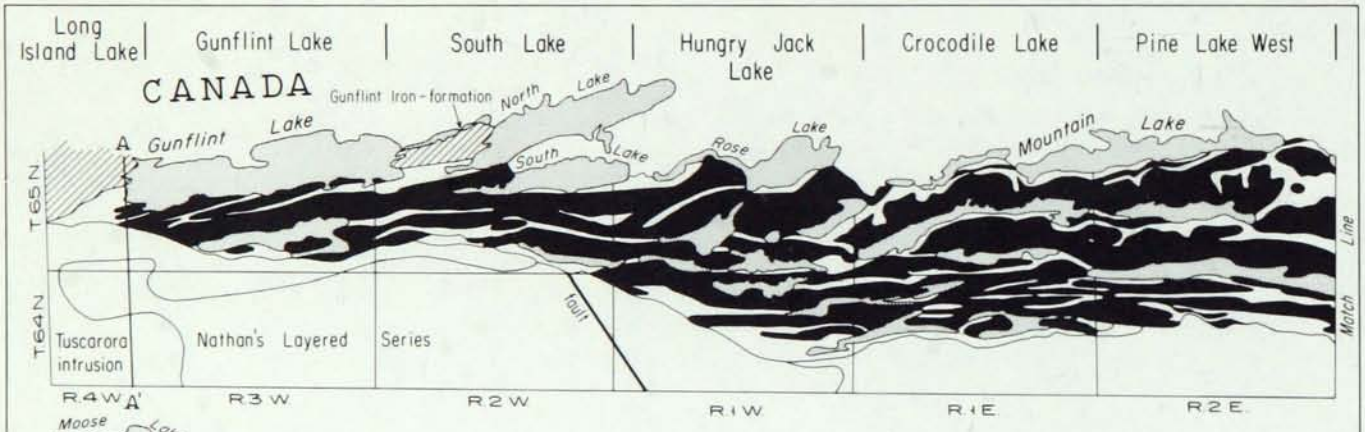

Figure V-56. Generalized geologic map showing the distribution of diabasic intrusions (black) in northern Cook County (modified from Grout and Schwartz, 1933).

steep-sided ridges that trend in several directions but predominantly at nearly right angles to the sills and to the regional strike (fig. V-56).

The hypabyssal rocks of this area have been collectively referred to as the "Logan intrusions," a widely used term that never has had formal status. Tanton (1931) recognized crosscutting relationships in the Thunder Bay district, and Grout and Schwartz (1933, p. 59) suggested that the sills and dikes in Minnesota possibly had been intruded over a period prior to and throughout Keweenawan time. In the Thunder Bay district, Geul (1970) distinguished two major petrogenetic types: (1) Early Mafic intrusions, consisting of diabase and porphyritic diabase of tholeiitic composition; and (2) Pigeon River intrusions, consisting of equigranular olivine diabase (table V-35). This paper primarily concerns the diabase intrusions in northwestern Cook County, which appear to be equivalent to the Early Mafic intrusions of Geul (1970) in the adjacent Thunder Bay district. They are characterized by having a predominantly sill-like form (see fig. V-56), a unique bulk composition ($TiO_2>3$ percent, $K_2O\geqslant1$ percent, $P_2O_5>0.3$ percent, and $MgO\leqslant7$ percent; see table V-35), and a distinctive internal stratigraphy consisting of chilled margins, plagioclase-pyroxene-oxide diabase having minor granophyric segregations, and diabase porphyry zones. Because other fine-grained mafic rocks occur in the area, we propose that the informal term "Logan intrusions" be restricted to those tabular bodies in Cook County having the characteristics enumerated above.

In Minnesota, the Logan intrusions, as defined here, have been studied recently in the Long Island Lake (Morey and others, 1969, Minn. Geol. Survey open-file map), Gunflint Lake (Grant, 1970), South Lake (Morey, 1965, unpub. Ph.D. thesis, Univ. Minn.), and Hungry Jack Lake (Mathez, 1971, unpub. M.S. thesis, Univ. Arizona) 7.5-minute quadrangles (table V-34). To the east, the thinning of the sills and the occurrence of large dikes create a distinctly different outcrop pattern. Both the previous work (Bayley, 1893; Daly, 1917; and Grout and Schwartz, 1933) and the current work (M. G. Mudrey, Jr., oral comm.) indicate that the Pigeon Point sill and the large dike-sill complexes in the Pigeon Point-Grand Portage area may be equivalent to the Pigeon River intrusions of Geul (1970). They both contain olivine and have similar bulk compositions (table V-35). Thus, the Logan intrusions as defined here have their principal occurrence in the northwestern part of Cook County.

STRATIGRAPHIC SETTING

The Logan intrusions in northwestern Cook County were thought by Grout and Schwartz (1933, p. 8) to be older than the Duluth Complex because they appear to be truncated by the basal contact of the complex. Recent mapping has shown that neither the stratigraphic nor structural setting is this simple, and the Logan intrusions of this report are undoubtedly both older than and younger than rock units presently assigned to the Duluth Complex.

In the Long Island Lake quadrangle (fig. V-56), the nearly flat-lying basal contact of the Tuscarora intrusion (Davidson, this chapter) of the Duluth Complex crosscuts the steeply-dipping Animikie strata and the Logan intrusions, and accordingly the Logan intrusions in this area are clearly older than the basal unit of the Duluth Complex, as was suggested earlier by Grout and Schwartz. However, at the western edge of the Gunflint Lake quadrangle, and from there eastward, a gabbroic layered series assigned to the Duluth Complex in the Gunflint prong (Phinney, this chapter; Nathan, 1969, unpub. Ph.D. thesis, Univ. Minn.) is structurally concordant with the Rove Formation and the Logan intrusions (fig. V-56). This concordant sequence dips steeply to the south; however, about 3,000 feet north of the present basal contact of the layered series the sequence dips only 5° to 10° S. This flexure appears to define the hinge line of a northeastward-trending monocline (fig. V-57). In

Table V-34. Location, thickness, and characteristics of Logan intrusions referred to in this report.

Sill	Location	Thickness	Reference	Remarks
Rose Lake				
A	E. ½ sec. 21 and W. ½	72	Mathez, 1971, unpub.	
B	sec. 22, T. 64 N., R. 1 W.	80	M.S. thesis, Univ. Arizona	
South Lake				
C	SE¼ sec. 27, T. 65 N., R. 2 W.	185	Jones, file report	Sills B and C are
D	E½ sec. 26, T. 65 N., R. 2 W.	680	Minn. Geol. Survey	separated by less than
E		360		20 feet of Rove Fm.
F	SE¼ sec. 26, NE¼ sec. 35, T. 65 N., R. 2 W.	920		
Long Island Lake				
G		125	Morey and others,	Sills separated by 75
H	secs. 24 & 26, T. 65 N., R. 4 W.	75	1970	feet of Gunflint Iron-formation

Table V-35. Chemical composition, in weight percent, of Logan intrusions, other rocks of similar composition, and related rocks of different composition.

	1	2	3	4	5	6	7	8
SiO_2	46.60	47.20	50.04	47.50	51.17	49.00	46.60	49.88
TiO_2	4.90	2.79	3.76	3.74	3.63	3.40	1.13	1.19
Al_2O_3	14.10	14.23	11.70	12.94	13.78	13.10	16.80	18.55
Fe_2O_3	2.22	5.61	2.28	3.94	1.70	2.87	2.43	2.06
FeO	11.30	9.86	13.51	11.52	10.97	13.00	10.60	8.37
MnO	0.21		0.15	0.22	0.18	0.22	0.20	0.09
MgO	5.80	7.21	4.20	5.62	5.37	5.60	9.60	5.77
CaO	12.10	8.32	7.16	8.38	9.32	7.45	10.30	9.70
Na_2O	2.27	2.57	3.47	2.39	2.79	2.52	1.82	2.59
K_2O	0.32	0.75	1.03	1.07	0.79	1.16	0.29	0.68
H_2O+		0.94	1.28	1.31		0.49	0.57	1.04
H_2O-			0.07	0.68		0.29		
P_2O_5			0.47	0.69	0.41	0.38	0.13	0.16
CO_2			0.25			0.14	0.10	
S			0.11			0.14	0.02	
	99.82	99.48	99.48	100.00	100.11	99.80	100.59	100.18

1. Calculated composition for sill A, Rose Lake. Calculation based on a mode of 49% plagioclase, 38% augite, 6% ilmenite, 3% titanomagnetite, and 4% granophyre. Mineral compositions and densities (from Deer and others, 1963): plagioclase, v. 4, p. 116, table 16, no. 6; augite, v. 2, p. 117, table 17, no. 18; ilmenite, v. 5, p. 29, table 5, no. 4; titanomagnetite, v. 5, p. 73, table 12, no. 7; and granophyre (from Taylor, 1964), p. 42, table 13, no. 16. Higher TiO_2 values than in analyses 3-6 can be attributed to the fact that incorrect proportions of ilmenite and titanomagnetite were assumed (no data are available on actual relative proportions). Higher CaO values than expected for Logan magma compositions may result from the fact that the mode of plagioclase does not reflect the actual amount of sodic plagioclase (Mathez, 1971, p. 54).
2. Calculated composition of sill A, Rose Lake (Mathez, 1971, p. 74). Calculation is based on all modes and provides an estimate of H_2O content and oxidized condition of the Rose Lake samples
3. Chilled diabase near top of sill D, South Lake, Table V-34
4. Diabase, Northland sill at Duluth (Schwartz and Sandberg, 1940, p. 1145, table 1, no. 8)
5. Average composition of differentiated lava (1955E) erupted on the east rift zone of Kilauea, Hawaii in 1955 (Wright and Fiske, 1971, p. 26, table 6)
6. Chilled diabase, Devon Township, Ontario, Early mafic intrusions of Geul (1970, p. 14, table 3, no. 2)
7. Olivine diabase dike, Devon Township, Ontario, Pigeon River intrusions of Geul (1970, p. 14, table 3, no. 5)
8. Composite sample of five olivine diabases from the Pigeon Point sill, northeastern Cook County, Minnesota (Grout and Schwartz, 1933, p. 41, table 8, no. 2)

addition, the Animikie strata, the Logan intrusions, and the layered series of Nathan (1969, *op. cit.*) are cut by several northwestward-trending faults having small, apparently vertical displacements which parallel well developed fracture systems in rocks of Early Precambrian age. Faults of this fracture system do not appear to transect the younger troctolitic rocks of the Tuscarora intrusion in the Long Island Lake quadrangle to the west. This implies that the Logan intrusions and the layered series of Nathan were emplaced, tilted, and faulted prior to intrusion of the Tuscarora intrusion. It is not clear if the sills predate the layered series. Nathan (1969, *op. cit.*) found no evidence that the sills had been metamorphosed by the rocks of the layered series. He also noted that the mineralogy of the Logan sills was complex, and that it resembled the mineralogy of a unit which, on petrographic evidence, is considered to be a late phase of the layered series. Present evidence thus indicates that the Logan intrusions of this report may be contemporaneous with or younger than the layered series of Nathan, but that they are clearly older than the Tuscarora intrusion (a local example of layered troctolitic rocks), one of the major intrusive units of the Duluth Complex (Davidson and Phinney, this chapter). It is also evident that Keweenawan igneous activity and deformation are temporally related.

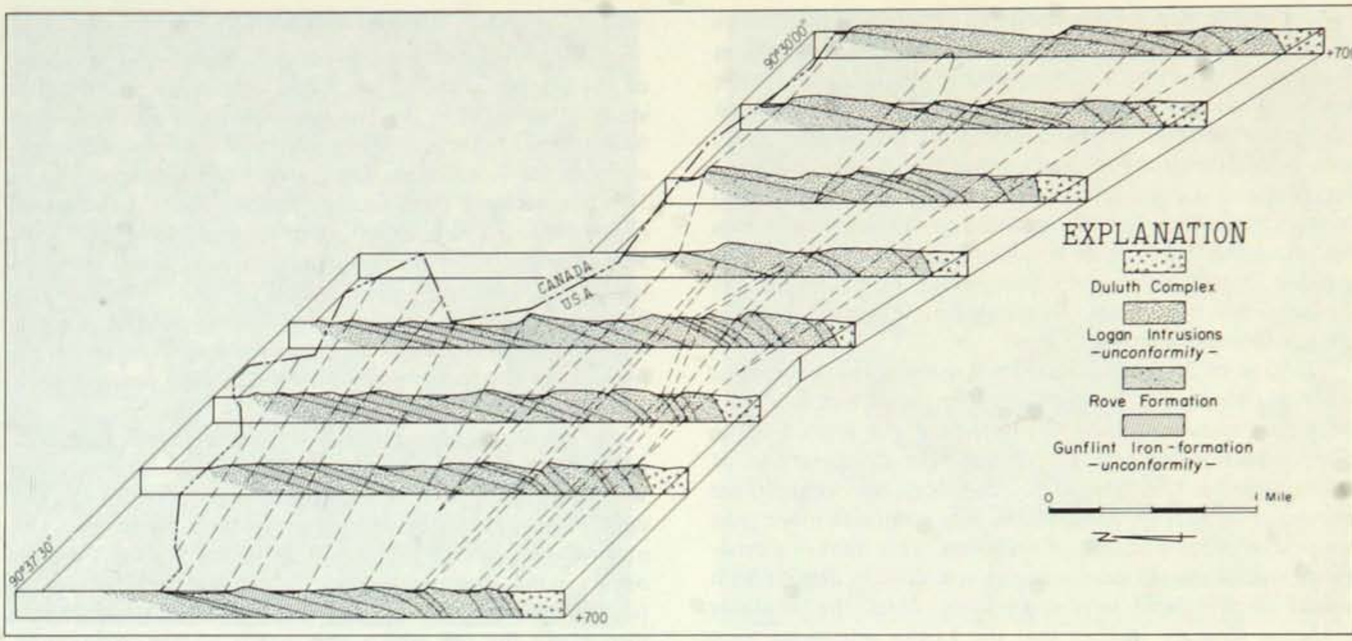

Figure V-57. Structural block diagram of part of Ts. 64 and 65 N., R. 2 W. illustrating large-scale branching and merging of individual sill units.

STRUCTURAL SETTING

The Logan intrusions are sill-like throughout most of Cook County, their emplacement having been controlled dominantly by bedding and to a lesser extent by joints and possibly fault planes in the Animikie strata. Branching and merging of individual sills along strike is common. Such branching results in isolated elongate islands of country rock between igneous masses (fig. V-57). Some of these islands are relatively thin septa of limited lateral extent, whereas others have a considerable thickness and can be traced along strike for several miles. Individual sills also thicken and thin up- and downdip and cause a marked deviation from a strictly planar form. Obviously these features complicate the outcrop pattern and make any subsurface interpretation difficult. This has been particularly well documented in the Hungry Jack Lake quadrangle where diabase is interlayered with Rove strata which pinch out along strike as well as downdip (fig. V-58).

Joints and possibly fault planes also have controlled the sill geometry. For example, a thin sill in the Gunflint Lake quadrangle abruptly terminates against a vertical fracture along which there has been no apparent movement inasmuch as the bedding in the underlying strata is not offset. Similarly, several sills in the South Lake quadrangle have basal contacts which are characterized by step-like offsets. Generally, individual offsets are no more than several feet but together they may aggregate as much as 20 feet over a strike distance of several hundred feet. The offsets are not fault-controlled, but are vertical fracture planes in the Rove Formation. Similar non-faulted, but step-like, contacts have been discussed and illustrated by Grout and Schwartz (1933, p. 16, fig. 6d). Thus, to judge from the available data, the sills form in three dimensions a relatively simple box-work pattern characterized by a lack of any appreciable cross-cutting relations.

Despite evidence of extensive intrusive activity (for example, igneous rocks comprise 80 percent of a 2,700-foot-thick section in the South Lake quadrangle), there is only minimal evidence of disruption of the host Animikie strata. For example, xenoliths of Rove Formation are restricted to contact zones in the sills and are relatively small (less than foot-size; Mathez, 1971, *op. cit.*).

Subsequent deformation involved tilting, development of the regional flexure described above, and faulting on an east-west axis. The faulting (normal, north side down) largely is confined to two zones, both of which parallel the present north shore of Lake Superior. The northern or inland belt is developed in Canada and intersects the International boundary in the vicinity of Gunflint Lake. The second or island belt of faulting coincides with the present shore of Lake Superior and is exposed in Minnesota in the Grand Portage-Pigeon Point area. Here the faults and associated fracture system are occupied by olivine diabase (Pigeon River intrusions of Geul, 1970; table V-35). From field descriptions (Grout and Schwartz, 1933), Logan intrusions as defined in this report occur in the Grand Portage-Pigeon Point area, but are apparently of minor extent, and the unique outcrop pattern of northeastern Cook County reflects the tectonic setting of the olivine diabase dikes and sills. High-angle faulting is present in the northwestern part of Cook County, but repetition of the sills and lithologic units of the Rove Formation makes interpretation of the extent and sense of movement equivocal. Extensive deformation of the Rove Formation is restricted to thin zones above sills and to thin septa between sills, particularly in the zone of deformed rocks related to the regional flexure described above. Therefore, the latter deformation must have occurred after emplacement of the Logan sills in this area. A detailed study of the Gunflint Iron-formation in the Long Island Lake quadrangle (Morey and Papike, oral comm.) has shown that the presence or absence of sills does not affect the thickness or continuity of recognizable stratigraphic units. This suggests that passive emplacement was accomplished by inflation of the Animikie section, with no appreciable assimilation.

In view of the preponderance of igneous rocks over host Animikie strata (fig. V-56), the minor extent and moderate degree of metamorphism is striking. In the Rove Formation, metamorphic mineral assemblages characteristic of the hornblende hornfels facies (see Morey, this volume) are restricted to narrow zones that never comprise more than about one percent of the sill thickness. This implies a process of successive sill emplacement at a shallow depth which would permit rapid heat dissipation. Thus, the available evidence strongly implies that the Logan intrusions were passively emplaced along several types of pre-existing, nearly horizontal structures in the Animikie rocks, and were fed through an evolving fracture system in the underlying Lower Precambrian rocks.

PETROGRAPHY

The following sections on the petrography and petrology of the Logan intrusions are based primarily on a detailed study of two sills in the Hungry Jack Lake quadrangle by Mathez (1971, *op. cit.*), but the interpretations are consistent with the mineralogic and textural relations observed in 115 thin sections from four sills in the South Lake quadrangle (Jones, 1964, unpub. open-file report, Minn. Geol. Survey) and 10 thin sections from two sills in the Long Island Lake quadrangle (Morey and others, 1969, *op. cit.*). Exceptional mineralogic or textural features are discussed with respect to specific sills. The contact metamorphism effects of the Logan intrusions on the Animikie strata are discussed by Morey, this volume.

In the course of mapping, several stratigraphic units that are definable at a scale of 1:24,000 have been recognized within the Logan sills. These consist of (fig. V-59A-D): (A) upper and lower chilled margins which grade into (B) plagioclase-pyroxene diabase; (C) diabase porphyry which occurs in the upper parts of the thicker sills; and (D) quartz-feldspar granophyre, a ubiquitous interstitial constituent throughout the sills which also occurs as late crosscutting veins, stringers, and irregularly shaped masses.

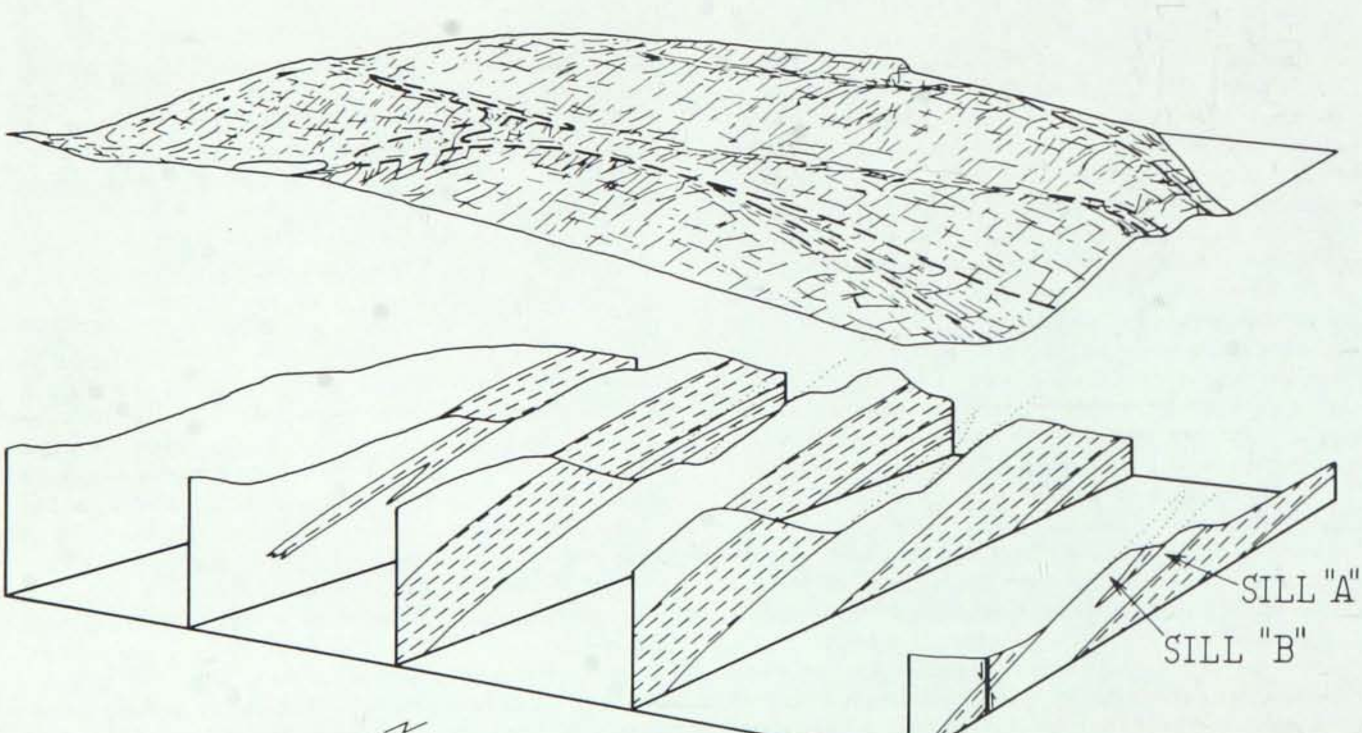

Figure V-58. Perspective diagram and sketch at the present land surface at Rose Lake. The block is 4,000 feet on a side. The surface distribution of the Rove is shown by the heavy dashed lines. In the cross-section the dotted lines show the hypothetical extension of diabase above the present land surface and the dashed pattern is the Rove Formation.

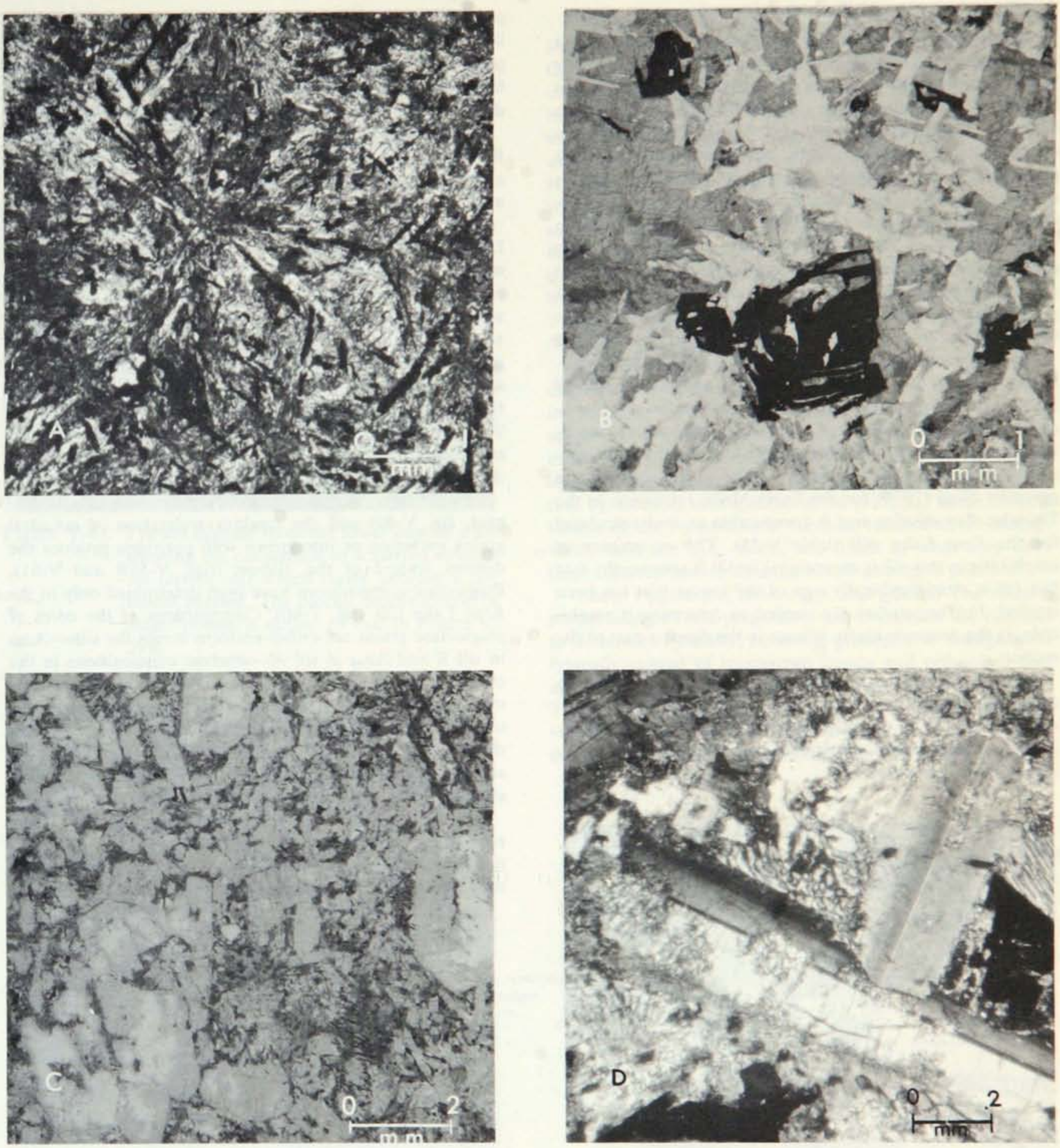

Figure V-59. Characteristic textures of recognizable units in the Logan intrusions. A, chilled diabase; the matted texture results from patchy and radial concentrations of feathery plagioclase (pl), augite (cpx), and ilmenite; B, pyroxene-plagioclase diabase; note the preponderance of pyroxene (cpx) in upper left, the large area of skeletal ilmenite (ilm) in center, and abundant granophyre (grn) interstitial to plagioclase (pl), lower right; C, diabase porphyry; note the bimodal grain-size distribution of the plagioclase (pl) and the similarity of the diabase groundmass to that in Figure V-59B; and D, except for a higher percentage of granophyre and plagioclase alteration, late-stage interstitial granophyre; note that the rims of plagioclase commonly are corroded and that sericitic alteration commonly fans out of the apices of interstitial granophyre.

Chilled Margins

Phenocrysts of calcium-rich andesine, as much as one cm in length, are characteristic of the chilled margins. The modal volume of phenocrysts varies widely; field estimates range from less than one to 10 percent. In chilled margins having a high phenocryst content the phenocrysts are oriented parallel to sill contacts, but more commonly the phenocrysts are randomly oriented. Fine-grained needles of opaque oxides, plagioclase, and pyroxene give the groundmass in the chilled diabase a felty appearance. Microlites in radial clusters are common. In the Rose Lake sills, quartz or potassium feldspar and shreds of biotite occur as isolated clots in the felty matrix. Apparently, all the primary phases present in the diabase nucleated in the chilled margins during quenching or devitrification (fig. V-59A).

In a 75-foot-thick sill in the Long Island Lake quadrangle (table V-34, sill H), the chilled margin consists of fine-grained plagioclase, skeletal opaque oxides, and feathery to bladed actinolite. A chemical composition calculated from a mode indicates a melt with 0.60 weight percent water. This value is in the range of analyzed values given by Geul (1970) for the Early Mafic intrusions in the Thunder Bay district and is comparable to that calculated for the Rose Lake sills (table V-35). The occurrence of amphibole in this sill is exceptional and it is noteworthy that this sill is stratigraphically one of the lowest that has been studied. Further studies are needed to determine if amphibole is the common mafic silicate in the deepest part of the section.

No systematic studies of the grain-size distribution across the chilled margins have been made, but field observations indicate that even within sills only 5 to 10 feet thick a recognizable diabasic texture is developed within a few inches of the contacts.

Diabase

Mineralogy

The primary mineralogy of the diabase is relatively simple: 40-50 percent plagioclase, 30-40 percent augite-pigeonite, 5-10 percent magnetite-ilmenite, and minor amounts of potassium feldspar, quartz, apatite, early olivine, and late biotite. Alteration of this primary mineral assemblage is discussed below.

Compositional and modal variations across sill A at Rose Lake are shown in Figure V-60. The limited compositional data on minerals obtained by electron microprobe analysis (Mathez, 1971, *op. cit.*) are summarized below.

Olivine. The minor olivine that has been found in the Logan intrusions is restricted to scattered subhedral grains in diabase near the bases of sills (fig. V-61). Minor olivine in sill D at South Lake (table V-34) was found in one thin section 80 feet above the base of the sill; it has a composition of Fo_{60}. It has been commonly assumed that olivine may be partly obliterated by alteration, and thus its modal abundance would be underestimated; however, igneous textures are preserved in altered diabase, and hydrous minerals that form pseudomorphs after olivine can be seen rarely, and accordingly it is clear that olivine is a minor early constituent in the diabase.

Plagioclase. The high plagioclase content (45-50 percent, fig. V-60) and the random orientation of euhedral grains enclosing or intergrown with pyroxene produce the distinct texture of the diabase (figs. V-59B and V-61). Plagioclase compositions have been determined only in the Rose Lake sills (fig. V-60). Compositions of the cores of plagioclase grains are rather uniform across the sills—An_{52} in sill B and An_{48} in sill A—whereas compositions in the chilled margins consistently are about five mole percent more sodic (fig. V-60). Normal zoning is in the range of 4 to 8 mole percent anorthite, but becomes more extensive in the upper parts of the sills (fig. V-60). Oscillatory zoning also is found. The borders of plagioclase grains are clouded and corroded next to interstitial granophyre (fig. V-59D).

Euhedral plagioclase phenocrysts in fine-grained diabase near chilled margins contain rectangular cores (fig. V-62) consisting of a typical diabase mineral assemblage. This is discussed in the section on textures.

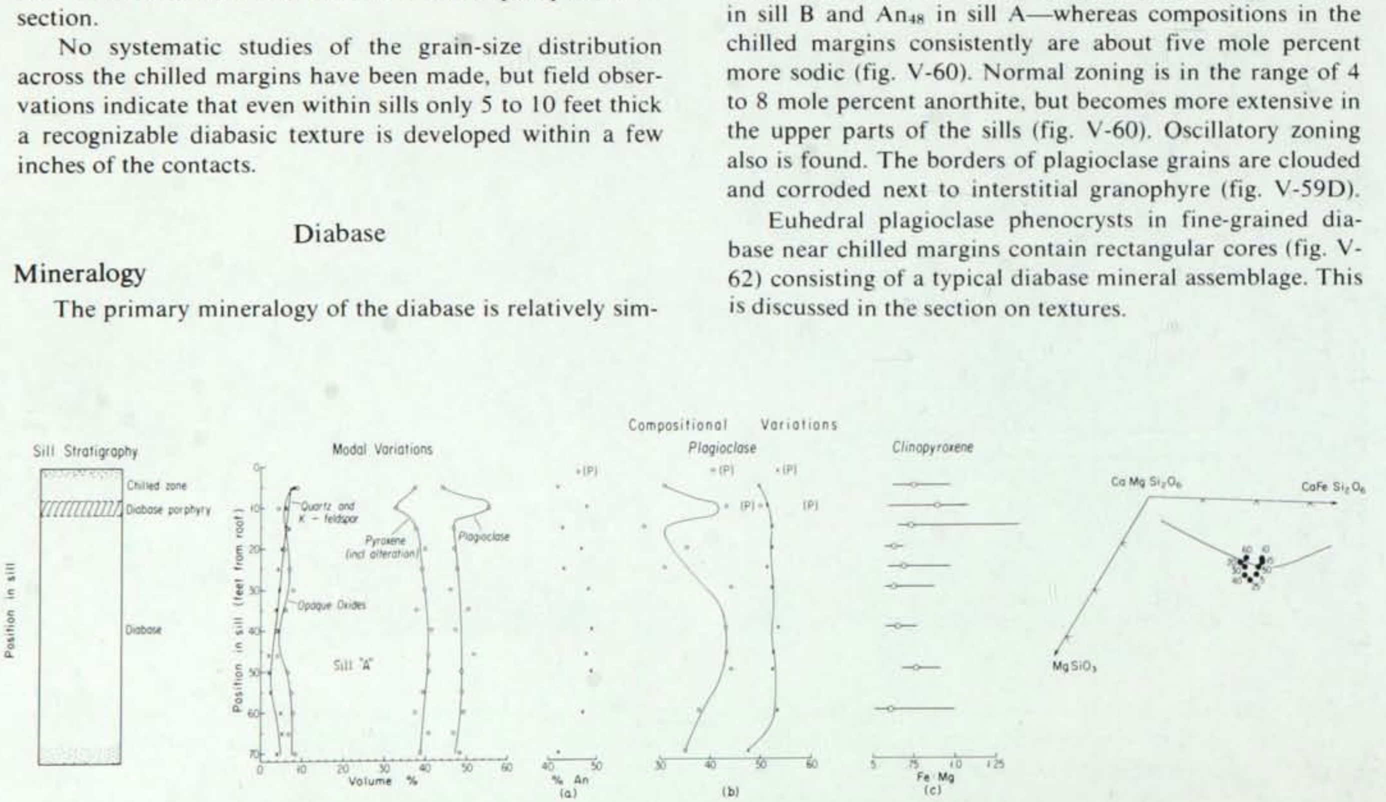

Figure V-60. Stratigraphic, modal, and compositional variations across sill A at Rose Lake. a, averages of all plagioclase analyses; (p) denotes composition of phenocrysts; (b) maximum (x) and minimum (o) plagioclase determinations; (c) total Fe:Mg for pyroxenes. The lines express the range from next-to-highest to next-to-lowest determinations. Points are median values.

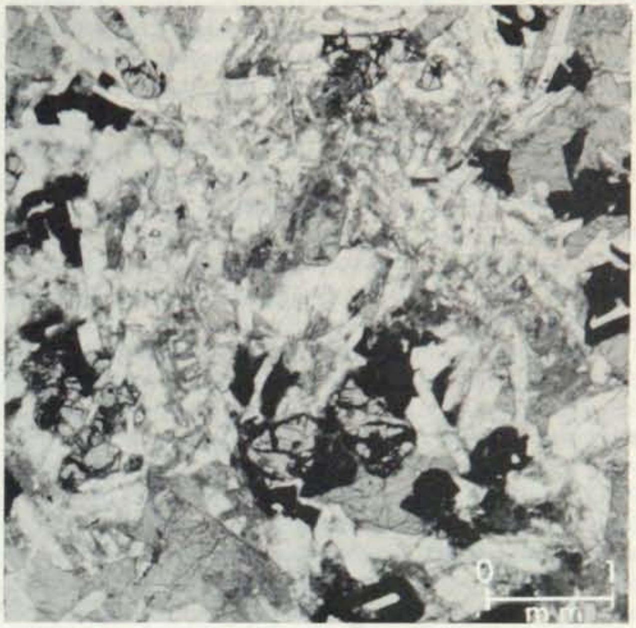

Figure V-61. Typical diabase from the lower part of a sill showing the occurrence of scattered subhedral olivine (ol) which rarely exceeds a few modal percent. The patchy concentration of major phases is similar to that in Figure V-59B.

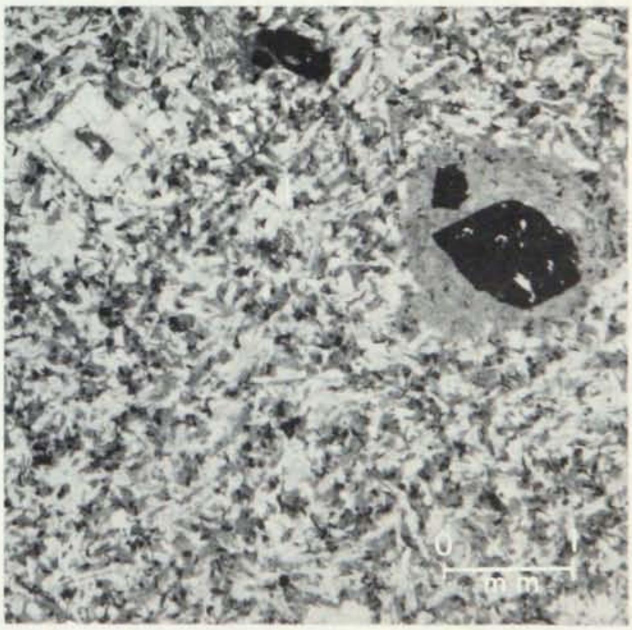

Figure V-62. Fine-grained diabase with cored plagioclase phenocrysts. Diabase of this grain size (*cf.* figs. V-59A, B) is developed within a few feet of sill contacts. Note the euhedral nature of the cored plagioclase phenocrysts. The opaque inclusion rimmed by amphibole (amp) is anomalous and probably noncognate.

Pyroxene. Pyroxene in the Rose Lake sills is characterized by a variable birefringence within individual grains. Other optical features, such as hourglass extinction, indicate strong compositional zoning (figs. V-63 and 64). This has been confirmed by electron microprobe analyses which show compositional variations in single grains of as much as 10 mole percent end-member compositions (fig. V-60). However, all the pyroxenes in the Rose Lake sills occur within the compositional range of augite. Optically recognizable exsolution lamellae are found in the coarser parts of these thin sills, and are parallel to (100), suggesting exsolution of low-calcium pyroxene. Pigeonite, which was thought to be the common pyroxene in the Logan intrusions (Grout, 1928), has not been recognized in the Rose Lake sills (Mathez, 1971, *op. cit.*, p. 57).

Magnetite-ilmenite. A skeletal habit of the opaque oxides (fig. V-59B) is the striking feature of the sills listed in Table V-34. The cubic form of many of the skeletons suggests that ulvospinel may have been the initial phase to crystallize; if so, subsolidus exsolution of ilmenite-titanomagnetite has occurred and ilmenite (qualitative determination by electron microprobe) is the principal oxide. Modal studies of the proportion of ilmenite relative to titanomagnetite have not been made, and much of what may have been exsolved titanomagnetite is now altered to goethite.

Sodic Plagioclase, Orthoclase, and Quartz. The relative proportions of sodic plagioclase and orthoclase are difficult to determine optically; however, granophyric intergrowths typically consist of intergrown quartz and orthoclase that are interstitial to larger grains of plagioclase (fig. V-59D). The granophyre assemblage increases from about five to 10 volume percent from the bases to the tops of sills (fig. V-60). No detailed data are available on the complex plagio-

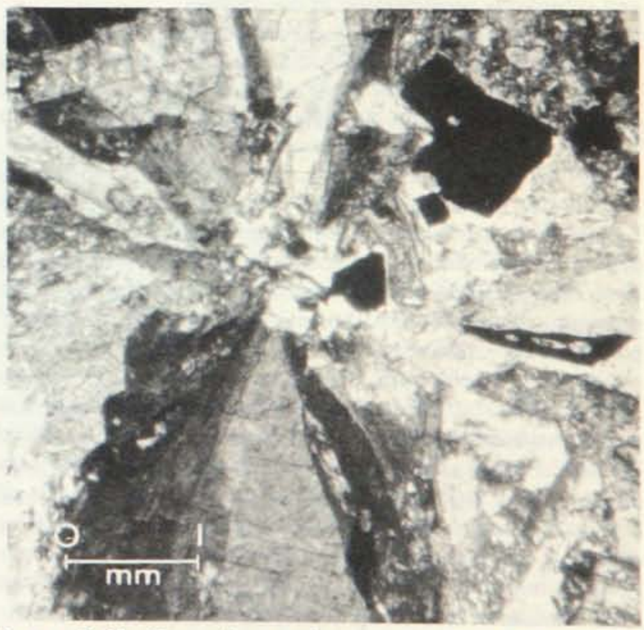

Figure V-63. Hour-glass extinction in zoned pyroxene. Note the patchy birefringence within individual sectors.

Figure V-64. Optically continuous augite (cpx) enclosing several grains of twinned plagioclase. Birefringence of the augite is variable as in Figure V-63.

clase-alkali feldspar textural and compositional relationships, but the work of Ernst (1960) on the Endion sill at Duluth has shown that detailed study can provide insight into the control of intensive parameters on crystallization and subsolidus equilibration of the diabase.

Although quartz most commonly occurs in granophyric intergrowths, it also occurs as discrete grains set in a matrix of layered silicates. This latter occurrence is related to hydration reactions involving earlier silicates.

Alteration

A systematic study of the alteration minerals in the Logan intrusions has not been made. Sericite, chlorite, talc, serpentine, calcite, rutile, hematite, goethite, and epidote have been reported, but detailed textural and compositional data are not available to adequately assess the validity of proposed alteration reactions.

Sericitic alteration of plagioclase is ubiquitous (figs. V-59C and D), and in many cases calcic cores of plagioclase grains have been preferentially altered over more sodic borders. Plagioclase phenocrysts in diabase porphyry generally have a sericitic border surrounded by a thin rim of cloudy feldspar—either albite or oligoclase—and it is possible that sodium liberated in the plagioclase-sericite reaction has migrated outward to form a sodic rim. Plagioclase associated with interstitial granophyric intergrowths is extensively altered (fig. V-59D).

Uralitization of pyroxene is common, and ranges from minor alteration along cleavage planes to a complete transformation to amphibole and an opaque oxide. Exsolution and oxidation of the primary opaque oxides have produced magnetite-ilmenite intergrowths, and where alteration is severe, magnetite has altered to goethite and hematite, as described above.

A colorless amphibole was found in some sections without associated relict pyroxene. It occurs in felty masses interstitial to plagioclase in a diabasic to poikilitic texture similar to that of the pyroxene. This occurrence suggests that it is possible that amphibole crystallized in parts of some sills directly from the melt before or during crystallization of granophyre.

There are two general characteristics of the alteration pattern. First, the upper and lower parts of some sills are not extensively altered (Mathez, 1971, *op. cit.*). Second, even where alteration is severe, it tends to be restricted to small clustered volumes less than a centimeter across. The first observation suggests that in the early stages of crystallization, the melt was not saturated with respect to volatiles, whereas the second suggests that when a water-rich gas phase ultimately separated from the melt it occurred within a crystal mush which inhibited diffusion over large volumes. Thus localized, small volumes of gas reacted with immediately adjacent anhydrous phases to produce patchy alteration products.

Textures

The diabase in the Logan intrusions is characterized by a fine- to medium-grain size. Plagioclase, for example, rarely exceeds five millimeters in diameter, even in sills as much as 1,000 feet thick. The grain size is directly related to sill thickness, but aside from casual field observations no systematic studies of the grain-size distribution have been made.

Perhaps the most distinctive textural characteristic of the Logan intrusions is the diabasic intergrowth of plagioclase and pyroxene (figs. V-51B and V-61). However, a variety of textures occurs in the sills, reflecting complexities involving the rate of crystallization, relative amounts of crystallizing phases, flow during crystallization, and crystal segregation by flow or contrasting densities. Thus, pyroxene occurs as isolated grains interstitial to euhedral plagioclase, as optically continuous oikocrysts enclosing plagioclase, or as grains enclosed within plagioclase (fig. V-64). The diabase, therefore, does not afford unambiguous textural evidence pertaining to the paragenetic relationship of plagioclase and pyroxene. However, it is clear from the occurrence of plagioclase phenocrysts in the chilled margins that plagioclase was on the liquidus when the magma was intruded. For those sills which contain a layer of diabase porphyry, a significant amount of plagioclase must have crystallized before pyroxene. The extent of plagioclase crystallization before intrusion is discussed below in the section on the origin of the porphyry layers. Aside from this problem, the prevalent diabasic texture in the main part of the sills indicates a cotectic relationship between plagioclase and pyroxene and only minor crystal segregation during crystallization. The latter interpretation also is apparent from the modal variations across the sills (fig. V-60).

The textures also indicate the simultaneous crystallization, at some stage, of opaque oxides, plagioclase, and pyroxene. Euhedral or skeletal crystals of opaque oxides are enclosed by plagioclase and pyroxene, but the latter also are found between the skeletal ribs. In addition, euhedra, skeletons, and needles of opaque oxides crosscut grains of pyroxene and plagioclase. Thus, in all cases, it is not possible to conclude from the textural evidence whether an opaque oxide phase preceded or followed pyroxene.

Quartz and potassium feldspar in granophyric intergrowth clearly are interstitial to plagioclase, pyroxene, and ilmenite and were the last phases to crystallize (fig. V-59D). Several textural features referred to above are similar to those found in some lunar basalts. Skeletal opaque oxides, regularly and irregularly zoned pyroxene, and cored plagioclase (figs. V-59B and V-62; Roedder and Weiblen, 1971) are common in both the diabases and lunar basalts and indicate similar cooling histories. Cored plagioclase, produced in synthetic melts by a limited range of undercooling and fast cooling rates (Lofgren, 1972), attests to the hypabyssal character of these rocks.

Diabase Porphyry

The essential difference between diabase and diabase porphyry is the amount and grain size of the plagioclase. In the Rose Lake sills, plagioclase comprises about 35 percent of the rock and the average grain size is 1.4×0.7 cm. The diabase porphyry most commonly occurs in the upper parts of sills (fig. V-60), where thicknesses of as much as 140 feet have been observed in some of the thicker sills. Contacts between diabase and diabase porphyry are sharp, but evidence indicative of intrusive relationships has not been observed. Planar orientation of plagioclase is common. The amount of late-stage granophyre in the diabase porphyry appears to be greater than in the diabase and alteration generally is more severe (fig. V-59C).

The presence of plagioclase phenocrysts in many chilled zones suggests that this phase was on the liquidus at the time of emplacement. Therefore, it is likely that at least some porphyry zones resulted from gravitational segregation in which the plagioclase either floated or sank (Grout and Schwartz, 1933). The relationship of this mechanism to the origin of the diabase porphyry zones in the Rose Lake sills is considered in light of modern data on densities and viscosities and alternate hypotheses of multiple intrusions and flow differentiation by Mathez (1971, *op. cit.*).

Granophyre

Granophyric intergrowths of quartz, sodic plagioclase, and orthoclase have been shown to be a minor, texturally late constituent of diabase (fig. V-59D), occurring most extensively in the upper parts of sills and in the diabase porphyry zones. Separate bodies of granophyre occur in both the diabase and the country rocks and range in size from dikelets to mappable intrusive masses. The granophyre in these bodies is texturally and mineralogically similar to the interstitial granophyre. The areally restricted amount of mappable granophyre in the Hungry Jack Lake and South Lake quadrangles (Mathez, 1971, *op. cit.*, p. 54) suggests that it comprises a relatively small proportion of the total sill complex. This in turn suggests that the origin of these masses does not require atypical assimilation of country rocks by the Logan magmas.

Inclusions

Inclusions in the Logan intrusions are of two types, hornfels derived from the Rove Formation, and igneous inclusions of an anorthositic affinity. The Rove inclusions are most abundant in the upper and lower parts of sills and in chilled margins, whereas the anorthositic inclusions more typically occur in the upper parts of sills. The inclusions range in size from a few centimeters to tens of centimeters and may be either angular or rounded. No detailed studies of the mineral assemblages in the inclusions are available, but field descriptions of Rove inclusions suggest that they are mineralogically similar to hornfels in contact zones (Morey, this volume). The absence of extensive reaction rims around the Rove inclusions suggests that the country rock was not assimilated to any great extent during the emplacement of the magma.

Detailed descriptions of anorthositic inclusions are summarized by Phinney (1968, p. 137). Most of the inclusions he described occur in the Pigeon Point sill. There, Grout and Schwartz (1933) concluded that they were not cognate, but represented material derived at depth which could be distinguished from diabase porphyry on the basis of texture and mineral compositions. Similar inclusions occur in the South Lake and Hungry Jack Lake quadrangles, but no detailed descriptions are available. Because the diabase porphyry represents layers in which plagioclase was concentrated in place, noncognate inclusions derived from a deep source are distinguishable by their more primitive interstitial assemblages and deformation textures.

Bulk Composition

One wet chemical analysis of the chilled phase of the type of intrusions described in this paper is listed in Table V-35. Also included in the table are bulk compositions calculated from the modes and mineral compositions of Rose Lake sill A. Several other analyses of rocks with similar and contrasting compositions are listed for comparison. The compositions of the Logan intrusions resemble the differentiated lavas of Kilauea, Hawaii, quartz tholeiite of Iceland and the North Shore Volcanic Group (Green, this chapter), diabase sills in the vicinity of Duluth, and the Early Mafic intrusions of Geul (1970). The bulk compositions of all these rocks are consistent within a reasonable sampling and analytical error for each element, and they are clearly distinct from Geul's Pigeon River intrusions and the Pigeon Point sill (nos. 7 and 8, table V-35). It can be seen from a magnesium variation diagram (fig. V-65) that, to a first approximation, the Logan intrusion bulk composition can be reached by the crystallization of olivine, augite, and plagioclase from a tholeiitic melt just prior to the crystallization of ilmenite.

The similarity between analyses of chilled diabase and analyses of medium-grained diabase from the upper parts of sills (table V-35) supports the contention based on textural evidence that sills as much as 500 feet thick solidified without appreciable gravity segregation of mineral phases. This condition is necessary, of course, if the analyzed compositions of chilled diabase and the calculated compositions of diabase from modal and mineral data are to agree. The possible significance of minor differences in silica, alkalis, and water is discussed below.

Crystallization

The petrographic data summarized above indicate the following crystallization sequence. Plagioclase, on the liquidus at the time of intrusion, was followed closely by pyroxene and opaque oxide in indeterminate order. When the magmas were about 90 percent crystallized, these phases were followed by potassium feldspar, quartz, and apatite. The magma probably was undersaturated with respect to water at the time of intrusion, but saturation was exceeded during the course of crystallization, and a separate water-rich phase reacted with the early-formed anhydrous minerals during the late stages of crystallization. However, the initial water content of the magma was sufficiently great in some cases for actinolite rather than clinopyroxene to crystallize in the chilled margins of the stratigraphically deepest sills.

The distribution of plagioclase, clinopyroxene, and ilmenite (fig. V-60) indicates that gravity segregation of these phases did not occur to any appreciable extent in the Logan sills. However, segregation of granophyre (fig. V-60) and hydrous phases toward the upper parts of the sills defines a stratification which can be explained by the upward migration of water from incremental volumes of crystal mush into adjacent magma so as to maintain a uniform chemical potential as crystallization proceeded inward (Hamilton, 1965).

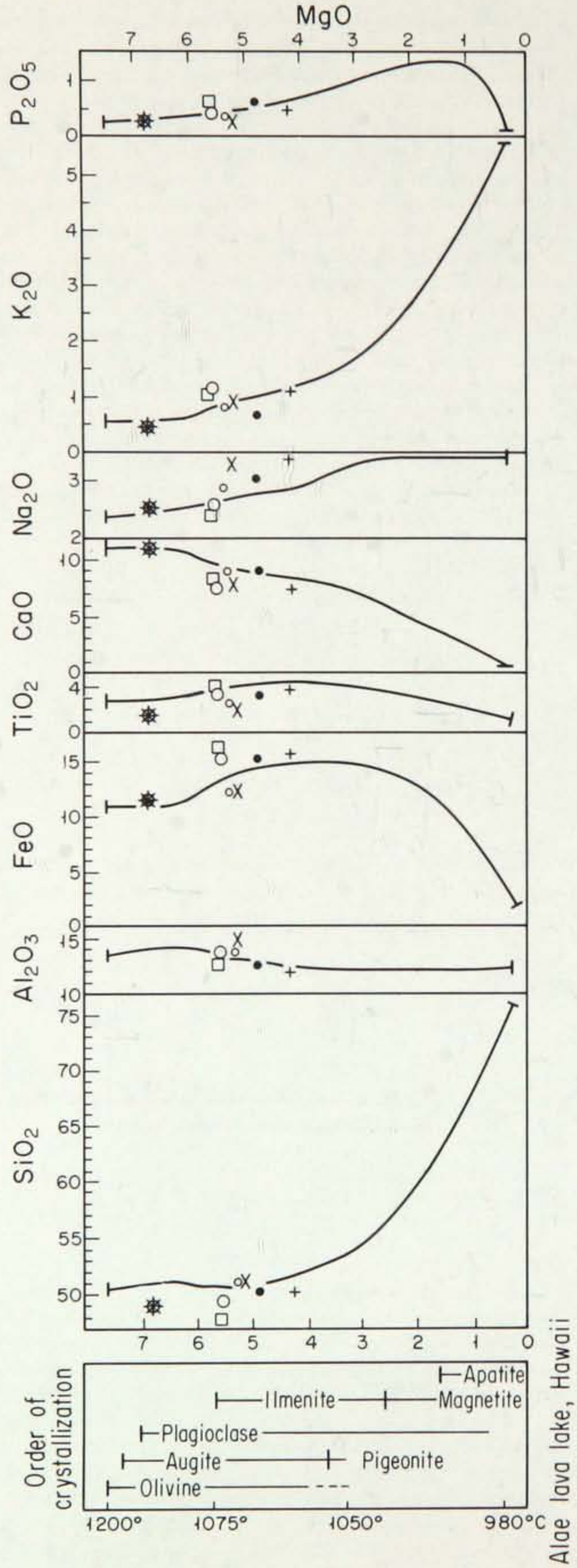

Figure V-65. Magnesium variation diagrams showing the differentiated nature of Logan magma relative to that of the liquid line of descent for Alae lava lake, Hawaii. The approximate correlation between liquid temperature and composition is shown at the bottom along with the approximate temperature of first appearance of minerals on the liquidus. Other compositions shown include: +, Logan intrusions, sill A at Rose Lake; ■, Northland sill at Duluth; O, Early Mafic intrusions of Geul (1970); X, average of four quartz tholeiites, North Shore Volcanic Group, analysis no. 7, Table V-4 (Green, this chapter); *, average of four olivine tholeiites, North Shore Volcanic Group, analysis no. 1, Table V-4 (Green, this chapter); ·, average of 7 (quartz) tholeiites, Thingmuli volcano, analysis no. 8, Table V-4 (Green, this chapter); o, Hawaiian differentiated lava, analysis no. 5, Table V-4 (Green, this chapter).

Early plagioclase, which crystallized near the base of Rose Lake sill A, is characterized by rims that are less sodic than are the plagioclases in the chilled margins; however, in the upper part of the sill, the plagioclase rims are more sodic than are the plagioclase phenocrysts in the chilled margins (fig. V-60). The uniform composition of the plagioclase cores suggests that plagioclase first crystallized throughout a magma having a fairly uniform bulk composition, whereas the sodic rims indicate that the bulk composition of the interstitial melt changed as solidification proceeded. This change most likely resulted from the concurrent migration of water and alkalis (Richter and Moore, 1966) and strongly implies that the alkalis migrated upward during crystallization.

The amount of water in the magma also increased as crystallization proceeded inward from the sill contacts, and this increase most likely affected the crystallization sequence. Inasmuch as the crystallization of plagioclase relative to clinopyroxene is depressed by increased water pressure (Yoder, 1968), clinopyroxene may have been the first silicate to crystallize in the central part of each sill and this may account for some of the ambiguous textural relations between plagioclase and pyroxene (figs. V-59B, V-61, and V-64). Similarly, a higher water content would also favor a higher partial pressure of oxygen and would result in the increased crystallization of opaque oxides (Roeder and Osborn, 1966). Thus, the trend toward slightly more pyroxene and oxide at the expense of plagioclase in the upper parts of the sills is compatible with an increase in the amount of water during crystallization (fig. V-60). In effect, the limited data suggest that the unique stratification of the sills reflects the relative mobility of various components in the interstitial melt.

INTERPRETATION

The rocks of the Duluth Complex (Phinney, Davidson, and Bonnichsen, this chapter), the North Shore Volcanic Group (Green, this chapter), and the Logan intrusions have been interpreted since the earliest work as recording a single but complex large-scale igneous process. The exposed rocks are clearly a part of the province associated with the Midcontinent Gravity High (Thiel, 1956); and the present consensus is that rifting was the essential large-scale process (Lidiak, 1964; King and Zietz, 1971; Green, this chapter) that transformed the Precambrian continental crust into a faulted terrane thickened by a substantial volume of mafic intrusive rocks and overlain by a thick succession of volcanic rocks. A close relationship between tectonic and igneous activity is characteristic of the process of rifting (Meissner and Berckhemer, 1967; Mueller, 1970), and our study of the available data on the Logan intrusions suggests that this relationship is fundamental to understanding the mode of origin of the Logan magmas, their relationship to other igneous rocks of the Late Precambrian in northeastern Minnesota, and their emplacement, cooling history, and subsequent deformation.

Rifting provides a mechanism for developing magma chambers at intermediate and shallow depths in the crust (Illies, 1970, p. 9). Such chambers are postulated by Phinney (this chapter), who has shown that a sequence of lava flows of the North Shore Volcanic Group can be derived by differentiation of a magma of the composition of an early flow. Similarly, the composition of the Logan intrusions indicates that they too are a product of differentiation (fig. V-65). Sufficient data are not available to postulate a specific parent magma composition, but Figure V-65 indicates that the Logan magmas are similar to the magmas produced by differentiation of oceanic tholeiite at shallow depths.

The average olivine tholeiite of the North Shore Volcanic Group (Green, this chapter) is a logical candidate for a parent magma, but it has a significantly higher alumina content than Hawaiian tholeiite (fig. V-65). However, the Logan magmas could be produced from the olivine tholeiite by the crystallization of olivine and a significant amount of plagioclase relative to clinopyroxene. This crystallization scheme would be favored at low pressures (Green and others, 1967; Kornprobst, 1970).

Based on the field relations and the absence of amphibole in the primary igneous assemblage, Mathez (1971, *op. cit.*, p. 77) concluded that the Logan intrusions were emplaced under a load pressure between 0.5 and one kilobar (depths of 2-4 km). The relatively fast cooling rates indicated by the diabasic texture suggest that the lower value is more reasonable.

The observations thus far provide us with a tentative model of (1) derivation of a parent magma at depths of at least 15 to 35 km (Green, this chapter), (2) differentiation in magma chambers at some shallower depth (Phinney, this chapter) involving crystallization of olivine, plagioclase, and some clinopyroxene, and (3) separation of the Logan magmas prior to the crystallization of ilmenite.

In this model, different units of the Duluth Complex represent the cumulus fraction and the Logan intrusions and North Shore Volcanic Group the separated fraction of parent magmas.

A comprehensive study is needed to elucidate definite genetic relationships between specific rock units. Mass balance calculations by the method of Wright and Doherty (1970) are currently being attempted. Presently, the number of variables appears to make the problem indeterminate. For example, the Logan intrusions plus certain layered troctolite units of the Duluth Complex might represent the complete products of one episode of igneous activity; on the other hand, accumulation of anorthositic gabbro in the upper part of an intermediate chamber and separation of a significant volume of lava prior to the separation of a Logan intrusion magma might also be part of the same episode. Interpretation of the tectonics and stratigraphy based on field relations and geophysical data places restraints on these possibilities, however. Clearly, minor and trace element chemistry will also be very useful in resolving the problem of Late Precambrian petrogenesis.

The specific location of postulated intermediate chambers from which the Logan magmas separated is not known, but a decrease in relative abundance of exposed igneous rocks eastward from the Long Island Lake-Hungry Jack Lake area (fig. V-56) implies that the intermediate magma chambers for the Logan intrusions were near this area rather than in the Grand Portage area as suggested by

Grout and Schwartz (1933). White (1966a, p. E19) has inferred from gravity and magnetic data that the first period of deformation in the western Lake Superior region consisted of a northeastward-trending ridge of pre-Keweenawan rocks. The axis of this ridge projects northeastward through the Beaver Bay Complex (Green, this chapter), and effectively separates rocks of the North Shore Volcanic Group into two distinct basins. This structure projects northeastward into northwestern Cook County, which is coincidentally the possible location of the postulated intermediate magma chambers for the Logan intrusions. White (1966a) was unable to determine if this ridge was the result of arching or block faulting. It is interesting that arching followed by block faulting is considered the normal sequence of tectonic events in rifting (Illies, 1970). Such deformation establishes a mechanism for tensional structures in layered strata, and permits passive emplacement of magmas in voids which thin away from structural highs.

The geometry and outcrop pattern of the Logan intrusions are consistent with emplacement away from an arch with a northward-trending axis. However, these rocks now define the north limb of the Lake Superior syncline. The change in tectonics from early arching along a north-south axis to later tilting and faulting along an east-west axis may reflect evolutionary changes in the development of the overall structure associated with the Midcontinent Gravity High, particularly adjustments to pre-existing crustal structure as rifting progressed.

The fracture system associated with the deformation of the Logan intrusions is now occupied by dikes and sills of olivine diabase (Pigeon River intrusions of Geul, 1970; table V-35). The composition of the olivine diabase which fills the fracture system of the late deformation (table V-35) indicates transport of parental magma from a relatively deep source without significant differentiation in intermediate chambers. In contrast to arching, faulting provides a mechanism for this type of hypabyssal dike and sill emplacement.

The diabasic rocks of Cook County, Minnesota can be divided into at least two distinct groups (the earlier Logan intrusions of this report and the later olivine diabase dikes and sills), but it is more difficult to define their stratigraphic relationships relative to other Keweenawan igneous rocks. Books (1968) has suggested that the Lower-Middle Keweenawan boundary be redefined on the basis of magnetic polarity (see Green, this chapter). DuBois (1962), Palmer (1970) and Robertson and Fahrig (1971) have shown that the Logan intrusions of this report (or the Early Mafic intrusions of Geul, 1970) and some associated dikes have a reverse polarity, whereas the olivine diabase dikes (or the Pigeon River intrusions of Geul, 1970) have a normal polarity. Thus, the Logan intrusions would be Lower Keweenawan whereas the olivine diabase dikes would be Middle Keweenawan. This classification scheme has many advantages in a geologic terrane characterized by generally similar rock types, but unfortunately the time significance of these subdivisions is not well documented. On the basis of analogous pole positions with rocks whose ages are known, DuBois (1962, p. 65) suggested that the reversely magnetized rocks may be as much as 1,400 m.y. old. Likewise, because the pole positions of the olivine dikes are similar to those of the Duluth Complex, DuBois (1962, p. 65) concluded that they were emplaced about 1,100 m.y. ago—the presumed age of the complex (Silver and Green, 1963). In contrast, Robertson and Fahrig (1971) concluded that the different pole positions resulted from the relatively rapid movement of the paleomagnetic pole at about 1,100 m.y. because K-Ar ages (990-1,055 m.y.) from dikes are statistically indistinguishable from those of the sills (730-1,060 m.y.). On the other hand, Hanson and Malhotra (1971) have dated one Logan sill at 1,300 m.y., and this value is consistent with DuBois' original interpretation. Obviously, additional data are needed before the validity of either explanation can be determined.

The available data from Cook County and the Thunder Bay district suggest that it might be possible to distinguish Keweenawan rocks of various ages on the basis of their petrologic characteristics. However, unique sequences of igneous activity may be repeated in time, as demonstrated in the Hawaiian Islands (Powers, 1955). The Lester River, North, and Endion sills in the vicinity of Duluth (Schwartz and Sandberg, 1940) may be examples of this in the Keweenawan of Minnesota. They have chemical compositions similar to those of the Logan intrusions of this report (table V-35), but they intrude lava flows of Middle Keweenawan age (Green, this chapter). Thus, it remains to be conclusively demonstrated that the Keweenawan of Minnesota contains a simple record of two discrete magmatic events. Present data can only suggest that the synthesis of more comprehensive data on the hypabyssal rocks of Minnesota will provide an intriguing insight into critical processes associated with the geologic evolution in space and time of the Midcontinent Gravity High.

COOK COUNTY FISSURE VEIN DEPOSITS

M. G. Mudrey, Jr. and G. B. Morey

Several types of fissure vein deposits are present in the Thunder Bay district, Ontario, and similar deposits of apparently less economic significance occur in northern and eastern Cook County. The deposits occupy fractures that crosscut rocks of the Animikie Group and associated sills and dikes of diabasic gabbro and are composed of minerals containing lead, zinc, copper, and small amounts of silver.

The deposits occur within a triangular area about 100 miles long and 50 miles wide (fig. V-66). The oldest known rocks are Early Precambrian in age and were folded, sheared, and metamorphosed during the Algoman orogeny about 2,700 m.y. ago (Hanson and others, 1971b). A profound erosional unconformity separates the older rocks from the overlying Middle Precambrian Animikie Group. Subsequent to the Algoman orogeny, an unknown amount of the Animikie Group was removed by erosion prior to deposition of Upper Precambrian sedimentary rocks (Sibley Series), which took place about 1,376±33 m.y. ago (Franklin and Kustra, 1970). These sedimentary rocks are overlain disconformably by a thick sequence of interbedded basic and felsic flows of Keweenawan age—the North Shore Volcanic Group in Minnesota (Green, this chapter) and the Osler Series in Canada.

The Animikie Group and the Keweenawan flows were intruded by a series of sill-like gabbroic rocks, called "Logan intrusions" in Minnesota (see Weiblen and others, preceding paper) and "Early Mafic intrusions" in Canada (Geul, 1970). These sill-like bodies are characterized by reverse paleomagnetic polarity, and hence are assigned to the Lower Keweenawan (Green, this chapter; Books, 1968). Normally polarized olivine diabase dikes, named the Pigeon River intrusions by Geul (1970), crosscut the sills in Canada. In Minnesota, similar intrusive relations are particularly well exposed above Pigeon Falls on the Pigeon River. The hiatus between emplacement of the Logan intrusions and the Pigeon River intrusions is of unknown duration. Although Hanson and Malhotra (1971) reported a 1,300-m.y.-age for a "Logan sill" near Whitefish, Ontario, potassium-argon ages for the sills to the east range from 1,055 to 990 m.y. (Robertson and Fahrig, 1971, p. 1357) and appear to be indistinguishable from the potassium-argon ages of the dikes (730-1,060 m.y.). The older age also reflects the time of emplacement of the main part of the Duluth Complex (1,150±15 m.y., Silver and Green, 1963, p. 107), and perhaps of other small gabbroic bodies, such as the Crystal Lake gabbro (Geul, 1970; Franklin, as cited in Wanless and others, 1966, p. 54).

STRUCTURAL SETTING

In this region, the Middle Precambrian and younger rocks dip southward and define the northwest limb of the Lake Superior syncline (White, 1966a and b). The homoclinal dip in rocks of the Animikie Group is interrupted locally by zones of tight folds adjacent to some of the sill-like intrusions (Tanton, 1931) and at other places by broad gentle folds whose axes trend northward.

A fracture system consisting of faults and joints that trend predominantly east-northeast, northwest, and north is well developed along the north limb of the syncline. Most of the fractures are in two nearly parallel belts (fig. V-66). One belt contains a swarm of olivine diabase dikes exposed on numerous islands in Lake Superior in an area extending from Victoria and Spar Islands, near the International boundary at Pigeon River, northeastward for a distance of 50 miles beyond the once-famous Silver Islet. The other belt is approximately 2 miles wide and is on the mainland 20 miles northwest of the island belt; it extends from the vicinity of Thunder Bay, Ontario southwestward toward Gunflint Lake for a distance of about 90 miles. The island belt has been called the "gray argillite" belt, and the mainland belt has been called the "black slate" belt (Oja, 1967, p. 211).

The fracture zones within the two belts appear to define hinge lines that are related to subsidence of the Lake Superior syncline, for bedding in the Animikie Group adjacent to the mainland dips 5° SE. (Goodwin, 1960) and dips 15° SE. adjacent to the island belt (Grout and Schwartz, 1933), and the south sides of most faults have moved downward with respect to the north sides. The east-northeast-trending fractures, which are parallel to the axis of the syncline, cut the Logan intrusions and localized emplacement of the Pigeon River intrusions, especially in the island belt. It seems likely, therefore, that the Lake Superior syncline is a tensional feature that developed after emplacement of the Logan intrusions and prior to the emplacement of the Pigeon River intrusions and related rocks such as the Duluth Complex and the Crystal Lake gabbro of Geul (1970).

ECONOMIC GEOLOGY

The mineral deposits are closely associated spatially with Keweenawan igneous rocks and they are interpreted as having been derived from the same gabbroic magma source or sources (Tanton, 1931, 1935). In the mainland belt, all but the northward-trending fracture set is mineralized; however, most veins are in east-northeast-trending fractures, which were opened preferentially at the time of mineralization (fig. V-66). On the other hand, most of the veins in the island belt are in north- or northwest-trending fractures, suggesting that the east-northeast-trending fractures were filled by the Pigeon River intrusions prior to ore deposition. At least some fissure vein deposits developed after emplacement of the Pigeon River intrusions, for some veins cut dikes of the Pigeon River intrusions.

Description of Veins

Fissure vein deposits composed of minerals containing lead, zinc, copper, and silver were first discovered in the Thunder Bay district in 1846. Since discovery, the district has been a sporadic producer of silver, and ore valued at more than $4,770,000 has been shipped; most of the yield was prior to 1900. Similar deposits have been found in Cook County, but as yet none has proved commercial.

The structure of the veins and the texture of the ore minerals are typical of Lindgreen's (1933) epithermal class. Filling of open space was the dominant mode of formation. The veins consist of several types, including (1) simple veins, (2) composite veins, and (3) shatter zones containing brecciated country rock cemented by vein material. Simple veins, ranging in width from a fraction of an inch to 10 feet, occur most commonly in the igneous rocks, whereas com-

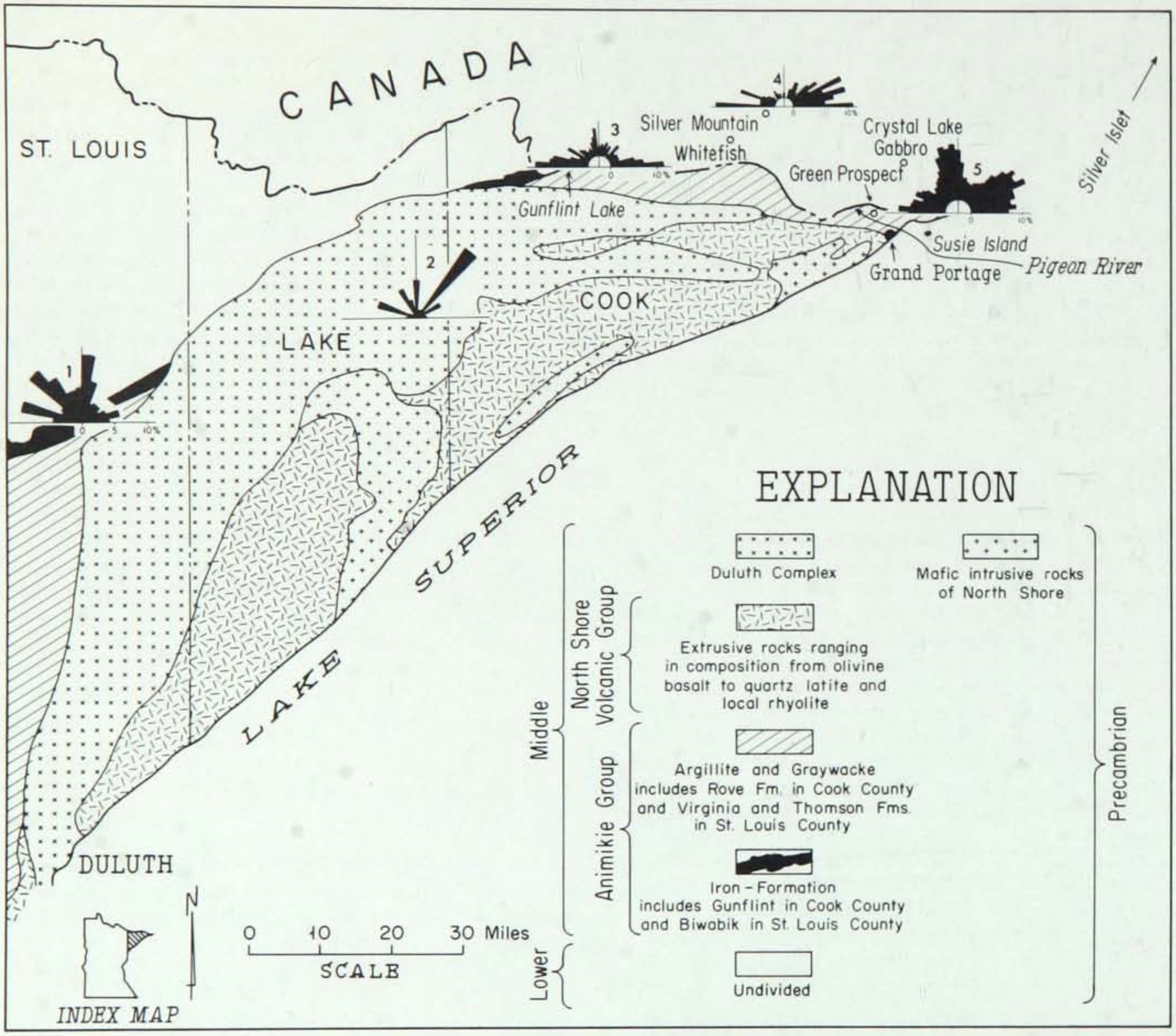

Figure V-66. Map of northeastern Minnesota showing relation of measured joints to areal geology: area 1, 290 joints in the Biwabik Iron-formation (White, 1954, p. 56, fig. 12); area 2, 106 joints in the Duluth Complex (Davidson, 1969a and b, p. 8, fig. 3); area 3, 261 joints and faults in Rove Formation and Logan intrusions (unpub. data of G. B. Morey from the 7.5-minute South Lake quadrangle); area 4, 121 mineralized veins in the Gunflint Iron-formation, Rove Formation, and diabasic rocks of the Thunder Bay district, based on data reported by Tanton (1931); and area 5, 577 joints in Rove Formation, Logan intrusions, Pigeon River intrusions of Geul (1970), North Shore Volcanic Group, and the Pigeon Point sill.

posite or brecciated veins are typically in sedimentary strata. Most of the latter vein types are less than 20 feet wide; a few veins are more than 100 feet wide and consist either of anastomosing veins of various trends and variable widths or of irregular fillings in fault-breccias.

Most of the veins contain variable proportions of the gangue minerals calcite, barite, quartz, and fluorite, and many veins contain only gangue minerals. The principal ore minerals include argentite, galena, and sphalerite; less common ore minerals are arsenides, sulfosalts and other copper and nickel sulfides. Small quantities of native silver have been reported from several veins (Tanton, 1931; Grout and Schwartz, 1933).

Commonly, successive crusts of different gangue minerals were deposited, giving rise to a well-defined paragenetic sequence (fig. V-67). The mineralogically different layers on opposite walls may be either symmetrical or asymmetrical, and filling commonly is incomplete, leaving vugs of various sizes. The vugs are partially filled either with other gangue minerals or with ore minerals of economic importance that were deposited during a later stage of deposition.

The Silver Islet mine in Canada, which is the largest silver producer in the district with a yield of more than $3,250,000, contains veins showing two distinct stages of hypogene deposition (fig. V-68). The silver occurs as bonanza-like lodes that apparently were deposited during both stages. Tanton (1931, p. 101) reported ". . . a mass rich in the secondary phase known as the second bonanza . . . was 5 feet wide, 50 feet long in its upper part, and tapered out downward . . ." from 150 to 250 feet below the level of Lake Superior. This body yielded 800,000 ounces of silver. In addition, supergene alteration took place near the present surface of the ore body (within 10 feet of the surface), giving rise to a third mineral association. The mineralogy and paragenesis of the second hypogene episode is essentially identical to that of many other base metal-bearing veins in the district, including some in Minnesota.

Several other unique sulfide-bearing deposits, such as on Susie Island south of Pigeon Point, have been described from Minnesota by Schwartz (1924, 1925, 1928). On Susie Island, calcite and barite are the principal gangue minerals (fig. V-69), but unlike other fissure vein deposits in the district much of the calcite contains minute inclusions of chalcocite and exsolved bornite (fig. V-69). Most of the other sulfides, such as pyrite, chalcopyrite, bornite, chalcocite (II, fig. V-69) and covellite are later than the gangue minerals. Evidently pyrite was the first sulfide formed, for

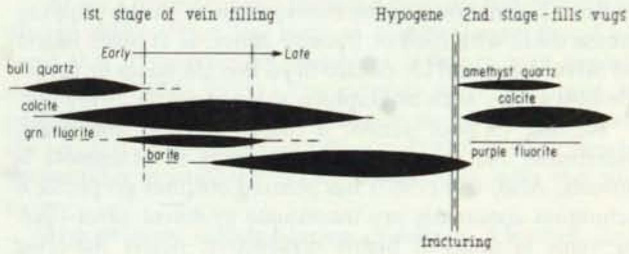

Figure V-67. Paragenetic sequence in veins from the mainland belt containing only gangue minerals (based on data from Ingall, 1887 and Tanton, 1931).

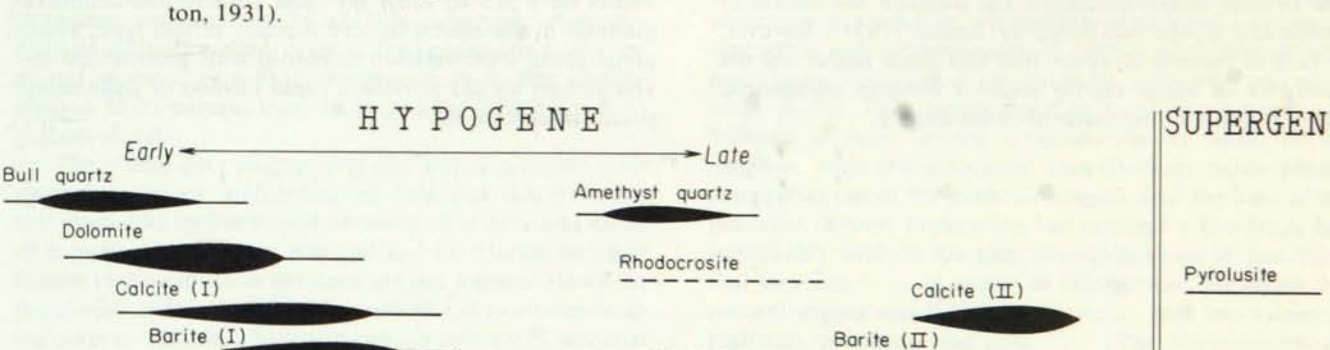

Figure V-68. Inferred paragenetic sequence in the Silver Islet mine (based on data from Tanton, 1931).

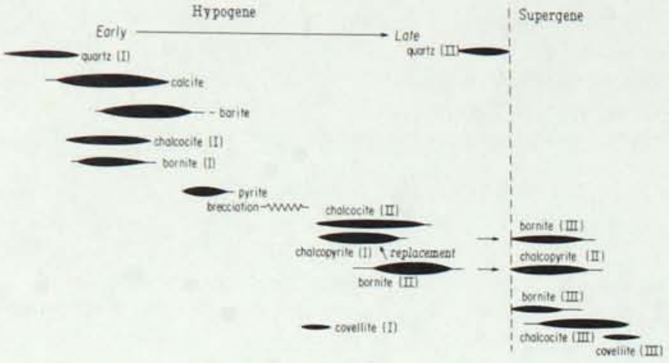

Figure V-69. Paragenetic sequence of the Susie Island vein (based on data from Schwartz, 1928).

it has been brecciated and replaced by a complex assemblage of either chalcopyrite, chalcocite, or bornite. Schwartz (1928) concluded that all the above minerals were hypogene because of their complete independence of visible channels, cleavage, and mineral boundaries. However, supergene sulfides also are present. Chalcopyrite replaced bornite along cracks and cleavage planes. Supergene bornite replaced hypogene chalcocite and chalcopyrite, and in turn was replaced by chalcocite and covellite. Hand picked ore assayed as much as 6.22 percent copper (Grout and Schwartz, 1933, p. 64).

In general, the vein deposits seem to have been formed during two distinct periods, separated at least locally by minor fracturing and shearing. In some Canadian prospects, such as near Silver Mountain, the intimate association of fluorite and galena was noted by Tanton (1931); however, the lack of fluorite at Silver Islet and Susie Island and the abundance of barite clearly imply a complex paragenetic history for all the vein material of the district.

Guides for Exploration

Geologists have been puzzled by an apparent lack of economically valuable fissure vein deposits in Minnesota as compared to Canada. The apparent discrepancy can be explained in several ways.

First, the mainland belt, which contains most of the commercial silver deposits in Canada, predominantly transects the Gunflint Iron-formation and the lower part of the younger Rove Formation in Minnesota. Inasmuch as this part of the section is truncated by the Duluth Complex in the vicinity of Gunflint Lake, near the International boundary, the potentially mineralized area along this belt in Minnesota is fairly small. Similarly, the island belt traverses only a small part of Minnesota—in the Grand Portage-Pigeon River area—and the potentially mineralized area here too is small. Inasmuch as scattered vein deposits occur outside the limits of these two belts in Ontario (Tanton, 1931), however, detailed mapping in Minnesota might be rewarding. The known deposits were localized by structural controls, and similar structural traps should be sought in Minnesota. Such traps include: (1) synformal structures in the Rove Formation; (2) the intersections of steeply-dipping diabase dikes with fault or fracture zones, as at Susie Island and Silver Islet; and (3) calcite-filled breccia zones or favorable host rocks, such as graphitic shale and iron-formation.

Because of poor access, a rugged terrane, and heavy underbrush, the search for vein deposits in Minnesota is difficult. Also, Oja (1967) has pointed out that geophysical techniques apparently are inadequate to detect silver-bearing veins in areas of highly conductive, nearly flat-lying argillaceous strata, and that magnetic surveys are no better than visual investigations of aerial photographs to detect fault zones. However, Oja has stated that geochemical techniques have proved to be the most positive and definitive methods in the search for ore deposits of this type. Thus, aerial photo interpretation combined with geochemical investigations should provide a rapid method of delineating potential target areas.

MAGMATIC SULFIDES AND ASSOCIATED FISSURE VEIN DEPOSIT AT THE GREEN PROSPECT, COOK COUNTY

M. G. Mudrey, Jr.

Several tabular gabbroic bodies, previously thought to belong to the Logan intrusions, contain copper- and nickel-bearing sulfides that were segregated near the base as primary magmatic phases. Most of the known sulfide-bearing rocks are olivine-bearing dikes that are in the so-called island belt in Ontario; Geul (1970) assigned these rocks to the Pigeon River intrusions. This geologic terrane extends southwestward from Pigeon River into Minnesota for a distance of at least 10 miles, and in this area scattered occurrences of copper and nickel sulfides have been reported from the Green prospect (fig. V-66) and other localities (Schwartz, 1925; Grout and Schwartz, 1933).

At the Green prospect there is a primary sulfide-bearing olivine diabase that has been so highly fractured and altered that it superficially resembles a fissure vein deposit (Schwartz, 1924, 1925). The primary sulfides consist of pentlandite, pyrrhotite, and chalcopyrite, and judged from the textural observations of Schwartz (1925, p. 263) clearly the sulfides crystallized contemporaneously with the silicates.

The primary sulfide-bearing diabase is fractured, and brecciated fragments of diabase within the "vein" are markedly altered. The gangue minerals are principally xonotlite ($Ca_5Si_5O_{14}(OH)_2$) and prehnite. Secondary copper sulfides, particularly chalcopyrite, in the "vein" are most abundant at fracture intersections. In addition, supergene chalcopyrite and violarite replace some of the pentlandite in the unaltered diabase. A sulfide concentrate from this locality assayed 32.63 percent iron, 18.26 percent copper and 0.52 percent nickel.

The above data suggest that the Green prospect consists of a primary, sulfide-bearing rock that was fractured and altered by hydrothermal solutions. The time and mode of formation of the vein material and its relation to other fissure vein deposits in the area are not known. However, the olivine diabase is similar in composition to olivine-bearing dikes in Canada, which also contain substantial amounts of copper-nickel sulfides and have been known for at least 50 years (Tanton, 1935). Typically, all dikes are composed of diabasic gabbro, which locally grades into a fine-grained, pyroxene-rich phase, which Tanton (1935) tentatively identified as norite. Both phases are mineralized with sulfides. Pyrite, pyrrhotite, pentlandite, and chalcopyrite are intergrown and are disseminated in the rock as fillings in interstitial voids. The same intergrown sulfides also occur in the norite as nodules or globules as much as 2 inches in diameter. The nodules or globules themselves are distributed as stringers of variable length and width.

In the older literature, the olivine diabase dikes were considered part of the Logan intrusions (Tanton, 1931, 1935). Geul (1970), however, has shown that the dikes are petrologically distinctive, and he assigned them to a new unit, the Pigeon River intrusions. The Pigeon River intrusions occupy fractures that are later than the Early Mafic intrusions of Geul (1970), which in Minnesota are called Logan intrusions (see Weiblen and others, this chapter); the Pigeon River intrusions appear to be coeval with, or younger than, the Crystal Lake gabbro of Geul (1970). In the same way as the dikes, emplacement of the Crystal Lake gabbro was structurally controlled by fractures and a synclinal trough in argillite of the Rove Formation.

The Crystal Lake gabbro was once thought to be a "Logan" dike-sill combination, but is now recognized as a distinctive intrusive body consisting of olivine gabbro, troctolite, anorthositic gabbro, and fine-grained diabase (McRae and Reeve, 1968; Geul, 1970). There is a trend toward a less mafic composition with increasing distance above the base of the intrusion. Mainwaring (1968, as cited in Geul, 1970) concluded that two distinct ore mineral assemblages are present: (1) a high-temperature magmatic assemblage, consisting of pyrrhotite, pentlandite, chalcopyrite, cubanite, magnetite, and ilmenomagnetite; and (2) a low-temperature assemblage consisting of mackinawite, marcasite, and nickeloan pyrite. The copper-nickel-iron sulfide minerals are believed to have formed syngenetically as liquid sulfide droplets, with the associated iron-titanium oxide phases segregating out of the gabbroic magma near the base of the intrusion. Recent exploration has outlined a low-grade but remarkably uniform ore zone averaging about 90 feet thick that contains ". . . in excess of 40,000 tons averaging 0.4 percent copper and 0.20 percent nickel, plus low values in platinum, paladium, and gold. . . ." (*The Northern Miner*, March 26, 1968).

Inasmuch as the olivine diabase at the Green prospect and other bodies of diabase in northeastern Minnesota are similar in many respects to the sulfide-bearing rocks exposed across the International boundary in Canada, these rocks should be evaluated for possible low-grade concentrations of copper and nickel.

PUCKWUNGE FORMATION OF NORTHEASTERN MINNESOTA

Allen F. Mattis

The Lower Keweenawan Puckwunge Formation of northeastern Minnesota is intermittently exposed for a distance of 25 miles from Stump Lake in T. 64 N., R. 2 E. east to Lucille Island, south of Pigeon Point in Lake Superior (fig. V-70). In 1893, N. H. Winchell designated exposures in the NE¼ sec. 25, T. 64 N., R. 3 E. as the type section of the Puckwunge conglomerate. Thirty-six feet of coarse and fine conglomerate is exposed along a small, northward-flowing intermittent stream. The conglomerate is overlain by 90 feet of sandstone (Winchell and others, 1899). Believing he was in the valley of Puckwunge Creek (now called Stump River), Winchell named the formation the Puckwunge conglomerate; but according to Grout and others (1959), the outcrop is actually located near Portage Brook. Because both sandstone and conglomerate are present, the unit is now known as the Puckwunge Formation.

At Nopeming, just west of Duluth, 25 feet of conglomerate and quartzite is exposed intermittently over a distance of half a mile in secs. 17 and 20, T. 49 N., R. 15 W. (fig. V-70). These rocks resemble the Puckwunge Formation of northeastern Minnesota lithologically, and in the past have been correlated with the Puckwunge. In the same way, two nearby conglomerate outcrops in secs. 1 and 15, T. 48 N., R. 16 W. were previously correlated with the sedimentary rocks at Nopeming; however, Morey (1967a) reassigned these two exposures to the Upper Keweenawan Fond du Lac Formation. The age and correlation of the quartzite and conglomerate at Nopeming have been in question since 1889, when Winchell's publication of the Short Line Park well record (Winchell, 1889) suggested that the unit may be an interflow sandstone associated with lavas of the North Shore Volcanic Group. The exposures in northeastern Minnesota and at Nopeming will be described separately in this report because they are 130 miles apart and their correlation is questionable.

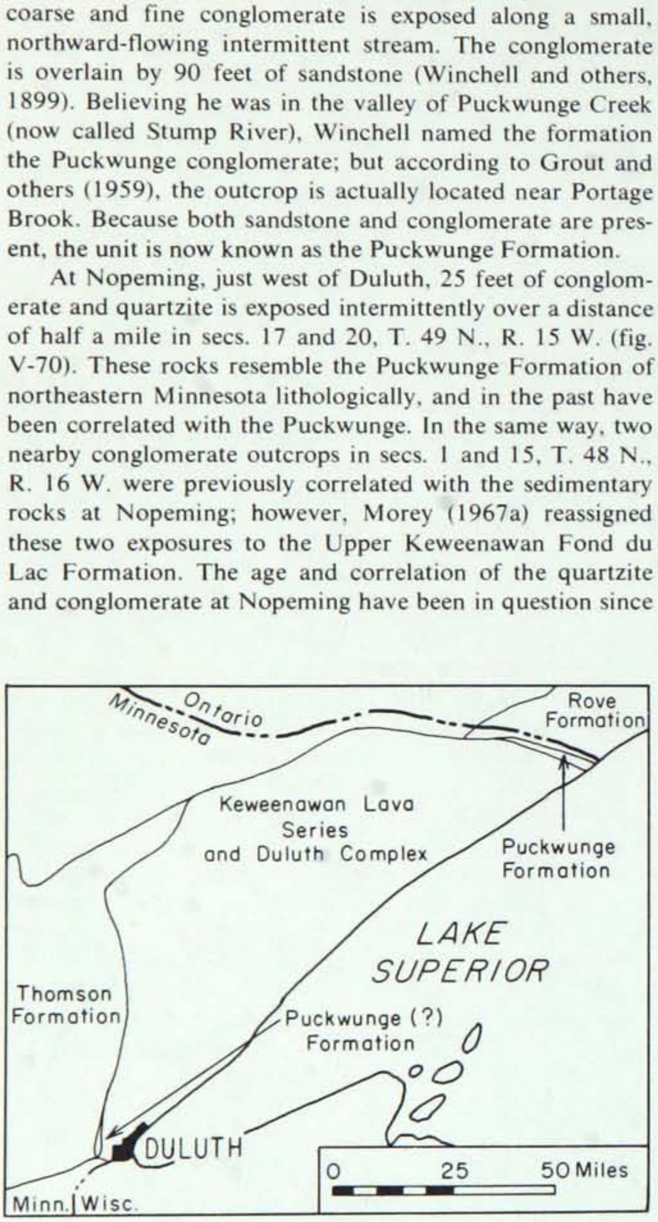

Figure V-70. Geologic map of northeastern Minnesota showing the location of the Lower Keweenawan Puckwunge Formation.

THE PUCKWUNGE FORMATION OF COOK COUNTY

Stratigraphy and Petrology

In extreme northeastern Minnesota, the Puckwunge Formation unconformably overlies the Middle Precambrian Rove Formation. The contact between the two formations is not exposed; however, a conglomerate, believed to be a basal conglomerate, is exposed at two localities. Because the Rove Formation was not deformed during the Penokean event (Morey, 1969), the relationship between the two formations is a disconformity rather than an angular unconformity.

The Puckwunge Formation appears to be overlain unconformably by Keweenawan lavas of the North Shore Volcanic Group. Both the basal lava flows and the Lower Keweenawan sedimentary rocks dip gently south-southwest; the Puckwunge Formation is exposed beneath the lava flows at the base of a steep northward-facing slope. However, the presence of angular quartzite inclusions in the basal lava flow on Grand Portage Island indicates that lithification of the Puckwunge Formation preceded extrusion of the lavas; therefore, an unconformity is indicated.

Two major igneous units intrude the Lower Keweenawan sedimentary rocks. The Logan intrusions transect the bottom of the formation, apparently following the contact with the underlying Rove Formation, and the Duluth Complex intrudes the sedimentary rocks. With the exception of a small outcrop near Stump Lake in T. 64 N., R. 2 E., the Duluth Complex forms the western boundary of the Puckwunge where it truncates the formation near Devilfish Lake in T. 64 N., R. 3 E.

Stratigraphically, the formation can be divided into two members, a basal conglomerate and an overlying sandstone. Because the lower contact is not exposed, the total thickness of the formation is not known. Approximately 30 feet of sandstone is exposed on Lucille Island. The formation appears to thicken to the west, with a maximum exposed thickness of nearly 200 feet being reached near the type locality in T. 64 N., R. 3 E.

The basal conglomerate is exposed at two localities. Thirty-six feet of conglomerate is exposed at the type locality in the NE¼ sec. 25, T. 64 N., R. 3 E., where the conglomerate (fig. V-71) was studied by Winchell (1897). The pebbles in the conglomerate are very well rounded, and average one and one-half inches in diameter. In order of decreasing abundance, they consist of white quartz, dark-gray quartz, red quartz and quartzite, jasper and iron-formation, and white chert. Fifteen feet of basal conglomerate exposed on Grand Portage Island contains flat chips and pebbles of argillite and slate, probably derived from the underlying Rove Formation, and rounded pebbles of white quartzite (Grant, 1894). Much of the sand-size matrix in this conglomerate has been replaced by carbonate; Nelson (1942, unpub. M.S. thesis, Univ. Minn.) reported a carbonate content of 48 percent.

The sandstone (fig. V-72) is thicker and much more extensively exposed than the basal conglomerate. Cross-bedding and rare ripple marks are present in the massive, thick-bedded, light-gray sandstone. It is fine grained and is typically composed of well sorted, subrounded to well rounded grains. Examination of thin sections indicates that the sandstone contains an average of 80 percent quartz (65 percent unit quartz, 15 percent polycrystalline quartz), 3 percent feldspar (orthoclase dominant), 1 percent granitic rock fragments, 7 percent carbonate cement, 5 percent silica cement, and 3 percent chlorite-sericite matrix. Zircon, apatite, epidote, and tourmaline are the common nonopaque heavy minerals.

Figure V-71. Basal conglomerate of Puckwunge Formation (NE¼ sec. 25, T. 64 N., R. 3 E., Cook County).

Figure V-72. Cross-bedded sandstone (light-colored) of Puckwunge Formation exposed beneath a basal flow (dark-colored) of the North Shore Volcanic Group (SE¼ sec. 8, T. 63 N., R. 6 E., southwest of village of Grand Portage).

Provenance and Sedimentation

The lithology of the conglomerate pebbles and the composition of the sandstone indicate that the Puckwunge Formation was derived from the pre-existing Middle and Lower Precambrian rocks of the region. Granite, quartzite, iron-formation, and slate appear to have been major rock types in the source area. Paleocurrent analysis of cross-beds, troughs, and ripple marks indicates a southwesterly direction of sediment transport, with the source area lying to the northwest (fig. V-73). The presence of cross-bedding and ripple marks in a well sorted quartz-rich sediment composed of well rounded grains suggests a shallow-water depositional environment. The Puckwunge Formation was

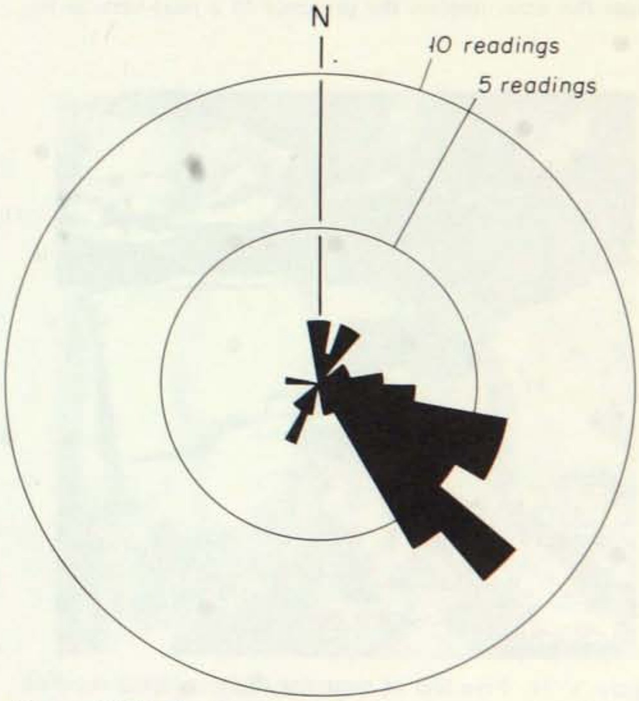

Figure V-73. Summary of paleocurrent data in the Puckwunge Formation of northeastern Minnesota (diagram based on 51 crossbeds, 4 troughs, 1 ripple mark).

probably deposited during the northward transgression of a sea into the region. The petrology, depositional environment, and regional stratigraphy suggest that the Puckwunge Formation is correlative with the nearby Lower Keweenawan Sibley Formation of Ontario.

THE SEDIMENTARY ROCKS OF NOPEMING, MINNESOTA

Stratigraphy and Petrology

At Nopeming, just west of Duluth, 25 feet of quartzite and quartz- and quartzite-pebble conglomerate is exposed beneath Keweenawan lava flows in secs. 17 and 20, T. 49 N., R. 15 W. (fig. V-74). These sedimentary rocks appear to unconformably overlie the Middle Precambrian Thomson Formation. The contact between these beds and the Thomson Formation is not exposed, but appears to be an angular unconformity. The nearest outcrop of the underlying Thomson Formation, 350 feet west of the quartzite, has an attitude of N. 85° E., 84° S., whereas the attitude of the quartzite is N. 10° W., 20° E.

Both the basal lava flows and the underlying sedimentary rocks dip gently eastward. Small-scale load structures in the upper 6 inches of the quartzite suggest that the sediments were probably unlithified at the time of lava extrusion. Pillow structures suggest that the basal lava flow may have been extruded into the same body of water in which the sediments were deposited.

In thin sections it can be seen that the lowermost exposed conglomerate and quartzite are more intensively recrystallized than the middle and uppermost quartzite. The recrystallization, as well as gravity and magnetic profiles across the area, implies the presence of a near-vertical intrusive dike beneath the quartzite and conglomerate (Mattis, 1972, unpub. M.S. thesis, Univ. Minn.).

Eight feet of conglomerate is interbedded with quartzite near the bottom of the exposed sedimentary rocks. The conglomerate is composed of 10 percent pebbles and 90 percent quartzite matrix. The pebbles are well rounded, and rarely exceed one and one-half inches in diameter. The majority of them are composed of white quartz, but a small number of white quartzite pebbles also are present.

The quartzite may be subdivided into two units. The upper unit is a light and dark banded, very fine-grained metasiltstone that ranges in thickness from 6 to 24 inches. This unit contains a parting lineation due to alignment of grains during deposition by a current. The lower unit consists of buff-colored, cross-bedded, medium- to coarse-grained quartzite, which appears to be identical to the conglomerate matrix. The grains in the quartzite are well sorted and well rounded. The quartzite is composed almost entirely of quartz, with unit quartz being dominant over polycrystalline quartz. The principal cement is silica, but carbonate also is present. Zircon is the most abundant nonopaque heavy mineral; apatite and tourmaline are present in minor amounts.

Provenance and Sedimentation

Because of the very mature character of the conglomerate and quartzite at Nopeming, petrography provides little information about the source area of the sediments. However, the quartz-rich composition of the sediments suggests that coarse-grained, felsic rocks (granite?) probably were present in the source area. Paleocurrent analysis of crossbedding and parting lineation indicates a north-northwesterly direction of sediment transport, with the source area lying to the southeast (fig. V-75). Igneous and metamorphic

Figure V-74. Five feet of quartzite (light-colored) exposed beneath basal flow (dark-colored) of the North Shore Volcanic Group (SW¼ sec. 17, T. 49 N., R. 15 W., near Nopeming, St. Louis County).

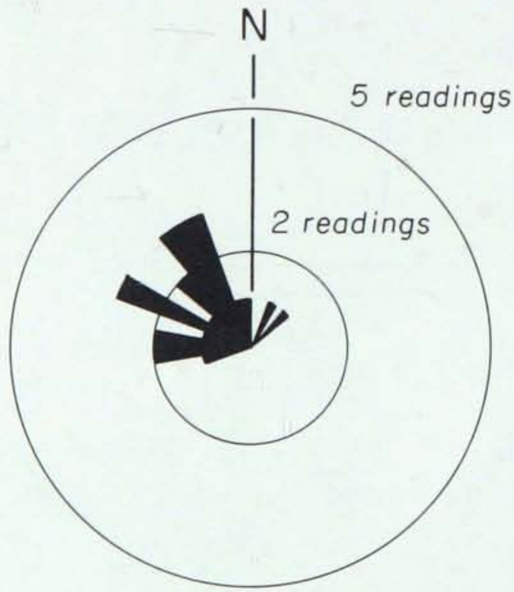

Figure V-75. Summary of paleocurrent data in the sandstones of Nopeming (diagram based on 14 crossbeds, 3 troughs, and 5 parting lineations).

crystalline rocks of Penokean age probably were the major rock types in the source area. The mature, cross-bedded sedimentary rocks appear to have been deposited in a shallow-water environment.

The age and correlation of the quartzose strata at Nopeming are in question. Small-scale load structures in the upper few inches of the quartzite suggest that the sediments may have been unlithified at the time of lava extrusion. In the Short Line Park well two and one-half miles south of Nopeming, quartzite and conglomerate appearing to be identical to the sediments at Nopeming were found to be interbedded with the lower 140 feet of lava flows (Winchell, 1889). The lowermost 84 feet of lava does not contain interbedded sedimentary rocks and lies directly on the Thomson Formation. From this evidence, it appears that the sediments at Nopeming cannot be correlated with the Puckwunge Formation of northeastern Minnesota, but instead are interflow sediments in the North Shore Volcanic Group.

KEWEENAWAN GEOLOGY OF EAST-CENTRAL AND SOUTHEASTERN MINNESOTA

Campbell Craddock

The Keweenawan rocks of Minnesota are part of a narrow province that extends from Kansas northeastward to Lake Superior and from there probably southeastward into Lower Michigan (Craddock, this chapter). The stratified rocks consist of a lower thin quartzite unit, a middle thick sequence of basaltic lava flows, and an upper thick sequence of detrital sedimentary rocks. These layered rocks define a regional syncline, and along much of the length of the province the axial zone is the site of a horst tens of miles wide. Numerous mafic igneous intrusive bodies cut the Keweenawan and older rocks throughout the region; such intrusions are abundant in northeastern Minnesota, in Wisconsin, and in Ontario.

The area described in this paper is south of the west end of Lake Superior, and includes the segment of the Keweenawan province which lies in east-central and southeastern Minnesota (fig. V-76). It is bounded on the south by the Iowa border, on the east by the Wisconsin border, and on the north and west by the contact between Upper Keweenawan sedimentary rocks and older Precambrian rocks. Keweenawan rocks crop out intermittently from Lake Superior southward to near Pine City (lat. 45°50′). Except for the volcanic rocks along the St. Croix River near Taylors Falls, Keweenawan rocks do not crop out in Minnesota south of Pine City. In this area they are covered by Paleozoic and younger deposits, and are known only from scattered wells and geophysical surveys.

IMPORTANCE OF THE AREA

The Keweenawan rocks in this part of Minnesota are of special interest for three reasons. First, the Midcontinent Gravity High, which is the most striking feature on the Bouguer gravity map of the United States (Woollard and Joesting, 1964), passes through southeastern Minnesota. Thiel (1956) and Craddock and others (1963) showed that this gravity feature is closely related to the geology of the Keweenawan province. The value of geophysical methods in deciphering the structure and composition of basement rocks is beautifully demonstrated here, where these Precambrian rocks plunge southward beneath a cover of younger strata.

Secondly, these rocks have been, and may again be, of considerable value as a mineral resource. Although none is active at present, several quarries have produced dimension stone from the Keweenawan sandstones. Traces of copper mineralization are common in outcrops of the Keweenawan volcanic rocks, and the future discovery of mineable copper is possible. The volcanic rocks in Minnesota are the southwest continuation of the lava sequence in the Keweenaw Peninsula of Michigan, which has produced copper for more than a century.

Finally, the Keweenawan sedimentary rocks are an important reservoir for fluids. In the Twin Cities area, the Hinckley Sandstone is a valuable aquifer, and many wells have been drilled into it in the search for adequate groundwater supplies. In addition, a pressing need exists in the metropolitan area for an underground reservoir for summer storage of natural gas, permitting deliveries to local utilities at a steady rate the year round. Because the Paleozoic formations are thin and close to the surface, the deeper Keweenawan sandstones offer a better chance for developing a storage facility.

PREVIOUS WORK

Although geologic observations of east-central and southeastern Minnesota occur in the narratives of explorers who visited the area as early as the seventeenth century, the first systematic work was a geologic survey of Wisconsin, Iowa, and Minnesota conducted in 1847-1850 for the U.S. Treasury Department by Dr. David Owen. Minnesota became a state in 1858, and in 1872 the legislature charged the University of Minnesota with conducting ". . . a thorough geological and natural history survey of the state." The initial survey took ten years, and the results were published in six volumes (Final Reports) between 1884 and 1901. Geologic work prior to this survey is admirably summarized in the first volume by N. H. Winchell, state geologist and survey director. In this survey all Keweenawan rocks were assigned to the Cambrian System.

Other early workers made important and enduring contributions to Keweenawan geology. Irving (1883) included the outcrops in Pine County in his study of copper-bearing rocks of the Lake Superior district. Berkey (1897-1898) made a detailed study of the geology of the Taylors Falls area, with emphasis on the Keweenawan volcanic rocks. Geologic maps and cross-sections of the Keweenawan area of east-central Minnesota were published by Hall (1901a), who showed the trace of the Douglas fault. Grout (1910a and b) studied the same area and described the petrography, chemical composition, and copper content of the rocks. Van Hise and Leith (1911) summarized the work of the U.S. Geological Survey on the Precambrian rocks of the Lake Superior region. Finally, Thwaites (1912) described the Keweenawan sandstones of Wisconsin and Minnesota and argued for their Precambrian age.

Subsequent reports on the surface and subsurface areal geology include those of Harder and Johnston (1918) on east-central Minnesota, Schwartz (1936) on the Twin Cities area, Stauffer and Thiel (1941) on the Keweenawan and Paleozoic sedimentary rocks of southeastern Minnesota, Thiel (1944, 1947) on southern and northeastern Minnesota, respectively, and Schwartz (1949) on the Duluth area.

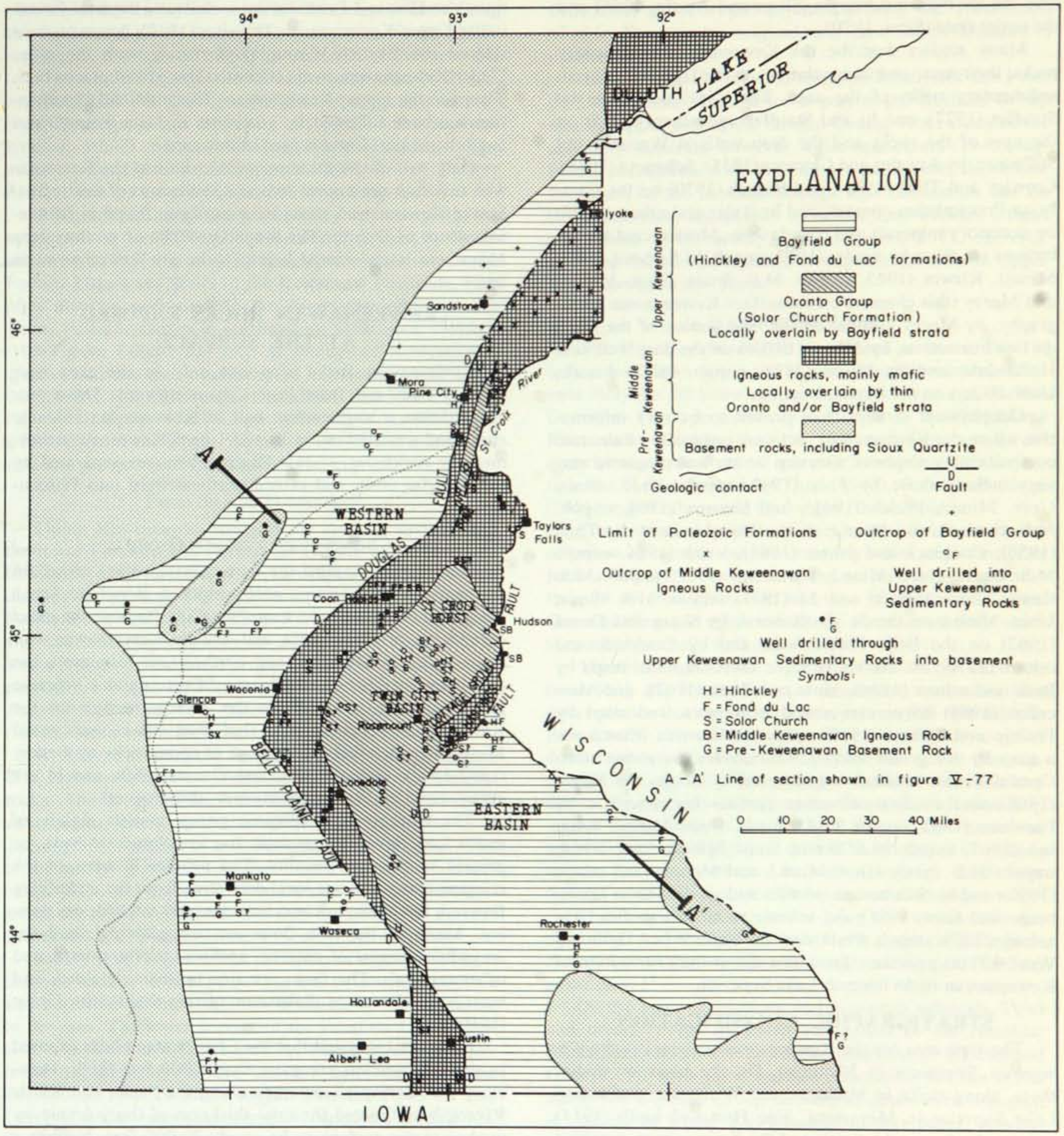

Figure V-76. Geologic map of Precambrian rocks, east-central and southeastern Minnesota. Compiled by Craddock, 1972, from various sources including Kirwin, 1963, unpub. M.S. thesis, Univ. Minn.; Craddock and others, 1963; Bath and others, 1964; Philbin and Gilbert, 1966; Sims and Zietz, 1967; Craddock and others, 1970; Mooney and others, 1970a and b; and Morey, this chapter.

Grout and others (1951) summarized the Precambrian stratigraphy of Minnesota, and Goldich and others (1961) discussed the Precambrian geology of the state in light of radiometric age determinations. Recent geologic maps portray the St. Paul quadrangle (Sloan and Austin, 1966) and the entire state (Sims, 1970).

Many studies describe the Keweenawan sedimentary rocks, their ages, and their relations to the Upper Cambrian sedimentary rocks of the area. These include papers by Stauffer (1927a and b) and Stauffer and others (1935) on the ages of the rocks and the deep wells at Waconia and Stillwater, by Atwater and Clement (1935), Schwartz (1935), Crowley and Thiel (1940), and Raasch (1950) on the Cambrian-Precambrian contact, and by Tyler and others (1940) on accessory minerals and correlations. More recent contributions are those by Snider (1962, unpub. M.S. thesis, Univ. Minn.), Kirwin (1963, unpub. M.S. thesis, Univ. Minn.), and Morey (this chapter) on subsurface Keweenawan stratigraphy, by Morey (1967a) on the type section of the Fond du Lac Formation, by Austin (1970b) on the deep well near Hollandale, and by Myers (1971, unpub. Ph.D. dissert., Univ. Wisc.) on the Bayfield Group.

Geophysical surveys have proved to be very informative where the Keweenawan rocks are covered by Paleozoic or Quaternary deposits. Gravity or surface magnetic surveys include those by Frey (1939, unpub. Ph.D. thesis, Univ. Minn.), Welch (1941), and Sharma (1964, unpub. M.S. thesis, Univ. Minn.) on the Douglas fault, by Thiel (1956), Craddock and others (1963), Veith (1966, unpub. M.S. thesis, Univ. Minn.), Barazangi (1967, unpub. M.S. thesis, Univ. Minn.), and Li (1971, unpub. M.S. thesis, Univ. Minn.) on the St. Croix horst, by Sloan and Danes (1962) on the Belle Plaine fault, and by Craddock and others (1970) on the entire state. Aeromagnetic maps by Bath and others (1964), Sims and Zietz (1967), and Marcellus (1968) cover east-central Minnesota, and maps by Philbin and Gilbert (1966) cover southeastern Minnesota; a map by King and Zietz (1971) covers the entire state. Crustal seismic studies were reported by Cohen and Meyer (1966), and shallow refraction profiles are described by Farnham (1967, unpub. Ph.D. thesis, Univ. Minn.), Johnson (1967, unpub. M.S. thesis, Univ. Minn.), Volz (1968, unpub. M.S. thesis, Univ. Minn.), and Mooney and others (1970a and b). Sub-bottom profiles and six drill holes (Zumberge and Gast, 1961) and seismic refraction studies (Anzoleaga, 1971, unpub. Ph.D. dissert., Univ. Wisc.; Halls and West, 1971a) provide information about the distribution of Keweenawan rocks beneath Lake Superior.

STRATIGRAPHIC CLASSIFICATION

The type area for the Keweenawan sequence is the Keweenaw Peninsula in Michigan. On the basis of studies there, along strike in Michigan and Wisconsin, and across Lake Superior in Minnesota, Van Hise and Leith (1911) suggested a three-fold division of the Keweenawan stratified rocks. The Lower Keweenawan consists of a few hundred feet of conglomeratic quartzite or sandstone, the Middle Keweenawan of tens of thousands of feet of basaltic lava flows, and the Upper Keweenawan of thousands of feet of detrital sedimentary rocks. At that time, the Bayfield Group (Lake Superior Sandstone) was considered Upper Cambrian and was thought to be laterally continuous with the Cambrian sandstones of the lower St. Croix valley. Although they expressed doubts about its Upper Cambrian correlation, Van Hise and Leith did not include the Bayfield Group in the Upper Keweenawan. Thwaites (1912), however, considered the Bayfield Group conformable with the older Upper Keweenawan strata (Oronto Group) and placed both groups in the Upper Keweenawan. This modified classification has been followed by geologists in Minnesota (Grout and others, 1951), and is used in this paper.

Only two of the three major divisions of the Keweenawan sequence are known to occur in the area of this report. Lower Keweenawan rocks have not been found in Minnesota south of Duluth. The Sioux Quartzite of southwestern Minnesota is considered herein to be pre-Keweenawan in age.

KEWEENAWAN ROCKS EXPOSED AT THE SURFACE

Keweenawan rocks crop out only in the area near Taylors Falls and from Pine City northward. These outcrops define a southeastern belt of Keweenawan volcanic rocks and a northwestern belt of Upper Keweenawan sedimentary rocks (fig. V-76). The two belts are separated by the Douglas fault, and extend northeastward into Wisconsin.

Chengwatana Volcanic Group

Hall (1901a) applied the term "Chengwatana series" to the sequence of volcanic rocks exposed along the Snake River east of Cross Lake near Pine City. Because the exact correlation of these rocks and Middle Keweenawan volcanic rocks in Michigan and northeastern Minnesota has not been established, the name "Chengwatana volcanic group" is used informally in this volume to describe the Keweenawan volcanic rocks that crop out in east-central Minnesota. Detailed descriptions of these rocks by Berkey (1897-1898), Hall (1901a), and Grout (1910a and b) are summarized by Morey and Mudrey (this chapter).

The Chengwatana volcanic group consists mainly of mafic lava flows, but at least five conglomerate beds are present in the type locality. The pebbles in most of the conglomerates indicate local derivation from the underlying flows. A few tuffs, volcanic breccias, and ash beds are present. Many of the lava flows are amygdaloidal, and the amygdules consist of chlorite, zeolites, calcite, quartz, and other minerals. The flows are fine to coarse grained, and most were classed as diabase or olivine diabase by Grout (1910b).

Hall (1901a) stated that the Chengwatana beds exposed in the type section are more than 4,000 feet thick. However, considering all outcrops along the Snake and Kettle Rivers, he estimated the total thickness of the volcanic sequence at the surface to be nearly 20,000 feet. Neither a lower nor an upper contact can be seen. The lowest beds exposed are along the Douglas fault, but the actual faulted contact is not exposed in Minnesota. The upper contact with Upper Keweenawan sedimentary rocks is preserved just across the St. Croix River in Wisconsin, but it does not appear to extend into Minnesota. The Keweenawan volcanic rocks are overlain by flat-lying Upper Cambrian sandstones at several localities along the St. Croix River.

Bedding in the volcanic rocks appears to define a complex asymmetrical syncline. Near the Douglas fault along the Snake and Kettle Rivers, the beds strike N. 10-20° E. and dip as much as 67° E. Dips diminish to the east, and near the mouth of the Kettle River the beds strike N. 20° E. and dip 10-20° W. This syncline is interpreted as a second-order flexure and not the main axis of the Lake Superior syncline, which must lie some miles to the southeast. The main synclinal axis probably passes just west of Taylors Falls, where the flows strike N. 20° W. and dip 15° W.

Because the Chengwatana volcanic group comprises the upper part of the "Middle Keweenawan," as used by Grout and others (1951), it may be roughly correlative with the Portage Lake Lava Series of the Keweenaw Peninsula. Part of it may be equivalent to the upper part of the North Shore Volcanic Group in northeastern Minnesota. Goldich (1968) gave 1,000-1,200 m.y. as the age of Keweenawan igneous activity, and most reported radiometric ages fall in the 1,040-1,115 m.y. range. Chaudhuri and Faure (1967) reported an age of $1{,}075 \pm 50$ m.y. for the Nonesuch Shale, a formation in the Upper Keweenawan. The Chengwatana volcanic group is probably about 1,100 m.y. old.

Fond du Lac Formation

Upper Keweenawan strata in the area are divided into the lower Fond du Lac Formation and the upper Hinckley Sandstone; both formations are considered correlative with the Bayfield Group. The type section of the Fond du Lac Formation is along the St. Louis River just west of Duluth, and scattered outcrops may be traced as far southwest as Mora. The Fond du Lac Formation comprises the western part of the Upper Keweenawan outcrop belt in the area. This formation has been described by Stauffer and Thiel (1941), Grout and others (1951), and Morey (1967a).

At its type section, the Fond du Lac Formation consists of a basal conglomerate overlain by finer grained detrital rocks. The conglomerate contains pebbles and cobbles of vein quartz, chert, quartzite, graywacke, and slate in a coarse-grained matrix of angular quartz and feldspar. Most of the overlying beds are reddish-brown sandstone, but siltstone, shale, and mudstone units are also present. The sandstones are arkosic or subarkosic; they contain 36-68 percent quartz and 5-29 percent feldspar (Morey, this chapter).

Morey (1967a) placed the thickness of exposed strata at the type locality at about 300 feet, but he considered this a small fraction of the maximum thickness of the formation in the area. The lower contact of the Fond du Lac Formation is exposed at the type locality, where it rests with definite unconformity on the strongly deformed Thomson Formation. Map relations suggest that the Fond du Lac Formation also rests unconformably upon the Middle Keweenawan volcanic rocks exposed just southwest of Duluth, a mile north of the St. Louis River, but the actual contact cannot be observed and has not been penetrated by drilling. The Fond du Lac Formation is overlain by the Hinckley Sandstone with structural concordance, but the contact cannot be seen in outcrop anywhere in the area.

The Fond du Lac Formation has a homoclinal structure, and there is no direct evidence that the beds have undergone any folding. In the outcrop area, the strike is northeastward and the dip 3-12° SE. In the type locality, the beds strike about N. 60° E. and dip 5-12° SE. How much of this dip may be original and how much tectonic is unknown.

Thwaites (1912) mapped the type locality of the Fond du Lac Formation as the lower part of the Bayfield Group and the upper part of the Oronto Group. After study of the accessory minerals, however, Tyler and others (1940) suggested correlation with the Orienta Sandstone, the lowest formation of the Bayfield Group. This correlation has received tentative acceptance by subsequent workers (Grout and others, 1951; Morey, 1967a; Myers, 1971, *op. cit.*). Both the Oronto and Bayfield Groups are considered Upper Keweenawan in this paper.

Hinckley Sandstone

The Hinckley Sandstone is exposed at the surface in only a few places along rivers from Hinckley northward to near Holyoke; all known exposures are in Pine and Carlton Counties. The type section is just north of the Grindstone River near Hinckley, but the best exposures are in old quarries along the Kettle River near Sandstone. This formation has been described by Thiel (1947), Grout and others (1951), and Tryhorn and Ojakangas (this chapter).

The formation consists of thin-bedded to massive units of yellow to salmon and red or nearly white sandstone. The rock is well sorted and mature, and on the average has more than 98 percent quartz and less than one percent feldspar. Cross-bedding and ripple marks are common, and the sandstones are generally well indurated.

The thickest sequence exposed is the 101-foot-thick section along the Kettle River at Sandstone, but subsurface data indicate that the formation is at least 500 feet thick in Pine County. The Hinckley Sandstone is believed to grade downward into the Fond du Lac Formation, but this contact is not exposed. The youngest beds of the Hinckley Sandstone probably are along the Douglas fault, but they cannot be observed in contact with the volcanic rocks and are overlain only by glacial drift. The Hinckley Sandstone appears to pass southward beneath Cambrian sandstones, but this relationship cannot be demonstrated in any outcrop.

The formation is nearly flat-lying and tectonically undisturbed. At Hinckley the bedding is approximately horizontal, but to the northeast it is inclined southeastward by as much as 5°.

On the basis of lithology and accessory minerals, Tyler and others (1940) suggested correlation of the Hinckley Sandstone with the Devils Island Sandstone of the Bayfield Group. This correlation appears reasonable, but the actual age of the Bayfield Group remains in doubt even though it is considered Upper Keweenawan in this paper. Atwater and Clement (1935) emphasized the conformable character of the entire Keweenawan sequence and the importance of the Keweenawan-Upper Cambrian unconformity, and concluded that the Bayfield Group is Precambrian. Raasch (1950), however, argued that the Bayfield Group may be Lower or Middle Cambrian.

KEWEENAWAN ROCKS IN THE SUBSURFACE

Keweenawan rocks do not crop out south of the exposures near Mora, Pine City, and Taylors Falls. Where the Keweenawan rocks are covered by Paleozoic, Cretaceous, and Quaternary deposits, their distribution has been mapped by gravity, magnetic, and seismic surveys (fig. V-76). A few score deep wells provide control points for the geologic map and some information on the composition of the Keweenawan rocks.

Chengwatana Volcanic Group

These rocks have been traced by their gravity and magnetic anomalies southward across east-central and southeastern Minnesota into Iowa. These mafic igneous rocks have been encountered in 11 wells, and data are given by Kirwin (1963, *op. cit.*), Sims and Zietz (1967), and Morey and Mudrey (this chapter). All these wells are located on the St. Croix horst, but the volcanic rocks also occur at greater depths in basins to the northwest and southeast (Mooney and others, 1970a). The wells at Hudson, Wisconsin, and Vermillion, Minnesota penetrated 446 and 526 feet of volcanic rocks respectively, but no well was drilled entirely through them into older rocks. Estimates of the maximum thickness of these volcanic rocks beneath the St. Croix horst, based on geophysical surveys, are 18,400 feet (Craddock and others, 1963), 25,000 feet (Sims and Zietz, 1967), and 33,000 feet (Veith, 1966, *op. cit.*).

In the subsurface, both on the St. Croix horst and in the adjacent basins, the Chengwatana volcanic group mainly is overlain, probably conformably, by the Solor Church Formation. Where the latter is absent, the volcanics commonly have a thin weathered zone at the top, and are overlain by Upper Cambrian strata. Locally, the volcanics may be overlain by the Fond du Lac Formation or the Hinckley Sandstone, but this cannot be demonstrated in present well records.

Solor Church Formation

For many years all the reddish sedimentary rocks encountered in the wells of the area beneath the Upper Cambrian and Hinckley strata have been assigned to the Red Clastic Series or the Fond du Lac Formation (Grout and others, 1951). Kirwin (1963, *op. cit.*) studied all available well samples and records and divided these red beds into four units. Morey (this chapter) reports information from more recent wells and defines the new Solor Church Formation; it is equivalent to Kirwin's lower three units. It is best known from wells on the St. Croix horst, but a few wells and seismic refraction lines suggest it occurs in the flanking basins as well.

The formation consists chiefly of immature, poorly sorted sandstone, siltstone, and mudstone; beds of sandy and silty limestone also occur and some are oolitic. Most of the framework grains in the detrital rocks are quartz, but feldspar and basaltic rock fragments together make up 10-60 percent of the grains in samples from the type well near Lonsdale.

The type well near Lonsdale penetrates 1,930 feet of Solor Church Formation without reaching older rocks. On the basis of magnetic anomalies (Sims and Zietz, 1967) and the thickness of strata penetrated in the Lakewood Cemetery well in Minneapolis, the formation may be 3,500 feet thick beneath southwest Minneapolis. The Solor Church Formation probably conformably overlies the underlying volcanic rocks, but evidence for this is slight and inconclusive. A few shallow wells encounter thin red bed successions resting upon weathered basalt, but these red beds may well be younger than the true Solor Church Formation even though they show a strong lithologic resemblance. On the St. Croix horst, the Solor Church Formation is overlain unconformably by Upper Cambrian strata, by the Hinckley Sandstone, and possibly by the Fond du Lac Formation. In the flanking basins, the Solor Church Formation appears to be overlain by the Fond du Lac Formation, but this contact has not been intersected by drilling.

Seismic refraction studies suggest that the Solor Church Formation may be concordant with the overlying Fond du Lac Formation in the flanking basins, but the evidence is not compelling. On the St. Croix horst, strike and dip information on the Solor Church Formation is lacking, but geophysical surveys indicate that it occupies an asymmetrical basin on the axis of the Lake Superior syncline in the Twin Cities area. This basin may be considered the southwestward-plunging counterpart to the Upper Keweenawan strata preserved in the northeastward-plunging Ashland-Gordon syncline in Wisconsin. The lower sedimentary rocks in the North Branch graben probably consist of Solor Church Formation.

Kirwin (1963, *op. cit.*) correlated the two lower units he defined in the basin with the Freda Sandstone of the Oronto Group. Morey (this chapter) assigns the Solor Church Formation to the Upper Keweenawan, and tentatively correlates it with the Oronto Group.

Fond du Lac Formation

In most previously published papers, all Precambrian red beds in the area have been assigned to the Fond du Lac Formation, but this formation will be used here in the restricted sense suggested by Kirwin (1963, *op. cit.*) and defined by Morey (this chapter). Wells and geophysical surveys show that the Fond du Lac Formation comprises an important part of the basin west of the St. Croix horst. Strata similar in composition and position in the eastern basin are assigned to the Fond du Lac Formation, but these rocks are not known to crop out in either Minnesota or Wisconsin. The Fond du Lac Formation is generally absent on the St. Croix horst, but a few wells near the western border contain beds that may represent this formation.

The Fond du Lac consists of poorly sorted reddish sandstone and minor shale; the rocks are very similar to the type section near Duluth. The sandstones are classified as arkose and feldspathic sandstone, and feldspar content averages about 25 percent. Rock fragments are rare except locally near the faults bounding the St. Croix horst.

The thickest Fond du Lac section recognized by Kirwin (1963, *op. cit.*) is 912 feet in a well at Mankato; the underlying 377 feet to the bottom would now be assigned to the Solor Church Formation. A well at Rochester penetrated 2,033 feet of red beds between the Hinckley Sandstone and basement granite; well cuttings are not available, but at least the upper part of this sequence probably is Fond du Lac. Gravity studies (Craddock and others, 1963; Li, 1971, *op. cit.*) suggest a maximum thickness of about 11,000 feet for Upper Keweenawan sedimentary rocks in the western basin, and seismic profiles (Mooney and others, 1970b) indicate that the Fond du Lac comprises more than 5,000 feet of the total thickness.

Seismic studies suggest that the Fond du Lac overlies the Solor Church in the deeper parts of the flanking basins, but no wells penetrate that contact. In the outer parts of the basins, it lies on Sioux Quartzite and older Precambrian rocks. Near Glencoe, it may lie on Middle Keweenawan volcanics as it appears to do west of Duluth. Several wells in the western part of the St. Croix horst encounter beds of probable Fond du Lac; at Coon Rapids, these strata overlie 60 feet of conglomerate that may be either Solor Church or possibly a unit interbedded in the Chengwatana volcanic group.

The Fond du Lac Formation is overlain by the Hinckley Sandstone, Upper Cambrian strata, or Quaternary deposits. Grout and others (1951) considered the Hinckley-Fond du Lac contact gradational.

Direct evidence on the structure within the flanking basins is lacking. The few refraction profiles and a comparison with the outcrop area suggest that the structure of the Fond du Lac Formation is essentially homoclinal with gentle dips toward the St. Croix horst. Flexures and minor faults may exist within the basins, especially near the margins of the horst.

The Fond du Lac Formation, as used here in reference to rocks in the subsurface, is considered approximately equivalent to the surface formation and Late Keweenawan in age. The lower part of the Fond du Lac Formation in the subsurface may contain beds that do not crop out at the surface.

Hinckley Sandstone

Atwater and Clement (1935) questioned the existence of Hinckley Sandstone in the subsurface of southeastern Minnesota and placed these beds in the overlying Mt. Simon Sandstone, but Schwartz (1935) defended the validity of the formation beneath the Twin City basin. Crowley and Thiel (1940) separated the Hinckley and the Mount Simon on the basis of the percentage of feldspar and argued for the presence of both formations in a number of wells throughout the area. Kirwin (1963, *op. cit.*) accepted the existence of the Hinckley in the subsurface, but emphasized the need for more study of the problem; his interpretations are followed here.

In the subsurface, the Hinckley is a pale-brown to light-gray, fine- to medium-grained, well sorted quartzose sandstone characterized by frosted grains and abundant quartz overgrowths. On the average, the feldspar content is only 0.12-0.31 percent (Crowley and Thiel, 1940), but some higher feldspar contents have been found recently (Tryhorn and Ojakangas, this chapter).

The formation is absent in many wells, but it reaches thicknesses of 470 feet at Pine City, 99 feet at Rosemount, and 106 feet at Rochester. In the flanking basins, the Hinckley Sandstone is considered to lie conformably on the Fond du Lac Formation. Morey (this chapter) emphasizes the important unconformity at the base of the Hinckley where it overlies the Solor Church on the St. Croix horst. The contact between the Hinckley Sandstone and the overlying Mt. Simon Sandstone has been interpreted as an unconformity by most workers, but evidence is indirect. The strong similarity of the two formations leaves their relationship in doubt.

The Hinckley Sandstone is nearly flat-lying and appears to be structurally concordant with the Fond du Lac Formation below and the Mt. Simon Sandstone above.

In the western basin, the Hinckley is probably the subsurface continuation of the Upper Keweenawan Hinckley Sandstone that crops out to the north. The stratigraphic position of the Hinckley beneath the Twin City basin and in the eastern basin remains in doubt. These strata may correlate with the type Hinckley Sandstone, or they may be the locally developed basal part of the Upper Cambrian sequence.

GEOLOGIC STRUCTURE

Knowledge of the structure of the Keweenawan rocks in east-central and southeastern Minnesota has advanced greatly during the past twenty years as a result of geophysical surveys and new deep wells. Principal structural features are shown on the geologic map (fig. V-76) and the section (fig. V-77). Important facts and inferences about these features are summarized in this section.

St. Croix Horst

The St. Croix horst is bounded by the Douglas and Hastings faults, and can be traced from the Belle Plaine fault northeastward at least to Bayfield County, Wisconsin. The lava flows within the horst define a broad synclinal flexure, the continuation of the Lake Superior syncline. The Solor Church Formation in the horst occupies this flexure, and the Hollandale embayment and the subsidiary Twin City basin in the overlying Paleozoic strata are also centered on this axis. Upper Keweenawan sedimentary rocks also occur in a narrow belt near North Branch; this structure is probably a graben or half-graben, but the dip reversal in the volcanic rocks along the Kettle River suggests that it may be a synclinal flexure on the limb of the main syncline (Sharma, 1964, *op. cit.*).

The horst is bounded by dip-slip faults, but their displacement is hard to establish. The Douglas fault dips about 45° SE. in Douglas County, Wisconsin, but is not exposed in Minnesota. Gravity gradients suggest that the fault is steep, and the surface dips of the fault may have been modified by gravitational bending of an originally steeper interface. Because volcanic rocks are at or near the surface east of the trace, the throw can be approximated by finding the depth to the volcanics west of the fault. Estimates of the throw at places along the Douglas fault include 11,000 feet (Welch, 1941; Craddock and others, 1963), 8,200 feet (Veith, 1966, *op. cit.*), 10,000 feet (Mooney and others, 1970b), and 11,500 feet (Li, 1971, *op. cit.*). Throw on the Hastings or Lake Owen fault has been calculated at 10,000 feet (Craddock and others, 1963), 9,000 feet (Veith, 1966, *op. cit.*; Barazangi, 1967, *op. cit.*; and Li, 1971, *op. cit.*), and 8,000 feet (Sims and Zietz, 1967). The Hastings fault probably continues southward from the end of the Hudson-Afton horst, with reduced throw, and its trace has been estimated from aeromagnetic maps. Estimates of the throw of the Cottage Grove fault range from zero (Veith, 1966, *op. cit.*) to 800 feet (Barazangi, 1967, *op. cit.*) to 2,000 feet (Sims and Zietz, 1967).

Flanking Basins

The figures cited above also are estimates of the total sedimentary section in the deepest parts of the flanking basins. These basins are markedly asymmetrical, with a steeply-inclined inner fault boundary and a gently-inclined outer depositional boundary. The western basin is known in the outcrop area and has been traced by geophysical surveys southward into Iowa. Seismic refraction profiles suggest that this basin contains Hinckley Sandstone, Fond du Lac Formation, Solor Church Formation, and rocks of the Chengwatana volcanic group. The eastern basin was unknown from surface outcrops and was discovered by geophysical surveys. The overlying Paleozoic strata define the gentle River Falls syncline, with an axis parallel to that of the basin. No deep wells penetrate the deeper parts of the eastern basin in Wisconsin, so its rocks are unknown; no samples exist from the deep well at Rochester. Seismic profiles can only suggest the presence of Fond du Lac, Solor Church, and Chengwatana volcanic rocks.

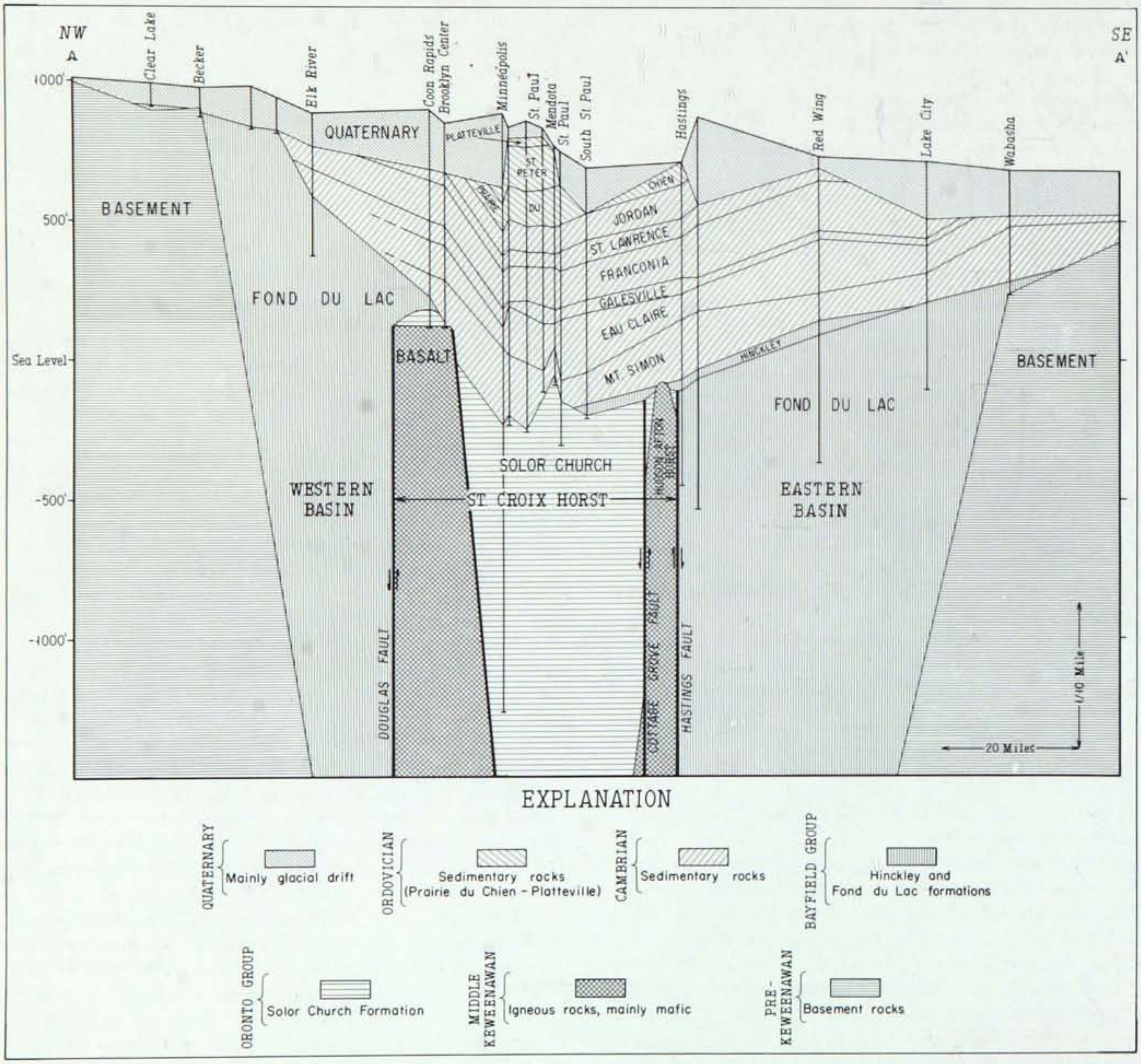

Figure V-77. Geologic section showing structure of Keweenawan province (modified from Kirwin, 1963, unpub. M.S. thesis, Univ. Minn. and Morey, in press).

Belle Plaine Fault

Sloan and Danes (1962) established the existence of the Belle Plaine fault from geologic mapping and gravity surveys. The fault cuts Lower Ordovician rocks at the surface and has a throw of 700 to 1,000 feet. The steep gravity gradient across the fault, however, indicates that the throw in the Precambrian rocks is much greater. Morey (this chapter) reports that the wells at Waseca penetrate conglomerates in the Fond du Lac Formation, with clasts that appear to be derived from Chengwatana volcanics and the Solor Church Formation; a suitable source for these pebbles must have existed nearby. Thus, the major movement on the fault probably was accomplished by Fond du Lac time, but significant movement postdates the Early Ordovician.

The mechanical classification of the fault is unknown, and three interpretations are possible. First, the fault may be an ancient, pre-Keweenawan fracture which acted as a transform fault during Middle Keweenawan time and perhaps as a dip-slip fault later. This model explains the apparent offset of the transcontinental rift associated with Keweenawan volcanism (fig. V-76). Secondly, the fault may be simply a post-Middle Keweenawan strike-slip fault that offsets Middle Keweenawan and older rocks by tens of miles. This model explains the bending of structural trends in the St. Croix horst near the Belle Plaine fault as a drag phenomenon. Finally, the fault may be a steeply-dipping, dip-slip surface similar to the Douglas and Hastings faults. The lateral displacement suggested by the geologic map (fig. V-76) may be only apparent. In this model, the fault represents a boundary between crustal blocks whose relative displacements were essentially vertical.

Structure Near Iowa Border

The Midcontinent Gravity High is developed very clearly between Austin and Albert Lea near the Iowa border (Craddock and others, 1970). Because no subsurface information was available at the time of my study, I (Craddock and others, 1963) attempted to account for the gravity anomaly by a crustal root of mafic rocks rather than a horst of volcanic rocks. Subsequent aeromagnetic profiles (Philbin and Gilbert, 1966), seismic refraction work (Mooney and others, 1970a and b) and strata penetrated by a deep well (Austin, 1970b) indicate that a horst also is present in the subsurface at this latitude.

GEOLOGIC HISTORY

The Keweenawan rocks of the area accumulated on a continental surface of low to moderate relief underlain by a complex succession of Middle and Lower Precambrian igneous and metamorphic rocks. The Barron Quartzite of northwestern Wisconsin and the Sioux Quartzite of southwestern Minnesota are classified as Upper Precambrian; although their exact stratigraphic position remains in doubt, they are considered here as pre-Keweenawan formations. Conglomeratic quartzites assigned to the Lower Keweenawan occur in Michigan, Wisconsin, and northeastern Minnesota, but these rocks have not been found in Minnesota south of Duluth.

The Keweenawan history of east-central and southeastern Minnesota began with the eruption of great volumes of basaltic lava during Middle Keweenawan time. This area was a segment of a transcontinental rift which seems to have been the source of the lava flows. The older flows are not exposed at the surface, but geophysical surveys suggest that the volcanic pile is more than 30,000 feet thick beneath the St. Croix horst. Subsidence was probably concurrent with these fissure eruptions, so that the Lake Superior syncline had its inception in Middle Keweenawan time.

Subsidence continued during the Late Keweenawan, and thousands of feet of detrital sedimentary rocks of the Oronto Group were deposited in the axial zone of the Lake Superior syncline. In this area, these beds are assigned to the Solor Church Formation, a widely distributed unit known only from the subsurface. Initially, much of the Solor Church was derived from Middle Keweenawan volcanic rocks along the edges of the Lake Superior syncline, but later the pre-Keweenawan basement complex became a more important source of detritus.

Except for its overall synclinal form, information on the structure of the Solor Church Formation is lacking. However, to the northeast in Wisconsin and Michigan, rocks of the Oronto Group are folded and cut by reverse faults. These structures, together with the nearly vertical dip of the bedding in the southeastern limb of the Lake Superior syncline, suggest that a compressive phase of deformation affected the Oronto and older rocks. As rocks of the Bayfield Group are not affected by this deformation, the compressive phase can be dated as post-Oronto but pre-Bayfield.

During or after the compressive deformation, the axial horsts began to rise relative to the adjacent crustal blocks. For the St. Croix horst this displacement probably was accomplished by actual uplift of the horst and simultaneous subsidence of the flanking basins; total relative uplift is about 10,000 feet. Material eroded from the rising horst, along with detritus washed in from the older Precambrian complex outside the Lake Superior syncline, helped fill the developing flanking basins. Uplift of the horst was slow and incremental, and for millions of years the boundary faults were the loci of low fault-line scarps. As uplift of the horst diminished, the upper beds of the Bayfield Group were deposited across the faults and on the horst, only to be locally flexed and offset by renewed fault movement and largely eroded from the horst. If patches of true Fond du Lac and Hinckley exist on the horst, their presence suggests that uplift of the horst in those localities since late Bayfield time has been modest. Deposition of the Hinckley Sandstone probably was completed long before the beginning of Paleozoic time, but substantiating evidence is both indirect and inconclusive.

The Late Cambrian marine transgression ended a long period of emergence and weathering of the basement rocks and was followed by widespread deposition of Paleozoic sedimentary rocks on the Keweenawan rocks. The Twin City basin is defined by Paleozoic strata and overlies the Lake Superior syncline as preserved in the St. Croix horst (fig. V-77). This basin is probably a primary depositional feature, but it may imply limited subsidence since the Middle Ordovician. Faults in the Paleozoic strata above the main Keweenawan boundary faults have throws of hundreds of feet and show that movements of the horst took place after the Early Ordovician (Morey and Rensink, 1969). Similar faults in southwestern Iowa indicate uplift of the horst after Late Pennsylvanian time.

POSSIBILITIES FOR STORAGE OF NATURAL GAS

The discovery and development of large geologic structures suitable for summer storage of natural gas would bring substantial benefits to consumers in the Twin Cities area. Such storage reservoirs markedly increase the efficiency of the field-to-customer gas delivery system, and their importance rises as potential shortages place upward pressure on natural gas prices. Consideration of the structure and stratigraphy of the area suggests some possibilities for the discovery of suitable structures for gas storage.

The Paleozoic strata contain formations with satisfactory porosity and permeability, but in the northern part of the area they are either absent or too thin for gas storage. These rocks thicken southward, however, and some possibilities exist south of the Twin Cities. Favorable structures may be found near the margins of the St. Croix horst, just inside the boundary faults. An attractive prospect is the probable arching of the thick Paleozoic section above the narrow horst near the Iowa border.

Prospects for developing a storage facility in the Solor Church Formation are limited. The formation appears to have an overall synclinal structure, although smaller scale flexures and fault structures may exist. Compositionally, the formation is an immature sedimentary sequence that has relatively unfavorable porosity and permeability. The Solor Church Formation is most fully preserved beneath the metropolitan Twin Cities area, where it may be impractical to develop a storage reservoir. Some possibilities exist further south on the St. Croix horst if a satisfactory structure can be located.

The deep basins flanking the St. Croix horst have great potential and deserve careful exploration. The sedimentary sequence is thousands of feet thick, but these rocks are poorly known because of limited drilling and the scarcity of well cuttings. However, sedimentary formations suitable for gas storage and for capping a reservoir are almost certainly present. The regional structure in the basins appears gently homoclinal, but local flexures and fault structures are likely to exist. Upward bending of these strata adjacent to the major faults may have produced some suitable structural traps.

ACKNOWLEDGMENTS

Thanks are extended to the National Science Foundation, the Minnesota Geological Survey, and the Graduate Schools of the Universities of Minnesota and Wisconsin for financial support provided to my colleagues and myself for our geologic and geophysical studies of the Keweenawan province. I appreciate the help given by Sharon Cook, Frank Komatar, and Roger Cooper in the preparation of this paper.

KEWEENAWAN VOLCANIC ROCKS IN EAST-CENTRAL MINNESOTA

G. B. Morey and M. G. Mudrey, Jr.

Upper Precambrian (Keweenawan) lavas that contain small amounts of native copper and other associated copper-bearing minerals underlie about 700 square miles in Pine, Carlton, and Chisago Counties in east-central Minnesota (fig. V-78). This area borders the State of Wisconsin and lies within the drainage system of the St. Croix River and its several tributaries. In this part of east-central Minnesota, the bedrock is covered by a relatively thick mantle of Pleistocene materials. The topography generally is flat, and the area is poorly drained except near wide but shallow valleys that formed in both pre- and post-Pleistocene time. Consequently, known exposures are limited to stream valleys and a few artificial exposures.

STRATIGRAPHIC SETTING

In early reports on the geology of the Lake Superior region (for example Van Hise and Leith, 1911) it was assumed that the lava flows, associated interflow sediments, and most of the gabbroic intrusions represented a more or less short-lived igneous event, which was defined as com-

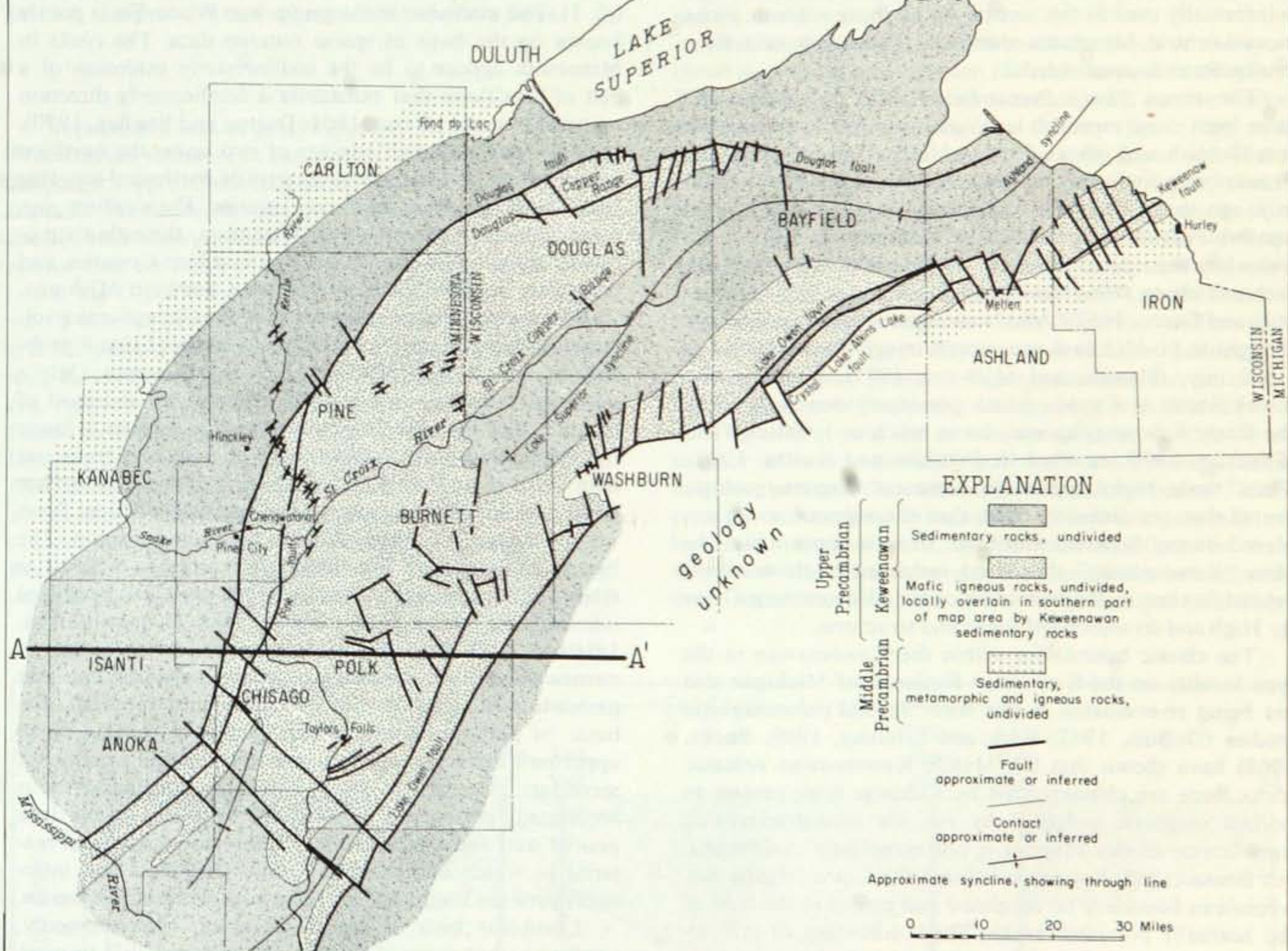

Figure V-78. Generalized pre-Paleozoic bedrock map of east-central Minnesota and adjoining northwestern Wisconsin showing inferred distribution and structure of Keweenawan flows and clastic strata. Compiled by G. B. Morey (1972) from unpublished data for Carlton, Pine, Kanabec and Isanti Counties, from Sims and Zietz (1967) for Anoka and Chisago Counties, and from Dutton and Bradley (1970) for Wisconsin.

prising the middle part of Keweenawan time. This igneous event was thought to have been preceded by and followed by deposition of dominantly clastic sediments. Hence, the Keweenawan System was divided into three units, Lower, Middle, and Upper. The assumption that all Keweenawan events occurred within a relatively short span of geologic time led to the conclusion that all the volcanic rocks were more or less contemporaneous. Therefore, the lava flows exposed in east-central Minnesota were presumed to correlate with similar-appearing rocks cropping out on the north shore of Lake Superior (see Green, this chapter). However, because the lava flows in east-central Minnesota are separated from those on the north shore of Lake Superior by a major structural discontinuity, the Douglas fault (pl. 1 and fig. V-78), direct evidence for such a correlation is lacking. Accordingly, as early as 1901, Hall referred to the lava flows along the Snake River in Pine County as the "Chengwatana series," a name taken from a small village east of Pine City that now is a historic site. Because correlation still is equivocal, the name "Chengwatana volcanic group" is informally used in this section for all those volcanic rocks in east-central Minnesota shown on Plate 1 as unit P€9, "volcanic rocks, undivided."

The terms "Late Precambrian" and "Keweenawan" have been used more or less synonymously in Minnesota (see Goldich and others, 1961, table 2). However, the Late Precambrian includes that period of time from about 1,600 m.y. ago to the time of Croixan deposition about 600 m.y. ago, whereas most of the Middle Keweenawan igneous activity has been dated at about 1,100 to 900 m.y. ago (Goldich and others, 1961; Robertson and Fahrig, 1971; Chaudhuri and Faure, 1967). Also, some igneous rocks previously thought to be Middle Keweenawan in age may be as old as 1,300 m.y. (Hanson and Malhotra, 1971). Similarly, the Sibley Series in Canada which previously was assigned to the Early Keweenawan may be as much as 1,400 m.y. old (Franklin, 1970, as cited in Franklin and Kustra, 1970). Thus, these rocks appear to represent discrete geologic events that are distinctly older than the volcanic rocks emplaced during Keweenawan time. If so, it appears that the term "Keweenawan" should be restricted to those events related in time to the formation of the Midcontinent Gravity High and its associated rocks and structures.

The classic boundaries within the Keweenawan in the type locality on the Keweenaw Peninsula of Michigan also are being re-evaluated at this time. Several paleomagnetic studies (DuBois, 1962; Beck and Lindsley, 1969; Books, 1968) have shown that the Middle Keweenawan volcanic rocks there are characterized by a change from reverse to normal magnetic polarity. As yet, the time-stratigraphic significance of this reversal is not completely understood, but Books (1968) has suggested that the Lower-Middle Keweenawan boundary be redefined and placed at the base of the normally polarized rocks. Thus, according to this definition, the Lower Keweenawan would contain a substantial thickness of igneous rocks which differ from the overlying Middle Keweenawan lavas in lithology, metamorphic grade, and magnetic properties (White and others, 1971). Similarly, White (1972) has suggested that the base of the Upper Keweenawan in Michigan be placed at the top of the Copper Harbor Conglomerate. Thus, the Middle Keweenawan would contain a substantial thickness of sedimentary strata.

Clearly, no Upper Precambrian volcanic, igneous, or sedimentary unit in the Lake Superior region can be assigned *a priori* to any part of the Keweenawan without additional substantive chronologic, paleomagnetic, or lithologic evidence. Such evidence from the Chengwatana volcanic group is lacking, but because the rocks appear to be continuous with those on the Keweenaw Peninsula in Michigan, the group is assigned to the less definitive Keweenawan, undivided.

GENERAL GEOLOGY

Available geologic and geophysical data indicate that the northwestern boundary of the Chengwatana volcanic group is marked by a substantial structural discontinuity, named the Douglas fault (fig. V-78). Upper Cambrian sedimentary rocks overlie the flows to the south and southeast (pl. 1). The extension of the group into Wisconsin is poorly known on the basis of sparse outcrop data. The rocks in Minnesota appear to be the southwesterly extension of a belt of lava flows that extends in a northeasterly direction across Wisconsin (Grant, 1901; Dutton and Bradley, 1970). In Wisconsin, this belt consists of two units, the northern one of which is truncated by a series of northward-trending faults in Douglas and Bayfield Counties. The southern segment appears to extend across Wisconsin, through a structurally complicated area in Ashland and Iron Counties, and thence to the Keweenaw Peninsula of northern Michigan.

The most complete description of the Chengwatana volcanic group is by Hall (1901a), who concluded that it probably is more than 20,000 feet thick in Minnesota. Only a small part of the section is exposed, however, and most of that is in Pine County. The exposed section consists of many individual flow units which range in thickness from less than 10 to more than 1,000 feet. Some of the thicker flow units may actually be composed of several individual flows whose separation is made difficult by poor exposures. The lateral continuity of individual flow units is difficult to establish, but some of the thinner flow units can be traced intermittently along strike for more than 25 miles (Grout, 1910a,b). Each flow unit is characterized by a regular sequence of textures, consisting of a basal aphanitic unit that passes transitionally upward into an ophitic (mottled), diabasic, or porphyritic unit, which in turn is overlain by an uppermost vesicular unit that contains variable amounts of secondary minerals. In many flows, the vesicular crust is brecciated, presumably a result of breaking as the lava poured out, and now consists of rubbly or fragmental material in which both vesicles within fragments and interstices between fragments are filled with secondary minerals.

Lenticular beds of conglomerate or, less commonly, sandstone, that range in thickness from less than 3 to more than 300 feet are intercalated with the flows of many localities. The conglomerate is variable in lithology; it consists of varying proportions of pebble- to boulder-size clasts of quartzite, quartz porphyry, red granite, gray granite, diabase, and diabase porphyry indurated by a matrix of dark-red or dark reddish-brown lithic sandstone. Locally, the conglomerates are crudely graded, cross-bedded, and ripple marked, and have many features, including cut and fill structures, indicative of alluvial deposition.

Beds of tuff and tuff-breccia commonly occur between individual flow units at several localities in Pine County. The breccia consists of angular fragments of amygdaloidal basalt enclosed within a tuffaceous matrix. At most places the tuff is fine grained and the sand-size fragments are angular; some tuffs contain rounded grains and show other evidence of deposition and reworking in a shallow-water environment. Thus, it is difficult to distinguish between an epiclastic sandstone and a reworked volcanogenic tuff. In all tuff and tuff-breccia units the matrix material has been appreciably altered, and now consists of intergrown very fine-grained quartz, epidote, chlorite, and lesser amounts of calcite and zeolites of various types.

North-northeastward-trending dikes averaging about 5 feet in width cut the lava flows and the conglomerates at several localities. The dikes have a well developed diabasic texture, and probably are cogenetic with the lava flows.

Structure

Because of a lack of good exposures, much of the Upper Precambrian structure in east-central Minnesota is obscure. Although Upper Cambrian sedimentary rocks overlie the flows to the south, several geophysical studies (Sims and Zietz, 1967; Mooney and others, 1970a and b) have demonstrated that the flows comprise part of a large basalt block —the St. Croix horst of Craddock and others (1963)— which extends beneath the Paleozoic strata into southeastern Minnesota. This basalt block is flanked by basins filled with Keweenawan strata, and these components collectively are referred to as the Midcontinent Gravity High (see papers by Craddock and Morey, this chapter). The block is bounded on the west by the Douglas and Pine faults and on the east by the Hastings fault, which is thought to be the southeasterly extension of Wisconsin's Lake Owen fault (Craddock and others, 1963; Dutton and Bradley, 1970). Both the Douglas and Pine faults are inferred to be nearly vertical and to have displacements of as much as several thousand feet in the southern part of the area. In central Pine County, the two faults merge into a single fault that has a total displacement in excess of 8,000 feet (Mooney and others, 1970a and b). Although previous interpretations (Craddock and others, 1963; White, 1966a and b) have inferred that the St. Croix horst is an upthrown block, the structure more recently has been interpreted as a continental-size rift (Hinze and others, 1971; King and Zietz, 1971). The structural configuration (fig. V-79) of the horst (Li, 1971, unpub. M.S. thesis, Univ. Minn.) and the stratigraphic evidence that the horst stood as a positive area during at least part of the time of Keweenawan sedimentation (Morey, this chapter) are consistent with the latter interpretation.

In Wisconsin, the flows on top of the horst define what appears to be a broad syncline whose axis more or less parallels the bounding faults. The axis of this inferred syncline defines the Lake Superior syncline of Dutton and Bradley (1970), and appears to project northeastward into the northward-plunging Ashland syncline of White (1966a and b).

The southwestward projection of the trace of the synclinal axis would pass between the rocks exposed at Taylors Falls in Chisago County and those exposed on the Snake River and its tributaries in Pine County (fig. V-78), and judging from the aeromagnetic data of Sims and Zietz (1967), would bisect a basin on the top of the St. Croix horst in the Minneapolis-St. Paul area. The rocks in this basin, called the pre-Paleozoic Twin Cities basin by Sims and Zietz (1967), appear to be folded about a southwestward-plunging axis. The subcrop pattern of Upper Precambrian basalts suggests that the Lake Superior syncline and the pre-Paleozoic Twin City basin were never connected; instead, these areas appear to represent sedimentary accumulations in separate basins that subsequently were folded.

The structure of the area is complicated further by a number of northwestward- and westward-trending faults that are younger than the major bounding faults. These younger faults appear to be nearly vertical and generally to have small displacements, but they are significant in that they break and offset the major boundary faults into several short segments. Thus, the nearly vertical major boundary

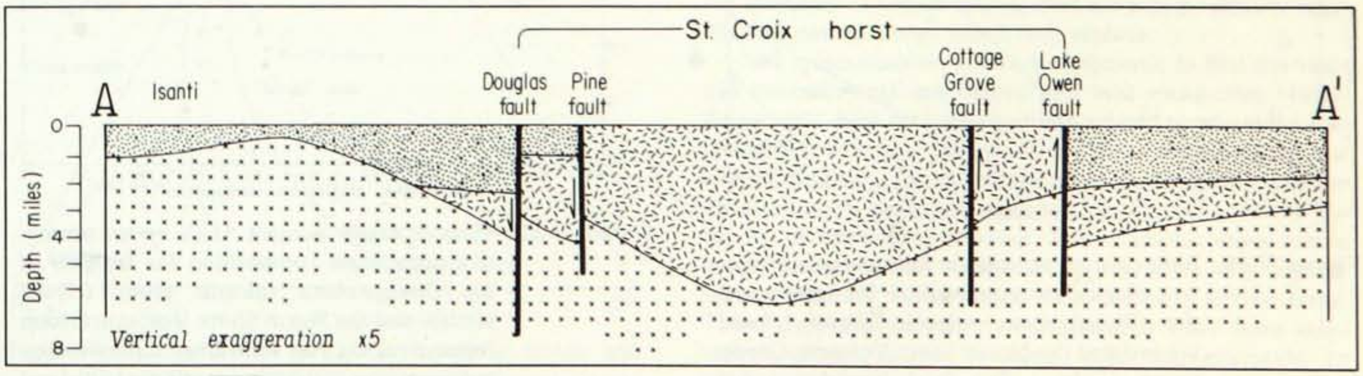

Figure V-79. Inferred geologic section across the St. Croix horst (based on the data of Li, 1971, unpub. M.S. thesis, Univ. Minn.). Location of cross-section shown in Figure V-77.

faults appear to have surface traces that are curved. In addition, movement on all the faults subsequent to Croixan (Late Cambrian) deposition has resulted in the Paleozoic strata having a complex outcrop pattern, particularly in Wisconsin (see Dutton and Bradley, 1970).

Petrology of the Flows

The mineralogy of the flows is poorly known (see Grout, 1910a), and thus the flows are difficult to classify. However, the available chemical analyses (summarized in Ruotsala and Tufford, 1965) indicate that the lavas have intermediate to mafic compositions. Unfortunately, all the analyses were made prior to 1920, and their reliability is uncertain. However, a comparison of the old analyses of rocks from the North Shore Volcanic Group with the more recent analyses presented by Phinney (1970) and Green (this chapter) suggests that the older analyses probably have systematic errors in which Na_2O is underestimated and Al_2O_3 is overestimated. The highly altered nature of the Chengwatana rocks also contributes to uncertainties in the analytic data. Nevertheless, these data, when tabulated with the data of Phinney (1970) and Green (this chapter) by the method of Irvine and Baragar (1971) show some interesting trends.

When normative color indexes and normative plagioclase compositions are considered, there appears to be little difference between the North Shore Volcanic Group and the Chengwatana volcanic group (fig. V-80). Rocks in both sequences are predominantly basaltic; dacite and rhyolite are subordinate. The lack of andesitic rocks may be explained in part by sampling bias; however, Sandberg (1938) also noticed this gap in the North Shore Volcanic Group, and it may be real.

Rocks from both areas define a subalkaline trend (fig. V-81), and fall between clearly defined tholeiitic and calc-alkaline affinities (fig. V-82). The trend of rocks of the

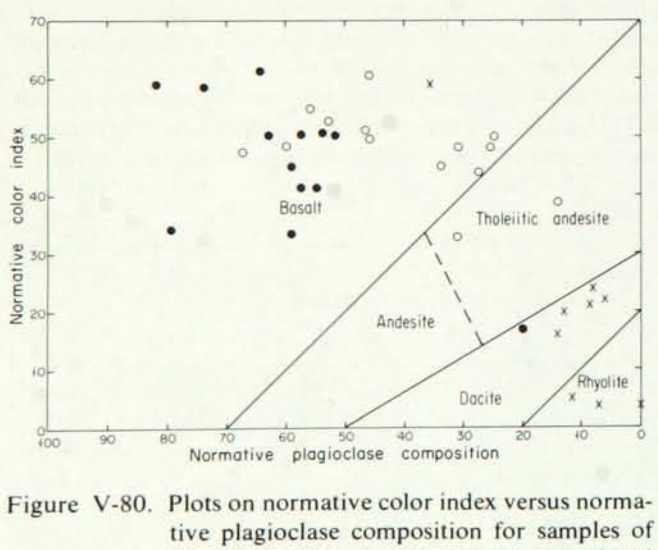

Figure V-80. Plots on normative color index versus normative plagioclase composition for samples of the Chengwatana volcanic group (closed circles) and the North Shore Volcanic Group (open circles). The field boundaries are those of Irvine and Baragar, 1971 (plots in location equivalents).

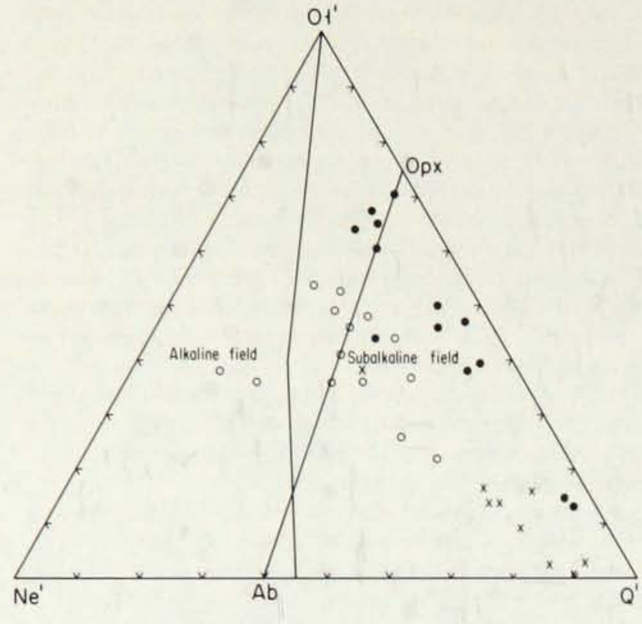

Figure V-81. Ol'-Ne'-Q' projection of samples from the Chengwatana volcanic group (closed circles) and North Shore Volcanic Group (open circles). The fine line is Yoder and Tilley's (1962) "critical plane of silica saturation" and the heavy line is the dividing line for alkaline and subalkaline rocks proposed by Irvine and Baragar, 1971 (plots in percent cation equivalents).

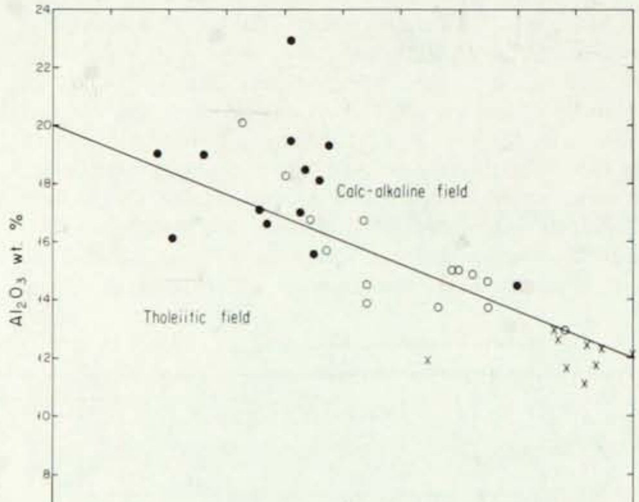

Figure V-82. Plots of weight percent Al_2O_3 versus normative plagioclase composition for samples of the Chengwatana volcanic group (closed circles) and the North Shore Volcanic Group (open circles). The solid lines separate contrasting suites of predominantly tholeiitic and calc-alkaline volcanic rocks (after Irvine and Baragar, 1971).

North Shore Volcanic Group is "average" in terms of alkali ratios, whereas the rocks from east-central Minnesota appear to be more potassium-rich (fig. V-83); this difference may be real or it may be the result of underestimating the Na_2O content.

Comparisons of the Upper Precambrian analyses with data of various igneous suites, as summarized by Irvine and Baragar (1971), indicate that differentiation trends in the Chengwatana volcanic group resemble those of the 1.2 billion-year-old Coppermine River flood basalts in Canada and the mid-Atlantic tholeiitic rocks. The latter similarity is consistent with Green's conclusion (this chapter) that rocks of the North Shore Volcanic Group resemble the Tertiary plateau lavas in eastern Iceland. The similarity to Icelandic lavas and mid-Atlantic tholeiites also is consistent with the hypothesis that the Keweenawan rocks are part of an ancestral rift feature. However, because of analytic uncertainties, it cannot be concluded that rocks of the Chengwatana volcanic group differ in essential chemical characteristics from those of the North Shore Volcanic Group. Nevertheless, the present data suggest that the source of the Chengwatana group may have been somewhat different from that of the North Shore Volcanic Group, an observation consistent with that of White and others (1971) who have suggested that the Upper Precambrian lavas accumulated in several separate basins rather than in a single large one, and with that of Green (this chapter) who concludes that the North Shore Volcanic Group is not correlative with the volcanic rocks exposed along the south shore of Lake Superior.

ECONOMIC GEOLOGY

Since at least 1865, it has been known that small amounts of native copper occur in rocks of the Chengwatana volcanic group (Taylor, 1866). Most commonly, the copper is concentrated in stratiform layers defined by the amygdaloidal portions of individual flow units. The following generalizations apply to this kind of occurrence: (1) both the footwall and hanging-wall are composed of massive basalt; (2) fragmental amygdaloid, the most common host rock, grades downward and laterally into nonfragmental amygdaloid and the nonfragmental parts tend to be very lean or barren of copper; (3) the fragmental layers pinch and swell, and commonly not all parts of the fragmental layer are copper-bearing; (4) because the distribution of fragmental amygdaloid is irregular and because copper is not uniformly distributed, it is difficult to determine either the size of potential ore bodies or meaningful values of ore tenor within these bodies.

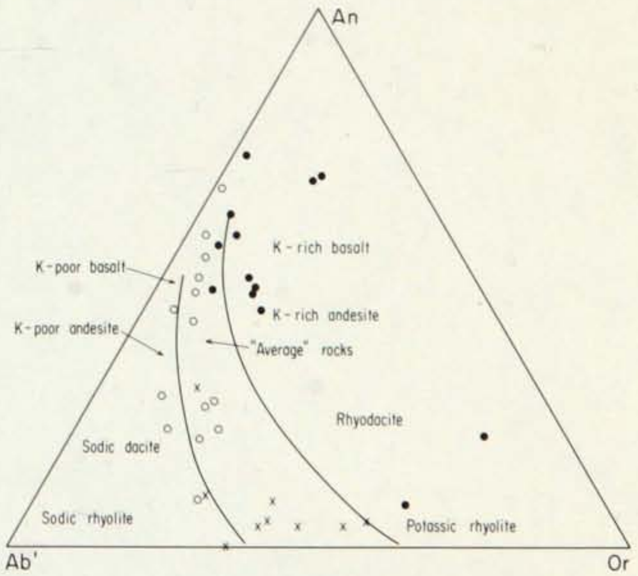

Figure V-83. An-Ab'-Or projection used by Irvine and Baragar (1971) to distinguish K-poor, "common," and K-rich variants of subalkaline rocks (plots in percent cation equivalents).

Native copper also occurs in a conglomerate at one locality in east-central Minnesota. The distribution of copper in the conglomerate is not co-extensive either laterally or vertically with the geometry of the conglomerate layer. Rather, the copper tends to be concentrated along certain layers within the bed, and tends to reflect the bedding configuration.

Native copper is, for practical purposes, the only potential ore mineral, although other minerals of economic importance, including native silver, chalcocite, and chalcopyrite, are present in trace amounts. The native copper occurs primarily as small to large disseminated grains, but some of it fills amygdules, interfragmental openings in the amygdaloidal layers, and irregular and discontinuous fractures in the host rocks. Commonly the native copper replaces paragenetically early secondary minerals, which like the copper, occur primarily in amygdules and interfragmental spaces. Paragenetically these include early secondary quartz, calcite, chlorite, epidote, and prehnite. Quartz, calcite, and chlorite are most abundant as void-filling minerals, but epidote and prehnite are abundant locally. Epidote and chlorite also may replace the amygdaloidal host rock itself, but the extent of such replacement is erratic both within and between flow tops.

Typically, native copper is later than epidote and in places—for example, along the Kettle River—is intimately intergrown with prehnite. Laumontite is fairly abundant along the Snake River, where it appears to have formed later than the native copper; locally, malachite and azurite are abundant as vesicle fillings that are paragenetically later than the native copper which they replace.

The mineralization is clearly epigenetic in that the various minerals were introduced into host rocks after burial. Apparently, host-rock permeability related to alteration was one of the principal factors controlling localization of the copper minerals. The distribution and abundance of copper do not always show a discernible correlation with the intensity of alteration, however. For example, some native copper occurs within rocks that appear to be only slightly altered, whereas it is absent in other intensely altered areas. Thus, copper mineralization appears to have been controlled, at least in part, by factors other than alteration. In Michigan, where the geology is similar to that in east-central Minnesota, there are strong indications that areas of synclinal folding are particularly favorable loci for the deposition of commercial quantities of ore. In addition, the ore deposits in Michigan show a relation to the regional zoning of secondary minerals in amygdaloids. A majority of the large ore bodies lie within a stratigraphic interval characterized by the presence of epidote and prehnite (Stoiber and Davidson, 1959). In addition, the ore deposits lie within, but close to, the boundary of an area containing abundant amygdaloidal quartz. Neither the small-scale structural features nor the amygdaloidal mineralogy have been evaluated in detail in east-central Minnesota. However, to judge from the available strike and dip data, the flows in Minnesota have been folded, and in addition, Berkey (1897-1898) and Grant (1901) reported the presence of quartz, prehnite, and epidote in the vicinity of Taylors Falls and along the upper St. Croix River where copper was once extracted.

In conclusion, it should be emphasized that there are many areas on the Keweenaw Peninsula of Michigan that satisfy the above criteria but do not contain mineable quantities of copper. This suggests that the probability of finding commercial ore deposits in Minnesota is small; but inasmuch as the area is large and much of it seems to have the basic geologic criteria favorable for ore deposition, it should receive some consideration for exploration.

SEDIMENTATION AND PETROLOGY OF THE UPPER PRECAMBRIAN HINCKLEY SANDSTONE OF EAST-CENTRAL MINNESOTA

A. D. Tryhorn and Richard W. Ojakangas

The Upper Precambrian Hinckley Sandstone, an orthoquartzite, is exposed at only a few localities in east-central Minnesota, but is present in the subsurface at least as far south as the Twin Cities (fig. V-84). It was first described by N. H. Winchell (1886); the type section was exposed in a quarry on the Grindstone River, but it is no longer accessible. The maximum exposed thickness is about 100 feet in a quarry at Sandstone, but 500 feet of the sandstone was penetrated in a nearby well (Grout and others, 1951, p. 1061). The unit is generally buff colored; yellow and red staining is present locally. Beds range in thickness from a few inches to a few feet. Large-scale cross-bedding is common and current ripple marks are present. Minor arkosic and conglomeratic facies are present locally near its base.

The formation lies conformably (?) on the Upper Precambrian Fond du Lac Formation (see Morey, this chapter) and unconformably on Middle Keweenawan lava flows. The basal contact of the formation is not exposed, as most of the rocks are poorly exposed in the Hinckley area. However, conglomerate exposed east of Hinckley contains quartzite clasts, and may be near the base of the formation where it rests on volcanic rocks. Drill hole data reveal that the Hinckley is overlain, probably disconformably, by the Upper Cambrian Mt. Simon Sandstone.

The Hinckley Sandstone has been correlated with the Devils Island Sandstone of the Bayfield Group in Wisconsin (Atwater and Clement, 1935; Tyler and others, 1940). Seismic velocity data (Mooney and others, 1970a, p. 5070-5072) and the general geographic distribution of these rocks are consistent with this correlation. The westernmost outcrops of the Devils Island Sandstone are near Cornucopia, about 70 miles from the northernmost Hinckley outcrops at Holyoke. For an account of regional relationships and stratigraphic problems concerning the Upper Precambrian rocks, the reader is referred to Atwater and Clement (1935) and Craddock (this chapter).

PETROGRAPHY

Previous petrographic work on the Hinckley Sandstone includes studies of the light minerals by Atwater and Clement (1935) and Crowley and Thiel (1940), and studies of the heavy minerals by Tyler and others (1940).

The sandstone is medium to coarse grained. The grains are generally moderately to well rounded; sorting varies from poor to moderate (fig. V-85). The rock is weakly to strongly cemented by silica. Carbonate cement occurs in the upper part of the formation in some wells of the Twin Cities area and a local, sparse clay fraction is composed of admixed illite and kaolinite (G. B. Morey, 1972, oral comm.). Outcrop samples are stained by iron oxides which were deposited both before and after deposition of quartz cement.

The mineralogy was determined by counting 600 points in each of 21 thin sections on traverses perpendicular to bedding. The main framework grains were then recalculated to 100 percent and plotted on a triangular diagram (fig. V-86); clay, cement, opaques, and miscellaneous grains were removed. The average framework composition of the Hinckley Sandstone is about 96 percent quartz, 2 percent feldspar, and 2 percent felsic volcanic rock fragments, metamorphic rock fragments, and chert. Sample H-10 (table V-36), from the matrix of the previously mentioned basal (?) conglomerate, contains 12 percent feldspar and is a feldspathic sandstone. In addition, each counted quartz grain was assigned to one of six categories, as follows: (1) unit (common) quartz without inclusions; (2) unit quartz with regular inclusions; (3) unit quartz with acicular inclusions; (4) unit quartz with irregular inclusions; (5) fine polycrystalline (metamorphic) quartz grains; and (6) reworked quartz grains with abraded overgrowths (fig. V-87). Heavy accessory minerals include dominant leucoxene and ilmenite as well as well rounded zircon, tourmaline, rutile, and garnet. The ZTR maturity index (Hubert, 1962) is estimated at greater than 90, showing that the Hinckley Sandstone is quite mature.

PROVENANCE AND SEDIMENTATION

The mineralogic maturity of the rock, the rounding of the grains, and the presence of some reworked quartz grains having abraded overgrowths imply that older sandstones provided much of the detritus now found in the Hinckley Sandstone. The most likely sedimentary source rock for this detritus is the underlying subarkosic to arkosic Fond du Lac Formation (Morey, 1967a). It extends further north than the Hinckley, and may have had a much greater areal extent to the west. In the type area near the western tip of Lake Superior, a 300-foot-thick section of the Fond du Lac Formation is exposed, but seismic evidence suggests that the formation is as much as 7,000 feet thick in the vicinity of Hinckley and more than 2,000 feet thick near the western tip of Lake Superior (Mooney and others, 1970a).

Fond du Lac sandstones consist of 36-68 percent quartz, 5-29 percent feldspar (mostly orthoclase, with microcline and plagioclase), 1-10 percent rock fragments—largely chert and quartzite, plus basalt, felsite, iron-formation and schist —and a trace to 21 percent mica (Morey, 1967a). The heavy mineral suite includes leucoxene, apatite, tourmaline, zircon, magnetite, ilmenite, and garnet (Tyler and others,

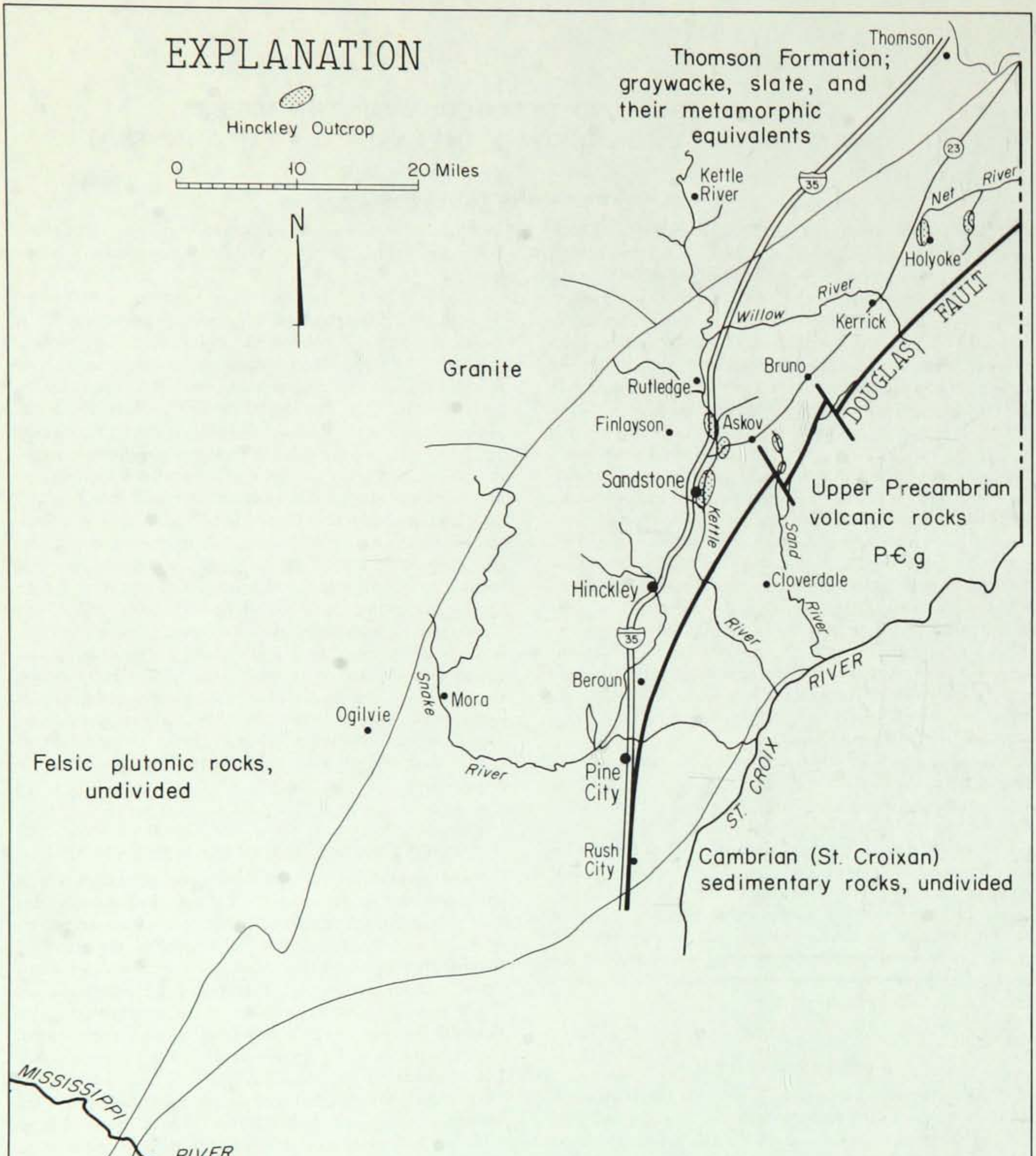

Figure V-84. Distribution of Upper Keweenawan sedimentary rocks in east-central Minnesota (after Sims, 1970). Areas in which Hinckley Sandstone is exposed are shown by stippled pattern.

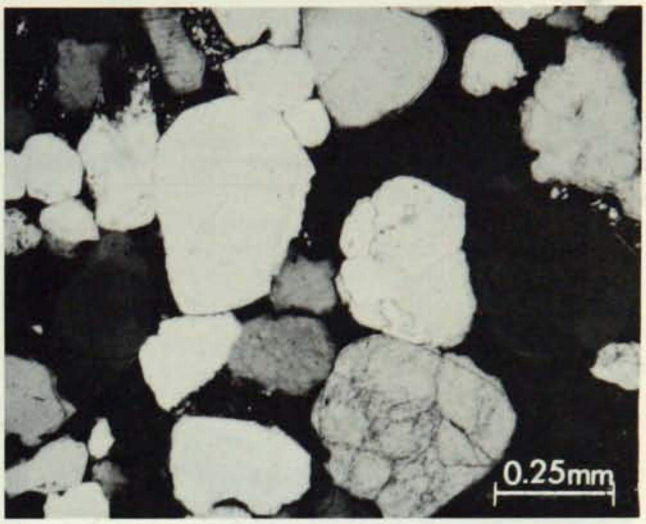

Figure V-85. Photomicrograph of a representative sample of Hinckley Sandstone (sample PH-13, table V-36; crossed nicols).

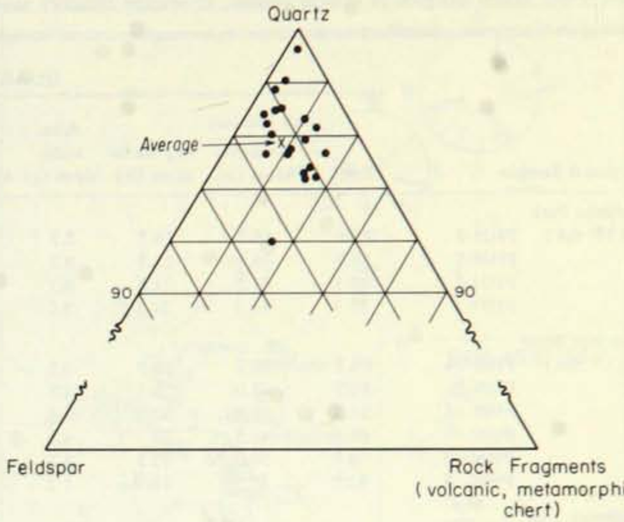

Figure V-86. Compositional diagram of framework constituents of 20 samples of quartzose Hinckley sandstones (see table V-36).

1940, p. 1511). Extensive weathering, erosion, and reworking of the sand-size fraction of the Fond du Lac Formation, resulting in the elimination of most of the labile components, could have yielded the quartzose Hinckley Sandstone. According to Myers (1971, unpub. Ph.D. dissert., Univ. Wisconsin), the sandstones of the Bayfield Group in Wisconsin probably were formed in a similar way by the recycling of the underlying Freda Sandstone.

The original sources for the Hinckley and Fond du Lac sandstones probably included many rock types. The unit or common quartz may have been derived from quartz-bearing plutonic rocks, coarse schists and gneisses, felsic volcanic rocks, and quartz veins. The minor abundance of composite (polycrystalline) quartz from metamorphic sources in the Hinckley is expectable in quartz-rich sandstones inasmuch as such grains probably are selectively destroyed during abrasion (Blatt and Christie, 1963). The types of inclusions in quartz have been shown to be related to certain source rocks (Keller and Littlefield, 1950). Because about a quarter of the quartz grains in the Hinckley have regular inclusions, much of the unit quartz was probably derived from metamorphic sources. The presence of acicular and irregular inclusions in about 12 percent of the quartz grains suggests igneous sources. The orthoclase is indicative of a granitic terrane, but chert, metamorphic rocks, and volcanic rocks also were present in the original source area. On the basis of abraded feldspar overgrowths, Morey (1967a) has suggested that sedimentary rocks were also an important source for the Fond du Lac detritus.

Paleocurrent patterns obtained at five localities by measuring a total of 114 cross-bed dip directions, trough axes, and ripple marks are shown in Figure V-87. As the formation generally dips less than five degrees, corrections for tilt were not necessary. The average paleocurrent direction is from northwest to southeast (arithmetic mean = 134°); but directions differ from one outcrop area to another (fig. V-88). The mean is similar to the paleocurrent trends determined by Hamblin (1965) for the Hinckley as part of a regional study, by Myers (1971, *op. cit.*) for the partially correlative Bayfield Group to the east in Wisconsin, and by Morey (1967a) for the subjacent Fond du Lac Formation. This southeasterly paleocurrent trend suggests that the sources for the sediments in this region during Late Precambrian time were to the north, west, and northwest.

All the postulated plutonic, metamorphic, volcanic, and sedimentary source rocks are present in Middle and Lower Precambrian terranes to the west and north. During Fond du Lac time, detritus was deposited in a fluvial-deltaic environment, as documented by Morey (1967a). Hamblin (1965) has shown that the configuration of the Keweenawan sedimentary basin, based on the total thickness of Upper Precambrian sedimentary rocks, and on sediment dispersal patterns for the entire Lake Superior region, was similar in orientation and shape, although somewhat larger than, the present basin of Lake Superior. The Fond du Lac Formation and the correlative Orienta Sandstone in Wisconsin form a fluvial-deltaic complex that apparently nearly filled the western part of the basin with immature detritus.

The greater textural and mineralogic maturity of the Hinckley Sandstone, and apparently of the correlative Devils Island Sandstone in Wisconsin as well, requires that a near-shore, high-energy environment succeeded the fluvial-deltaic environment. Whether this change from one environment to another required any great length of time is unknown, but Morey (this chapter) reports that the Fond du Lac-Hinckley boundary is transitional; this could be interpreted as meaning that no large time span is represented by the transition. Fond du Lac sedimentary rocks exposed west and north of the area covered by the present Hinckley Sandstone could have been weathered and transported

Table V-36. Modal analyses, in volume percent, of selected Hinckley sandstone samples.[1]

		QUARTZ								ROCK FRAGMENTS					
Section & Sample		Total	No inclusions (a)	Reg. inclusions (b)	Acic. inclusions (c)	Irreg. inclusions (d)	Composite (meta) (e)	Sed. reworked (f)	Feldspar (g)	Chert	Volcanic (h)	Opaques (i)	Miscellaneous (j)	Clay matrix	Silica cement
Holyoke Park															
(35′ thk)	PHH-7	84.6	53.1	18.5	2.7	6.2	4.1		2.0	P	2.5	P		2.7	7.0
	PHH-5	78.7	46.5	21.7	3.2	3.7	3.2	P	2.8		1.7	8.5	T	P	7.8
	PHH-3	82.3	41.5	21.8	4.3	7.0	7.5	T	1.1	T	1.9	1.2	T	P	12.7
	PHH-1	87.3	41.3	24.3	7.2	7.0	7.3	T	2.5	T	P	T		T	8.8
East Net River															
(55′ thk)	PHN-24	86.2	46.7	22.3	3.3	9.2	4.5	T	1.3	T	2.5	P	T	1.3	7.3
	PHN-21	90.2	45.0	26.3	4.8	6.6	7.5		2.1	1.1	1.7	T	T	P	3.8
	PHN-18	87.0	36.8	29.5	6.3	9.8	4.5		2.0	T	P	1.5	T	1.0	7.5
	PHN-15	80.6	43.5	24.7	4.1	3.5	4.7	T	2.2	1.2	1.5	P		2.8	11.0
	PHN-12	78.8	54.2	12.5	3.3	4.0	4.8		4.2	1.0	1.7	2.3	T	1.2	9.8
	PHN- 9	82.0	52.5	18.0	1.2	3.1	7.2		2.1	P	1.8	T		T	12.5
Highway 35W															
(10′ thk)	PH-13	81.6	36.3	20.0	6.0	15.8	3.5	T	1.7	T	P	P		1.7	13.2
	PH-14	86.0	52.8	16.8	1.7	10.2	4.5		1.2	T	1.6	5.0	T	1.2	4.7
Sandstone (quarry)															
(20′ thk)	PH-11	83.7	34.8	22.7	11.7	8.2	6.0	T	1.2	P	T	T			13.7
	PH-10	85.0	47.3	23.7	5.7	3.3	5.0		2.2	P	1.0	T	T	P	9.8
	PH- 9	83.3	51.8	19.7	6.2	1.3	4.3		1.5	T	T	1.2		P	13.0
	PH- 8	86.3	56.0	18.3	5.2	2.0	4.8		1.2	T	P		T	T	11.3
Highway 23															
(25′ thk)	PH- 1	81.0	42.8	20.2	6.2	7.7	4.1		2.3		1.0	P		2.3	12.7
	PH- 4	84.5	23.8	39.0	7.0	10.5	4.0	T	1.7	T	1.8	5.2	T	1.8	4.5
	PH- 5	86.6	38.5	23.3	6.7	14.6	3.5		2.4		2.0	1.0		P	7.5
	PH- 6	84.1	37.8	24.5	5.2	10.8	5.8		2.3	T	T	2.5		P	9.3
Hinckley	H-10[k]	66.7	31.4	7.9	10.9	4.5	12.0		12.0	T	6.2	1.8	T	9.7	3.2
Averages, excluding sample H-10		84.0	44.2	22.4	5.1	7.2	5.1	T	2.0	T	1.3	1.7	T	P	9.4

T = 0-0.49 percent; P = 0.50-0.99 percent
[1] 600 points per sample, counted on random traverses perpendicular to bedding
(a) Unit (common) grains with straight to strongly undulose extinction; some semi-composite grains; coarse polycrystalline grains with non-sutured boundaries; all without vacuole trails or microlites larger than 10 microns
(b) Unit grains with regular inclusions; like (1), but with microlites of tourmaline, biotite, and zircon
(c) Unit grains with acicular inclusions; like (1), but with needle-like inclusions of rutile and minor tourmaline
(d) Unit grains with irregular inclusions; like (1), but with fluid inclusions or abundant bubble trails, opaques, etc.
(e) Composite (metamorphic) polycrystalline grains with either straight or sutured boundaries, with or without inclusions
(f) Sedimentary reworked grains of unit quartz with abraded overgrowths beneath later unabraded overgrowths
(g) Orthoclase, with minor plagioclase and microcline
(h) Dominantly felsic volcanic fragments; minor intermediate volcanic fragments and metamorphic fragments
(i) Dominantly hematite-limonite, some ilmenite and leucoxene
(j) Includes zircon, mica, unknowns
(k) Matrix of quartzite pebble conglomerate, excluding pebbles

southeastward to the slowly encroaching shoreline. As the shoreline transgressed onto the fluvial-deltaic complex, additional reworking ensued.

Although the Hinckley is apparently quite similar in most respects to the overlying Upper Cambrian marine sandstones, there is no paleontologic evidence that the Hinckley was deposited under marine conditions. Furthermore, all the paleogeographic data (Hamblin, 1965) indicate a closed or nearly closed basin during Late Keweenawan time. As the basin stabilized tectonically, it was filled by fluvial-deltaic products. A large lake probably formed in the basin and slowly transgressed higher onto the older fluvial-deltaic deposits. Therefore, reworking of the feldspathic sands of the Fond du Lac Formation in a stable, shallow-water lacustrine environment is proposed as the most suitable sedimentation model for the development of the Hinckley Sandstone.

ACKNOWLEDGMENTS

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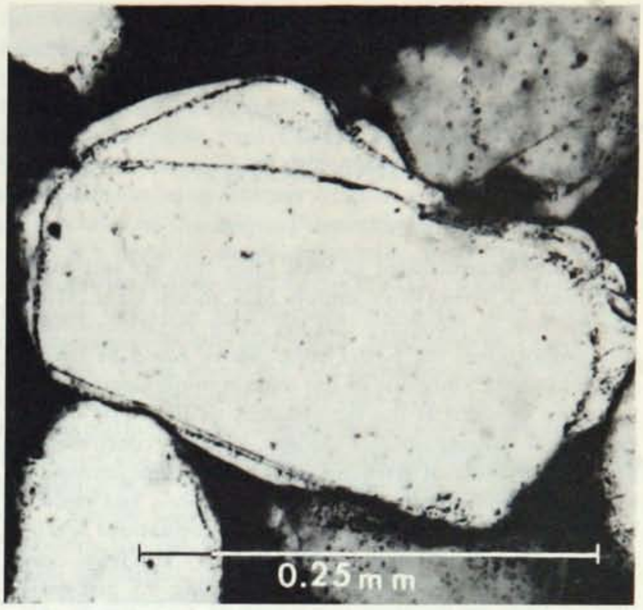

Figure V-87. Photomicrograph of multi-cycle quartz grain with abraded quartz overgrowth beneath present quartz overgrowth (sample PH-13, table V-36; crossed nicols).

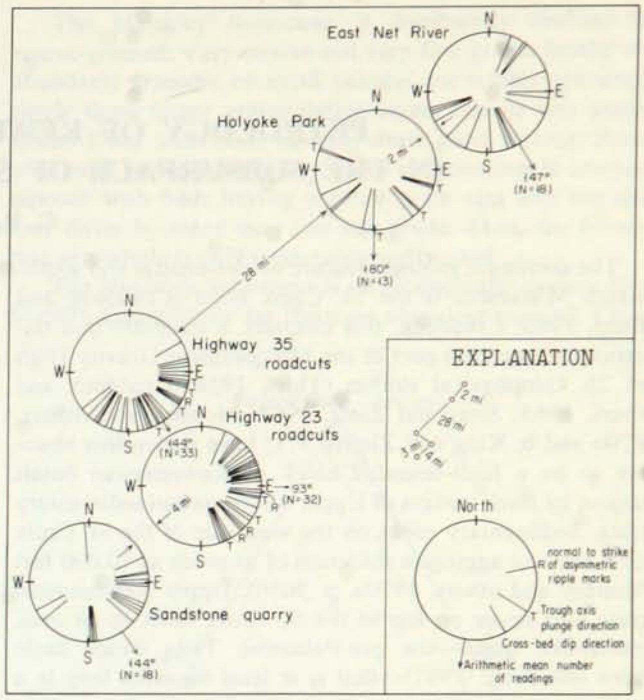

Figure V-88. Summary of paleocurrent data from the Hinckley Sandstone.

PETROLOGY OF KEWEENAWAN SANDSTONES IN THE SUBSURFACE OF SOUTHEASTERN MINNESOTA

G. B. Morey

The dominant geologic feature of east-central and southeastern Minnesota is the St. Croix horst (Craddock and others, 1963; Craddock, this chapter), a structure that underlies the northern part of the Midcontinent Gravity High (pl. 2). Geophysical studies (Thiel, 1956; Craddock and others, 1963; Sims and Zietz, 1967; Mooney and others, 1970a and b; King and Zietz, 1971) have shown this structure to be a fault-bounded block of Keweenawan basalt flanked by thick wedges of Upper Keweenawan sedimentary strata. Sedimentary rocks on the west side of the St. Croix horst have an aggregate thickness of as much as 10,000 feet (Mooney and others, 1970a, p. 5056). Upper Keweenawan strata also occur on top of the St. Croix horst in an oval, graben-like basin—the pre-Paleozoic Twin Cities basin (Sims and Zietz, 1967)—that is at least 60 miles long in a northeasterly direction and 30 to 35 miles wide at its widest point. These clastic rocks range in thickness from a few tens or hundreds of feet on the west flank to at least 3,000 feet under the central part of the basin. A narrow wedge of basalt—the Hudson-Afton horst (Sims and Zietz, 1967)—defines the east side of the basin and separates Keweenawan strata within the pre-Paleozoic Twin Cities basin from that in the flanking basin on the east side of the horst (fig. V-79).

The majority of exposures of Keweenawan sedimentary rocks are found in Wisconsin and Michigan along the south shore of Lake Superior (Thwaites, 1912; Hamblin, 1965). Scattered outcrops of Keweenawan strata in Minnesota are limited to a small area in east-central Minnesota (figs. V-78 and V-84) where the sedimentary sequence is divided into a lower part called the Fond du Lac Formation and an upper part called the Hinckley Sandstone. These rocks extend in the subsurface southward into southeastern Minnesota where they are overlain by as much as 800 feet of Paleozoic strata. Although the subsurface occurrences have been recognized for nearly 100 years (Winchell and Peckham, 1874, p. 79; Winchell, 1876, p. 187-189), the rocks were not named until Hall and others (1911, p. 53) used the term "Red Clastic Series." They regarded this name as a temporary, but convenient, term for those red beds of uncertain age penetrated in the deep wells of southeastern Minnesota, rather than as a formal designation. Nevertheless, a dual system of nomenclature using the name "Fond du Lac Formation" for surface exposures and "Red Clastic Series" for subsurface occurrences evolved over the years in Minnesota. In general, these terms have been used interchangeably. For example, Grout and others (1951, p. 1058) referred to both the surface and subsurface strata as the "Fond du Lac beds," whereas Tyler and others (1940, p. 1507) used "Red Clastic Series" for the same rocks. In recent years, however, it has been demonstrated that the two terms have been applied to rock units that are not totally equivalent. Kirwin (1963, unpub. M.S. thesis, Univ. Minn.) was the first to recognize that, by using variations in mineral proportions, the Red Clastic Series could be divided into four units. Only one of the units is mineralogically like the rocks exposed at the type locality of the Fond du Lac Formation. Subsequently, I (Morey, in prep.) confirmed the presence of three of Kirwin's units and recommended that the name "Red Clastic Series" be abandoned and replaced by a formal nomenclature. Thus, three formations are recognized in the subsurface (fig. V-1): (1) Hinckley Sandstone, a buff to tan sandstone containing 95 percent or more quartz; (2) Fond du Lac Formation, consisting of intercalated moderate red shale and sandstone containing variable amounts of quartz, orthoclase, microcline, sodic plagioclase, and lesser amounts of granitic and aphanitic rock fragments; and (3) Solor Church Formation, consisting of reddish-brown mudstone, siltstone, and sandstone composed of variable amounts of quartz, plagioclase of intermediate composition, and aphanitic igneous rock fragments. The latter two formations are contiguous with units exposed at the surface in east-central Minnesota, but so far as is known, the Solor Church Formation is confined to the subsurface.

DESCRIPTIVE STRATIGRAPHY

Stratigraphic analysis (Morey, in prep.) of the Upper Keweenawan strata indicates that in the flanking basins the Solor Church Formation is overlain by the Fond du Lac Formation, which in turn is gradationally overlain by the Hinckley Sandstone (fig. V-77). On top of the St. Croix horst, the Solor Church Formation overlies basaltic rocks in the pre-Paleozoic Twin Cities basin and is locally overlain by the Hinckley Sandstone; at places, the two formations are separated by a regolith as much as 100 feet thick. Either the Fond du Lac Formation was not deposited on top of the horst or was removed by erosion prior to Hinckley deposition.

Hinckley Sandstone

The Hinckley Sandstone was first recognized by Winchell in 1884, but he did not name and describe it until 1886 (p. 337). The original type section at Hinckley is no longer exposed, but the formation crops out almost continuously for nearly 20 miles along the Kettle River in east-central Minnesota (fig. V-84). In outcrop, the sandstone is medium to very thick bedded, fine to coarse grained, and pale red to light pinkish or brownish gray in color (Tryhorn and Ojakangas, this chapter). A similar sandstone was traced southward in the subsurface to the Minneapolis-St. Paul (Twin Cities) area by Crowley and Thiel (1940), but has been recognized only locally south of there. In east-central Minnesota, the Hinckley Sandstone is more than 500 feet thick, but it thins progressively southward from the type locality and on the average is only about 150 feet thick near Minneapolis and St. Paul (Grout and others, 1951, p. 1061). South of the Twin Cities, it generally is less than 50 feet thick and is missing in many wells. How much of this thinning is due to the original depositional pattern and how much is a result of post-Keweenawan-pre-Croixan erosion is not known. Recognition of the Hinckley's subsurface distribution is further complicated by the fact that it rather closely resembles the Mt. Simon Sandstone of Late Cambrian age. Thus, where the two formations are in contact, it is difficult, particularly where only well cuttings are available, to distinguish the Keweenawan-Croixan boundary. In the past, it has been assumed that the two formations could be distinguished primarily by differences in feldspar content. Crowley and Thiel (1940) showed that the Hinckley Sandstone has less than 2 percent feldspar whereas the Mt. Simon Sandstone has from 2 to 5 percent feldspar. However, the presence of relatively feldspar-rich beds in the Hinckley Sandstone partly negates this criterion, and detailed petrographic data are needed from both formations before this problem can be resolved. However, several other criteria serve to distinguish the two formations in diamond drill core. For example, the Hinckley Sandstone lacks the red and green laminated mudstone and shale beds and small-scale planar cross-bedding that characterize much of the Mt. Simon Sandstone and it generally is more indurated than the Mt. Simon, as a consequence of having abundant quartz overgrowths.

On the top of the St. Croix horst the Hinckley Sandstone is uncommonly thin, and its base is marked by as much as 10 feet of conglomeratic sandstone. The conglomeratic clasts generally are pebble-size or smaller and consist of detritus derived from the underlying Solor Church Formation. West of the Douglas fault, the formation is much thicker, and it grades into the underlying Fond du Lac Formation. Where the contact has been penetrated in the subsurface, the buff to tan Hinckley grades downward into the underlying red Fond du Lac Formation. Because of this gradation, the formational boundary has been defined by a change in feldspar content. Using this criterion to distinguish the two formations, the lower several hundred feet of Hinckley Sandstone is distinctly red in color.

As in surface exposures, Hinckley Sandstone in the subsurface is mostly thick to very thick bedded, but locally it contains a few irregular mudstone laminae as much as 2 inches thick. The upper part of the formation contains vaguely cross-stratified beds that are as much as 10 feet thick. In thin section, the formation appears to be texturally mature (fig. V-89A), although some clay commonly is present as a detrital matrix. Most thin sections show less than 2 percent clayey matrix, but because of possible bias in sampling, only qualitative generalizations can be made. Some strata contain as much as 5 percent matrix; it is exceptional to have more than 15 percent matrix. Where abundant, the clayey detritus is concentrated in laminations and thin beds.

The Hinckley Sandstone is dominantly medium to coarse grained. Very coarse and very fine grains locally are abundant; granules or small pebbles are widely scattered; rarely these larger grains define layers one or two grains thick. Thus, individual samples show good sorting. However, well sorted beds of one grain size commonly are juxtaposed with beds having another grain size and the size may differ by more than one size grade. Thus, the formation as a whole is only moderately well sorted.

The Hinckley Sandstone is mineralogically mature (fig. V-89B). Quartz is by far the most abundant mineral. Clear

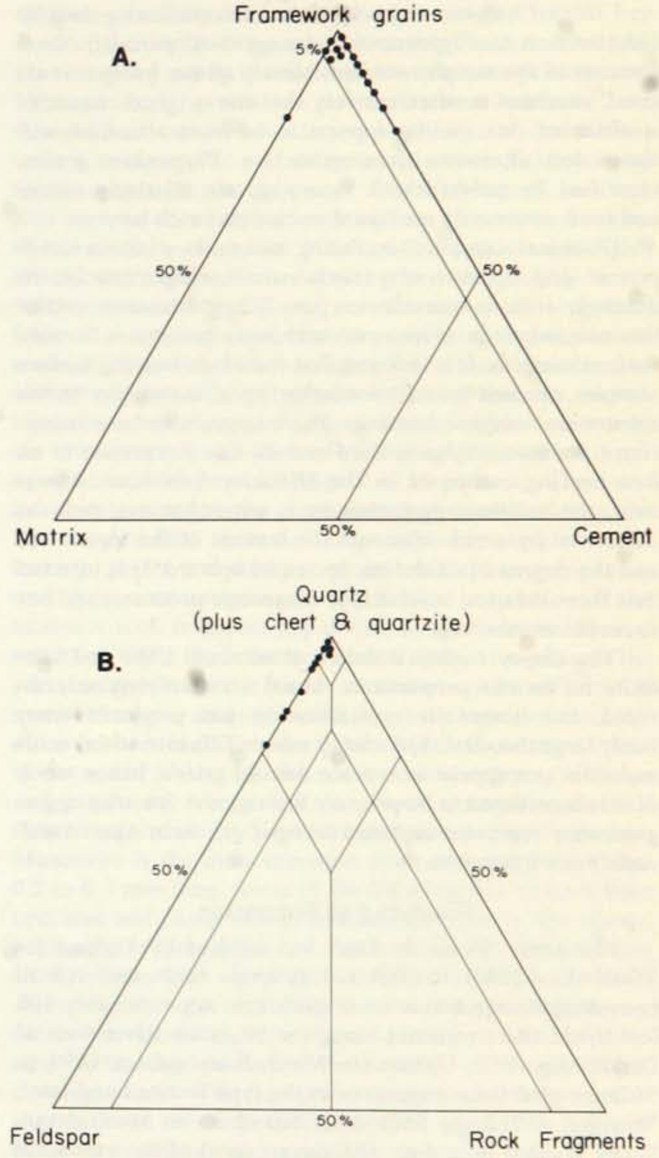

Figure V-89. Textural and compositional diagram of selected Hinckley Sandstone samples. A, texture (data based on 300 points per section); B, framework grain mineralogy (data based on classification of 100 framework grains per thin section).

quartz predominates and the majority of grains have strong undulatory extinction. Polycrystalline quartz grains are present in most thin sections, commonly comprising from 5 to 10 percent of the total quartz content. Authigenic quartz, in the form of detrital grain overgrowths, occurs throughout the Hinckley Sandstone. However, there is no apparent stratigraphic control on the distribution of overgrowths in the sand-rich layers; overgrowths may be present on most of the detrital grains in one thin section, whereas they may be rare in another section. The overgrowths also are clear quartz and can be distinguished from the detrital cores only where a thin concentration of vacuoles or dust occurs along the detrital grain boundaries.

Feldspar (albite, orthoclase, and microcline) generally constitutes a small proportion (range 0-10 percent; $\bar{X}=5$ percent) of the samples studied. Nearly all the feldspar is altered, much of it so extensively that the original character is obscured. Microcline appears to be most abundant and shows less alteration than orthoclase. Plagioclase grains, identified by polysynthetic twinning, are relatively sparse and most commonly are found in rare clay-rich layers.

Chemical cement—excluding silica—is a minor component and consists of various iron-bearing minerals. In outcrop, iron is present as pore-filling hematite and/or limonite, whereas an iron-rich carbonate dominates the subsurface samples. It is inferred that the oxide-bearing surface samples resulted from the weathering of carbonate. In the subsurface samples, however, there appears to be a transition from iron oxides in the Fond du Lac Formation to an iron-bearing carbonate in the Hinckley Sandstone. Moreover, the iron-bearing carbonate is somewhat oxidized—as evidenced by a red color—at the bottom of the formation, and the degree of oxidation decreases upward. It is inferred that this oxidation is related to diagenetic processes and not to recent weathering.

The clayey matrix consists of admixed illite and kaolinite in various proportions. Most is extremely finely divided, but diagenetic crystallization has produced some fairly large flakelets. The clayey matrix fills interstitial voids and does not appear to replace detrital grains; hence much of it is considered to be primary in origin. A few clay aggregates may represent replaced feldspar grains or altered volcanic rock fragments.

Fond du Lac Formation

The term "Fond du Lac" was applied by Upham (*in* Winchell, 1884a) to dark-red to pink shale and red to brown argillaceous to arkosic sandstone, approximately 400 feet thick, that crops out along the St. Louis River west of Duluth (fig. V-2). Upham (*in* Winchell and others, 1899, p. 567) denoted these exposures as the type sections and later, Thwaites (1912, p. 567) described them in some detail. Water ponded by a dam now covers most of the exposures that Thwaites described; consequently, I (Morey, 1967a) redescribed the type section as it is now exposed. The formation also crops out north of Mora in Kanabec County at various places along the valley of the Snake River.

In east-central Minnesota, the Fond du Lac Formation is overlain gradationally by the Hinckley Sandstone and throughout most of its outcrop area is inferred to overlie a variety of igneous and metamorphic rocks of Middle Precambrian age. However, the base of the formation is exposed only at the type locality, where it overlies the Thomson Formation of Middle Precambrian age. Unfortunately, the top of the formation is not exposed at the type locality, and the total thickness is not known; approximately 300 feet of strata are exposed, although the total thickness may exceed several thousand feet. The formation is much thicker in southeastern Minnesota where seismic data (Mooney and others, 1970a, p. 5079) indicate that it may be as much as 8,000 feet thick immediately adjacent to the Douglas fault (fig. V-77); however, only approximately 2,000 feet have been penetrated by drilling.

In both surface and subsurface occurrences the formation is characterized by beds of lenticular sandstone and interbedded mudstone. Both rock types are predominantly various shades of red, but streaks and mottles of white, light greenish gray, and pinkish gray are common. The sandstone is a poorly indurated rock that is both texturally and mineralogically less mature than is the overlying Hinckley Sandstone (figs. V-90A and B). Most of the samples studied are coarse to fine grained, fairly well sorted and skewed to the fine side. The mudstones contain more than 75 percent fine silt and clay and have either a laminated or a massive appearance; much of the laminated variety is fissile.

Several other rock types are recognizable at the type locality. Lenses of conglomerate having a diverse pebble composition occur near the base of the formation. The basal conglomerate is at least 60 feet thick, and predominantly consists of pebbles and cobbles as much as 6 inches in diameter. In addition, there are lesser amounts of chert, quartzite, graywacke, and slate. The matrix is mostly a coarse grit of angular quartz and feldspar, with some clay-size matrix material and dolomite cement. Pyrite and marcasite occurring either as concretions or individual grains are common in interstitial areas; locally these sulfides have been altered to limonite.

The conglomerate passes transitionally upward into a sequence of interbedded light greenish-gray to dark reddish-brown sandstone and mudstone. Within the gradational zone, which is several feet thick, the size of the conglomeratic clasts becomes progressively smaller and the amount of sand-size material increases; there is no distinct break between the quartz-pebble conglomerate and the sandstone beds. Because of the gradational nature of this contact, the quartz-pebble conglomerate—which previously was assigned to the Lower Keweenawan Puckwunge Formation—is now considered part of the Fond du Lac Formation (Morey, 1967a).

A second kind of conglomerate also is intercalated with the sandstone near the base of the formation. This conglomerate overlies the basal quartz-pebble conglomerate and consists of pebbles of highly altered basalt and basalt porphyry randomly distributed in a matrix of fine-grained, reddish-brown sandstone. The matrix is similar mineralogically to that of the sandstone associated with the basal quartz-pebble conglomerate. Similar conglomerates also occur higher up in the formation, where they make up a small but significant part of the sequence. Similar conglomerates occur in southeastern Minnesota where they are associated with pebble-size clasts of sandstone lithologically like that of the Solor Church Formation.

Intraformational conglomerates composed of pebbles of reddish-brown shale are common near the base of many sandstone beds. The clasts range in size from less than one-half to more than six inches in largest dimension. Most of the pebbles are irregular in shape and show little effects of abrasion. Generally, they comprise thin beds or lenses, but many isolated "floating pebbles" are scattered throughout most of the sandstone units.

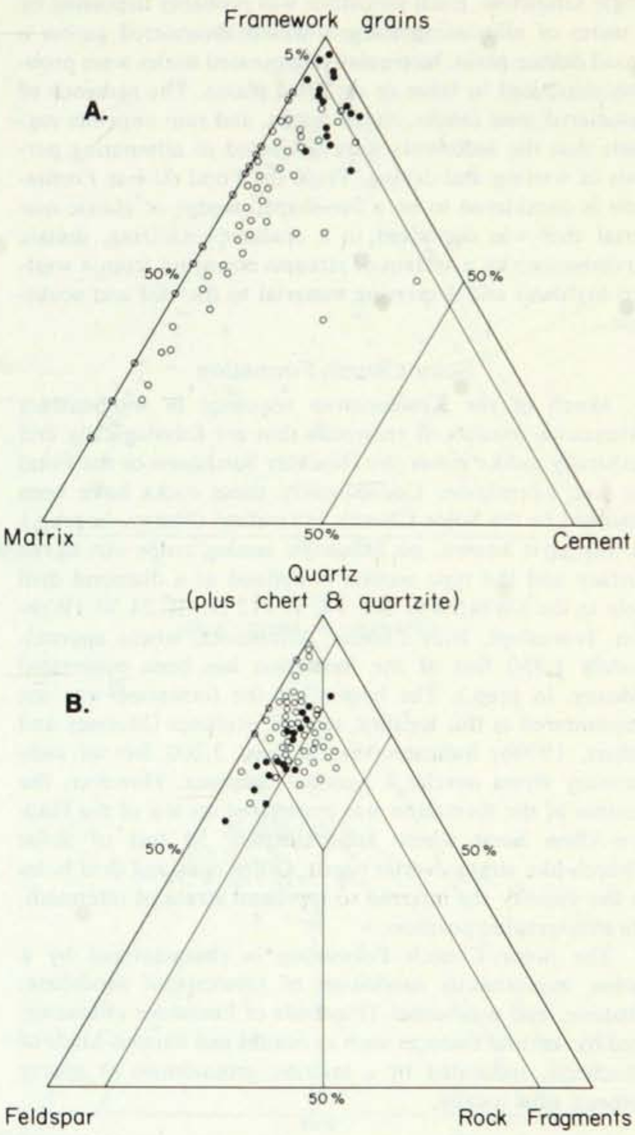

Figure V-90. Textural and compositional diagram of selected Fond du Lac samples. A, texture (data based on 300 points per section); B, framework grain mineralogy (data based on the classification of 100 framework grains per thin section). Open circles, type locality; closed circles, southeast Minnesota.

The mineralogic composition of the Fond du Lac Formation was studied in 50 thin sections from selected drill cores and in 19 thin sections from the type locality (Morey, 1967a). A comparison of data from the two localities (figs. V-90A and B) shows that, although there is a somewhat greater sand-size rock fragment component in the subsurface samples, the formation is everywhere nearly identical in texture and mineralogy.

Framework grains, in order of decreasing abundance, are quartz, feldspar, and rock fragments of igneous and, to a lesser extent, sedimentary and metamorphic origin. In addition, some large grains of detrital biotite and chlorite are present. Quartz comprises the bulk of the samples studied, averaging around 70 percent (range 45 to 95 percent) of the framework constituents. Angular to subrounded grains having strong undulatory extinction are predominant although many grains are polycrystalline and have both straight and curved boundaries, implying derivation from both a metamorphic and an igneous terrane. Feldspar content averages 20 percent (range 5 to 40 percent), and consists dominantly of orthoclase with lesser amounts of microcline and albite-oligoclase. Much of the plagioclase is highly altered and turbid. Some detrital grains have highly sericitized cores and abraded overgrowths of weakly sericitized feldspar, implying more than one cycle of deposition. In contrast, the orthoclase is only slightly altered and the microcline is clear and seemingly fresh. Rock fragments, including both sedimentary and igneous-metamorphic derivatives, are ubiquitous constituents of all the samples studied, averaging around 10 percent (range 2 to 15 percent). Chert and quartzite grains are the most abundant constituents at the type locality, but "granitic" fragments consisting of interlocking quartz and feldspar grains and aphanitic igneous rock fragments comprise an appreciable proportion of the sand-size rock fragment population in southeastern Minnesota (fig. V-91). Sedimentary rock fragments similar to the rocks of the Fond du Lac Formation may be of intraformational origin, although some may have been derived from older red beds including the Solor Church Formation. Mudstone and/or shale are the most common types; siltstone and sandstone are present in minor quantities. Detrital flakes longer than 0.03 mm are common in trace amounts. Muscovite is the most common mica and occurs as flakes 0.2 to 0.3 mm long. Some of the flakes appear to have been hydrated and partly altered to illite. Commonly, the altered micas grade into adjacent matrix material, complicating identification and classification. Chlorite, and to a lesser extent biotite, also are ubiquitous components. Both are somewhat altered and are characterized by partial replacement by the clay matrix and development of hematite along cleavage planes. The hematite is present as earthy-appearing, dark-red microspecks. Alteration appears to begin at flake terminations and gradually proceeds inward along cleavage planes. The biotite also seems to have been leached or hydrated—characterized by swelling and loss of pleochroism and birefringence—and partly altered to chlorite (?) or illite (?). All evidence suggests that the biotite flakes were predominantly modified after deposition.

Chemical cement averages 10 percent (range 2 to 20 percent) of the samples studied. Calcite and an iron oxide

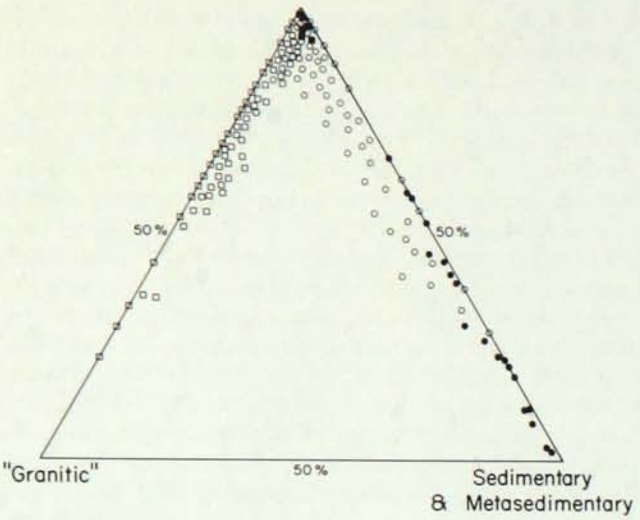

Figure V-91. Compositional diagram of rock-fragment components (recalculated to 100 percent in selected Keweenawan sandstone samples. •, Fond du Lac Formation, type locality; O, Fond du Lac Formation, southeast Minnesota; □, Solor Church Formation, sand-size grains; ▪, pebble- and greater-size clasts. The aphanitic component of the samples from the Fond du Lac type locality includes clasts derived from plutonic rocks of gabbroic composition.

presumed to be hematite are the major cementing agents. Hematite occurs as thin coatings on framework grains, as a stain on clay minerals, rock fragments, detrital biotite and chlorite, and as interstitial void fillings. Where hematite and calcite occur together, the calcite always is paragenetically later.

Matrix material—or detritus less than 0.03 mm in diameter—averages 15 percent (range 5 to 70 percent) of the samples studied. X-ray diffraction shows the matrix to consist of quartz, well ordered illite, and lesser amounts of kaolinite.

Sedimentologic Framework

Medium- to large-scale cross-bedding is the most abundant primary sedimentary structure in the Fond du Lac Formation. It is predominantly of the trough-type in which each set of cross-strata forms a wide, shallow, concave-upward channel that is u-shaped in plan view and wedge- or lens-shaped in longitudinal section. At the type locality, the large-scale trough-like units occur in channels that are cut into previously deposited sediments. The azimuths of the cross-bedding, as well as other directional sedimentary features such as trough axes, grain lineations, ripple marks, and flutes, are consistent with an eastward current movement (Morey, 1967a).

The above sedimentary structures, as well as the presence of mud cracks and rain imprints, indicate that the Fond du Lac Formation most likely was deposited by alluvial processes. Other features supporting this conclusion include: (1) sharp erosional basal contacts of sandstone units which channel-cut parts of the underlying strata; (2) an upward change in sedimentation units from thick-bedded, coarse-grained, locally conglomeratic sandstone at the base to thin-bedded, fine-grained sandstone and siltstone at the top; (3) common cut-and-fill structures; (4) poor sorting and bimodal size distribution of conglomerate within the sandstones; and (5) rapid lateral changes in both thickness and lithology.

The lateral extent of the sandstones suggests that more than one stream was operative during the deposition of a single sandstone. Each sandstone was probably deposited by a series of alluviating streams which meandered across a broad deltaic plain. Intercalated laminated shales were probably deposited in lakes or on flood plains. The presence of associated mud cracks, ripple marks, and rain imprints suggests that the sediments were subjected to alternating periods of wetting and drying. Thus, the Fond du Lac Formation is considered to be a fan-shaped wedge of clastic material that was deposited in a shallow, oxidizing, deltaic environment by a system of streams emerging from a western highland and dispersing material to the east and southeast.

Solor Church Formation

Much of the Keweenawan sequence in southeastern Minnesota consists of red rocks that are lithologically and texturally unlike either the Hinckley Sandstone or the Fond du Lac Formation. Consequently these rocks have been assigned to the Solor Church Formation (Morey, in prep.). So far as is known, no lithologic analog crops out at the surface and the type section is defined as a diamond drill hole in the SW¼SW¼ sec. 14, T. 112 N., R. 21 W. (Webster Township), Rice County, Minnesota, where approximately 1,930 feet of the formation has been penetrated (Morey, in prep.). The bottom of the formation was not encountered at this locality; seismic evidence (Mooney and others, 1970b) indicates that at least 3,200 feet of sedimentary strata overlie a basaltic basement. However, the bottom of the formation was penetrated on top of the Hudson-Afton horst where approximately 50 feet of Solor Church-like strata overlie basalt. Other diamond drill holes in the vicinity are inferred to represent strata of intermediate stratigraphic position.

The Solor Church Formation is characterized by a rather monotonous succession of intercalated sandstone, siltstone, and mudstone. Thin beds of limestone characterized by textural features such as oolites and various kinds of allochems, indurated by a micritic groundmass or sparry cement, exist locally.

Clastic Rocks

Clastic rocks in the Solor Church Formation characteristically are dense and well indurated. Accordingly, it was not possible to determine in detail many of the textural parameters associated with individual grains. However, an examination of grain-size, rounding, and sphericity in approximately 150 thin sections permits some qualitative conclusions. All the samples examined are texturally immature in that they contain more than 5 percent matrix material (fig. V-92A). Framework grains range in size from silt to medium-grained sand, but fine sand is most abundant. However, the detritus cannot be characterized simply as "fine grained" inasmuch as the grain-size distribution is markedly bimodal. Commonly, sand-size grains are embedded in a finer grained groundmass of sand-, silt-, or clay-size detritus. Therefore, variation in the overall grain-size, which is a degree of sorting, of an individual sample the size of a thin section is large.

As in the case of grain-size and sorting, marked variations and a lack of unimodal distribution characterize the

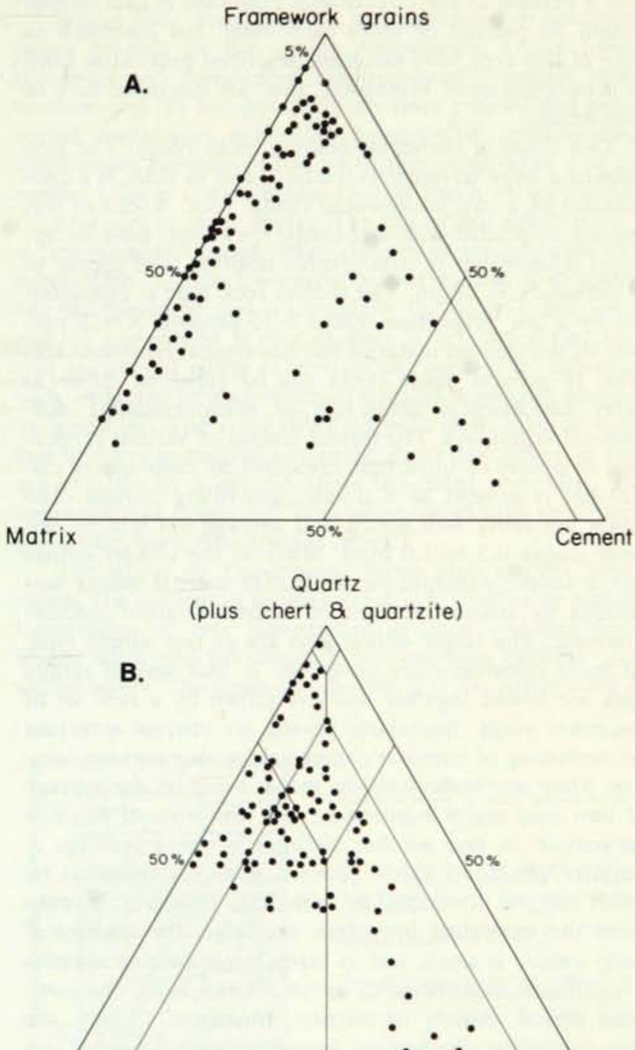

Figure V-92. Textural and compositional diagram of selected Solor Church samples. A, texture (data based on 300 points per section); B, framework grain mineralogy (data based on the classification of 100 framework grains per thin section).

grain shapes in these rocks. In general, rocks having a bimodal size distribution also have a bimodal shape distribution. Commonly, the larger grains are subrounded to moderately well rounded, whereas the smaller grains are angular to subrounded.

Sandstone and siltstone of the Solor Church Formation are mineralogically immature in that they contain appreciable amounts of feldspar and aphanitic igneous rock fragments. Of 150 thin sections examined in detail, approximately 2 percent are quartzite, 31 percent subarkose, 20 percent arkose, 22 percent lithic arkose, 14 percent feldspathic lithic arenite, 10 percent lithic subarkose and one percent lithic arenite (fig. V-92B). Part of the mineralogic variation results from differences in grain size; very fine-grained sandstone and siltstone contain more quartz and less feldspar and rock fragments than do their coarser grained counterparts. However, variations also are in part stratigraphically controlled in that there is an upward trend in the formation toward less labile material.

Hematite is a common cementing agent in the Solor Church Formation and imparts the red color to these rocks. It constitutes less than 1 percent of the rock, and was formed soon after the framework grains were deposited, as indicated by rinds of about the same thickness on all surfaces of the framework grains except at points of contact. Calcite cement is a rare to abundant constituent that occurs as small void-fillings between framework grains and as large irregularly shaped patches. Commonly the calcite in these patchy areas is optically continuous, and, where it occurs associated with hematite, is paragenetically later as indicated by hematite-rimmed framework grains embayed in calcite and by sharp boundaries between the two cementing materials.

Matrix material is a common constituent; it is not distributed homogeneously, but rather occurs either interstitially to the framework grains in small patchy areas or as thin and irregularly dispersed laminae. X-ray diffraction studies show that the matrix consists of mixed-layer illite/montmorillonite and trace amounts of chlorite.

Framework grains are, in order of decreasing abundance, quartz, feldspar, and aphanitic igneous rock fragments. In addition "granitic" rock fragments and irregularly shaped rock fragments having a schistose or slaty fabric are sparingly present in the upper part of the formation, as are sand-size detrital grains of hematite-stained carbonate.

Quartz comprises from 13 to 95 percent of the framework grains, and appears to increase in abundance stratigraphically upward. For example, it averages 84 percent (range 75-95 percent) near the top, 57 percent (range 40-90 percent) through the middle, and 13 percent (range 5-20 percent) near the bottom of the formation. Individual grains range in apparent size from 0.005 to 0.06 mm and have a mean apparent diameter of 0.15 mm. In general, the grains are larger and more rounded in the upper part of the formation. Most of the quartz is monocrystalline and has straight to slightly undulose extinction. Although relatively rare, polycrystalline quartz and/or grains having well developed undulose extinction can be found in the upper part of the formation. Similarly, quartz overgrowths were observed only in the upper part of the formation where they commonly are abraded, suggesting derivation from a preexisting sedimentary source.

Feldspar decreases in abundance stratigraphically upward; it averages 30 percent in the lower, 28 percent in the middle, and 15 percent in the upper part of the formation. All grains are slightly smaller than the accompanying quartz and typically are subangular to subrounded. Most of the feldspar is plagioclase (An_{25-45}) that is strongly altered and characterized by a turbid appearance and embayed edges that result from varying degrees of replacement by the matrix minerals. Microcline and lesser orthoclase are restricted to the uppermost part of the formation, where they are fresh and have sharp outlines in thin section.

Rock fragments are almost entirely lacking (less than 5 percent) in the upper part of the formation, but constitute as much as 40 percent ($\bar{X} = 64$ percent) of the lower part of the formation. Aphanitic igneous rock fragments consisting of lath-shaped plagioclase crystals in a fine-grained, nearly opaque groundmass comprise the bulk of the rock fragments. Their actual compositions have not been determined, but an absence of quartz, an apparent plagioclase composition of An_{40-50} and a dark groundmass suggest that they are rocks of basaltic affinity. In general, most of the rock fragments are only slightly smaller than the accompanying quartz and feldspar grains; an exception is the basal part of the formation where they are large and dominate the samples; many smaller detrital grains of epidote, pyroxene, hornblende, and various kinds of zeolites more or less fill the interstitial voids.

Chert and quartzite fragments having a simple mosaic fabric are ubiquitous regardless of stratigraphic position; however, they comprise only 1-2 percent of the framework grains. "Granitic" rock fragments, consisting of interlocking quartz and potassium feldspar grains and several kinds of metasedimentary rock fragments, are sparingly present in the upper part of the formation.

Sand-size detrital grains of calcite, although not common, occur in several zones near the top of the formation where, in general, they are spatially related to intercalated beds of limestone. The detrital grains consist of either clear sparry calcite crystals or micritic aggregates and have an apparent average diameter of around 0.2 mm. They are moderately rounded and have a moderate to high sphericity. In one sample, sand-size detrital grains consisting of broken oolites also were observed.

Other detrital material includes large grains of colorless muscovite and pale-green chlorite. Both occur as elongate flakes oriented subparallel to bedding. Chlorite also occurs as a crystal aggregate in subrounded fine sand- to silt-size grains. Well rounded to euhedral grains of magnetite commonly are disseminated throughout the red parts of all the samples examined, whereas magnetite is almost entirely lacking in the green parts. Tourmaline, zircon, and apatite are common and ubiquitous accessory mineral constituents.

Mudstone in the Solor Church Formation has either a laminated or massive fabric. Many mudstone units are fissile and therefore are true shales. The presence or absence of parting planes is a function of amount and distribution of the silt-size fraction; fissile and laminated rocks contain small amounts of silt concentrated in discontinuous layers less than a millimeter thick. Structureless beds, however, are very silty and the silt-size detritus is generally randomly distributed; these rocks commonly break with a flat-conchoidal fracture.

X-ray diffraction studies show that the clay-size fraction is predominantly illite and mixed-layer illite/montmorillonite, with trace amounts of chlorite. Hematite is a common constituent, imparting a deep red color to many beds, and calcite is locally very abundant.

Carbonate Rocks

Rocks containing more than 50 percent carbonate comprise a small proportion of the Solor Church Formation (only 9 percent of the 150 samples examined in thin section contain 50 percent or more carbonate); but inasmuch as rocks of this type have not been described previously from the Keweenawan of Minnesota, they are discussed here in some detail.

Two kinds of carbonate-bearing rocks occur. The first contains a large terrigenous fraction, and as such, is a continuation of a calcite-cemented clastic rock. Rocks of this type are associated with any clastic rock type, have no apparent relationship to stratigraphic position, and appear to be secondary in origin. The second rock type is characterized by a low proportion (range 5-30 percent; $\bar{X} = 10$ percent) of terrigenous material and has typical limestone textures. In general, these rocks can be classified either as sparry allochemical limestones or microcrystalline allochemical limestones. The former consist of various proportions of oolites or intraclasts cemented by clear sparry calcite that is present as a simple pore-filling cement. The oolites are fairly well sorted and average 0.6 mm in diameter (range 0.3 to 1.0 mm). Many of the smaller oolites have a simple structure—a nucleus of detrital quartz surrounded by concentric rims of hematite-stained calcium carbonate. The larger oolites also are of this simple type, but more commonly are composite in that several simple types are linked together and overgrown by a new set of concentric rings. Intraclasts having no internal structure and consisting of microcrystalline calcite also are very common. They are "roller-like" in shape, being on the average 2.3 mm long and 0.8 mm wide, and are oriented more or less parallel to one another, defining a crude bedding. A complete gradation exists between a rock dominated by oolites and one dominated by intraclasts. However, in rocks where the individual intraclasts are large, the amount of sparry calcite is small, and in many intraclast-rich samples it is difficult to distinguish a true micrite from one composed almost entirely of micritic intraclasts. Clearly, the microcrystalline allochemical limestones were deposited in a less energetic environment than were the sparry allochemical limestones.

Sedimentologic Framework

Both the Fond du Lac and Solor Church Formations contain a variety of sedimentary structures which, when taken together, are thought to indicate alluvial deposition. The most prominent attribute in both formations consists of alternations or cycles of coarse- and fine-grained groups of beds. Each cycle consists of a sequential arrangement of beds in which the grain size generally becomes finer upward; thus, each cycle generally is divisible into a coarse-grained facies below and a fine-grained facies above. Similar cycles have been called "fining-upward cycles" (Allen, 1963). Each cycle is defined by two criteria: (1) the presence of an erosional surface at the base, and (2) a fine-upward appearance defined by an upward reduction in grain size and/or a decrease in the level of hydraulic energy as indicated by the sedimentary structures. As defined, a complete cycle ranges in thickness from 10 to 50 feet. However, only a small number of cycles are complete in that all the attributes of an ideal fining-upward cycle are present (fig. V-93).

Where fully developed, the coarse-grained or sandstone facies consists of two major components: (A) a lower, coarse-grained component that commonly appears structureless, and (B) an upper, slightly finer grained and better sorted component exhibiting festoon-type cross-bedding. Each major component can be further subdivided. For example, a basal conglomerate (A_1) as much as several feet thick occurs in many cycles. The top of this subdivision is structureless and passes either abruptly or gradationally into subdivision A_2. The lowermost part of A_2 also is structureless but the upper part is irregularly flat-bedded. The lower part of component B (B_1) consists of cross-stratified beds that are present either as continuous cosets or as individual beds separated by thin layers of parallel laminated siltstone or mudstone. The cross-stratified beds in B_1 may be overlain by wavy to parallel laminated strata (B_2), which in turn are overlain by a current-ripple laminated layer (B_3). The appearance of ripple-drift laminations marks the uppermost part of the lower fine-grained facies.

The upper or fine-grained facies also can be subdivided. The lowermost subdivision (C_1) is typified by parallel laminations of very fine-grained siltstone and mudstone. Minor scour and fill structures are common between individual layers. The upper subdivision (C_2) has either wavy or

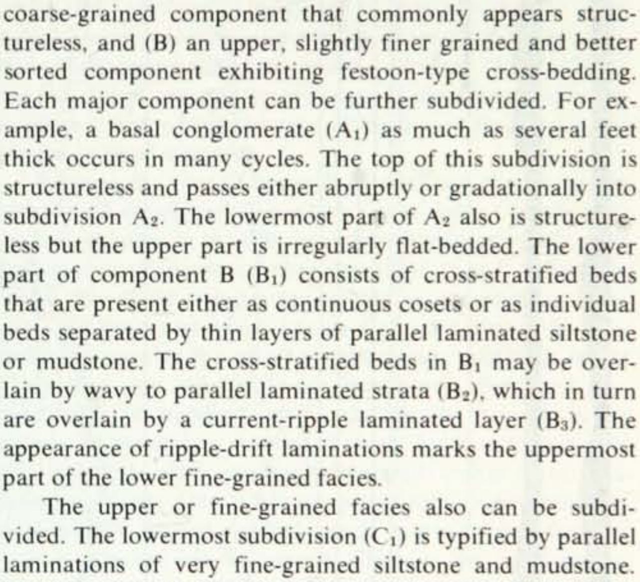

Figure V-93. Idealized minor cycle of "fining upward cycle" in the Solor Church Formation; see text for a discussion of symbols.

parallel laminations and individual siltstone layers may be current-ripple laminated or less commonly cross-laminated.

A second kind of cyclic repetition of lithologic types has been recognized in the Solor Church Formation and has been referred to as "major cycles" (Morey, in prep.) in order to emphasize the repetitive nature of lithologic types over stratigraphic thicknesses of as much as 600 feet. Each major cycle is characterized by a gradually increasing proportion of mudstone relative to siltstone and sandstone and each is terminated by an abrupt decrease over a short stratigraphic distance in the relative proportion of mudstone. These major cycles are particularly well developed in drill core from the type locality of the Solor Church Formation. Because most beds in the Solor Church Formation are relatively thin, all the detailed lithologic interrelationships cannot be graphically portrayed. Therefore, gross lithologic changes relative to stratigraphic position are shown by a moving average plot of the relative percentage of sandstone, siltstone, mudstone and/or shale, and limestone (fig. V-94). This plot was prepared by calculating the percentage of each lithology present within a 50-foot-thick section; additional percentages were determined in succeeding 50-foot sections that overlapped the preceding section by 25 feet. Each set of average values was then plotted at the center point of its particular section and a curve smoothed through these values.

The lower part of each major cycle is characterized by repetitive sandstone beds with little if any intervening shale and mudstone. Commonly these sandstones are coarse grained and exhibit both festoon-type cross-stratification and flat bedding. Many beds have sharp erosional boundaries, contain intraformational fragments, and are vaguely graded. Siltstones, where present, are thin and apparently discontinuous. Most of the characteristics described above typify the coarse-grained facies of the fining-upward cycles. However, the complexity of these deposits, the frequency of erosive contacts, and other evidences of reworking suggest a complex of river channels, which possibly were braided (Smith, 1970).

In contrast, the mudstone/shale-rich upper part of each major cycle contains thin beds of fine-grained siltstone and limestone, and the overall sedimentologic aspects of these rocks suggest that they were deposited in an environment where alluvial activity was minimal. However, the presence of oolitic limestones indicates that local shoal areas in which agitated conditions existed were at least fairly common.

Each major cycle is characterized by several other attributes: (1) fining-upward cycles are well developed in the sandstone-abundant part of the cycle and consist of a lower coarse-grained facies that is thicker and better developed than the upper fine-grained facies. However, the abundance of well-defined fining-upward cycles decreases as the proportion of mudstone increases and, where developed near the top of a major cycle, the fining-upward cycles consist of a very thin sandstone facies and a thick fine-grained siltstone facies. (2) In each major cycle, the average grain size and amount of the framework detritus relative to the amount of matrix material decrease as the sandstone proportion decreases. (3) Ordering in the structure of mixed-layer montmorillonite/illite inversely follows the sandstone

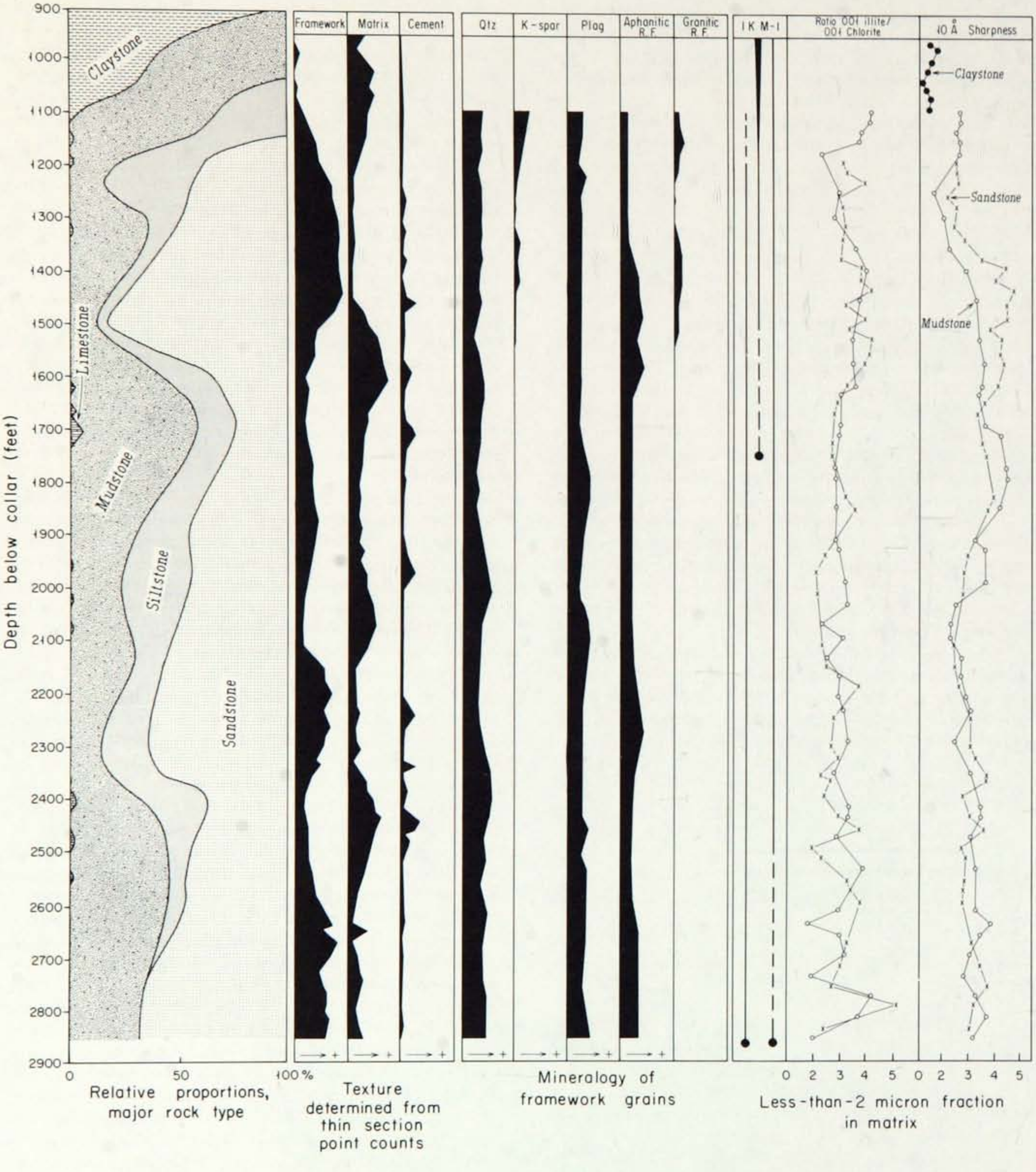

Figure V-94. Gross lithology and mineralogy of the Solor Church Formation as exemplified by a drill core at the type locality.

abundance trend; that is, as sandstone becomes more abundant in the stratigraphic section, ordering of the illite/montmorillonite structure decreases. This observation needs some elaboration inasmuch as there is no immediately apparent reason why ordering of the illite/montmorillonite structure should correlate directly with sandstone distribution. Weaver and others (1961) have suggested that the symmetry of the illite peak—as reflected by a sharpness ratio or the ratio of the 001-peak height at 10 Å to the peak height at 10.5 Å—reflects the degree of metamorphism. However, Hower and Mowatt (1966) have suggested that the tailing or lack of sharpness of the 001 illite peak ". . . appears to be an intrinsic feature of illites which in addition to being disordered are potassium deficient and have a high H_2O+ content as compared with true dioctahedral micas. . . ." Perry and Hower (1970) later suggested that where the geothermal gradient is highest—due to deep burial—a decrease in expandability of randomly interstratified illite/montmorillonite results, whereas Burst (1969) suggested that overburden pressures in places was sufficient to squeeze water from the expandable layers, thus leading to a more ordered structure.

Because the overall mineralogy does not change appreciably with stratigraphic position in the Solor Church Formation, changes in the illite/montmorillonite structure must have occurred sometime after deposition, and it is here inferred that these changes are related to compaction; however, the degree to which compaction took place was controlled in part by the original sedimentary fabric. It seems likely that clay minerals with their c-axis directions perpendicular to bedding would statistically have a better chance of becoming ordered through loading than would clay minerals with randomly oriented c-axis directions. However, Grim (1953) has suggested that the presence of isometrically-shaped detritus—such as quartz and feldspar—in a sediment has an adverse effect on the development of a fabric with a preferred clay mineral orientation. Thus, the fabric would exert a strong influence on the development of an ordered structure, and because the mudstones in the sandstone-rich part of each major cycle are a heterogeneous mixture of grain sizes, the clay minerals would be expected to have a more random orientation and thus a disordered structure. This hypothesis in part is confirmed by the presence of abundant shale—in which the fissility reflects a well ordered fabric—in the mudstone-rich parts of each major cycle.

ENVIRONMENTAL SYNTHESIS

Mode of Transport and Environment of Deposition

The sedimentologic features described above strongly imply that sedimentation of both the Fond du Lac and Solor Church Formations took place on an extensive flood plain that most likely consisted of several distributaries radiating outward from the mouths of major streams. Each fining-upward cycle records the creation, infilling, and eventual abandonment of an individual stream channel. The basal sandstones represent a coarse residuum or lag deposit concentrated during continuous stream action. The overlying cross-stratified and fining-upward division records the growth of a point bar complex (Harms and Fahnestock, 1965; Visher, 1965), and commonly shows an upward decrease in scale of cross-stratification. The scoured surface at the bottom of each cycle may indicate either lateral migration of a channel into an area previously characterized by overbank deposition, or a sudden change of channel course after the growth of relatively great alluvial relief by building up some previous channel belt (Allen and Friend, 1968, p. 53).

In the Solor Church Formation, the general trend toward a decreasing proportion of the coarse-grained facies relative to the fine-grained facies of each fining-upward cycle stratigraphically upward defines a major cycle. The bottom of each major cycle is characterized by non-cyclic sandstone deposits and by minor cycles that have abundant sandstone relative to mudstone and/or shale. This arrangement suggests that the start of each major cycle was characterized by deposition of sandstone from meandering or braided streams. Sedimentary structures such as cross-bedding, scour and fill, and flat bedding denote intensive reworking of the sediment in a complex of continuously shifting banks, pools, and channels. Thus, the beginning of each major cycle records the accumulation of channel deposits to the exclusion of those of the overbank environment, and these deposits are inferred to be the result of deposition by rivers of low sinuosity whose channels were unrestricted in their lateral wandering. However, sedimentation most likely outpaced subsidence of the flood plain and as deposition continued, the stream gradient decreased. Reduction in the stream gradient resulted in a steady decline in stream velocity and this in turn presumably resulted in more pronounced meanders and in low velocity flow in channels choked with shoals. As the meanders became more incised, there was a general decrease in the frequency of channel deposits and an increase in the amount of fine-grained sediment being deposited at any one place.

Although the fine-grained upper parts of each fining-upward cycle can be regarded as a top-stratum deposit, it is difficult to decide whether they represent genuine overbank flood deposits or whether they were deposited in the final stages of filling of an abandoned stream course, perhaps a meander cutoff or an oxbow lake. The presence of ripple-drift lamination favors the second alternative inasmuch as such features generally imply a permanently inundated environment with minor wave action. Furthermore, Wolman and Leopold (1937, p. 97) concluded that overbank deposits from individual channels comprise a small proportion of modern flood plain deposits. Thus, it seems likely that the mudstone-rich component of each fining-upward cycle represents sediments that originated from a possibly large number of channels of different ages and with differing courses across the alluvial plain.

It is concluded that most of the fine-grained rocks were deposited on mud flats and in shallow oxbow lakes in which clays and silts could settle under quiet conditions. The redness of the fine-grained rocks is interpreted as having developed in place by oxidation of a pre-existing iron-bearing mineral and not as a result of derivation from a lateritic source area. Therefore, the environment of the interchannel areas was oxidizing. The limestones associated with the siltstone and mudstone indicate the presence of a number of semi-permanent bodies of water from which calcium carbonate precipitated, perhaps under semi-arid conditions. Thus, after an alluvial plain had been abandoned by the rivers, it was rapidly invaded by other alluvial environments as the site was converted under subsidence once more into a low-lying area where little deposition took place.

Composition of Source Area

The petrographic data summarized above imply that the Keweenawan detritus was derived from several terranes characterized by (1) extrusive volcanic rocks of basic to intermediate composition, (2) plutonic and metamorphic rocks of "granitic" composition, and (3) older and contemporaneous sedimentary rocks (fig. V-95).

The Solor Church Formation overlies amygdaloidal basalt and basalt porphyry and the lower several feet of the formation contains pebble- and cobble-size clasts of these extrusive igneous rocks. Sandstones that are intercalated with, and transitionally overlie, the conglomerate consist almost entirely of aphanitic igneous rock fragments and intermediate to calcic plagioclase. In addition, angular to subrounded grains of pyroxene, hornblende, epidote and various kinds of zeolites are present in fairly abundant quantities. In appearance, the hornblende and pyroxene resemble phenocrysts, whereas other minerals are similar to amygdules in the extrusive rocks.

Much of the Solor Church Formation was derived from a basaltic terrane inasmuch as sand-size rock fragments having an aphanitic, amygdaloidal, or microdiabasic texture are everywhere abundant. However, the presence of quartz suggests that the source area did not consist entirely of basalt. The actual source of the quartz-bearing material is not known, but possibly was the Middle Precambrian metamorphic rocks now exposed in east-central Minnesota. Furthermore, the presence of sodic plagioclase, potassium feldspar, granitic rock fragments consisting of interlocking grains of quartz and feldspar, and polycrystalline quartz attests to an igneous or metamorphic terrane as a source toward the end of Solor Church deposition.

The Fond du Lac Formation contains detritus indicative of a mixed terrane consisting of a distal source—represented by the sand-size detritus—composed of igneous and metasedimentary components, and a proximal source—represented by the pebble- and cobble-size detritus—composed of aphanitic igneous rocks and sedimentary rocks lithologically like those in the Solor Church Formation. The degree to which the sand-size detritus in the Fond du Lac Formation is second-cycle sand cannot be completely evaluated. However, it is unlikely that the Solor Church Formation contributed an appreciable quantity of reworked sand-size detritus to the Fond du Lac sediments for several reasons: (1) the grain size of the Fond du Lac detritus, although bimodal in distribution, is on the average larger than that of the Solor Church Formation; (2) only minor amounts of plutonic igneous rock fragments occur in the Solor Church

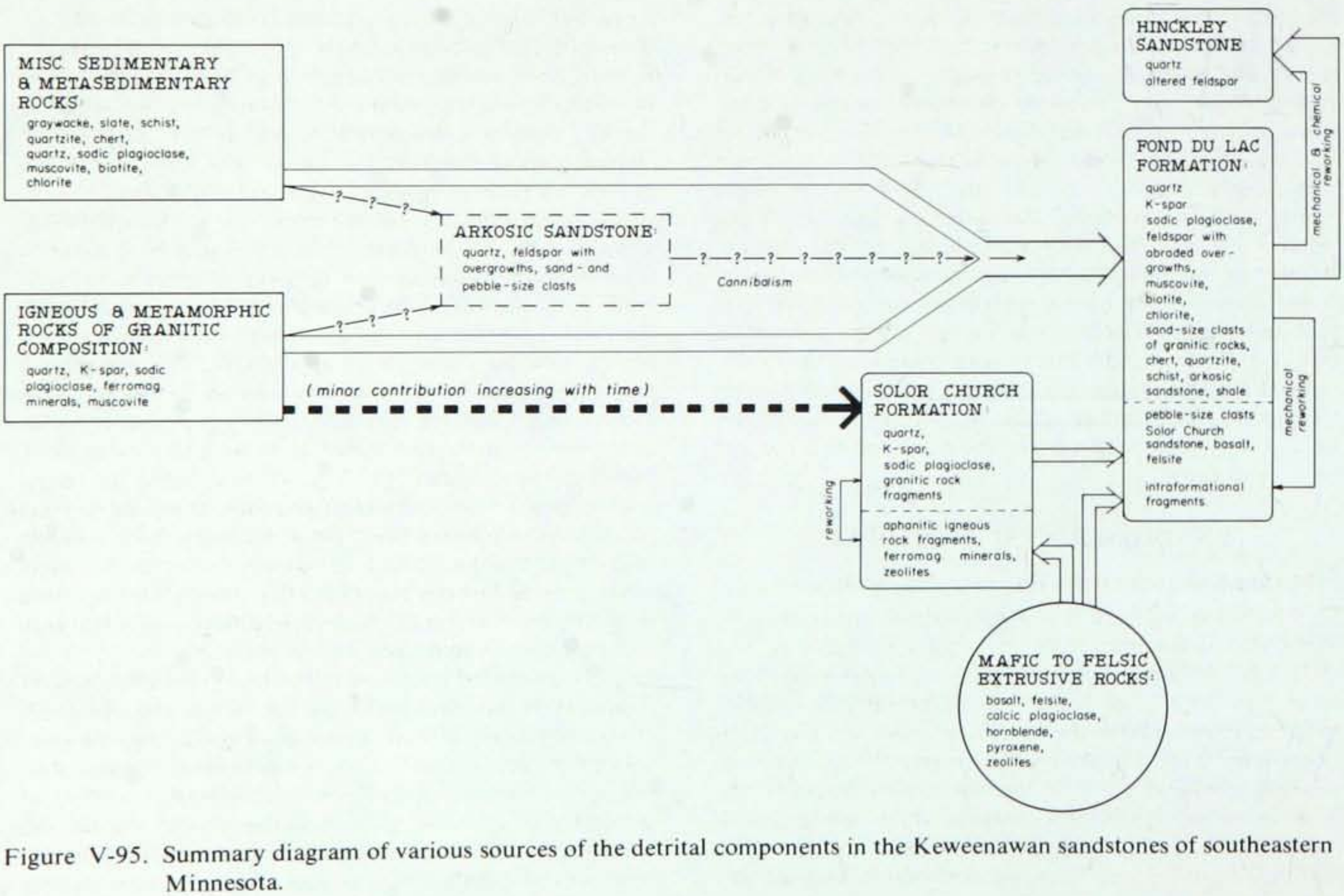

Figure V-95. Summary diagram of various sources of the detrital components in the Keweenawan sandstones of southeastern Minnesota.

Formation, whereas they comprise an appreciable component in the Fond du Lac Formation. Therefore, it seems likely that reworked sedimentary detritus in the Fond du Lac Formation was derived by cannibalism from previously deposited Fond du Lac-like material.

The source of the Hinckley detritus is even more difficult to determine. However, the physical and optical characteristics of the accessory minerals in the Hinckley Sandstone are very similar to those of the Fond du Lac Formation except that the less stable components—such as apatite, garnet, and leucoxene aggregates—are absent or rare (Tyler and others, 1940). This observation, together with the presence of trace amounts of highly altered feldspar and the apparently gradational contact with the Fond du Lac Formation, suggests that the Hinckley detritus was derived, at least in part, by chemical and mechanical reworking of Fond du Lac detritus.

The change in composition of the source area as reflected by a change from aphanitic to granitic components may have resulted from either (1) a gradual change in the geographic location and hence the composition of the source area, or (2) downcutting through a sequence of extrusive volcanic rocks into an underlying older "granitic" terrane. Inasmuch as data indicative of the sediment transport direction cannot be obtained from unoriented diamond drill cores, it is not possible to completely evaluate either possibility at this time. However, geophysical data (Craddock and others, 1963; Mooney and others, 1970a) indicate that crystalline plutonic rocks occur beneath the basalts throughout this general area. Therefore, it seems likely that the change in the relative proportion of the various kinds of rock fragments reflects a "reversed stratigraphic" effect. Thus, downcutting in the source area exposed successively older rocks in such a way that the derived sediments contain a greater proportion of older material toward the top of the vertical succession. It can be further inferred that ultimately the rate of subsidence did not keep pace with the rate of infilling and that the granitic source area became buried by its own debris; most likely, this debris was reworked several times before it was finally deposited. Such reworking would lead to the appearance of recycled feldspar in the Fond du Lac Formation and ultimately to a texturally and mineralogically mature sediment like that now found in the Hinckley Sandstone.

SUMMARY OF GEOLOGIC HISTORY

The St. Croix horst comprises the northern part of the Midcontinent Gravity High, the major tectonic feature of the Precambrian terrane in the Midcontinent region. It cuts discordantly across a prevailing east-west-trending fabric of the older Precambrian rocks, and King and Zietz (1971) have suggested that the entire feature is a major rift composed of a series of short, linear, echelon segments with offsets similar to transform faults characteristic of the present mid-ocean rift system. They concluded (p. 2204) that ". . . although this midcontinent rift did not develop into a new ocean . . ." perhaps because ". . . it did not persist as long in geologic time, or, more probably, the crust in this area may or may not have been free to move apart except in a very limited way at that time. Instead of a broad, thin layer, the upwelling mafic material formed a thick, layered block with marginal basins that filled with clastic material, as suggested by Lidiak (1964). . . ." Hinze and others (1971, p. 29) have suggested that this and other rifts were initiated by plate splitting and the subsequent upwelling along the base of the crust of low velocity-layer material derived from the upper mantle. They further postulated that ". . . uplift of the earth's surface and igneous activity along pre-existing zones of weakness is associated with the vertical rise of this material, while lateral movement results in thinning and rupture of the crust producing extensional rift grabens. . . . The convergent movement of adjacent mantle material into the void . . . places the crust under compression, thus accounting for . . . the subsequent development of sedimentary basins over rift zones."

The fact that several similar but distinct units comprise the structure underlying the Midcontinent Gravity High indicates that the evolutionary pattern of the structure may have been somewhat variable from place to place. Thus, even though it is not possible to outline a history for the entire Midcontinent Gravity High, it is possible to recognize in southeastern Minnesota most of the tectonic events suggested above. The first, or igneous phase, is recorded in the 20,000 feet of basalt and associated rocks that were extruded during the early stage of Keweenawan time (fig. V-96A). A hiatus of unknown duration separated the volcanism and clastic deposition, inasmuch as the Solor Church Formation is thickest in a depression on top of the St. Croix horst, indicating that sinking of the central part of the basalt mass continued after volcanism ceased (fig. V-96B). The magnetic pattern associated with the depression indicates that it in part may be fault-bounded and graben-like in shape (King and Zietz, 1971). However, the presence of Solor Church material on what are now the flanks of the St. Croix horst indicates that deposition was not always restricted to this basin, but once was much more widespread.

Although there is ample evidence that the Solor Church Formation was deposited by alluvial processes, the location of the source area and the direction of stream transport are unknown. The streams first had headwaters in areas underlain by volcanic rocks, but it is inferred that erosion and downcutting of the streams ultimately exposed an underlying pre-Keweenawan "granitic" terrane. Thus, a direction of stream transport perpendicular to the axis of the basin is inferred.

Variations in the sedimentary aspects of the fining-upward cycles within each major cycle of the Solor Church Formation can be attributed to fluctuating energy gradients within the depositional regime itself; however, the repetitive nature of the major cycles cannot be explained in this way. The overall coarse to fine grain size of the major cycles suggests a progressive change toward a quasi-equilibrium balance condition between rate of subsidence and rate of infilling. The sudden reappearance of abundant sandstone at the start of a new major cycle indicates an increase in current velocity coupled with greater competence, a decrease in deposition of overbank deposits, and a reduction in the sinuosity of individual channels. Consequently, channels deepened and flood plains received little or no sediment

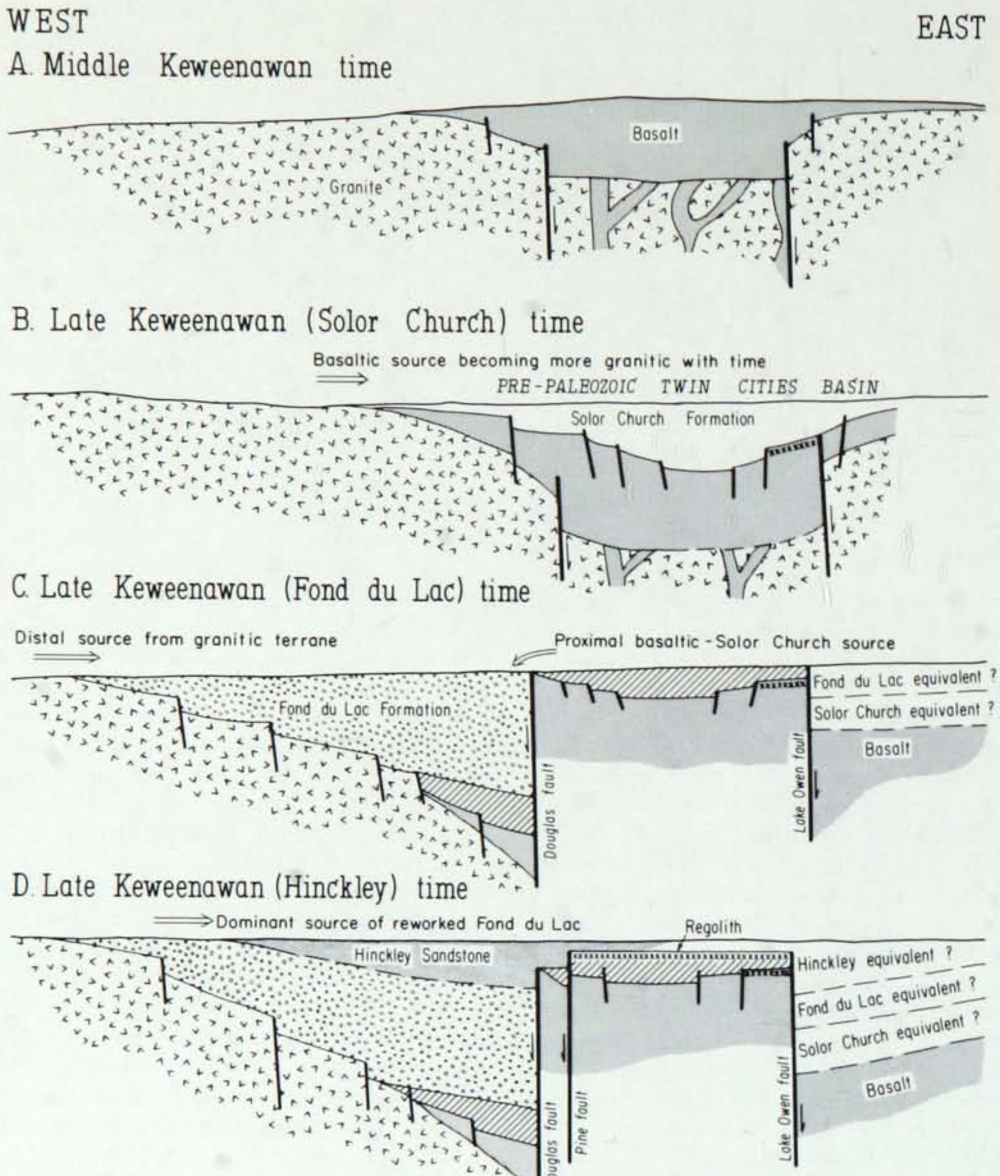

Figure V-96. Schematic east-west cross-sections showing the inferred evolution of the St. Croix horst from Middle Keweenawan (?) to approximately Middle Ordovician time. A, Middle Keweenawan time; B, Solor Church time; C, Fond du Lac time; D, Hinckley time; and E, Early Paleozoic time.

and were progressively eroded as new distributary channels developed.

Because the origin of the Solor Church Formation must be closely related to the development of the rift structure, it is inferred that each major cycle reflects episodic movements associated with the development of that structure. Each period of movement was followed by a period of relative quiescence during which alluvial deposition again progressed toward a quasi-equilibrium state.

The relationship between the Solor Church Formation, the Fond du Lac Formation, and the Hinckley Sandstone is complex and strongly dependent on the present geometry of the St. Croix horst. West of the Douglas fault, the contact between the Fond du Lac Formation and the overlying Hinckley Sandstone clearly is gradational, and cross-bedding data indicate that both formations had a dominant sediment transport direction from west to east (Morey, 1967a; Tryhorn and Ojakangas, this chapter). However, the contact between the Solor Church and Fond du Lac Formations is not well defined. The gradual increase of the "granitic" rock component in the Solor Church Formation toward values like those found in the Fond du Lac Formation indicates that the source area did not change appreciably. However, the presence of pebbles and cobbles of Solor Church-like material in the Fond du Lac Formation implies that lithification of the Solor Church Formation occurred prior to Fond du Lac deposition. Thus, whereas the source area did not change, there probably was a time break of considerable duration between the deposition of the two formations.

The wedge shape of the Fond du Lac Formation implies deposition in a half-graben-like basin bounded on the east by the St. Croix horst and on the west by an older granitic terrane (fig. V-96C). It is inferred that both areas contributed detritus to the Fond du Lac Formation—the horst being the proximal source and the granitic terrane the distal source—suggesting that the horst stood as a positive area during the time of Fond du Lac deposition. However, the lack of fanglomeratic material like that found in the Triassic basins of eastern North America suggests that fault scarps of considerable extent were not formed. Rather, it appears that subsidence more or less kept pace with infilling and that only periodically were conglomerates introduced into the basin. The lack of volcanic and Solor Church detritus in the Hinckley Sandstone, and the presence, at least locally, of sandstone on top of the horst, suggest that the horst was not a source of clastic detritus during the time of Hinckley deposition (fig. V-96D). However, the abnormally thin and irregular nature of the Hinckley Sandstone on top of the horst, and the presence in east-central Minnesota of basalts in juxtaposition with the Hinckley Sandstone (Grout and others, 1951; Sims, 1970) indicate that movement and concurrent erosion occurred on the horst after the end of Hinckley deposition (fig. V-96E). Many of the structures in the overlying Paleozoic rocks occur over inferred structures in the Keweenawan rocks (fig. V-96F). Among these structural features are the Hollandale embayment, the Twin City basin, the Hudson-Afton anticline (Thiel and Schwartz, 1941; Craddock and others, 1963) and the Vermillion anticline (Morey and Rensink, 1969). All these structures may be accounted for by minor rejuvenation of major faults in the Precambrian rocks and the effects of these renewed movements on the overlying Paleozoic strata.

CONCLUSIONS

The Keweenawan rocks in the subsurface of southeastern Minnesota comprise part of what has been called the St. Croix horst, a feature named by Craddock and others (1963, p. 6015) and described as a ". . . block . . . elevated thousands of feet, mainly in late Precambrian time. . . ." However, the petrographic data summarized above clearly place restraints on such a model. Rather than representing flows and sedimentary rocks that accumulated in a gradually sinking trough which was later uplifted along several boundary faults, the St. Croix horst appears to have stood as a positive area that supplied sediments to the flanking basins. Thus the "St. Croix horst" is not a horst in the sense that it is a "block that has been uplifted relative to the rocks on either side," but rather is a structure in the sense the term horst was first used—"a mass of earth-crust which is limited by faults and which stands in relief with respect to its surroundings" (Am. Geol. Instit., 1966, p. 140).

THE SIOUX QUARTZITE, SOUTHWESTERN MINNESOTA

George S. Austin

The Sioux Quartzite crops out at several localities in southwestern Minnesota, southeastern South Dakota, and extreme northwestern Iowa (fig. V-97). Drill hole and geophysical data have shown further that the formation occurs in the subsurface over a large area in southwestern Minnesota. In Minnesota, it is inferred to unconformably overlie rocks of Early and Middle Precambrian age and in turn is overlain by Cretaceous strata and Pleistocene drift; hence its age is imprecisely known. Goldich and others (1961) obtained a K-Ar age of 1,200 m.y. for intercalated sericitic argillite near Pipestone, Minnesota, but stated that this date probably represents a time of folding. Goldich and others (1966) argued that the time of deposition of the Sioux can be given only within the limits of 1,700-1,200 m.y. However, a recent well drilled at Hull, Iowa encountered alternating layers of rhyolite and quartzite. The rhyolite has an apparent Rb-Sr age of $1{,}470 \pm 50$ m.y. which, because of slight alteration, is regarded as a minimum age (Lidiak, 1971). The rhyolite bodies may represent flows, but they also may be intrusive sill-like bodies that are not temporally related to Sioux deposition. With either interpretation, however, the radiometric age indicates that the Sioux Quartzite was deposited at least 1,470 m.y. ago. Therefore the Sioux Quartzite is now assigned to the Late Precambrian, and it is tentatively correlated with other similar-appearing quartzites in the Upper Midwest (fig. V-1).

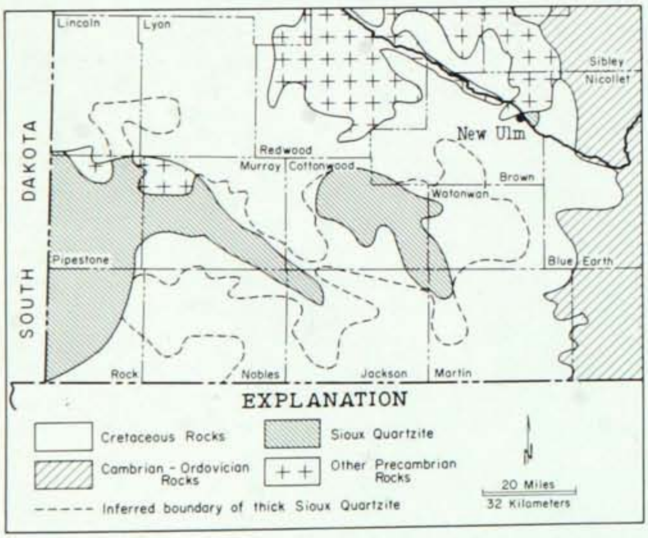

Figure V-97. Generalized geologic map of southwestern Minnesota.

AREAS OF OUTCROP

The Sioux Quartzite crops out in three areas in Minnesota—in northern Rock and southern Pipestone Counties, in the extreme southwestern part of the state; in northern Cottonwood, southwestern Brown, and western Watonwan Counties; and in Nicollet County, along the Minnesota River east of New Ulm.

In each of the outcrop areas of Sioux Quartzite the dominant structure is that of a gentle basin approximately 10 to 20 miles long and about half as wide (Baldwin, 1951, unpub. Ph.D. thesis, Columbia Univ.). Because of the paucity of exposures, it is not known whether the basins are depositional or tectonic. There is some evidence that faults may be present but none have been observed. Three dominant joint sets have been measured in the outcrop areas of the Sioux: N.75-35°W., N.10° W. to N.15°E., and N.50-70°E. Baldwin (1951, *op. cit.*) indicated that the joints are independent of the minor warps present on the limbs of the several basins.

Rock and Pipestone Counties

In Rock and Pipestone Counties of southwestern Minnesota, the Sioux Quartzite forms a structural basin, which Baldwin (1951, *op. cit.*) named the "Rock County structural basin" (fig. V-98). The longer axis trends north-northwest and the center of the basin is about two miles southeast of Jasper, Minnesota. Dips toward the center of the basin are generally less than 10°, but along the southwestern limb of the structure dips range from 7° to 20°. It is not possible to observe all beds at any single locality, and the sequence determined by Baldwin (1951, *op. cit.*) was pieced together from the scattered outcrops. About 3,150 feet of beds are present below the base of a distinctive conglomerate that Baldwin identified as the third or "upper conglomerate," which is 180 feet thick. About 2,100 feet of Sioux Quartzite is present above the upper conglomerate. Accordingly, within the Rock County structural basin the Sioux is at least a mile thick; neither the top nor the bottom is exposed. If a conglomerate cropping out along the Minnesota-South Dakota border southwest of Pipestone, Minnesota is a basal conglomerate as proposed by Baldwin (1951, *op. cit.*), the Sioux Quartzite in the Rock County structural basin is at least a mile and a half thick.

Cottonwood County

In the northern part of Cottonwood County and adjacent parts of Brown and Watonwan Counties (fig. V-99), the Sioux Quartzite crops out along an eastward-trending belt as much as 3 miles wide and more than 20 miles long. The exposures are part of the northern and western limbs

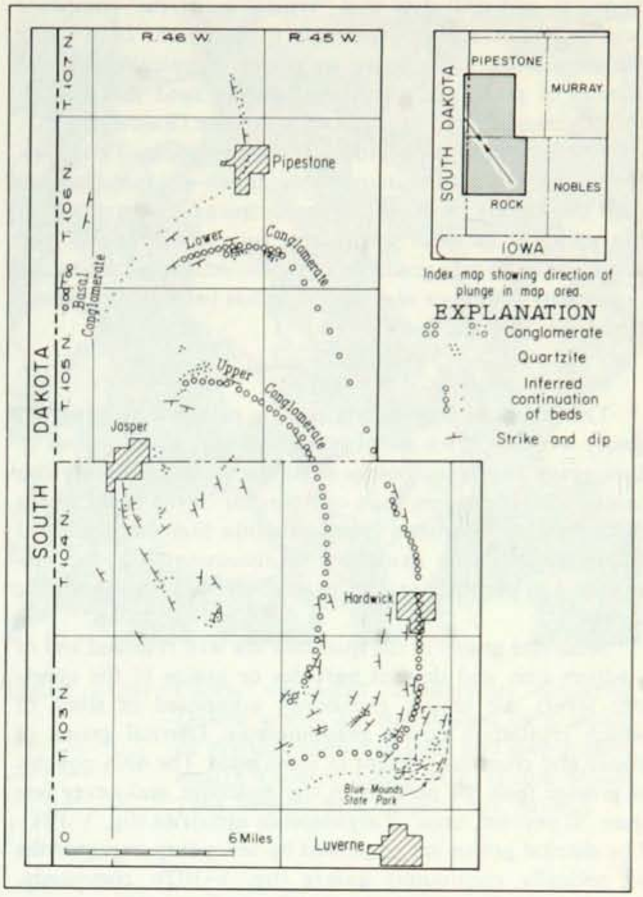

Figure V-98. Outcrop and structure of the Sioux Quartzite in the Rock County structural basin in extreme southwestern Minnesota. (After Baldwin, 1951, unpub. Ph.D. thesis, Columbia Univ.)

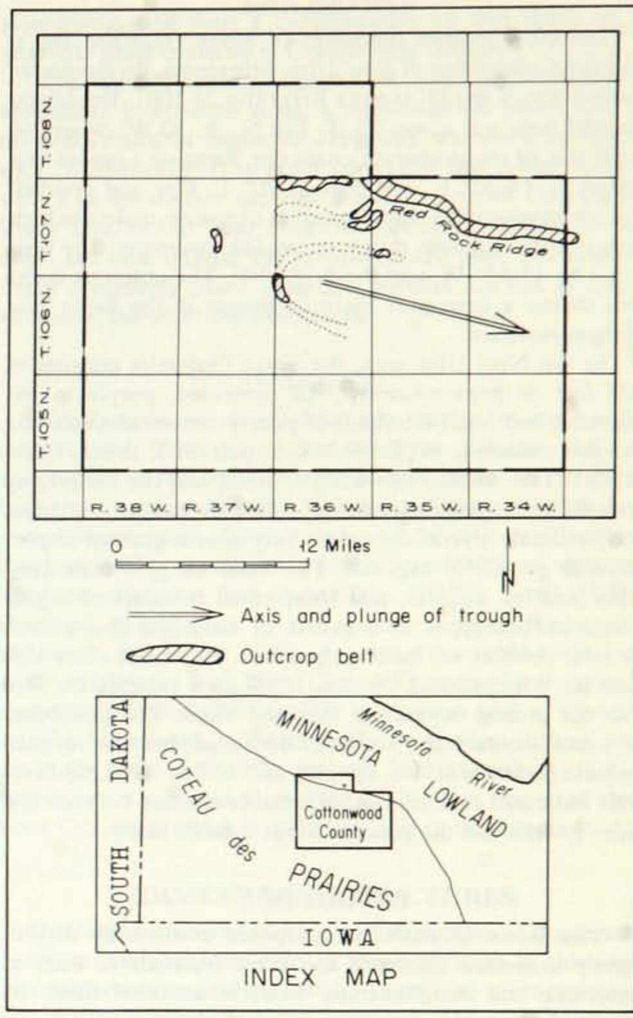

Figure V-99. Sioux Quartzite outcrop belt in Cottonwood, Brown, and Watonwan Counties (after Baldwin, 1951, unpub. Ph.D. thesis, Columbia Univ.).

of an elongate basin that plunges gently east or southeast (Baldwin, 1951, *op. cit.*). Dips range from 14° in the northwesternmost exposures to 3° in the east-southeasternmost ones. The Sioux dips about 6° S. on the north limb, and forms the leading edge of the Coteau des Prairies, a highland plateau occupying the southwestern corner of Minnesota above the lowland near the Minnesota River.

A sharp flexure in the outcrop belt attributed to either a fold or a fault striking northwestward separates the Sioux into two units (Baldwin, 1951, *op. cit.*). Outcrops are too sparse to determine the amount of dislocation on the structure, and it is not known whether the beds on opposite sides of the dislocation are stratigraphically equivalent. On the east side of the dislocation, some 1,500 feet of quartzite is exposed, a part of which is argillaceous. Many layers contain scattered coarser grains and in the upper 1,000 feet of the section the coarser beds contain fragments of mudstone. West of the dislocation, some 850 feet of generally medium-grained quartzite is poorly exposed. The lower 500 feet has some coarser layers, at least one of which contains mudstone particles. Baldwin (1951, *op. cit.*, p. 38) offered two possibilities for correlation: ". . . either the beds are offset less than a mile, with a stratigraphic displacement of less than 500 feet, and the beds are essentially equivalent, or else the offset is as much as 4 miles, with a stratigraphic displacement of 2,000 feet or more. . . . The latter possibility would mean that about 3,000 or 4,000 feet of beds are exposed in Cottonwood County; the former that about 2,000 feet of beds are exposed here. In any case no maximum thickness can be computed for the area, because neither the top nor bottom of the formation is seen." However, geophysical data indicate that the Sioux is thin or absent at the foot of the coteau, north and east of the outcrop belt (Austin and others, 1970).

New Ulm Area

The most eastern exposures of Sioux Quartzite are in Nicollet County east of New Ulm, Minnesota, on the northeastern side of the Minnesota River (fig. V-100). The lowest exposed beds are in sec. 27, T. 110 N., R. 30 W., where 50 to 60 feet of conglomerate crops out, forming a cuesta that strikes N. 15-20° E. and dips 15-20° E. One and one-half miles southeast of the conglomerate exposure, quartzite beds crop out that contain thin interbedded mudstone. The beds strike N. 45-85° W. and dip 5-38° NE. The outcrops in the area define a structural basin or trough in the Sioux that plunges eastward.

In the New Ulm area, the Sioux Quartzite consists of 700 feet of predominantly well cemented, purple to red quartzite and local interbeds of poorly cemented sandstone and red mudstone (Miller, 1961, unpub. M.S. thesis, Univ. Minn.). The contact between the Sioux and the underlying rock is not exposed. However, 360 feet west of the basal conglomerate a reddish-orange, very coarse-grained to porphyritic granite is exposed. The basal conglomerate contains pebbles, cobbles, and some small boulders of highly siliceous rock types in a matrix of sand and fine gravel. Granite pebbles are lacking; however, all writers since Upham (*in* Winchell and Upham, 1888) have rejected the idea that the granite is younger than the Sioux. The possibility of a fault contact also has been discussed, but was rejected by both Baldwin (1951, *op. cit.*) and Miller (1961, *op. cit.*) who indicated that an unconformable contact between the older granite and the younger Sioux is more likely.

PETROLOGY

The Sioux Quartzite is composed dominantly of red, tightly cemented quartzite and some intercalated beds of mudstone and conglomerate. Baldwin assumed from the exposures and drill data that "typical quartzite" composes most of the formation—perhaps three-fourths of it—and mudstone and conglomerate are minor. "Typical quartzite" consists of pink, silica-cemented quartz sand that is well sorted, rounded, and of medium grain size (Baldwin, 1951, *op. cit.*). The mudstone within the Sioux is identified as indurated mud of predominantly clay- to silt-size material that lacks the fissility of shale and the incipient recrystallization of argillite; it includes both claystone and siltstone as well as intermediate size gradations. Pipestone, or "catlinite," is a clayey mudstone or claystone that has been quarried near Pipestone, Minnesota.

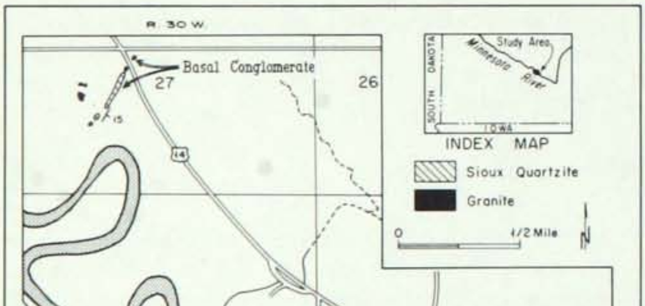

Figure V-100. Sioux Quartzite outcrop areas in the New Ulm area (after Miller, 1961, unpub. M.S. thesis, Univ. Minn.).

Quartzite

The quartzite is characteristically pink but ranges from nearly white to brick red. Less commonly, it is purplish or dark gray. The color results from finely disseminated iron oxide, comprising less than one percent of the rock, which coats the original quartz grains as a thin film. In darker red specimens, the iron oxide also is disseminated in the matrix, and in purplish or dark-gray quartzite it has a metallic luster.

Sand-size grains in the quartzite are well rounded and of medium size, and detrital particles or grains in the quartzite layers are almost exclusively composed of silica of which crystalline quartz predominates. Detrital grains of jasper and chert are present in some beds. The SiO_2 content is greater than 94 percent in the quartzite and never less than 70 percent, even in argillaceous quartzite (fig. V-101). The detrital grains are cemented by secondary overgrowths of optically continuous quartz (fig. V-102); commonly, there is some diaspore and sericite in the interstices.

Conglomerates

Of the three conglomeratic zones identified in the Sioux Quartzite, the coarsest (probably the basal) is exposed near New Ulm. It is composed of pebbles, cobbles, and small boulders as much as 13 inches in diameter that occur in several beds. The mixture of sand, fine gravel, and larger particles is poorly cemented locally, and many of the cobbles and boulders have weathered out. About half the detrital particles are composed of white vein quartz; the remaining clasts are composed of jasper, chert and cherty iron-formation, and fine- to medium-grained white quartzite, in order of decreasing abundance.

A second exposure of the basal conglomerate is about 6 miles southwest of the city of Pipestone in the SW¼ sec. 36, T. 106 N., R. 47 W. (fig. V-98; Baldwin, 1951, *op. cit.*). Here, the conglomerate is composed of pebble- and cobble-size clasts of mudstone, quartzite, jasper, chert, and felsite in a matrix of medium- to very coarse-grained quartz sand and argillaceous fragments.

A second conglomeratic zone is exposed intermittently from 3 miles south of Pipestone to Blue Mound State Park, and outlines the surface expression of what Baldwin (1951, *op. cit.*) called the Rock County structural basin. This conglomerate is at least 350 feet thick and is composed primarily of vein quartz with red jasper, pink to cream finely crystalline chert, and some quartzite pebbles. On the basis of lithologic similarities, Baldwin correlated the second con-

Figure V-101. Poorly cemented argillaceous quartzite. Quartz grains are surrounded by a matrix of predominantly sericite and fine-grained quartz. Crossed nicols, 50X. (From Miller, 1961, unpub. M.S. thesis, Univ. Minn.)

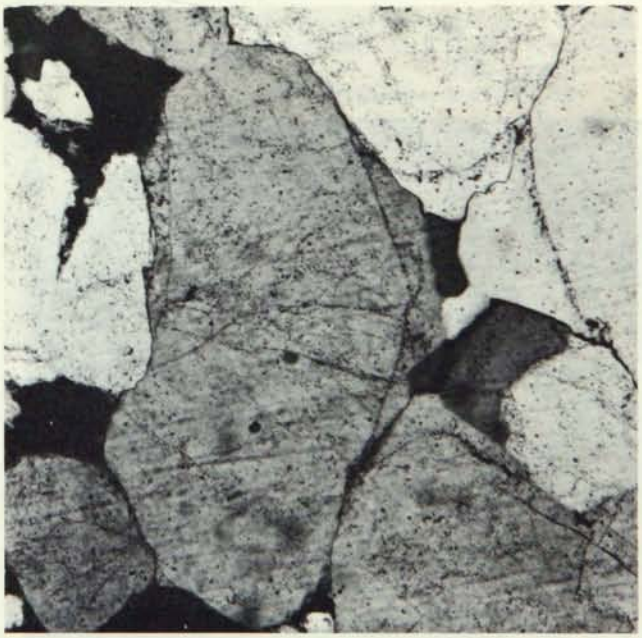

Figure V-102. Well cemented quartzite showing secondary overgrowths. Crossed nicols, 50X. (From Miller, 1961, unpub. M.S. thesis, Univ. Minn.)

glomeratic zone with a conglomerate 13 feet thick on a quartzite ridge centered in Cottonwood County.

The third or upper conglomeratic zone in southwestern Minnesota is exposed along a semi-elliptical path from 2 miles northeast of Jasper to 3½ miles southwest of Hardwick. Baldwin (1951, *op. cit.*) correlated this conglomerate with a conglomerate exposed 2 miles north of Garretson, South Dakota. At both localities, about two-thirds of the clasts are vein quartz; the remainder are pink and cream, finely crystalline chert. Quartzite pebbles are not as common as in the other conglomerates.

Mudstones

Argillaceous or clayey beds are present in most outcrops of the Sioux Quartzite, but they comprise only a minor part of the unit. The mudstones range in lithology from nearly pure claystone to silty mudstone to argillaceous siltstone, and these in turn grade into argillaceous quartzites. The mudstones consist principally of sericite, hematite, diaspore, and quartz (fig. V-103). Berg (1938) identified pyrophyllite in the mudstone from Pipestone, Minnesota, but Miller (1961, *op. cit.*) was unable to find it in the mudstones near New Ulm, Minnesota. The marked chemical difference between the mudstones and a "typical shale" is illustrated in Table V-38; the greatest differences are in the SiO_2 and Al_2O_3 content. The mineralogy of the mudstones is significant in determining the origin of the Sioux Quartzite, which is discussed later in this report.

SEDIMENTARY STRUCTURES

Bedding is observed in nearly all exposures of the Sioux Quartzite, varying from thin in the mudstones to thick in the conglomerates and quartzites. Torrential cross-bedding (fig. V-104) is common in the quartzites. The cross-bedding is inclined generally southward in Cottonwood County.

Ripple marks are particularly common in the quartzite. The ratio of wave length to amplitude ranges from 4:1 to 14:1, suggesting formation in water. The ripples in the Rock County structural basin appear to trend northeastward; in Cottonwood County, the trend is north, and the majority in the New Ulm area are close to N.60-70°W. Baldwin (1951, *op. cit.*) suggested that the currents or waves, which moved at right angles to the trends in ripple marks, were aligned generally northwestward.

Irregular polyhedral mud cracks from a few cm to 50 cm across, have been observed on bedding surfaces of both the mudstone and the quartzite. Mud cracks indicate that the clastic material was deposited periodically, with intervening times of emergence or quiescence. Baldwin (1951, *op. cit.*) suggested that during periods of aridity, salt was deposited in cracks in the sand surface. Later, during wet seasons, the salt was removed, leaving polyhedral cracks in the quartzite.

DEPOSITIONAL ENVIRONMENT

The quartzite and conglomerate of the Sioux Quartzite consist mainly of quartz with lesser amounts of chert, iron-formation, previously formed quartzite grains, and minor

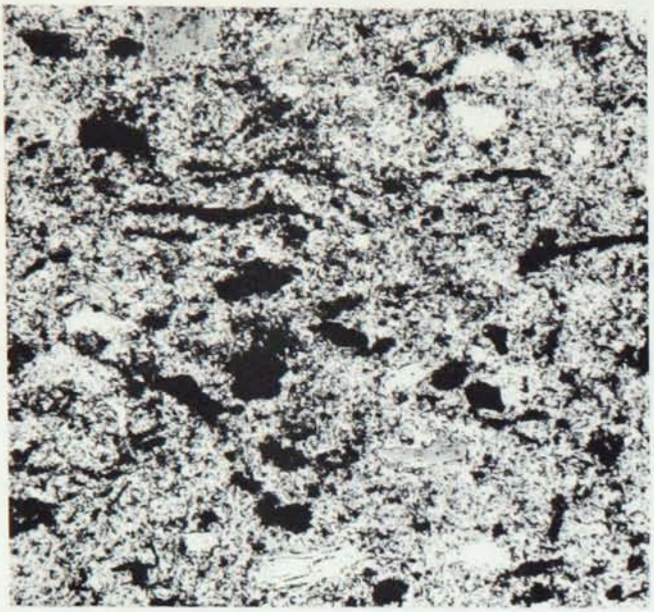

Figure V-103. Finest grained mudstone. Predominantly sericite and hematite with a few angular quartz grains. Crossed nicols, 128X. (From Miller, 1961, unpub. M.S. thesis, Univ. Minn.)

Table V-37. Chemical analyses, in weight percent, of the Sioux Quartzite.[1]

	1	2	3	4	5	6
SiO_2	97.58	99.14	97.82	84.52	94.56	98.68
Al_2O_3	0.31	0.28	2.12	12.33	3.19	0.66
Fe_2O_3	1.20	0.50		2.12	0.66	0.16
FeO					0.13	0.13
MgO	0.10			Tr	Tr	Tr
CaO	0.14	Tr		0.31	0.00	0.00
Na_2O	0.10			0.34	0.03	0.01
K_2O	0.03			0.11	0.01	0.05
P_2O_5					0.03	0.01
MnO		Tr			0.00	0.00
TiO_2					0.13	0.05
H_2O				3.31	1.03	0.01
Ignition loss	0.03					
	99.56	99.92	99.94	——	99.77	99.76

[1] Compiled by T. P. Miller (1961, unpub. M.S. thesis, Univ. Minn.):
Analyses 1-3, typical quartzite; Rothrock, 1944
Analysis 4, argillaceous quartzite; Winchell and Upham, 1884
Analysis 5, Grab sample of typical quartzite from New Ulm quartzite quarry. Analysis by University of Minnesota Rock Analysis Laboratory, Eileen Oslund, Analyst
Analysis 6, Grab sample of typical quartzite from quarry at Jasper, Minnesota. Analysis by University of Minnesota Rock Analysis Laboratory, Eileen Oslund, Analyst

Table V-38. Comparison of chemical analyses, in weight percent, of "average shale" and mudstone.

	"Average Shale"[1]	Mudstone[2] 1	2	3	4	5
SiO_2	58.01	49.01	48.20	57.43	58.25	50.40
Al_2O_3	15.40	35.17	28.20	25.94	35.90	33.30
Fe_2O_3	4.02	3.06	5.00	8.70		2.80
FeO	2.45	none				
MgO	2.44	0.23	6.00			0.17
CaO	3.11	0.05	2.60			0.60
Na_2O	1.30	0.06				4.10
K_2O	3.24	5.62				
H_2O	5.00	5.63	8.40	7.44	6.48	9.60
H_2O-		0.24				
TiO_2	0.65	0.44				
Li_2O		0.16				
P_2O_5	0.17					
CO_2	2.63					
SO_3	0.64					
BaO	0.05					
C	0.80					
MnO			0.60			
Ignition, less total H_2O		0.24				
	100.00	99.91	99.00	99.51	100.63	100.97

[1] Analysis of average shale taken from Clarke, 1924
[2] Analyses of catlinite from Pipestone, Minnesota:
1, Berg, 1938
2-5, Winchell and Upham, 1884

Figure V-104. Cross-bedding in typical Sioux Quartzite.

amounts of clay- and silt-size material. These constituents indicate that the area of deposition and the source areas, which were to the north, were tectonically stable. Miller (1961, *op. cit.*) suggested that the basal conglomerate was formed at the edge of a shallow transgressing sea, and Baldwin (1951, *op. cit.*) believed that the Sioux Quartzite as a whole was formed as an offshore deposit on a slowly sinking shelf. Both Baldwin (1951, *op. cit.*) and Miller (1961, *op. cit.*) indicated that the Sioux could have been laid down in a marine or nonmarine environment, but Baldwin suggested that the size and thickness of the unit imply that the basin was not a shoreline deposit but a shelf in the open sea with shifting currents. Miller stated that the mudstones, which contain sericite, hematite, diaspore, and quartz, originally may have formed from a bauxitic clay protolith which due to compaction and other factors recrystallized to their present composition. A bauxitic clay is consistent with the presence of the compositionally stable framework minerals in the quartzites and conglomerates. In either case the clay minerals represent the product of incomplete weathering and probably were washed from the eroding source areas into the basin during storms.

Cement in the Sioux Quartzite consists essentially of silica, with lesser sericite. Generally, the silica cement is in optical continuity with the original detrital quartz grains and forms secondary overgrowths that interlock with one another in irregular boundaries. The boundary between the original grains and the overgrowths commonly has a thin coating of iron oxide. Baldwin (1951, *op. cit.*) suggested that the silica was deposited from circulating silica-saturated ground waters; Miller (1961, *op. cit.*), however, believed that the concave-convex contacts between larger grains and the abundance of sutured contacts and solution effects in the finer grained quartzite and argillaceous quartzite in the New Ulm area suggest that some of the silica may have come from the solution of fines. Baldwin (1951, *op. cit.*) stated that the poorly cemented or sandy parts of the Sioux occasionally observed in outcrop and encountered in drill holes are related to weathering surfaces and joints. In a study of the weathered surface of the Sioux and the overlying Cretaceous sedimentary rocks, I (Austin, 1970a) came to the same conclusion, and suggested that in the New Ulm area weathering products within and on the Sioux lie below sedimentary rocks known to interfinger with rocks of Cenomanian (Late Cretaceous) age.

Baldwin (1951, *op. cit.*) suggested that the climate in the source area was humid with dry seasons. The absence of boulders containing feldspars is indicative of extreme chemical weathering and deep leaching. However, even a cool climate in the source area may have been present because plants that would deter deep weathering probably were absent. With the exception of pyrophyllite, reported by Berg (1938) from the mudstones at Pipestone, Minnesota, minerals indicative of temperatures above that of diagenesis have not been found. Baldwin (1951, *op. cit.*) doubted the identification of the pyrophyllite, and Miller (1961, *op. cit.*) found no pyrophyllite in the mudstone near New Ulm, Minnesota. Both Baldwin and Miller concluded that pyrophyllite is rare in the Sioux, if indeed it is present at all. Thus, both Baldwin and Miller considered the mineralogy of the Sioux to be of nonmetamorphic origin.

Chapter VI

PALEOZOIC AND MESOZOIC

PALEOZOIC LITHOSTRATIGRAPHY OF SOUTHEASTERN MINNESOTA

George S. Austin

The Lower and Middle Paleozoic rocks, exposed in southeastern Minnesota primarily along the Mississippi, Minnesota, and St. Croix Rivers (fig. VI-1), have been examined and discussed for more than 125 years. The publication of the latest geologic map of the State of Minnesota (Sims, 1970) was the culmination of more than 100 years of investigation. Recent work on the Paleozoic rocks of southeastern Minnesota has centered upon (1) producing updated geologic maps of the area, (2) producing a revised geologic column, (3) recognizing depositional and post-depositional events and environments, and (4) understanding the nature and variation of the biotic fraction of these units. Recently published geologic maps by the Minnesota Geological Survey at a scale of 1:250,000 cover approximately 80 percent of the Paleozoic sedimentary rocks (Sloan and Austin, 1966; Austin and others, 1970). A geologic column of these rocks has been published by the Minnesota Geological Survey (Austin, 1969), and the development of this lithostratigraphic column from the column shown on the 1932 state geologic map and correlative parts of the Wisconsin Paleozoic column are shown in Figure VI-2. The following discussion is a summation of the current lithostratigraphic knowledge of these units.

LITHOSTRATIGRAPHY

The Paleozoic rocks of southeastern Minnesota were deposited from a marine sea that occupied the Hollandale embayment (fig. VI-3), a shallow depression that extended northward from the Ancestral Forest City basin (Iowa basin) onto the cratonic shelf and into Minnesota and Wisconsin in Early and Middle Paleozoic time. The marine rocks that now remain within the embayment are bordered to the east by nearshore-facies Paleozoic rocks on the Wisconsin Arch, to the northeast by Precambrian rocks that constitute the Wisconsin Dome, and to the north and west by nearshore-facies Paleozoic rocks lying between the Hollandale embayment and the Precambrian rocks of the Transcontinental Arch. The embayment overlies older basins and horsts that are bounded by large-scale Precambrian faults (Sims and Zietz, 1967), but the pre-Mt. Simon rocks for the most part form an older, smaller basin which has been cut by large-scale faults. Relatively minor recurrent movements along these faults during Paleozoic time have resulted in a complex depositional and post-depositional history of the Paleozoic rocks in the Hollandale embayment. Many smaller Paleozoic basins, depositional barriers, and faults within the embayment probably have resulted from renewed activity along Precambrian structures (Craddock and others, 1963).

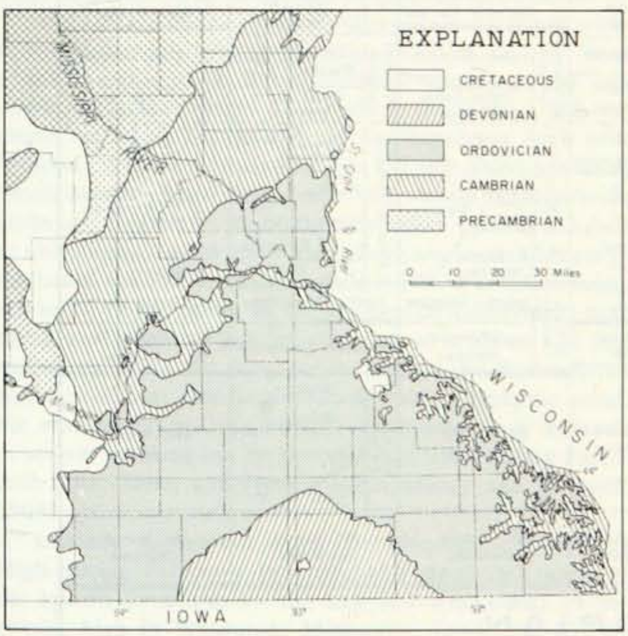

Figure VI-1. Generalized bedrock geologic map of southeastern Minnesota.

The Paleozoic stratigraphic nomenclature for southeastern Minnesota has developed primarily from the study of outcrops near the St. Croix and Mississippi Rivers, along the eastern border of Minnesota, and of deep cored holes within the Hollandale embayment. Correlations between the exposures and the deep cored holes across the embayment are made with comparative ease because of the continuity of thin lithic units across the embayment and the Wisconsin Arch (Berg and others, 1956; McGannon, 1960, unpub. Ph.D. thesis, Univ. Minn.; Austin, 1970b). Correlation between units in the area of the nearshore facies west of Mankato and west and north of the metropolitan Minneapolis-St. Paul area and rocks to the east and south is more difficult. The nearshore facies of these Paleozoic rocks in Minnesota initially differed significantly from correlative units within the embayment, and deep weathering prior to the advance of the seas from the west in Late Cretaceous time has increased the differences. In addition, the Paleozoic rocks are partly covered by lithologically similar rocks of Late Cretaceous age and by thick Pleistocene drift, which covers all previously-formed units, limiting examination of these rocks to outcrops along the Minnesota River. For these reasons the lithologic descriptions of the rock units given later in this report do not apply strictly to the rocks in the area containing the nearshore facies.

Pre-Mt. Simon Rocks

Several types of older rocks, mainly Keweenawan in age, lie directly beneath the Cambrian Mt. Simon Sand-

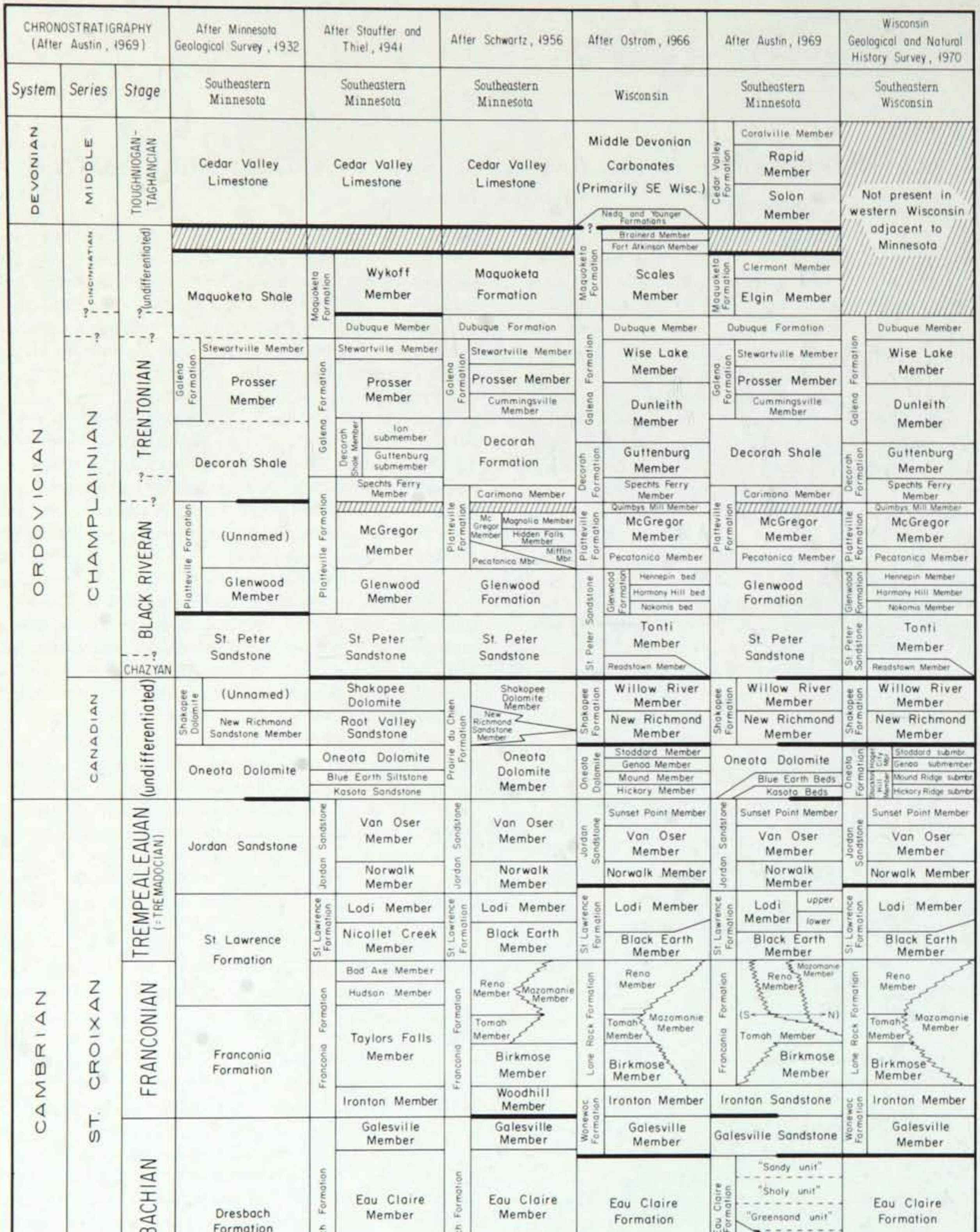

Figure VI-2. Comparison of some classifications of Paleozoic rocks in southeastern Minnesota and western Wisconsin from 1932 to 1970. Thick lines between units denote unconformities.

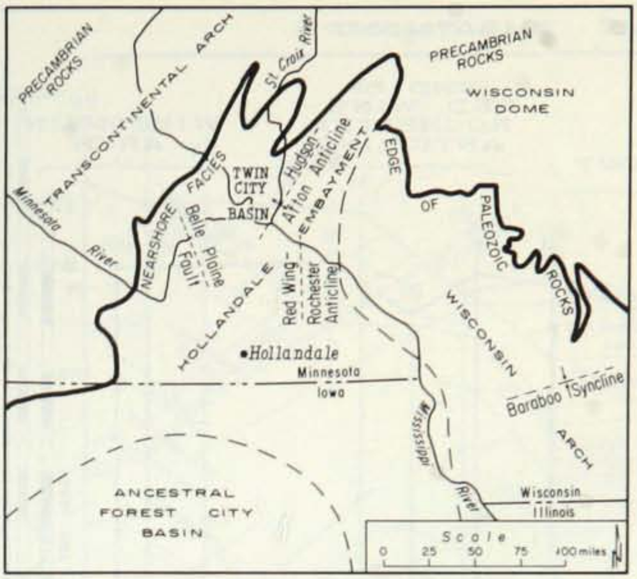

Figure VI-3. Map showing the major structural features in southeastern Minnesota and adjacent parts of Wisconsin and Iowa (modified from Austin, 1970b).

stone in southeastern Minnesota (see Morey, this volume). These are principally the Hinckley Sandstone and arkosic sandstones and shales, considered to be Late Keweenawan in age, and basalt and rhyolite flows of Middle Keweenawan age. Older granitic rocks directly underlie the Mt. Simon both west and east of the relatively thick section of Paleozoic rocks in the Paleozoic Hollandale embayment.

The surface on the Precambrian rocks is depressed within the Hollandale embayment (figs. VI-4, 5, and 6). West of the Paleozoic rocks in south-central Minnesota, the Precambrian surface ranges in altitude from 700 to 1,500 feet above sea level. To the east, on the Wisconsin Arch near Winona, Minnesota, the surface is 155 feet above sea level. North of the Twin City basin (fig. VI-3), Precambrian rocks are exposed or found directly below Pleistocene drift at altitudes of 700 to 1,000 feet above sea level, whereas in the basin, the Precambrian surface is 100 to 250 feet below sea level. In the center of the Hollandale embayment near the Iowa-Minnesota border, the surface is about 420 feet below sea level. The irregular surface on the Precambrian rocks is in part due to pre-Mt. Simon topographic relief, but mainly results from vertical displacements of as much as several hundred feet on Precambrian faults during Paleozoic time (Sloan and Danes, 1962; Morey and Rensink, 1969).

Isostatic adjustment along the Midcontinent Gravity High during Paleozoic time probably produced many of the known structures on the east and west flanks of the gravity high in Wisconsin, Minnesota, Iowa, and Kansas (Robert Miller, 1970, Northern Natural Gas Company, written comm.).

Paleozoic Formations

Mt. Simon Sandstone

The Mt. Simon Sandstone is composed of white, gray, pink, or locally yellow, medium-grained quartzose sandstone and some thin shale beds. Fine-grained sandstone near the top of the unit and coarse- to very coarse-grained sandstone toward the bottom are interbedded with medium-grained sandstone. Some siderite is present, particularly near the base. The presence of many cross-bedded zones and numerous fragments of inarticulate brachiopods, especially near the top, indicates a high-energy depositional environment. However, the thin grayish-green shale beds, which are especially common in the upper part of the formation and are interstratified with these high-energy deposits, indicate periodic lower energy environments of deposition.

Eau Claire Formation

The Eau Claire Formation consists of five different rock units in Minnesota. A "red shale phase" (Stauffer, 1927a and b; Stauffer and Thiel, 1941)—more properly classified as a red, silty, fine-grained quartzose sandstone and shale, or red, locally worm-bored siltstone—occurs along the western border of the Hollandale embayment and below the other rock types of the formation near the center of the embayment (Austin, 1970b). Where the red unit is absent, fine-grained, non-glauconitic quartzose sandstone and interbedded grayish-green fissile shale are found in the lowest part of the Eau Claire. Glauconitic, very fine- to medium-grained quartz sandstone with some thin grayish-green shale beds characterizes the middle unit of the formation.

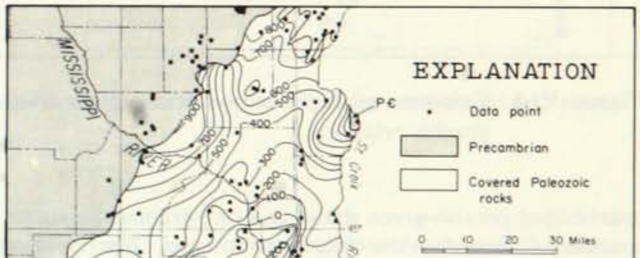

Figure VI-4. Structural contour map of the present altitude of the pre-Mt. Simon unconformity in southeastern Minnesota (in feet above sea level).

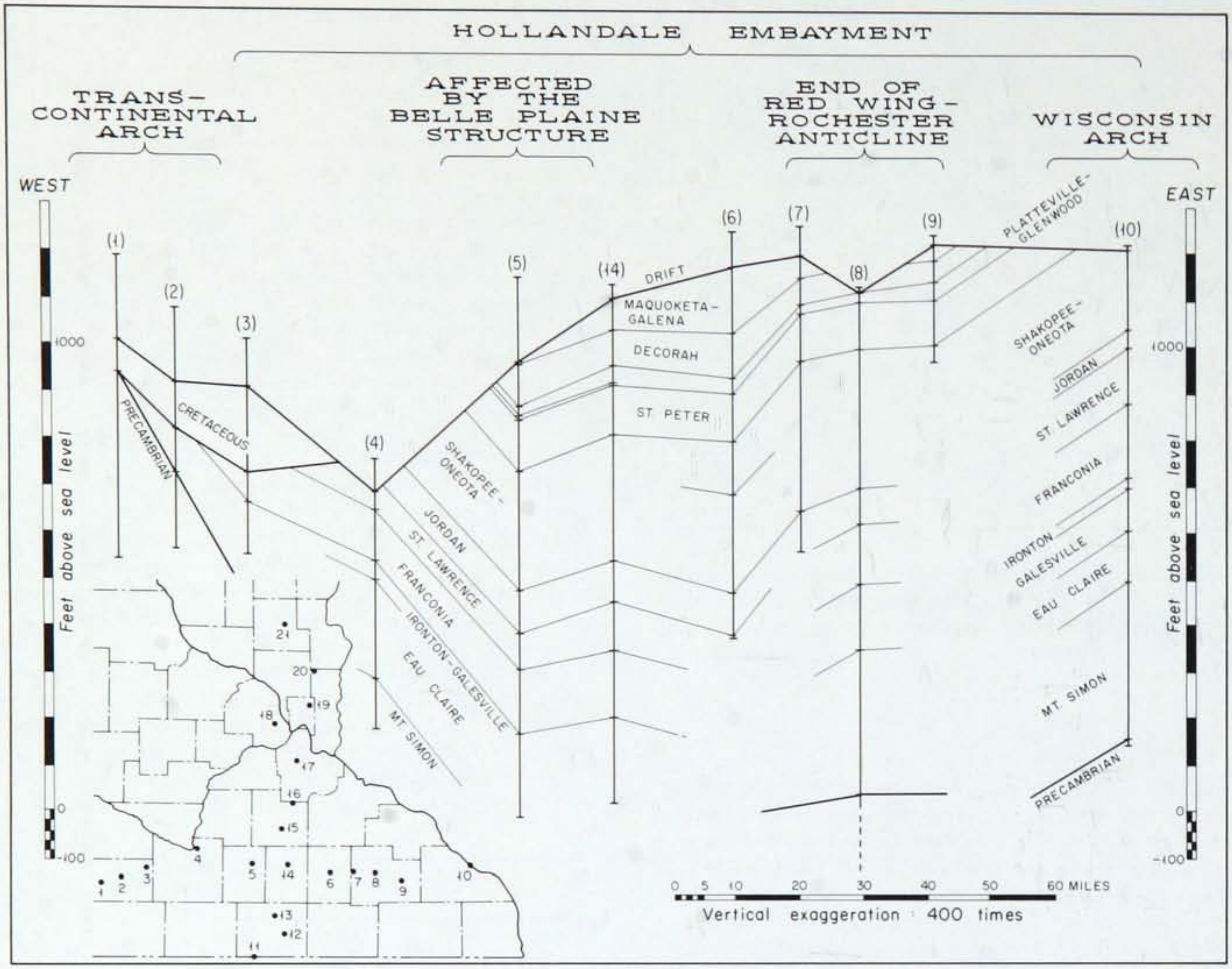

Figure VI-5. East-west geologic cross-section across southeastern Minnesota, showing the subsurface structural and stratigraphic relations.

Interbedded grayish-green shales and fine-grained quartzose sandstones comprise the fourth rock type. The thickest shale bed that has been noted in this unit is about 8 feet thick. The fifth and uppermost unit is a massive, light gray, locally glauconitic, fine-grained quartzose sandstone with some interbedded shale. West of the Minnesota River and north and west of the Twin City basin, the distinctive lithologies of the Mt. Simon, Eau Claire, Galesville, and Ironton formations blend, and this sequence may be composed of white quartzose sandstone in one locality and interbedded white sandstone and varicolored shale in another.

Galesville Sandstone

The Galesville Sandstone in Minnesota is a white to light gray, slightly glauconitic, well- to moderately well-sorted, mostly medium-grained quartzose sandstone, but is interbedded with fine-grained quartzose sandstone beds toward its base. The base is placed just below the first medium-grained sandstone and just above the massive fine-grained sandstone that is typical of the upper part of the Eau Claire Formation. In western Wisconsin, the Galesville lies disconformably on the Eau Claire (Ostrom, 1966) but this relationship has not been observed in Minnesota.

Ironton Sandstone

The Ironton Sandstone is a white, medium-grained, moderately well- to poorly-sorted quartzarenite that contains a significant amount of admixed silt-size material. The top few feet contain some glauconite and are typically stained yellowish brown in outcrop in contrast to the dominant white or light gray that characterizes most of the formation. The base of the Ironton is placed just above the non-silty, better-sorted, medium- and fine-grained sandstones of the Galesville and just below the silty, less well-sorted, medium-grained sandstones of the Ironton.

Berg (1954) and Berg and others (1956) identified an unconformity at the base of the Ironton which separates the regressive Galesville from the transgressive Ironton. The

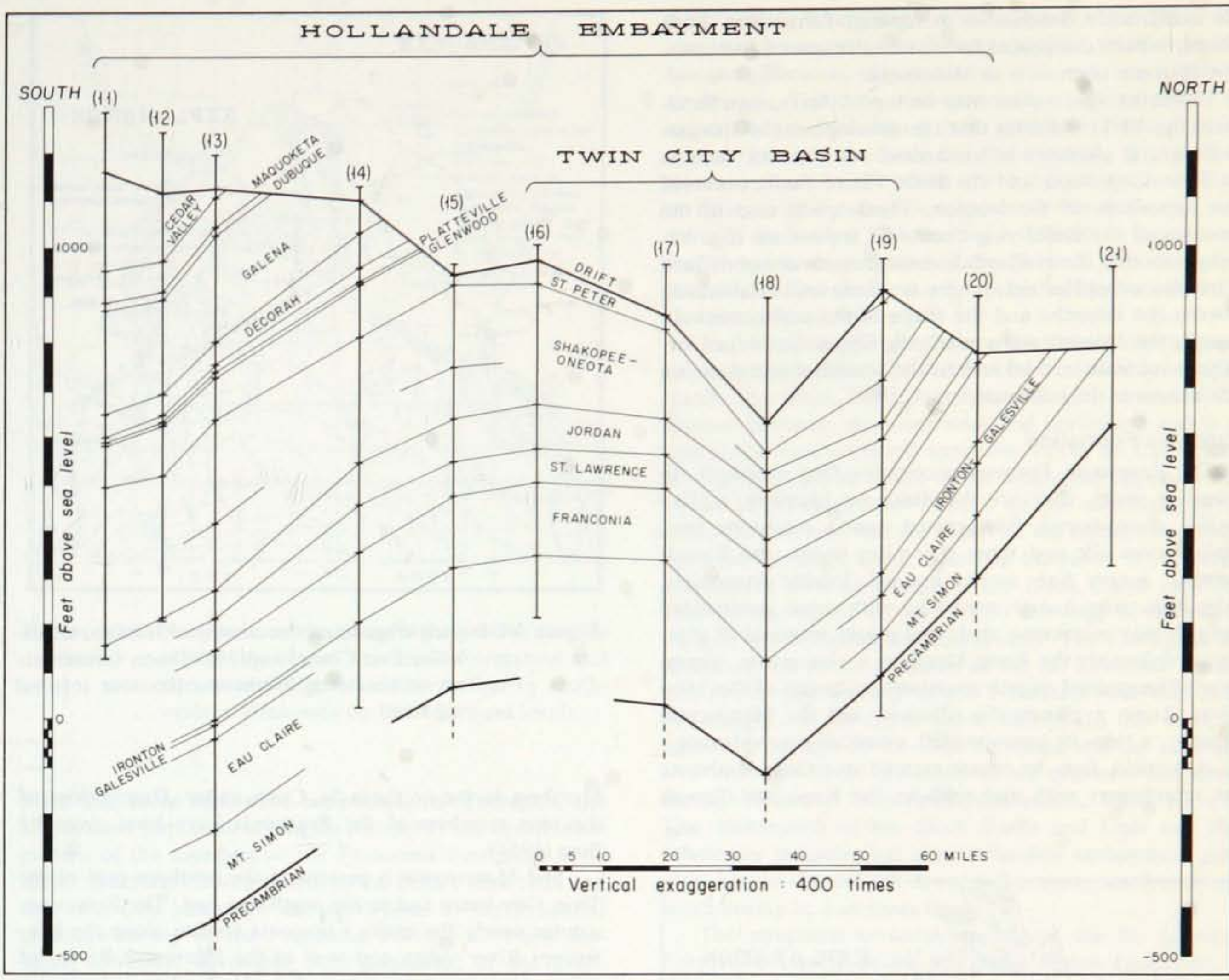

Figure VI-6. North-south geologic cross section across southeastern Minnesota, showing the subsurface structural and stratigraphic relations.

Ironton Sandstone is more properly classified as a lower energy sandstone that lies above the higher energy Galesville. In accord with this view, a reversal from regressive to transgressive deposition is present within the Galesville. Where the entire Galesville-Ironton succession is present, as near the center of the Hollandale embayment, the bottom part of the Galesville is regressive and the upper part is transgressive. Away from the center of the embayment, as on the Wisconsin Arch, only the upper transgressive Galesville is present as is an unconformity between the underlying Eau Claire Formation and the overlying Galesville Sandstone (Ostrom, 1966). The Ironton Sandstone was deposited in quieter and deeper water. Sand from the same source as the Galesville, and therefore of the same grain size, was deposited during Ironton time. The sandstone of the Ironton, however, is less well-sorted than that of the Galesville and commonly contains silt. Bottom currents of low energy did not sort grains as efficiently as those of Galesville time. The contact between Ironton and Galesville Sandstones may be sharp due to a rapid change in current activity. This abrupt current change in the St. Croix valley area may have resulted from movement along Precambrian faults in Cambrian time (G. B. Morey, 1969, oral comm.); this movement also produced the unconformity between the Galesville and Ironton noted by Berg (1954) and Berg and others (1956).

Berg (1954) considered the Ironton, which he called the Woodhill, a member of the Franconia Formation. In Wisconsin, Ostrom (1965, 1966, and 1967) placed both the Ironton and Galesville in the Wonewoc Formation, inasmuch as it is difficult and impractical to distinguish one from the other or to identify their contact in that area. Because the higher energy nature of the Galesville and the lower energy nature of the Ironton are more easily recognized in Minnesota, away from the Wisconsin Dome, the Minnesota Geological Survey prefers to retain the Galesville and Ironton Sandstones as separate formations. Both units have been designated formations because of their substantial lateral continuity in Minnesota.

The structural contour map on top of the Ironton Sandstone (fig. VI-7) indicates that the development of the major structural elements in southeastern Minnesota, such as the Twin City basin and the Belle Plaine fault, occurred after deposition of the Ironton. The isopach map of the Ironton and the underlying Cambrian formations (fig. VI-8) suggests that the Hollandale embayment was not defined in Ironton or earlier time; there is no obvious relationship between the isopachs and the shape of the embayment. In general, the Ironton and underlying Upper Cambrian formations increase in thickness to the southeast and decrease in thickness to the northeast.

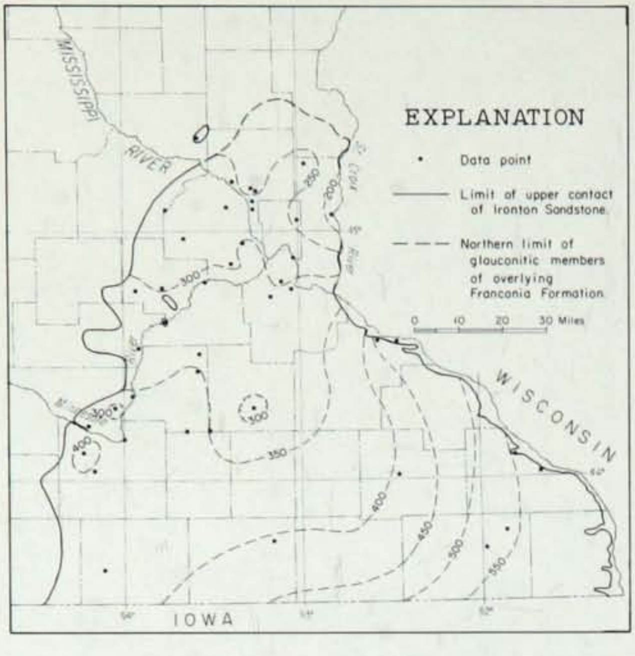

Figure VI-8. Isopach map of the combined Ironton, Galesville, Eau Claire, and Mt. Simon formations in southeastern Minnesota (contour interval 50 feet).

Franconia Formation

The Franconia Formation contains four members. In ascending order, they are the Birkmose Member, a glauconitic, worm-bored, fine-grained quartz sandstone containing some silt and some dolomitic layers; the Tomah Member, a very fine- to fine-grained, locally glauconitic, feldspathic, silty quartz sandstone with some interbedded greenish-gray micaceous shale and minor amounts of glauconitic dolomite; the Reno Member, a glauconitic, worm-bored, fine-grained quartz sandstone and, west of the Minnesota River, a glauconitic siltstone; and the Mazomanie Member, a thin- or cross-bedded, essentially non-glauconitic, dolomitic, fine- to coarse-grained quartzose sandstone that interfingers with and replaces the Reno and Tomah Members in the northern St. Croix valley. Descriptions of the four members of the Franconia have been given by Berg (1954).

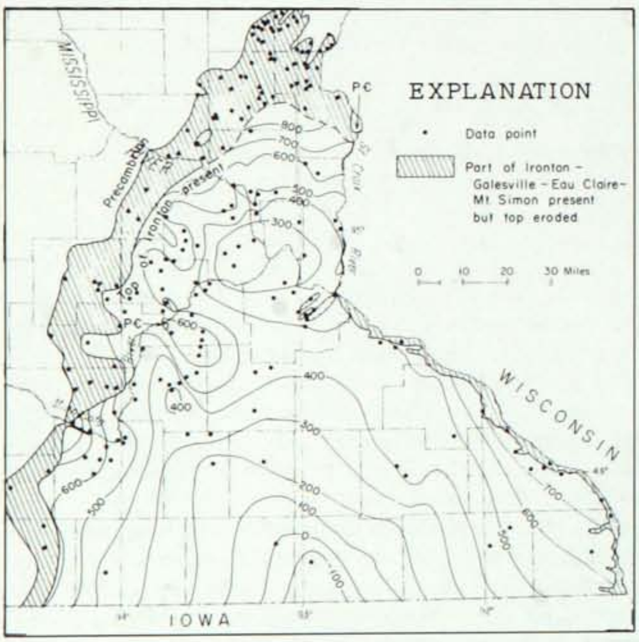

Figure VI-7. Structural contour map of the present altitude of the top of the Ironton Sandstone and the present eroded edge of the combined Ironton, Galesville, Eau Claire, and Mt. Simon formations (in feet above sea level).

The Mazomanie is present in the northern part of the Twin City basin and to the north and east. The Reno constitutes nearly the entire Franconia section along the Mississippi River valley and west of the Minnesota River but disappears toward the center of the Hollandale embayment. The Tomah is the most laterally persistent member of the formation and is at places the only member present near the center of the embayment. The base of the Franconia Formation is placed just above the highest essentially non-glauconitic, medium-grained sandstone of the Ironton and just below the commonly glauconitic, fine-grained sandstones of the Franconia.

The Franconia Formation ranges in thickness from about 100 to 200 feet, and is thinnest along the western border of the subdrift outcrop belt (fig. VI-9). Berg and others (1956) suggested that the regional distribution of the members of the Franconia Formation and the thickness of the unit are dependent on the position of the shoreline. The Franconia is thick to the north, where the shoreward Mazomanie Member predominates, and thins to the south, where the glauconitic Reno and Birkmose Members predominate. Outcrop and well data have shown that the Mazomanie predominates north of the center of the Twin City basin and that the Reno is the dominant facies along the Mississippi and Minnesota Rivers. Recent work by Austin (1970b) has suggested that the Tomah Member is the dominant unit near the center of the Hollandale embayment and that the glauconitic members predominate between the more shore-

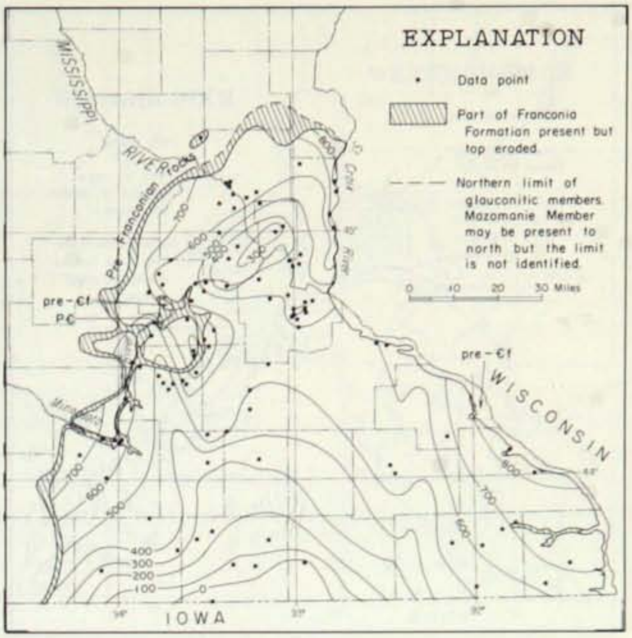

Figure VI-9. Structural contour map of the present altitude of the top of the Franconia Formation and the eroded edge of the formation in southeastern Minnesota (in feet above sea level).

ward and more basinward accumulations. Therefore, the embayment formation is roughly defined by the distribution pattern of the members of the Franconia Formation. The major structural elements, such as the Belle Plaine fault and the Twin City basin, are sharply defined by the structural contours on top of the Franconia Formation. These features, therefore, developed after deposition of the Franconia.

St. Lawrence Formation

The St. Lawrence Formation contains a variety of silty and sandy dolomitic rocks lying between the overlying Jordan Sandstone and the underlying Franconia Formation. The lower member of the St. Lawrence Formation, the Black Earth Member, is composed of glauconitic, argillaceous, silty or sandy dolomite; the dolomite content of the member exceeds 70 percent. The Black Earth is commonly highly resistant but does contain some less resistant silty beds. Flat-pebble conglomerates occur in the member and are particularly common near the base (McGannon, 1960, *op. cit.*; Twenhofel and others, 1935). The lower contact of the member is placed just above the sandstone beds of the Franconia Formation, which may be dolomitic and conglomeratic, and below the lowest, highly dolomitic, resistant bed of the Black Earth Member.

The upper or Lodi Member of the St. Lawrence Formation is composed of silty argillaceous dolomite; the dolomite content is less than 70 percent except locally at the top of the unit. The lower part of the Lodi contains mottled green, slightly glauconitic, argillaceous dolomite, which is less resistant than the Black Earth Dolomite, and buff dolomitic siltstones, or alterations of these types. The upper Lodi consists of unmottled buff or gray, thin- to thick-bedded, hard, argillaceous, silty dolomite, with some units of white or brown, friable to well-cemented sandy siltstone. The upper Lodi may contain a trace amount of glauconite. The contact between the Lodi and the overlying Jordan Sandstone is placed at the point where the dominant dolomite or siltstone of the Lodi is overlain by the dominant sandstone of the Jordan. This contact is difficult to locate where the fine-grained sandstone of the Norwalk Member of the Jordan overlies siltstones of the Lodi Member of the St. Lawrence Formation.

In Minnesota the St. Lawrence Formation ranges in thickness from 35 to 190 feet but the variation is not systematic. The Black Earth Member of the St. Lawrence Formation, however, thins eastward and northward and is absent at the most northerly exposures of the St. Lawrence in the valley of the St. Croix River, where the upper member of the St. Lawrence lies directly on the underlying Franconia Formation. Near the center of the embayment and to the west, the Black Earth makes up the entire St. Lawrence succession. West of Mankato, the Black Earth caps subdrift cuestas which extend to near the edge of the eroded margin of the Hollandale embayment. The Lodi generally makes up the entire thickness of the St. Lawrence Formation on the northern part of the Wisconsin Arch and in the Mississippi and St. Croix River valleys. In this area, the carbonate content of the Lodi decreases and the sand content increases north and south of an east-west line drawn through Winona, Minnesota (McGannon, 1960, *op. cit.*). The distribution of the Black Earth and Lodi and their lithologies indicate that the Hollandale embayment probably opened basinward toward the west, southwest, and south during St. Lawrence time.

The structural contours on top of the St. Lawrence Formation clearly define the Belle Plaine fault and the Twin City basin (fig. VI-10), indicating that these features were developed after deposition of the St. Lawrence.

Jordan Sandstone

The Jordan Sandstone in Minnesota contains three members. In ascending order, these are the Norwalk Member, a yellow, silty, fine-grained quartzose sandstone; the Van Oser Member, a white or yellow, coarse- to medium-grained orthoquartzite; and the Sunset Point Member, an argillaceous and dolomitic quartz sandstone with pebble-size clasts of dolomitic sandstone and thin beds of dolomite. The Van Oser is the thickest and most laterally persistent member of the Jordan Sandstone in Minnesota, and probably is the only member present in the Twin City basin. The Sunset Point Member occurs principally along the Mississippi River valley, and the Norwalk Member is confined to the fringes of the Hollandale embayment in Minnesota. The dolomitic Sunset Point Member is a precursor of the thick dolomites of Early Ordovician age in Minnesota. At the type locality in Madison, Wisconsin, the member is a dolomite and yields Cambrian fossils (Raasch, 1952). The Sunset Point in the Mississippi River valley is primarily a dolomitic sandstone but does contain some dolomite beds (Ostrom, 1965).

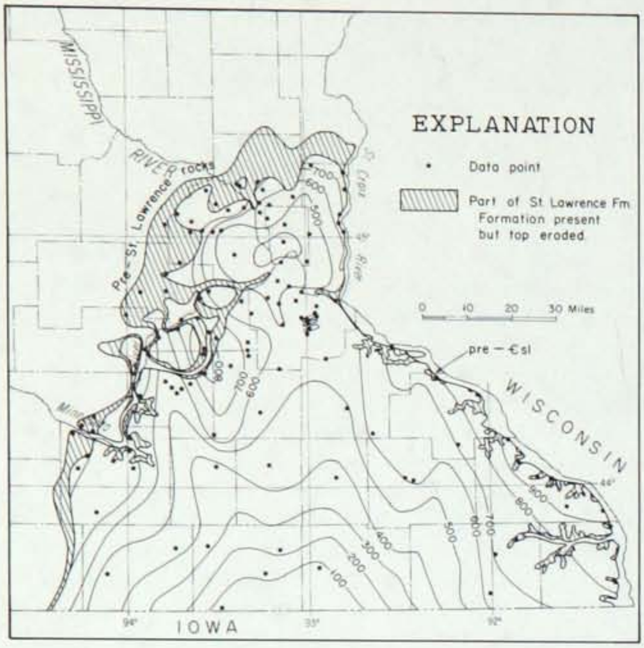

Figure VI-10. Structural contour map of the present altitude of the top of the St. Lawrence Formation and the eroded edge of the formation in southeastern Minnesota (in feet above sea level).

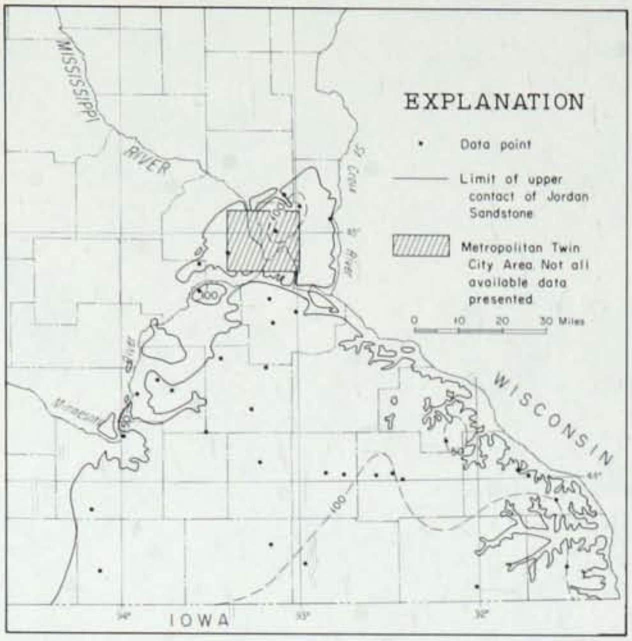

Figure VI-11. Isopach map of the Jordan Sandstone in southeastern Minnesota (contour interval 50 feet).

The Jordan Sandstone is an average of 85 feet thick north to south across the Hollandale embayment and thins from as much as 120 feet in extreme southeastern Minnesota to 50 feet to the west near Mankato, Minnesota (fig. VI-11). This east-west trend in sedimentation is similar to the trend in the underlying St. Lawrence Formation, and suggests that the embayment opened seaward to the west and southwest during the Trempealeauan Stage.

Structural contours on top of the Jordan in southeastern Minnesota (fig. VI-12) indicate that all major structural elements affected the altitude of the Jordan, and thus were developed after the Jordan was deposited. Sloan (unpublished data), in a detailed study of the present altitude of the top of the Jordan Sandstone in southeastern Minnesota, has defined many smaller structures, particularly along the outcrop belt in extreme southeastern Minnesota.

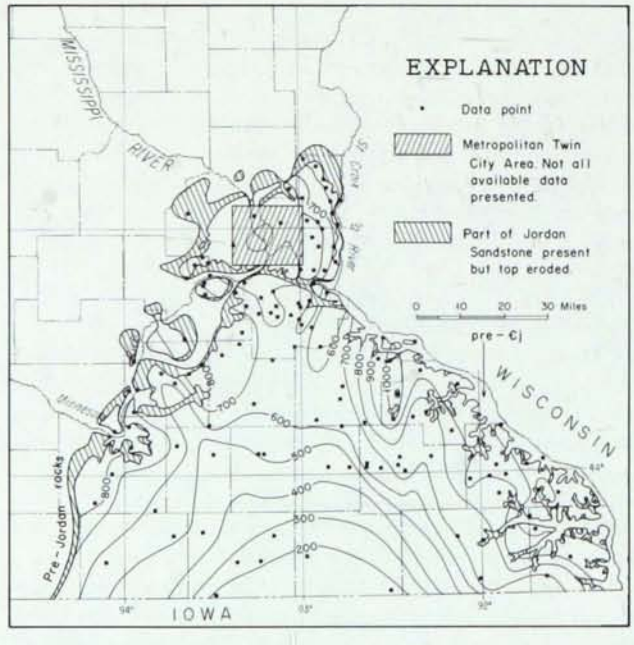

Figure VI-12. Structural contour map of the present altitude of the top of the Jordan Sandstone and the eroded edge of the formation in southeastern Minnesota (in feet above sea level).

Prairie du Chien Group

The Lower Ordovician rocks are difficult to delineate in southeastern Minnesota (Heller, 1956). Accordingly, where the underlying Oneota Dolomite cannot be separated from the overlying Shakopee Formation, it is advisable to identify the entire succession as the Prairie du Chien Group.

The Prairie du Chien Group thickens to the south in Minnesota, exceeding 350 feet near the center of the Hollandale embayment, and thins to the east, west, and north (fig. VI-13). Near the center of the Twin City basin, the Prairie du Chien is an average of 100 feet thick, and it thickens away from the basin.

The isopach contours follow a trend similar to the structural contour lines in the Twin City basin during Prairie du Chien time, suggesting that the basin was affecting deposition of the group (fig. VI-14). This is the first indication in the geologic succession of the development of the Twin City basin as a structural feature.

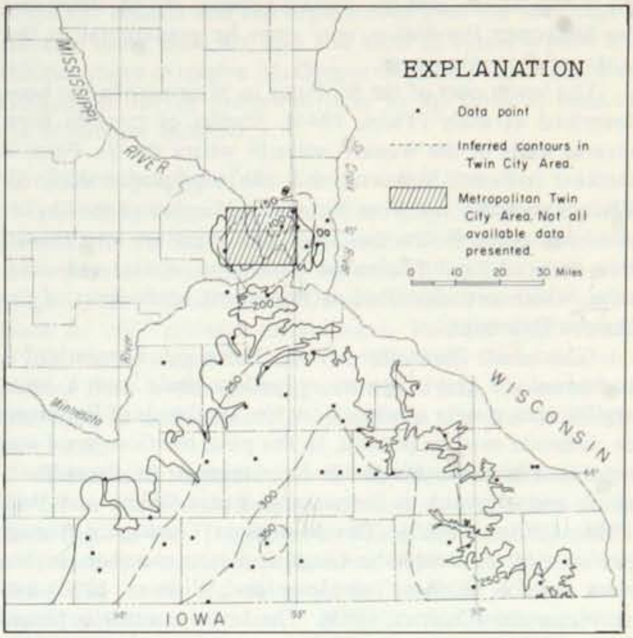

Figure VI-13. Isopach map of the Prairie du Chien Group in southeastern Minnesota (contour interval 50 feet).

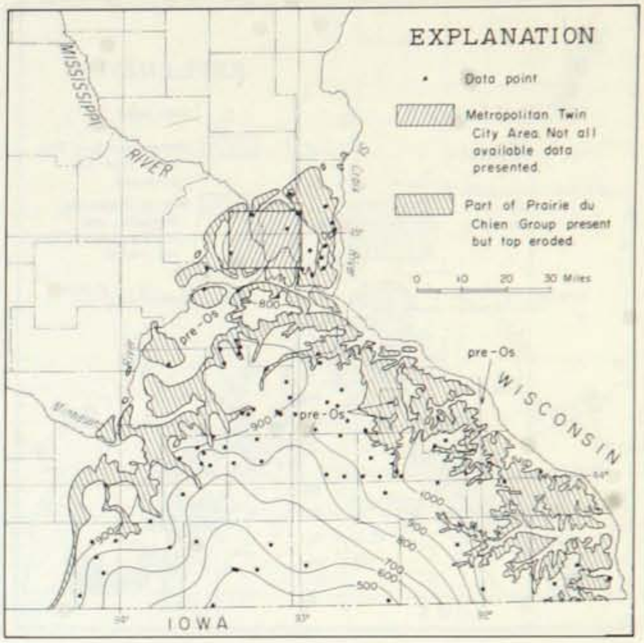

Figure VI-14. Structural contour map of the present altitude of the top of the Shakopee Formation and the eroded edge of the Prairie du Chien Group in southeastern Minnesota (in feet above sea level).

Oneota Dolomite. The Oneota Dolomite is primarily a thin- to thick-bedded, locally stromatolitic, light brownish-gray or buff, fine- to medium-grained dolomite with a silt-size dolomite matrix. Chert and some sand-size detritus occur locally near the base of the formation. A greenish-gray, shaly siltstone locally present beneath the dolomite is designated the Blue Earth siltstone beds of the Oneota Dolomite. A thin sandstone unit lithologically similar to the underlying Jordan Sandstone but containing Ordovician fossils (Powell, 1935) is present locally beneath the Blue Earth beds and is designated the Kasota sandstone beds of the Oneota Dolomite. The Sunset Point Member of the Jordan Sandstone and the Blue Earth siltstone and Kasota sandstone beds of the Oneota Dolomite illustrate facies changes across the Cambro-Ordovician boundary; whereas carbonate deposition began in the east, in Wisconsin, during latest St. Croixan time, clastic sediments from terrestrial sources continued to be deposited in the west, in Minnesota, into earliest Canadian time.

Shakopee Formation. The Shakopee Formation consists of two members (Davis, 1966a). The lower member, the New Richmond, consists of fine- to medium-grained quartzose sandstone and quartzitic dolomite and minor amounts of shale and pure dolomite. The upper boundary of the member commonly is marked by a thin zone of interbedded grayish-green shale, quartzose sandstone, and dolomite. The New Richmond in Wisconsin and extreme southeastern Minnesota lies on truncated Oneota (Ulrich, 1924; Ostrom, 1965; Davis, 1966a, 1966b). The New Richmond is thin and not distinguishable from the thin sandstone beds within the upper member of the Shakopee Formation west of the Red Wing-Rochester anticline. The upper member of the Shakopee Formation, the Willow River Dolomite, consists of thin- to thick-bedded dolomite, sandy dolomite, some interbedded quartzose sandstone, and some grayish-green shale. Chert and algal stromatolites commonly are present in the Willow River Member.

The Shakopee Formation and the underlying Oneota Dolomite separately and together as a unit (fig. VI-13) thicken toward the center of the Hollandale embayment. This accumulation of shallow-water carbonates is indicative of the more rapid deepening of the center of the embayment and indicates that the embayment opened seaward to the south more than to the west during Early Ordovician time. The New Richmond Member of the Shakopee Formation thins to the west and north from extreme southeastern Minnesota and represents a remnant of the east-west trend in sedimentation of the underlying Trempealeauan Stage in the Early Ordovician.

Rock Stratigraphy of the Black Riveran Stage

In Minnesota the Black Riveran Stage contains, in ascending order, the St. Peter Sandstone, the Glenwood Formation, and the Platteville Formation, but may not be separated from the overlying and underlying strata by unconformities. The lack of a convincing unconformity in the center of the Hollandale embayment and the change upward from the oolitic and stromatolitic dolomites of the underlying Shakopee Formation to shales and poorly sorted fine-grained sandstone with overlying well-sorted sandstone in the St. Peter have been taken as evidence that the seas

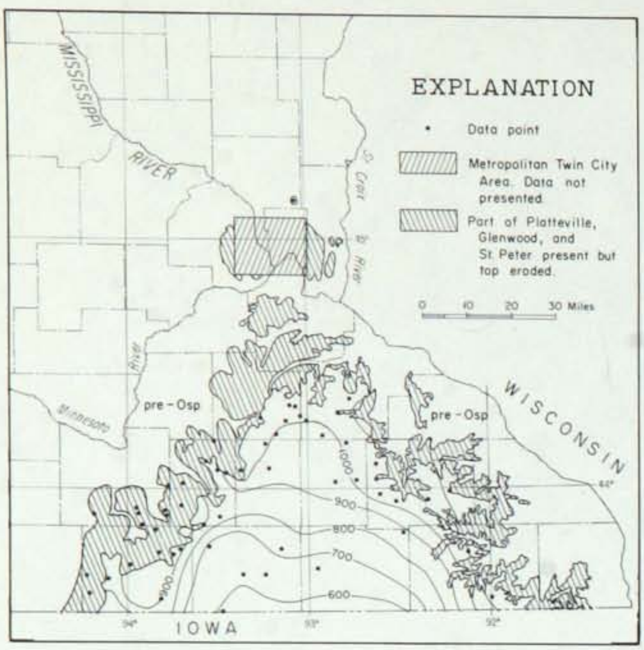

Figure VI-15. Structural contour map of the present altitude of the top of the Platteville Formation and the present eroded edge of the combined Platteville, Glenwood, and St. Peter formations in southeastern Minnesota excluding the metropolitan Minneapolis-St. Paul area (in feet above sea level).

may not have withdrawn entirely from the center of the embayment at the end of Canadian time (Austin, 1969, 1971). Therefore, the basal part of the St. Peter Sandstone in Minnesota may be of Chazyan age.

In Minnesota, strata of the Black Riveran Stage increase in thickness from south to north, from about 100 feet near the Minnesota-Iowa border to 190 feet in the Twin City basin.

Inasmuch as the Platteville and Glenwood Formations have maximum thicknesses of 35 and 18 feet respectively (Austin, 1969), the St. Peter Sandstone comprises the greater part of Black Riveran strata. The St. Peter thickens from south to north, and is as much as 155 feet thick in the Twin City basin (Austin, 1969).

Structural contours on top of Black Riveran strata (fig. VI-15) indicate that the Belle Plaine fault has affected the Platteville Formation and older strata and thus is younger than these formations. Data for the Twin City basin are given by Mossler in this chapter.

St. Peter Sandstone. The St. Peter Sandstone is a light yellow or white, medium-grained but locally fine-grained, massive-appearing, generally well-sorted orthoquartzite composed of rounded and subrounded grains. Because of its uniform grain size and low mineral content other than quartz, cross-bedding is rarely observed in outcrop. A few thin beds of green shale are present and the lowermost few feet locally may be silty and shaly. The large-scale relief on the erosional unconformity at the base of the St. Peter in Wisconsin (Ostrom, 1965, 1967) does not seem to be present in Minnesota, and the contact between the St. Peter and the Shakopee Formation may even be gradational in the center of the embayment.

The lower part of the St. Peter in Minneapolis has been described as shaly (Thiel, 1944). Studies of cuttings from several wells in the western suburbs where the St. Peter is thickest indicate, however, that the supposedly shaly St. Peter is probably the New Richmond Member of the Shakopee Formation. Below this shaly sandstone are intercalated thin beds of sandy dolomite, sandstone, shale, and dolomite, which are identified as shoreward equivalents of the Oneota Dolomite.

Glenwood Formation. The Glenwood Formation is composed of grayish-green or yellow shale and a basal argillaceous quartz sandstone. A few local beds of limestone or dolomite may be present. In the past, the Glenwood was separated from the Platteville Formation or St. Peter Sandstone and elevated to formational status (Weiss and Bell, 1956; Ostrom, 1969). The Minnesota Geological Survey prefers not to divide the Glenwood into members as has been done in Illinois (Templeton and Willman, 1963) and in Wisconsin (Ostrom, 1969). The lower contact is placed between the massive white sandstone of the underlying St. Peter and the argillaceous, poorly-sorted sandstone of the overlying Glenwood. The upper contact is placed above the uppermost thick shale bed and below the first massively-bedded, commonly sandy, dolomitic limestone of the Platteville Formation.

On the basis of the lateral variation in clay mineral assemblages, Parham and Austin (1967) suggested that the Glenwood Formation in southeastern Minnesota was derived from a positive area which lay to the west or southwest. Further, during this time the Wisconsin Dome was not contributing clastics to this area.

Platteville Formation. The Platteville Formation consists of three members. The lower or Pecatonica Member is a yellowish-brown, medium- to fine-grained dolomite or dolomitic limestone which may be sandy, particularly at the base. The sandy texture is a result of the presence of medium to fine-grained, rounded quartz sand. The Pecatonica Member in Minnesota commonly contains several corrosion zones. The middle or McGregor Member is a gray, light olive-gray, or buff, fine- to very fine-grained, thin-bedded, dolomitic limestone or dolomite with interbedded brown or olive-green shale. The McGregor has rippled bedding surfaces, which give it a characteristic crinkly bedding. The upper or Carimona Member is a medium-bedded, fine-grained, light olive-gray or buff limestone with interbedded olive-gray shale. A bentonite bed, generally 0.1 to 0.2 feet thick (the "Carimona bentonite"), occurs at or just above the Carimona-McGregor contact. In the past, the interval occupied by the McGregor Member in most of southeastern Minnesota has been divided locally into members, primarily in the Twin City basin (Weiss and Bell, 1956; Rassam, 1967, unpub. Ph.D. thesis, Univ. Minn.). In ascending order, these are the Mifflin, a very thin and crinkly-bedded limestone, which is 11 to 13 feet thick in the Twin City basin; the Hidden Falls, a very argillaceous, dolomitic limestone about 6 feet thick; and the Magnolia, a micro-granular dolomitic limestone or calcareous dolomite that is thicker bedded and less argillaceous than the McGregor. Because these units are thin and local in extent within the thicker, more extensive McGregor Member, the Minnesota Geological Survey prefers to refer to the units as beds of the McGregor Member.

Rock Stratigraphy of the Trentonian Stage

In Minnesota, the Trentonian Stage, containing the Decorah Shale below and the Galena Formation above, is not separated from rocks of the underlying Black Riveran Stage by an unconformity. On the basis of vertical variations in clay mineral assemblages, Parham and Austin (1969) have suggested that the Decorah Shale was deposited from a transgressing sea which had regressed but still covered southeastern Minnesota at the end of Platteville time.

Trentonian strata thicken to the south toward the center of the Hollandale embayment and thin in other directions (fig. VI-16). The Decorah Shale, the basal formation of Trentonian strata in Minnesota, is 95 feet thick in the Twin City basin and thins to the south to 55 feet near the Minnesota-Iowa border. It is 20 feet thick near the Mississippi River valley in extreme southeastern Minnesota, and is 40 to 45 feet thick in south-central Minnesota. Thus, the thickening of the Galena Formation toward the center of the embayment (fig. VI-17) is more pronounced than is shown by the combined Decorah and Galena formations.

Parham and Austin (1969) have suggested that the source of the clay minerals in the Decorah Shale was that part of the Transcontinental Arch lying west or southwest of the embayment. The same source is suggested for the clay minerals in the Glenwood Formation of the underlying Black Riveran strata. Therefore, the shaly formations of Trentonian and Black Riveran strata appear to be derived from the west or southwest rather than from the part of the Transcontinental Arch to the north of the embayment or from the Wisconsin Dome to the northeast or east.

Decorah Shale. The Decorah Shale is a greenish-gray or olive-gray, fissile, fossiliferous shale containing scattered limestone beds which are commonly coquinoidal in Minnesota. For practical purposes, the lower contact is placed above the lowest noncoquinoid carbonate bed of the Platteville and below the first thick shale bed of the Decorah. Parham and Austin (1969) have indicated, however, that locally the contact may be a shale-on-shale type, with Decorah Shale lying directly on Carimona shale. Because such a contact is difficult to identify in the field, it is preferable to place the contact at the lithologic break previously used. In Minnesota, the Decorah Shale has been divided with difficulty into members on the basis of lithologic differences (Agnew, 1956; Weiss and Bell, 1956). However, because the formation is essentially a shale in Minnesota, the Minnesota Geological Survey has assigned it the formal name Decorah Shale, and does not subdivide it into members.

Galena Formation. The Galena Formation in southeastern Minnesota contains three members. In ascending order, they are the Cummingsville, Prosser, and Stewartville Members. The Cummingsville is composed of interlayered thick beds of light olive-gray or buff limestone and

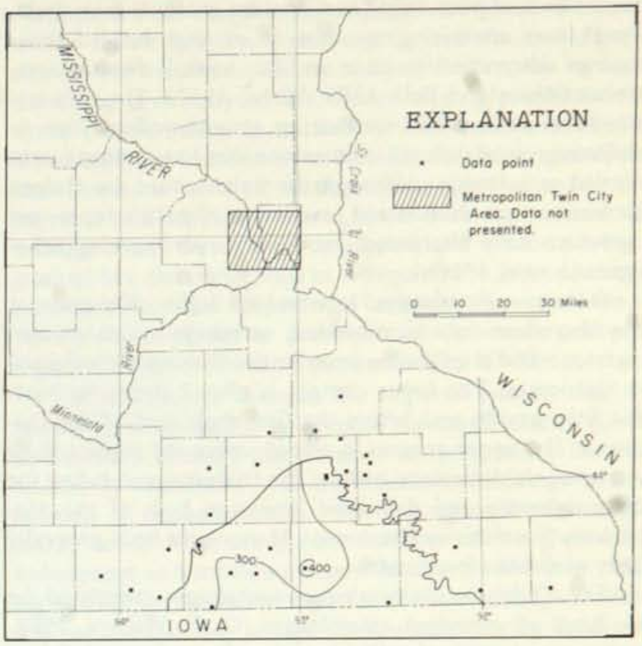

Figure VI-16. Isopach map of combined Galena and Decorah formations in southeastern Minnesota and excluding the metropolitan Minneapolis-St. Paul area (contour interval 100 feet).

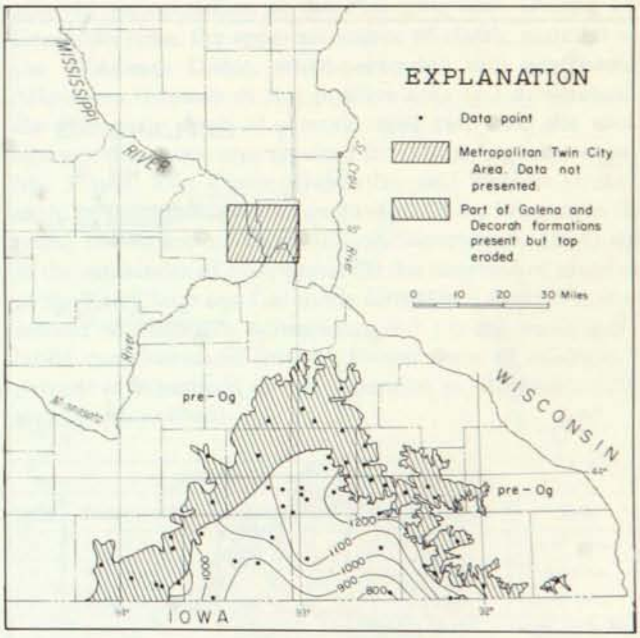

Figure VI-17. Structural contour map of the present altitude of the top of the Galena Formation and the present eroded edge of the combined Galena and Decorah formations in southeastern Minnesota (in feet above sea level).

thinner beds of both fissile and massive greenish-gray shale. The Prosser contains gray, thin- to medium-bedded limestone or dolomitic limestone and has a small detrital component (Weiss and Bell, 1956; Weiss, 1957). The Stewartville Member is a buff-weathering, grayish-yellow, fine- to medium-grained dolomitic limestone that has a conspicuous mottled appearance. Although the units within the Galena Formation are considered members, the lithotopes are known to have alternated, producing interfingering lithologies (Austin, 1970b).

Dubuque Formation. Interbedded light olive-gray or grayish-yellow, medium-bedded, crinoidal, fine-grained limestone and gray shale comprise the Dubuque Formation in Minnesota. The lower contact is placed above the mottled Stewartville and below the first shale bed of the Dubuque. The upper contact is placed above the highest shale and crinoidal limestone beds of the Dubuque and below the shaly dolomite and dolomitic limestone beds of the Maquoketa Formation. The lowest Maquoketa beds generally carry graptolite fragments.

The Dubuque Formation is tentatively correlated on the basis of conodont assemblages (G. F. Webers, 1969, oral comm.), with the Cincinnatian formations of Eden-Maysville age in Ohio and Kentucky, although a study of the ostracodes (Burr and Swain, 1965) indicates that this correlation is questionable.

Maquoketa Formation. Two members comprise the Maquoketa Formation (Bayer, 1965, unpub. Ph.D. thesis, Univ. Minn.). The lower, or Elgin, member contains flaggy limestones with nodular calcareous shales, shaly dolomite and calcareous shale beds, and coarsely crystalline

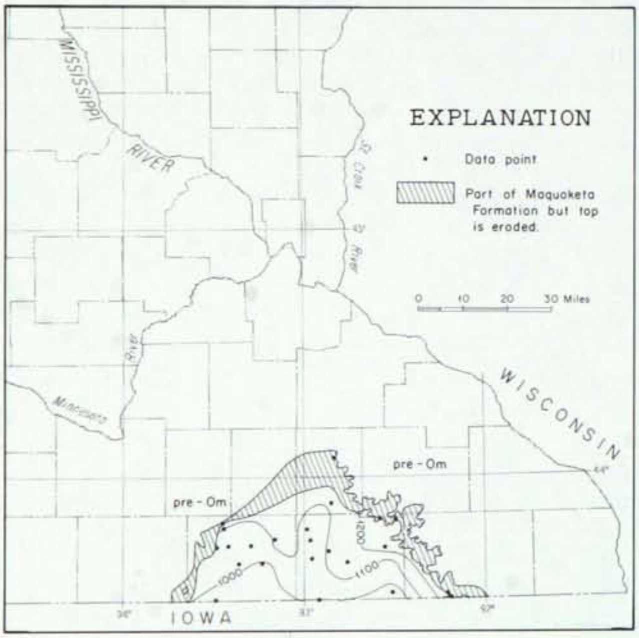

Figure VI-18. Structural contour map of the present altitude of the top of the Maquoketa Formation and the present eroded edge of the formation in southeastern Minnesota (in feet above sea level).

dolomitized limestone. The upper, or Clermont, member is a tan, sandy limestone. The remainder of the Maquoketa Formation in Minnesota was eroded prior to the deposition of the Cedar Valley Formation during Devonian time. The stratigraphy of the Maquoketa in Minnesota closely corresponds to that of the lower part of the Maquoketa in northern Iowa (Parker and others, 1959). In Minnesota, however, the Maquoketa does not contain the distinctive "depauperate beds" characteristic of the Maquoketa in Iowa (Glenister, 1957); nor is there in Minnesota a break in deposition between the Dubuque and Maquoketa Formations. In Minnesota, the Maquoketa Formation (fig. VI-18), which ranges in thickness from 50 to 90 feet, was deposited from a sea which entered the area before the underlying Dubuque Formation was deposited. Bayer (1965, *op. cit.*) indicated that the source of the detritus in the Elgin Member lay to the northeast or east, possibly as far away as the Taconic orogenic belt; however, the detritus in the sandy Clermont or upper member was derived from an uplift of the Transcontinental Arch in central Minnesota.

Cedar Valley Formation. The Cedar Valley Formation consists of three members in Minnesota (Kohls, 1961, unpub. Ph.D. thesis, Univ. Minn.). In ascending order, they are the Solon, Rapid, and Coralville Members, as defined by Stainbrook (1941). The Solon Member is transitional from a buff-gray, fine-grained, biogenic dolomite at the base to a light buff-gray, sublithographic dolomitic limestone toward the top. The Rapid Member is composed of gray, fine-grained, shaly dolomite with prominent microbedding and black streaks of finely-divided pyrite. The Coralville Member contains lithographic high-calcium limestone with microbedding and buff-gray, fine- to medium-grained dolomite and calcitic dolomite. Collinson and others (1967) have placed the Cedar Valley Formation of northern Iowa in the late Middle Devonian. The formation was deposited during the Tioughniogan and Taghanic Stages (Collinson and others, 1967).

The Cedar Valley was deposited in a shallow sea under slightly reducing conditions. As the sea transgressed across the erosion surface, debris from the underlying Maquoketa was incorporated into the basal part of the unit. As very little non-Maquoketa debris is present in the basal Cedar Valley, it is probable that the Cedar Valley sea advanced over a tectonically stable land mass with very low relief. A thickness of 305 feet in southeastern Minnesota, more than twice that in east-central Iowa, is indicative of greater subsidence in Minnesota than in southeastern Iowa (Kohls, 1961, *op. cit.*), and suggests that the Cedar Valley extended much farther beyond its present outcrop area (fig. VI-19).

Post-Cedar Valley Rocks

Non-marine rocks of Cretaceous and, perhaps, Tertiary age (Bleifuss, 1966, unpub. Ph.D. thesis, Univ. Minn.; see also this chapter) cover much of southeastern Minnesota in thin discontinuous beds that lie disconformably on rocks ranging in age from Cambrian to Devonian. To the west, these rocks interfinger with and are overlain by marine shales and sandstones of Late Cretaceous age (Sloan, 1964). Bedrock exposures of these rocks are commonly limited to deeper stream valleys because of the thick mantle of Pleistocene materials.

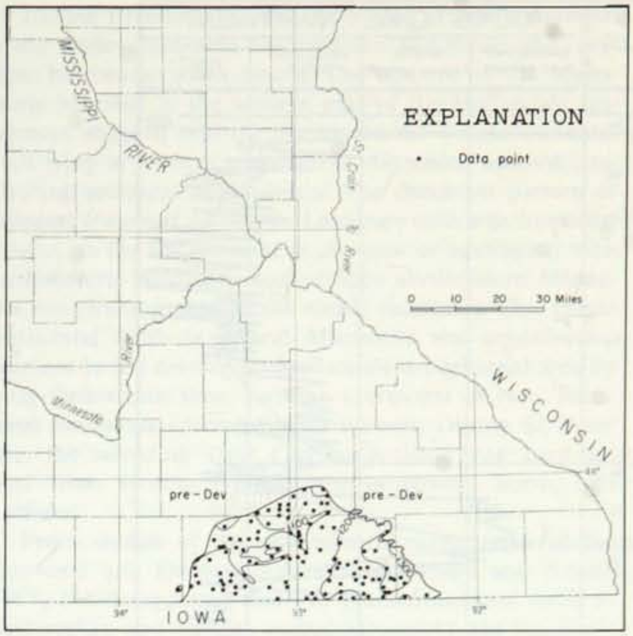

Figure VI-19. Bedrock topographic contour map of the present altitude of the eroded top of the Cedar Valley Formation in southeastern Minnesota (in feet above sea level).

SUMMATION OF THE LITHIC ENVIRONMENT DURING THE DEPOSITION OF PALEOZOIC STRATA IN SOUTHEASTERN MINNESOTA

Cyclic Sedimentation

Cyclic sedimentation in the Cambro-Ordovician rocks in the upper Mississippi valley has been described by Ostrom (1964), and I have applied this concept to the Paleozoic rocks of southeastern Minnesota (Austin, 1970b). Ostrom (1964) identified four recurrent lithotopes that characterize the rocks of the region: (1) well-sorted quartzarenite; (2) poorly-sorted unit of mixed lithologies; (3) shale or argillaceous sandstone; and (4) carbonate rock. He defined five successive episodes of submergence and emergence of the depositional shelf in the continental interior during early Paleozoic time, which are reflected in the lithologies of pre-Cincinnatian rocks (fig. VI-20). The cycles resulted from repeated emergence and submergence, the former caused by rejuvenation of tectonically positive parts of the craton, and the latter resulting from submergence of the Appalachian geosynclinal basin far to the south and east and the neighboring shelf area of the craton. In applying Ostrom's concept to the Paleozoic rocks of southeastern Minnesota, I defined nine cycles of recurrent lithotopes that characterize the rocks deposited during early and middle Paleozoic time (Austin, 1970b).

All Paleozoic rocks of southeastern Minnesota are shallow-water deposits, and the terms "transgressive" and "regressive" denote vertical changes in the sequence of grain size, sorting, and detrital/nondetrital ratio of the rocks. The recurrent lithotopes were caused by the cyclic variation of the amount, mineralogy, and grain size of the clastic influx into the basin, and reflect the cyclic depositional environments and, to a lesser degree, the environment and proximity of the source area. My approach to cyclic sedimentation in southeastern Minnesota resulted in (1) identifying incomplete cycles, (2) extending the cycles into Devonian time, (3) distinguishing the Hollandale embayment as a depositional area where both regressional and transgressional facies were developed, and (4) identifying a gradual but detectable shift in sedimentation from predominant sandstone and subordinate carbonate near the base of the Paleozoic cycles to predominant carbonate and subordinate sandstone in the upper cycles (fig. VI-21).

The differences between the regional pattern and the pattern in southeastern Minnesota are small and, in general, Ostrom's concept can be used in interpreting the succession in Minnesota. The differences in lithology and in position of unconformities from the regional pattern primarily result from the greater stability of the Hollandale embayment as a subsiding depositional area and from shifting sources for clastic material. In Minnesota, the paleoslope was not continuously toward the southeast during Paleozoic time, as was suggested by Ostrom. Further, the direction of sediment transport differs from the regional northeast-to-southwest pattern identified by Ostrom.

Direction of Sediment Transport

Studies indicate that the direction of sediment transport and of the paleoslope in southeastern Minnesota may have varied with time as a result of uplift and degradation of different parts of the Transcontinental Arch-Wisconsin Dome positive area and the development of structures that affected sedimentation in the Paleozoic seas. During Late Cambrian time, the apparent source of clastic material was the Wisconsin Dome, which extended into northeastern Minnesota. Erosion of this positive area and subsidence of the Wisconsin Arch at a more rapid rate than the southeastern Minnesota area resulted in: (1) thicker sequences of Mt. Simon, Eau Claire, Galesville, and Ironton strata in extreme southeastern Minnesota (fig. VI-6), Wisconsin (Ostrom, 1967), and northern Illinois (Buschbach, 1964) than in the remainder of Minnesota; (2) the decrease of grain size in the Eau Claire and Galesville formations from Wisconsin toward southeastern Minnesota; and (3) the reduction of labile constituents in the Mt. Simon from as much as 40 percent in Wisconsin to 1 to 7 percent in Minnesota (Thiel and Crowley, 1940).

	Depositional environments			
Cycles	Beach-nearshore	Nearshore shelf	Depositional shelf	Reef
5	St. Peter Fm.	Nokomis Mbr.	Harmony Hill Mbr.	Ottawa Group
4	New Richmond Fm.	?	Present, but unnamed	Shakopee Fm.
3	Jordan Fm.	Madison Mbr.	Blue Earth Mbr.	Oneota Fm.
2	Galesville Fm.	Ironton Fm.	Franconia Fm. and Lodi Mbr.	Black Earth Mbr. (Trempealeau Fm.)
1	Mt. Simon Fm.	"U. Mt. Simon"	Eau Claire Fm.	Bonneterre Fm.

Figure VI-20. Strata comprising Ostrom's five pre-Cincinnatian Paleozoic sedimentary cycles in the upper Mississippi River valley (after Ostrom, 1964).

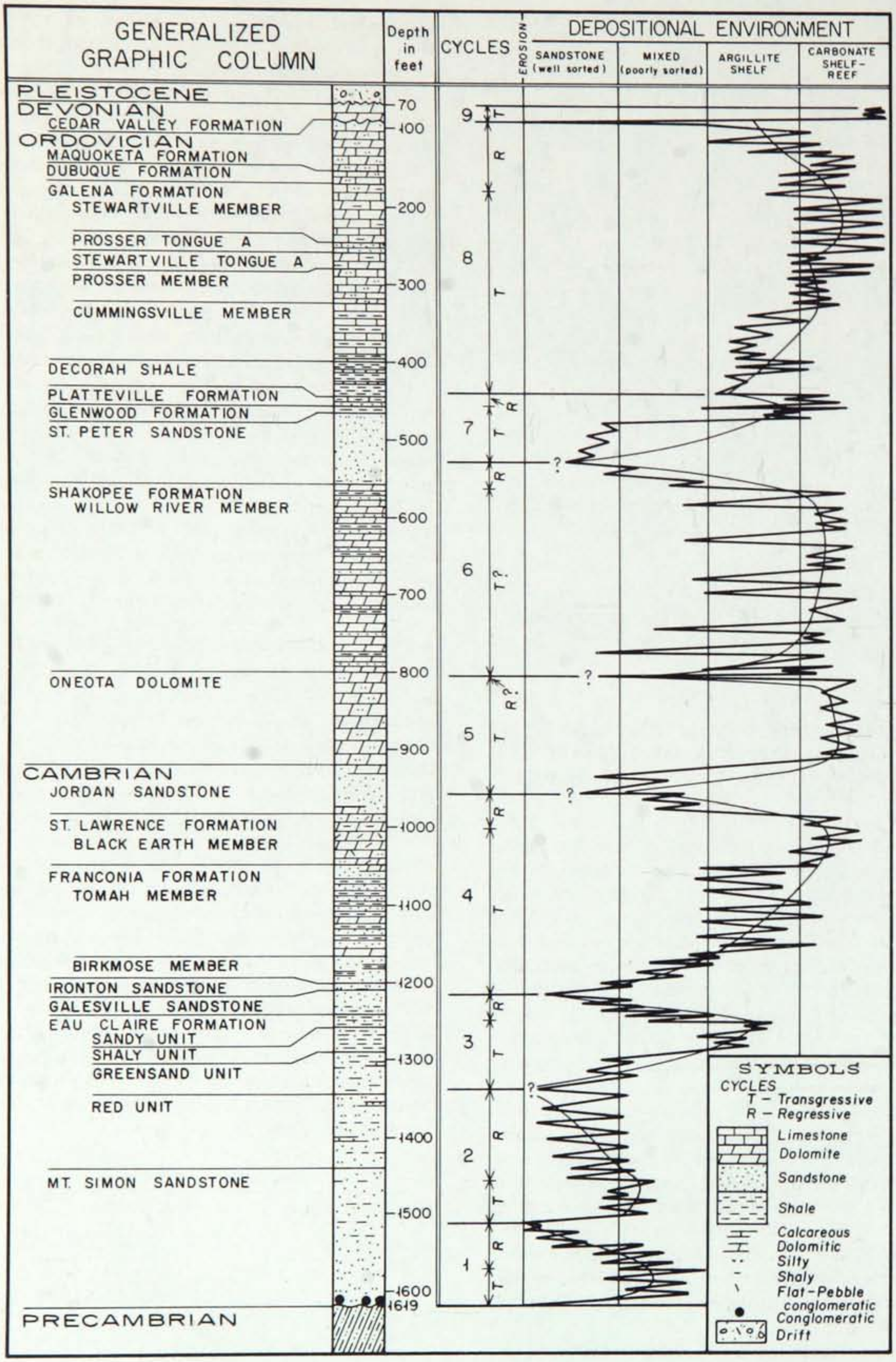

Figure VI-21. Generalized graphic log and cyclic depositional environments of rock units in a deep stratigraphic test well near Hollandale, Minnesota. Major cycles are outlined by the thin line superimposed on the irregular thick line which indicates minor fluctuations in the depositional environment (after Austin, 1970b).

During Franconian time, the source of clastic material in the embayment was northwestern Wisconsin and, perhaps, northeastern Minnesota. The absence of the Mazomanie Member in the western part of the Hollandale embayment suggests that the portion of the Transcontinental Arch lying in western and central Minnesota was not contributing sediment to the basin. The dominant pattern of sediment transport during St. Lawrence time was from east to west, as the sea widened to the west or southwest, with northwestern Wisconsin and perhaps northeastern Minnesota contributing most of the clastic sediments. The Transcontinental Arch in central Minnesota was contributing sediment to the developing Hollandale depositional area by Early Ordovician time, forming a crescent of New Richmond Sandstone around the embayment. During St. Peter time, the subsiding Twin City basin filled with sand derived from sources lying to the northwest, north, and northeast.

From studies of the clay mineral assemblages of the Glenwood and Decorah formations, Parham and Austin (1967, 1969) suggested that the Transcontinental Arch in southwestern and perhaps central Minnesota was the dominant positive area during Black Riveran and Trentonian time. The Decorah is present in the Twin City basin (Parham and Austin, 1969); however, this area was relatively far from the source of the clay minerals. This suggests that the Twin City basin was subsiding more rapidly than the surrounding area, and acting as a "sediment trap" during Decorah time. By the end of Middle Ordovician time, little detrital material was derived from the old positive area, but a local uplift of central Minnesota in Late Ordovician time resulted in the influx of sand into the upper or Clermont Member of the Maquoketa Formation.

After uplift and erosion of the Ordovician and older strata in the embayment, the sea returned during Middle Devonian time and deposited the Cedar Valley Formation on an eroded surface of low relief, which was subsiding relatively more rapidly to the north. Some time after deposition of the Cedar Valley Formation, the continuing uplift of the Transcontinental Arch with subsequent erosion defined the limits of the present expression of the Hollandale embayment.

Development of Large-Scale Structural Features in Southeastern Minnesota

The Hollandale embayment and its associated smaller structural features were not clearly defined until Early Ordovician time. The first indication of the embayment as a feature affecting sedimentation is given by the distribution of facies in the Cambrian Eau Claire Formation. The occurrence of the "red unit" (Austin, 1969) along the western border and as the basal unit near the center of the embayment, and its absence on the Wisconsin Arch, indicates that the Transcontinental Arch was emergent and contributing sediment to the embayment. From the color and grain size of this unit, the source of the sediments presumably was the Keweenawan redbeds, which now underlie the center of the embayment but which in Late Cambrian time stood as a positive area to the west. The decrease from western Wisconsin toward the embayment in the amount and grain size of the sand-size fraction of the other units within the Eau Claire suggests that the source of this clastic material was to the east. That the embayment affected sedimentation during deposition of the Franconia Formation is indicated by the semicircular pattern of the highly glauconitic members of the Franconia Formation around the center of the embayment. However, it was not until Early Ordovician time, when southeastern Minnesota was covered by the shallow Prairie du Chien sea, that the subsiding center of the embayment was defined by thickening of the units (fig. VI-13) and a semicircular ring of the New Richmond Member of the Shakopee Formation.

The Twin City basin was outlined in Early Ordovician time. The isopach contours of the Prairie du Chien clearly show that sedimentation in the Twin City basin was restricted during Early Ordovician time. Analysis of drill data suggests that the contact between the Shakopee Formation and the overlying St. Peter Sandstone of Middle Ordovician age may not be erosional, as is the case in Wisconsin, and that isopachs accurately define the thickness of the Prairie du Chien Group prior to deposition of the St. Peter Sandstone. During St. Peter time, the Twin City basin was subsiding more rapidly than the remainder of southeastern Minnesota, and accordingly the St. Peter Sandstone is thicker here than in other parts of the state.

Movements along the Belle Plaine fault caused faulting or folding of the Paleozoic units along the western margin of the Hollandale embayment. This feature affected the altitude of units at least as far away from the Minnesota River as Waseca, Minnesota. The age of the movement is definitely post-Black Riveran (fig. VI-16), and probably is post-Ordovician (Sloan and Danes, 1962).

Isostatic adjustments in Paleozoic time along Precambrian faults east and south of the Twin City basin produced the Hudson-Afton anticline and the Vermillion anticline (Morey and Rensink, 1969). The movements, at least along the Vermillion anticline, appear to have been recurrent and concurrent with sedimentation from Mt. Simon through Jordan time, but ceased before Early Ordovician time (G. B. Morey, 1970, oral comm.).

The time of the initiation of the Red Wing-Rochester anticline as a structural feature, which also affected sedimentation, is not clear. The New Richmond Member of the Shakopee Formation is less than 10 feet thick west of the anticline and increases to 65 feet east of the anticline, in extreme southeastern Minnesota. However, the lithologies of the underlying and overlying units apparently do not differ across the structure. Most likely, therefore, the Red Wing-Rochester anticline is a post-Ordovician feature that was produced by gentle warping of the Paleozoic strata during subsidence of the Hollandale embayment.