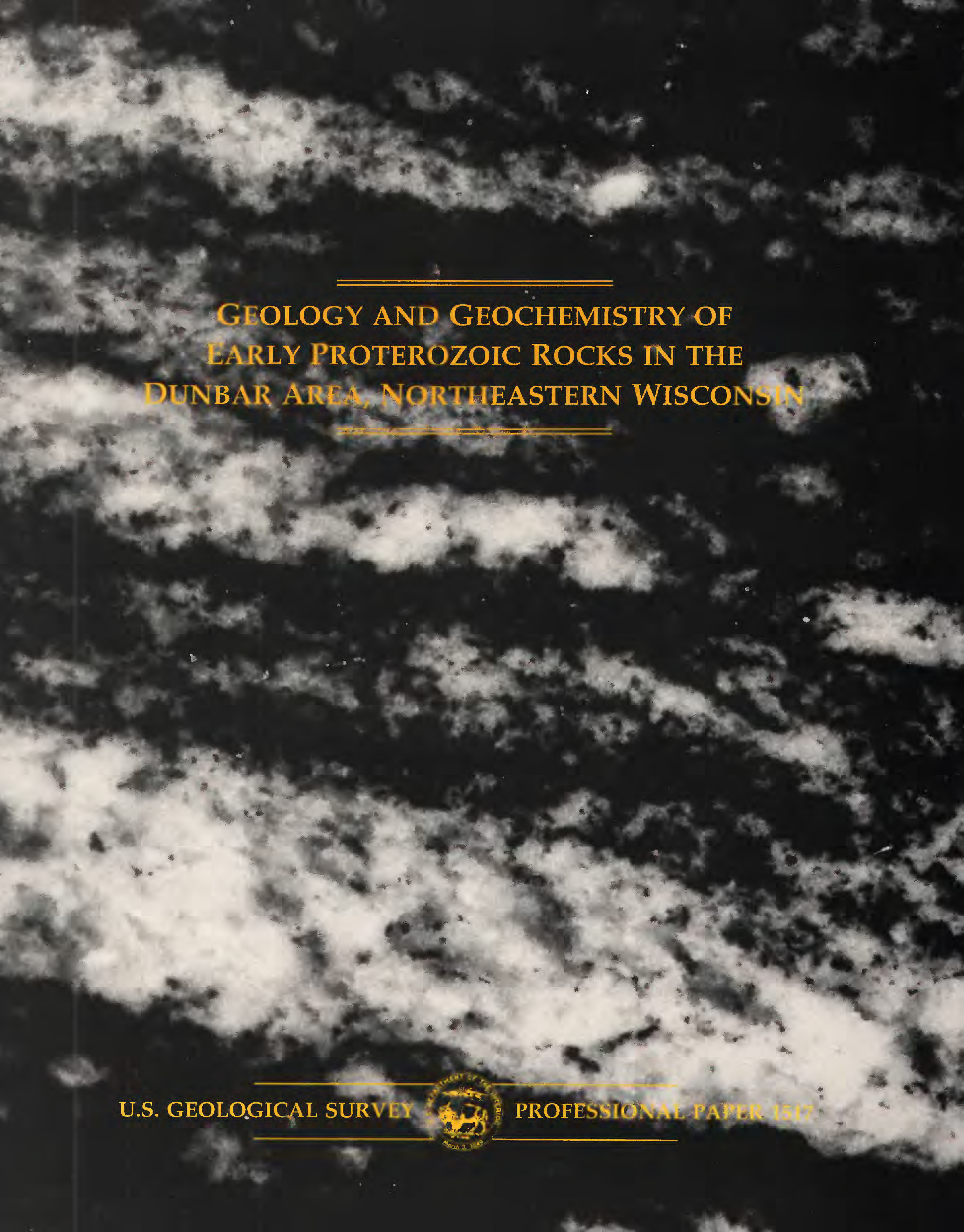

Geology and Geochemistry of Early Proterozoic Rocks in the Dunbar Area, Northeastern Wisconsin

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By P.K. SIMS, K.J. SCHULZ, *and* Z.E. PETERMAN

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GEOLOGY AND GEOCHEMISTRY OF EARLY PROTEROZOIC ROCKS IN THE DUNBAR AREA, NORTHEASTERN WISCONSIN

By P.K. SIMS, K.J. SCHULZ, and Z.E. PETERMAN

ABSTRACT

The Dunbar area, in northeastern Wisconsin, is immediately south of the Niagara fault zone, the tectonic boundary between the Early Proterozoic continental margin to the north and the Wisconsin magmatic terranes of the Penokean orogen to the south. The geology appears to be typical of the northern part of the Wisconsin magmatic terranes, an oceanic arc complex of volcanic and granitoid rocks that was accreted to the North American continent about 1,850 million years ago. East of the Dunbar area, the Niagara fault zone contains a dismembered ophiolite and may represent a paleosuture zone.

The complex structure of the Dunbar area has resulted from one or more northward-verging thrust faults that juxtaposed volcanic and granitoid arc rocks against sedimentary rocks and Archean basement of the continental margin. The present structure is an irregular dome, composed of several discrete bodies of granitoid rocks and a large surrounding area comprising volcanic rocks assigned to the Quinnesec Formation and a comagmatic volcanic-arc intrusion, the Newingham Tonalite. On the west margin, metasedimentary rocks having continental-margin affinities separate granitoid rocks of the dome and the Quinnesec Formation.

The granitoid rocks within the dome consist of, from oldest to youngest, the Dunbar Gneiss, the Marinette Quartz Diorite, the Hoskin Lake Granite, and the Bush Lake and Spikehorn Creek Granites; the latter two are possibly coeval. The Dunbar and the Marinette constitute the core of the dome. The Bush Lake and Spikehorn Creek Granites compose lateral protuberances (lobes) from the core. The Newingham Tonalite composes the southern part of the core and a prominent protuberance.

The Dunbar Gneiss, which constitutes a substantial part of the core, consists mainly of layered biotite gneisses that are predominantly granodiorite in composition and calc-alkaline in character. Its gneissic fabric probably resulted from deformation of an incompletely crystallized magma. The Marinette Quartz Diorite is a layered mafic intrusion of alkaline-calcic to calc-alkaline composition that was intruded as a generally concordant body along the contact between the Dunbar Gneiss and the Quinnesec Formation; its composition was partly controlled by crystal accumulation. The granite bodies intrude the Marinette and a narrow fringe of the Quinnesec Formation adjacent to the core. The Newingham Tonalite is a homogeneous body that intrudes the volcanic rocks of the Quinnesec Formation and within the core is in turn intruded by the Marinette Quartz Diorite. Crystalline rocks in the core of the dome were variably metasomatized by a source like the Hoskin Lake Granite, resulting in the addition of potassium feldspar and large-ion-lithophile elements.

The Newingham Tonalite has the compositional characteristics of the high Al_2O_3-type trondhjemite-tonalite, consistent with volcanic-arc intrusions found in subduction-related settings. In contrast, the Dunbar Gneiss, Marinette Quartz Diorite, and the three granite bodies in the dome show compositional affinities with syn- to post-collisional granites present in Holocene orogenic belts. Their chemical composition, particularly high Ta and Nb concentrations, suggests magma derivation from and (or) magma interaction with within-plate-type basaltic melts possibly generated from continental lithosphere during the collisional event (Penokean orogeny).

Uranium-lead zircon ages of these rocks range from 1,865 to 1,835 Ma, which interval presumably represents the time span of major rock-forming and structural events in the Dunbar area. The older rocks, the Dunbar Gneiss, Marinette Quartz Diorite, and Quinnesec Formation, about 1,865 Ma, are variably foliated and deformed, whereas the youngest dated rock, the Spikehorn Creek Granite, is virtually undeformed. Rubidium-strontium whole-rock and biotite ages are variably reset and are always 100 million years or more younger than zircon ages.

The Dunbar dome is about 200 square kilometers in area and is characterized by a consistent parallelism of structures in the cover rocks and in the margins of the crystalline core, and by intensely foliated and lineated rocks, indicative of high strain, along its north margin. The dome is interpreted as a large-scale fold-interference structure resulting from crossfolding; it was modified by intrusion of granitoid rocks.

A zone of prograde metamorphism extends beyond the north margin of the dome. An amphibolite-facies inner zone, ranging from less than 0.5 to at least 8 kilometers wide, occurs in the Quinnesec volcanic rocks and metasedimentary rocks adjacent to the core and has resulted from prograde metamorphism of regional greenschist-facies rocks. The amphibolite zone crosses the Niagara fault (suture) zone, passing into rocks of the continental-margin assemblage, indicating that the culmination of this thermal metamorphism postdated collision. The crystalline rocks in the northern part of the core of the dome were metamorphosed to amphibolite facies and metasomatized during this thermal event. The source of the heat and metasomatism was the Hoskin Lake Granite or a related deeper granitic body.

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Our geologic mapping and related studies have shown that the boundary zone between the Wisconsin magmatic terranes and the epicratonic continental-margin assemblage both is complex and is a tectonic collage of lithologies having affinities with continental as well as oceanic arc terranes. Presumably, rocks of the two disparate terranes were juxtaposed by northward-directed thrusting during the Penokean orogeny, about 1,850 Ma.

INTRODUCTION

The Dunbar area, in northeastern Wisconsin, is immediately south of the juncture between the two major crustal segments in the Early Proterozoic Penokean orogen in the eastern part of the Lake Superior region (fig. 1): (1) a continental-margin epicratonic assemblage to the north, and (2) a complex island-arc sequence to the south. The continental-margin assemblage, formally designated as the Marquette Range Supergroup (Cannon and Gair, 1970), is a generally southward thickening, few thousand-meters-thick accumulation of sedimentary and bimodal volcanic rocks that lies on an Archean crystalline basement (Morey and others, 1982). The island-arc sequence, termed the Pembine-Wausau terrane of the Wisconsin magmatic terranes (Sims and others, 1989), consists of mafic to felsic volcanic rocks, subordinate sedimentary rocks, and granitoid intrusive rocks largely of calc-alkaline affinity. The two major terranes are juxtaposed along the Niagara fault (Sims and others, 1985), a major ductile deformation zone interpreted as a paleosuture (Sedlock and Larue, 1985).

The Dunbar area provides a key to understanding the stratigraphic and tectonic evolution of Early Proterozoic rocks in the northern part of the Wisconsin magmatic terranes. These rocks are fairly well exposed, compared to other areas in northern Wisconsin, and include a moderate-sized gneiss unit that intrudes a widespread, thick succession of metavolcanic rocks. Granitoid rocks that generally are foliated intrude the gneiss and surrounding volcanic rocks. These rocks taken together appear to be roughly correlative with other bodies of less well exposed gneiss and granitoid rocks in northern Wisconsin (Morey and others, 1982), which have comparable isotopic ages of about 1,850 Ma (Sims and Peterman, 1980; Sims and others, 1989) as well as chemical similarities (Schulz, 1983) to rocks in the Dunbar area.

The Dunbar area lies a short distance south of the former iron ore-producing Menominee range, which was mapped previously by Bayley and others (1966) and Dutton (1971). Also, Cain (1964) made an earlier reconnaissance geologic map of the Dunbar area. As a consequence of this mapping, most of the major rock units were named and described prior to our study. Prinz (1959; 1965) named the Marinette Quartz Diorite and the Hoskin Lake Granite, important units in the northern part of the area, and Cain (1964) named the Dunbar Gneiss and the Newingham Granodiorite, major rock units to the south. The volcanic unit, the Quinnesec Formation, was named "Quinnesec schists" by Van Hise and Bayley (1900).

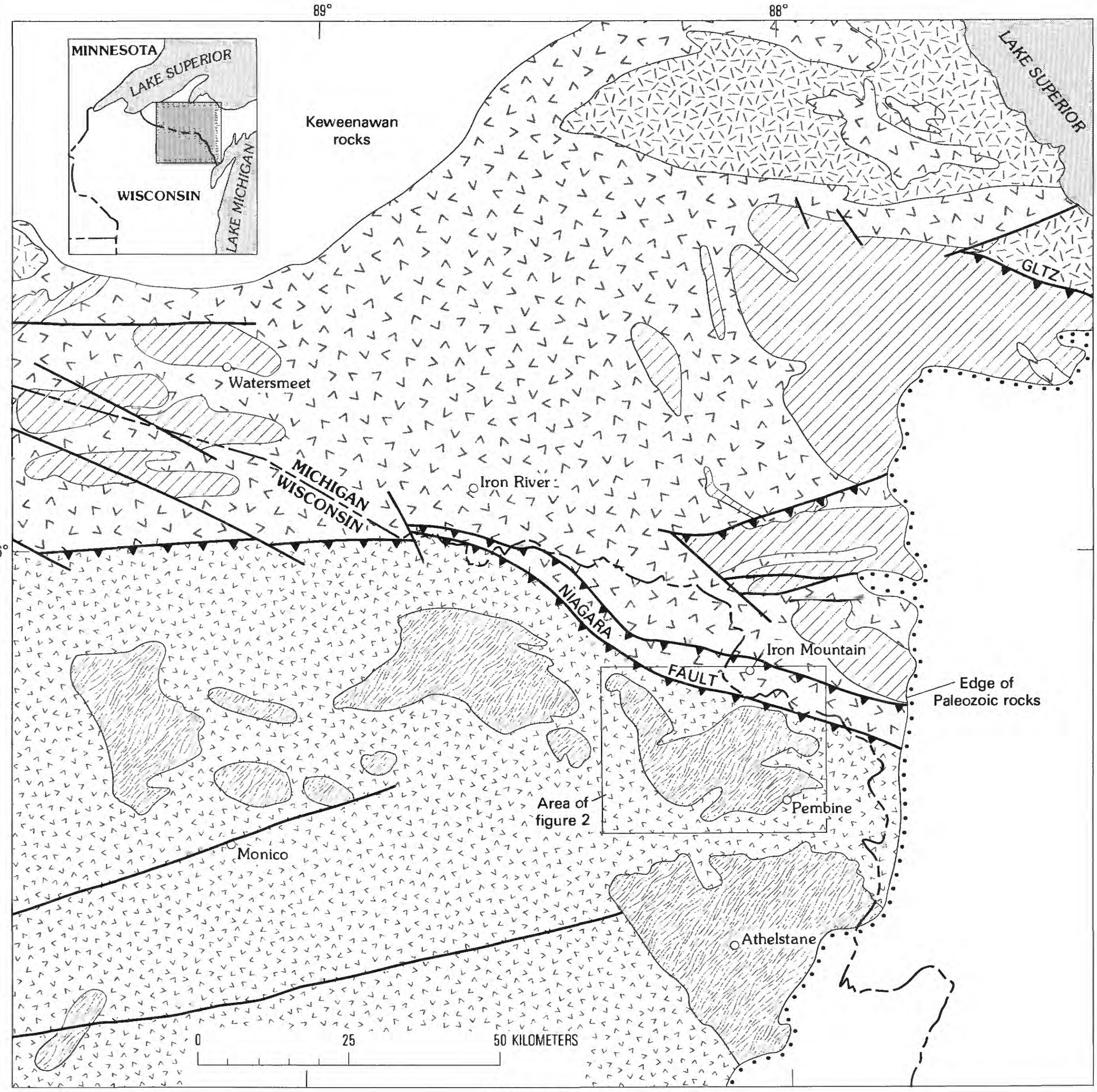
Geologic mapping of the rocks in the central part of the Dunbar dome has been difficult, because the Newingham Tonalite and the Marinette Quartz Diorite were partially metasomatized subsequent to their solidification, producing rocks that are similar mineralogically to the Dunbar Gneiss. This convergence in modal mineralogy of the three major lithotypes in the region and the lack of continuous outcrops led to uncertainties in rock designations, which were not resolved satisfactorily until abundant data on minor element chemistry became available. Accordingly, figure 2 of this report differs in some minor respects from earlier maps (Sims and others, 1985; Sims and Schulz, 1988).

The main purpose of this study was to gain a better understanding of the geologic evolution of this part of the Wisconsin magmatic terranes and its relationship to the adjacent continental-margin assemblage in Michigan. As part of the study, we mapped in detailed reconnaissance fashion the Dunbar and Dunbar NE 7½-minute quadrangles (Sims and Schulz, 1988) mainly during the summers 1980–1983, and the northern part of the Pembine 15-minute quadrangle (Schulz, unpub. data, 1990). Samples of representative rocks were collected from these areas as well as from previously mapped areas, for petrographic and chemical studies and isotopic age determinations. The locations of outcrops discussed in this report are shown on the geologic map of the Dunbar and Dunbar NE quadrangles.

This report describes the field relations, structure, geochemistry, and ages of major rock units in the Dunbar area. The geochemistry of the granitoid rocks and their relation to Early Proterozoic tectonic regimes are emphasized. We conclude that the major rock bodies in the Dunbar area were formed in two disparate tectonic environments that later were tectonically juxtaposed, presumably by large-scale thrusting.

ACKNOWLEDGMENTS

We acknowledge the assistance of M.G. Mudrey, Jr., particularly during early reconnaissance mapping, and G.L. LaBerge, who mapped a small area northwest of the Dunbar dome during 1982. In addition to analysts



EXPLANATION

WISCONSIN MAGMATIC TERRANES

EARLY PROTEROZOIC

Metavolcanic and subordinate metasedimentary rocks

Gneiss and granitoid rocks (1,890–1,835 Ma)

CONTINENTAL-MARGIN ASSEMBLAGE

EARLY PROTEROZOIC

Metasedimentary and metavolcanic rocks of Marquette Range Supergroup (2,100–1,850 Ma)

ARCHEAN

Greenstone-granite complexes (2,750–2,600 Ma)

Gneiss terrane (3,550–2,600 Ma)

Approximate contact

Thrust fault—Sawteeth on upper plate

Fault—Relative movement not known

GLTZ **Great Lakes tectonic zone**

FIGURE 1.—Generalized geologic-tectonic map of northern Michigan and northern Wisconsin.

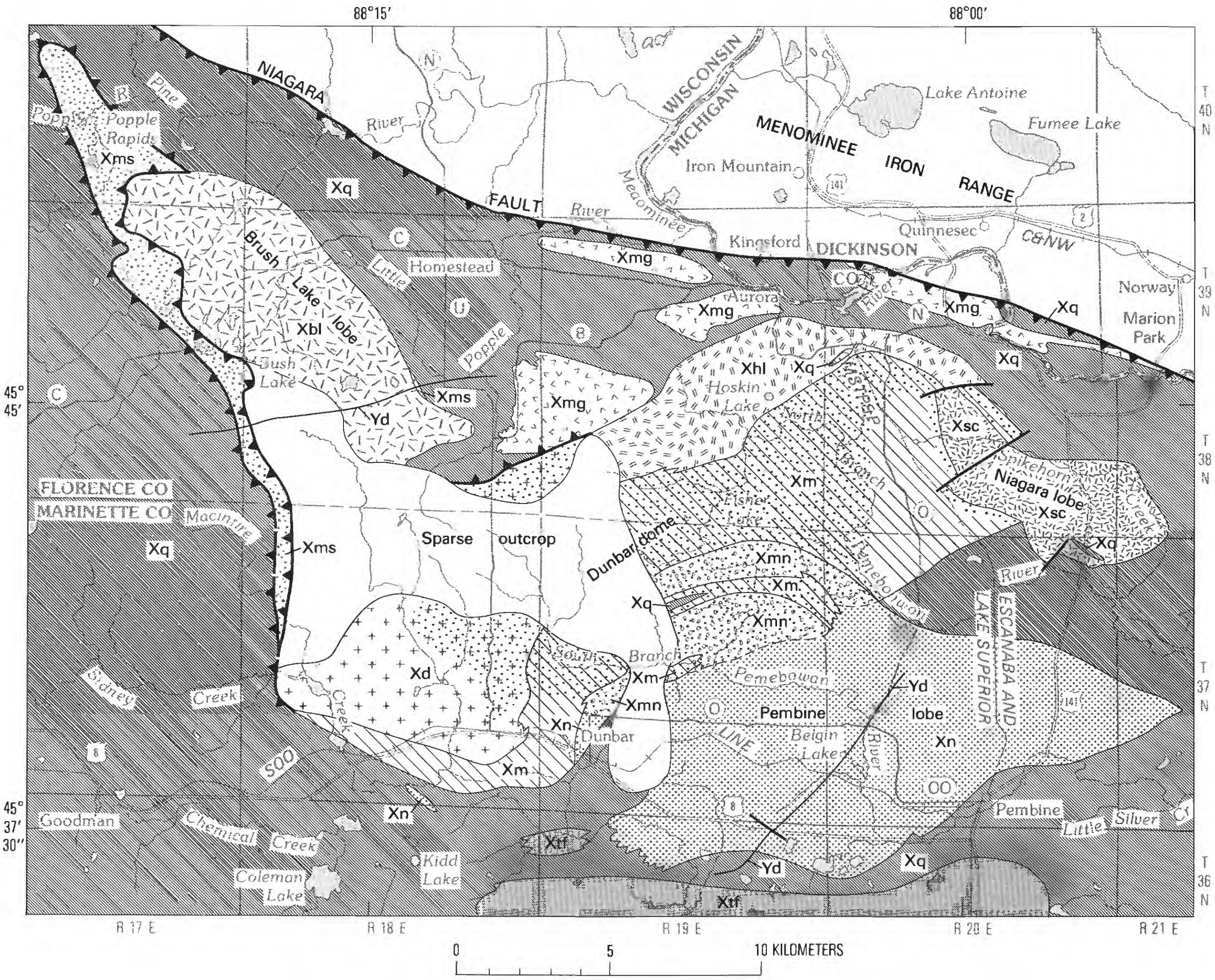
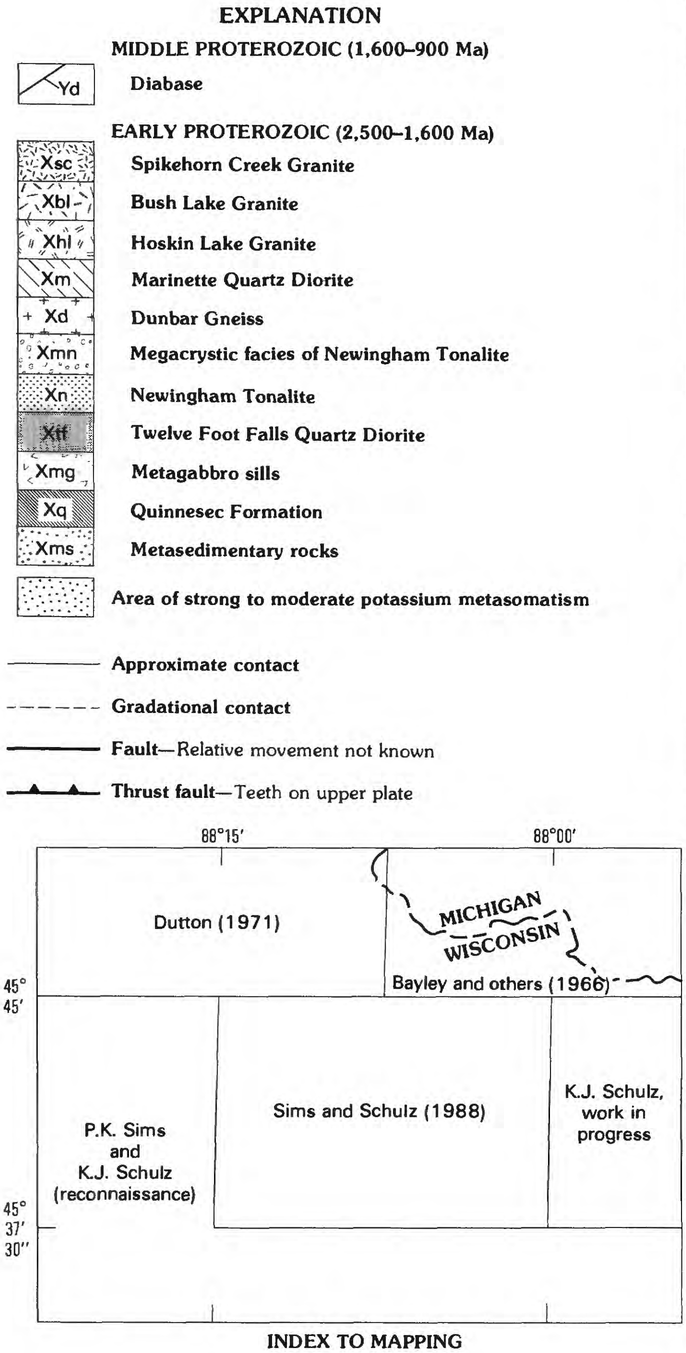


FIGURE 2 (above and facing page).—Geologic map of Dunbar area, northeastern Wisconsin. Compiled by P.K. Sims, 1988. Base from Iron Mountain, Mich.-Wis. 1:100,000 quadrangle.

credited in the text, Ross Yeoman provided X-ray fluorescence analyses of minor elements. Robert L. Jenkins and W.C. Prinz introduced us to the geology of the area in 1975. In this report, Sims is responsible for the geologic mapping, petrographic studies, and structural interpretation; Schulz is responsible for the geochemical data; and Peterman is responsible for the isotopic age data. Edward du Bray, Russell Tysdal, M.G. Mudrey, Jr., and H.J. Stein made many helpful comments on drafts of the manuscript.

ROCK UNITS

Two major groups of rocks are exposed in the Dunbar area. These are (1) Dunbar Gneiss, Marinette Quartz Diorite, and three granite bodies—Hoskin Lake, Spikehorn Creek, and Bush Lake Granites—which compose a complex domal structure, the Dunbar dome, and (2) the Quinnesec Formation and associated intrusive bodies—the Newingham Tonalite, the Twelve Foot Falls Quartz Diorite, and unnamed metagabbro—which surround the crystalline core (fig. 2). All these rocks were formed during the relatively short interval 1,865–1,835 Ma. Distinct chemical differences between the two groups of rocks suggest that they were formed in different tectonic environments. The Quinnesec Formation and associated intrusive bodies appear to have formed in an oceanic island-arc. Subsequent collision of the arc terrane with the continental-margin assemblage contributed to the generation and



juxtaposition of the gneiss and associated granitoids in the Dunbar dome.

The stratigraphic nomenclature used by previous authors is retained in this report with minor modifications. The formal terms Dunbar Gneiss, Marinette Quartz Diorite, Quinnesec Formation, and Twelve Foot Falls Quartz Diorite are retained. The formal term Hoskin Lake Granite is retained for the granite body in the type area, but the other two granite bodies—previously assigned to the Hoskin Lake Granite (Dutton and Linebaugh, 1967; Dutton, 1971)—have been renamed the Spikehorn Creek Granite and the Bush Lake Granite, respectively (Sims and Schulz, 1988). Also, the Newingham Granodiorite of Cain (1964) has been redescribed as the Newingham Tonalite (Sims and Schulz, 1988), because our mapping indicates that the original tonalite composition was subsequently modified in the southern part of the dome by potassium metasomatism. We have designated the hybrid tonalite informally as the megacrystic facies of the Newingham Tonalite; it is primarily granodiorite in composition.

ROCK TEXTURE TERMINOLOGY

The crystalline rocks in the Dunbar area have been variably deformed by ductile deformation, resulting in grain-size reduction by syntectonic recrystallization; this grain-size reduction was accompanied in places by growth of potassium feldspar porphyroblasts. According to the terminology of Wise and others (1984), many of the rocks are protomylonites or orthomylonites. These rocks have a mylonitic foliation, and generally contain variable percentages of concentrically zoned plagioclase phenocrysts that have survived destruction through crystal-plastic grain-size reduction. The resultant microstructure is a predominantly mortar structure or core and mantle structure (White, 1976) in which new, smaller grains occur adjacent to older host grain boundaries and within zones of localized high intragranular strain such as isolated shears. Hanmer (1982) has termed this type of new grain aggregates as type IP and type IM aggregates. Type IP aggregates result from recrystallization of the host plagioclase, and type IM aggregates result from recrystallization of the host potassium feldspar. A less common microstructure existing in the rocks and resulting from relatively lower bulk strains has been termed type IIP new grain aggregates (Hanmer, 1982). This type of aggregate is characterized by an intragranular distribution of new grains independent of host grain boundaries and shears, and results from microfracturing and rotation of domains within a host plagioclase grain. Bent plagioclase grains and undulose extinction in plagioclase, microcline, and quartz are other textural characteristics observed in the rocks.

DUNBAR GNEISS

GENERAL FEATURES

The Dunbar Gneiss is a succession of interlayered biotite gneisses and subordinate amphibolite exposed in the vicinity of Dunbar (Sims and Schulz, 1988).

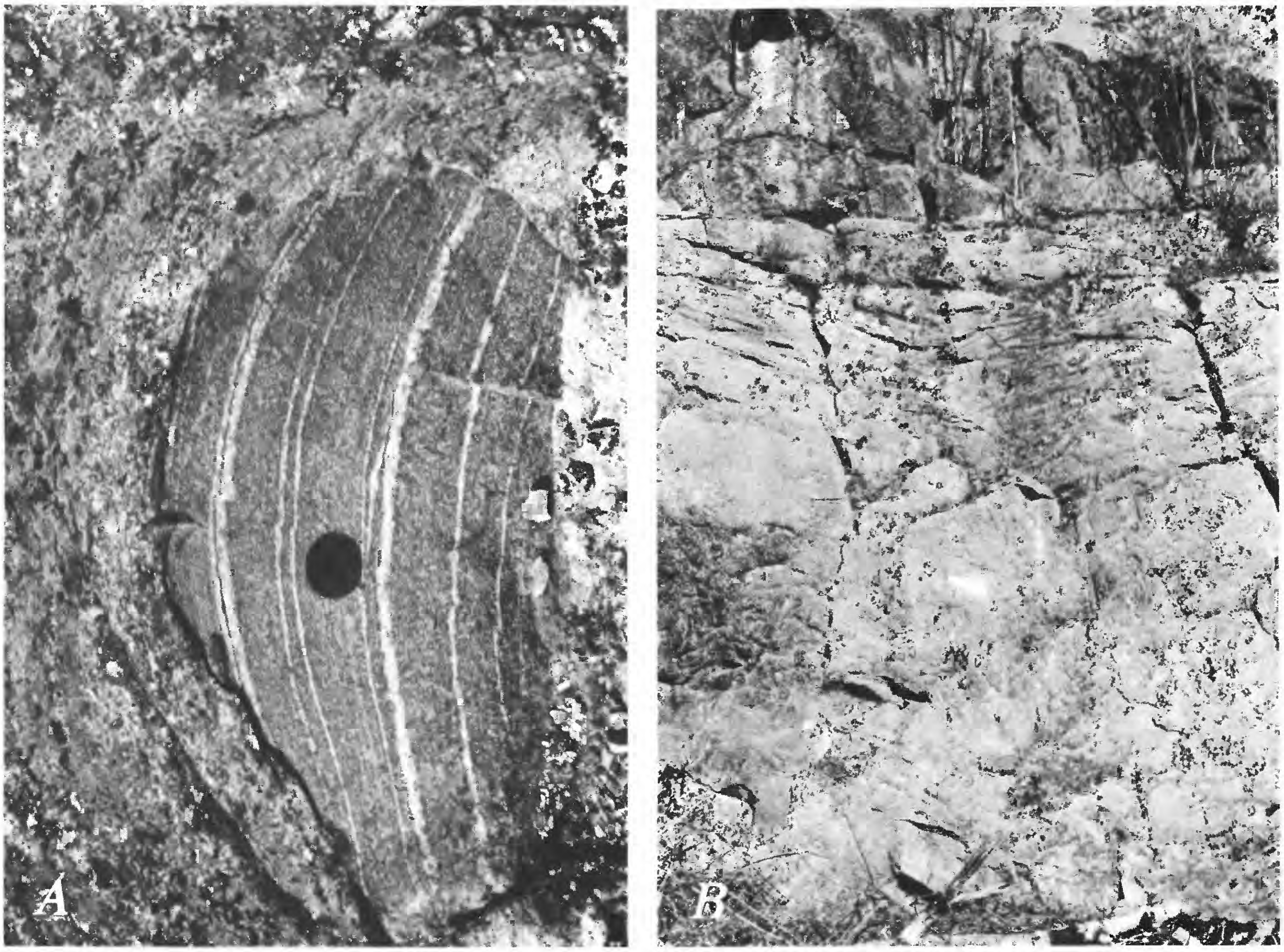


FIGURE 3.—Dunbar Gneiss outcrops. *A*, Thinly banded gneiss surrounded by late, coarse-grained pegmatite (sec. 26, T. 37 N., R. 18 E.). Lens cap (5 cm) for scale. *B*, Relatively massive (lower part of photo) to thickly banded gneiss. Rock face is about 5 m high (sec. 24, T. 37 N., R. 18 E.).

Some exposures in secs. 13, 14, 15, and 21, T. 37 N., R. 18 E. are large and spectacular. Other good exposures, including freshly blasted rock, are along Marinette County Highway U in secs. 13 and 14. Granite pegmatite and aplite intrude the Dunbar in the western part of the dome and not uncommonly compose 50 percent or more of outcrops. A few outcrops of the unit occur in sec. 36, T. 38 N., R. 18 E., near the north margin of the dome.

The gneiss is folded on northwest-trending axes (F_2). Foliation (S_1) is subparallel to layering, and an axial-planar foliation is developed only rarely.

The gneiss is intruded locally by dikes of the Marinette Quartz Diorite, but the contact between the two subconformable bodies is not exposed. In the western part of the dome, the gneiss is in contact with metasedimentary rocks (fig. 2); this contact could be tectonic. The contact of the gneiss with the surrounding Quinnesec Formation has been largely obliterated by the intrusive Marinette Quartz Diorite, which commonly intervenes between the two rock types.

DESCRIPTION

The Dunbar Gneiss consists predominantly of gray biotite gneiss units, which are layered at scales ranging from a few centimeters to several meters (fig. 3). The layering generally is expressed by differences in the amount and kind of the major minerals but also by differences in grain size. Typically, the rocks range from fine grained to the coarse side of medium grained and are inequigranular; layers containing megacrysts of plagioclase and (or) potassium feldspar as much as 2 cm in length are moderately common. Rarely, amphibolite layers are intercalated with the biotite gneiss. Leucogranite dikes cut the gneiss at several localities.

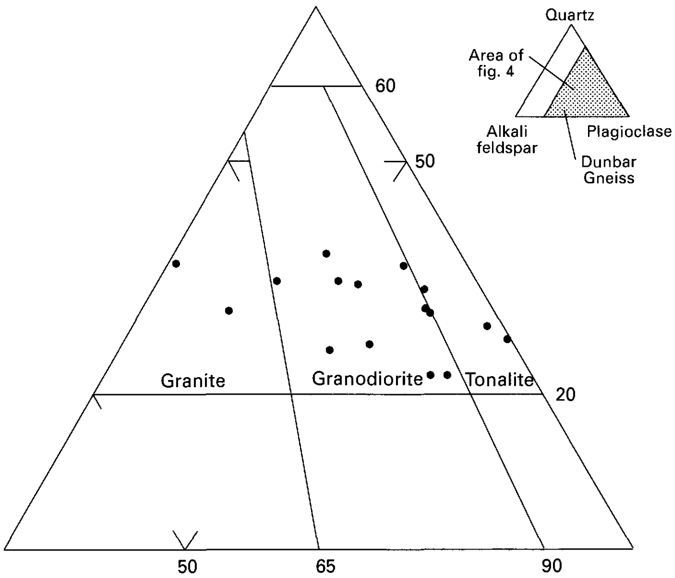


FIGURE 4.—Quartz–alkali feldspar–plagioclase diagram showing composition of representative rocks in the Dunbar Gneiss. Rock classification modified from Streckeisen (1976). Modes determined by 600 to 1,000 point counts on standard thin sections.

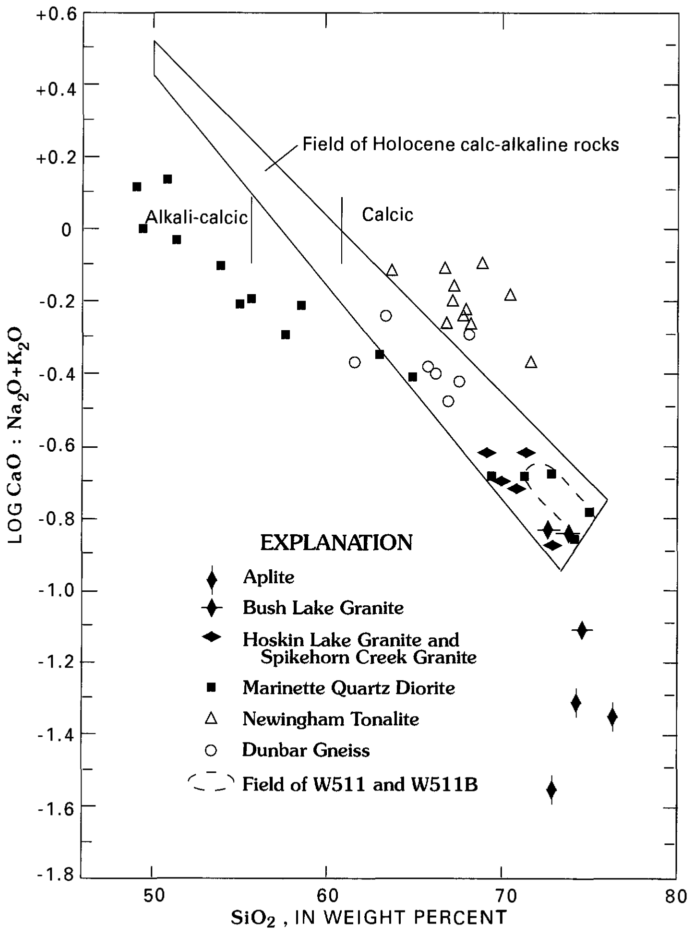


FIGURE 5.—Modified Peacock diagram showing rocks of Dunbar area. Field for Holocene calc-alkaline rocks (modified from Brown, 1982), included for comparison. Samples numbered W511 and W511B, high-silica, minimum-melt granites, discussed in text, p. 19.

The biotite gneisses are mylonitic rocks. They have a dominant xenomorphic granular (granoblastic) texture and a penetrative foliation expressed by oriented biotite and, less commonly, by elongate and flattened aggregates of quartz and plagioclase that are generally subparallel to compositional layering. The ductile flow textures in these rocks are superposed on and at places obliterate original magmatic textures expressed mainly by large, concentrically zoned plagioclase phenocrysts and by potassium feldspar phenocrysts.

The biotite gneisses have a moderate range in composition from layer to layer; their average and modal composition is granodiorite (table 1; fig. 4). Plagioclase (calcic oligoclase-andesine) and quartz are the principal minerals, potassium feldspar varies from 0 to about 30 percent, and biotite generally composes from 10 to 20 percent of the rock. Myrmekite and myrmekitic plagioclase are abundant. Dusky green hornblende is a local varietal mineral. Sphene (titanite) is the principal accessory mineral and composes as much as 2 percent of the rock. Plagioclase phenocrysts, as much as 2 cm in diameter, are anhedral to subhedral, and have a weak to moderate concentric zoning ranging from about An_{30} to An_{25}. Small patches of microcline (antiperthite) occur locally in the grains. The plagioclase grains have combined carlsbad-albite twins and lesser pericline twins and are commonly weakly altered to sericite and epidote-clinozoisite.

Potassium feldspar is dominantly microcline, which is fresh and either has grid twinning or is untwinned microperthite. Most potassium feldspar occurs as small interstitial grains, but larger poikiloblasts with myrmekitic plagioclase, biotite, and quartz inclusions are present locally. Biotite is dark reddish brown, generally fresh, with locally ragged outlines, and is weakly altered to epidote-clinozoisite and opaque oxides or to chlorite. Quartz occurs mainly in recrystallized aggregates showing undulose extinction; some aggregates (ribbon quartz) are elongate parallel to biotite.

GEOCHEMISTRY

Samples of Dunbar Gneiss (table 1) generally have calc-alkaline composition (fig. 5), are weakly peraluminous and corundum normative, and show approximately curvilinear compositional trends of decreasing

TABLE 1.—*Approximate modes and chemical analyses of representative samples of Dunbar Gneiss*

[Leaders (---), not determined; Tr, trace; <, less than. Major oxide analyses by Z.A. Hamlin, J.S. Wahlberg, A.J. Bartel, J. Taggart, and J. Baker. FeO, H_2O, and CO_2 analyses by Z.A. Hamlin, H.G. Neiman, G. Mason, and J. Ryder. Minor elements by X-ray fluorescence analysis by J. Storey, S. Donahey, B. Vaughn, and M. Coughlin; samples W43C, W602, W587, and W617 analyzed by Z.E. Peterman. Rb and Sr analyses for samples W143A, 679A, and 679B by isotope dilution method by K. Futa. Minor elements by instrumental neutron activation analysis by J.S. Mee and K.J. Schulz]

Sample No...	W43A	W43C	W144	W145	W143B	W143A	W679A	W679B	W140	W602	W587	W495A	W42	W588	W613	W617	W491	W742
Modal analyses (volume percent)																		
Plagioclase.......	37	45.5	38	50	51.5	47	28	43	52	51.5	49	52	45.8	46	40.8	60	42	37
Quartz.......	29	29.5	30	20.5	17	22	33	30	17	25.5	31.5	29	22	22	32.8	22	31	23
Potassium feldspar.	28	11.5	19.5	1	8.5	13	29	13	7	6.2	6	5	5.2	18.4	13.4	0	10.7	14
Biotite.......	6	12.5	11.5	27	19	14	9.5	13	20.5	16.3	13	12	23.6	13	12.6	18	14.3	14
Hornblende....	0	0	0	0	3.5	1.1	0	0	1.5	0	0	0	0	0	0	0	0	7.5
Sphene......	Tr	Tr	0.5	1	0.2	0.8	Tr	0.5	1.9	0.3	0	0.5	1.8	0.4	0.4	Tr	0	0.7
Apatite......	Tr	Tr	Tr	Tr	Tr	0.2	Tr	0.3	Tr	Tr	Tr	Tr	Tr	Tr	Tr	Tr	0.5	0.2
Opaque oxides....	Tr	0.5	0	Tr	0	0	Tr	0.2	0	0	0.5	0	0	Tr	Tr	0	0.5	0
Allanite.....	Tr	Tr	Tr	Tr	0.1	0.1	0.4	Tr	0.1	Tr	Tr	Tr	Tr	Tr	Tr	Tr	Tr	Tr
Other minerals.	Tr	0.5	0.5	0.5	0.2	1.8	0.1	Tr	Tr	0.2	Tr	1.5	1.6	0.2	Tr	Tr	1	3.6
Major oxides (weight percent) by X-ray fluorescence analysis																		
SiO_2..........	---	67.0	---	---	63.6	65.9	71.6	67.7	61.7	---	---	68.2	---	---	---	---	66.3	64.9
Al_2O_3.........	---	15.8	---	---	16.3	16	14.0	15.9	17.5	---	---	16.1	---	---	---	---	16.5	14.8
Fe_2O_3........	---	0.6	---	---	0.88	0.95	0.40	0.51	1.02	---	---	0.62	---	---	---	---	0.55	0.95
FeO..........	---	3.39	---	---	4.07	2.8	2.18	2.99	4.41	---	---	2.3	---	---	---	---	3.32	4.21
MgO..........	---	1.23	---	---	1.64	1.2	0.44	1.0	1.64	---	---	1.01	---	---	---	---	1.11	2.12
CaO..........	---	2.36	---	---	3.87	3	1.40	2.74	3.20	---	---	3.14	---	---	---	---	2.79	3.77
Na_2O........	---	3.87	---	---	4.21	3.9	2.75	3.55	4.22	---	---	3.90	---	---	---	---	4.14	2.53
K_2O..........	---	3.26	---	---	2.42	3.3	5.12	3.66	3.16	---	---	2.37	---	---	---	---	2.89	4.35
H_2O^+........	---	0.48	---	---	0.76	0.65	0.43	0.55	0.85	---	---	0.60	---	---	---	---	0.71	0.75
H_2O^-.........	---	0.02	---	---	0.06	0.07	0.14	0.14	0.02	---	---	0.04	---	---	---	---	0.04	0.11
TiO_2..........	---	0.71	---	---	0.82	0.53	0.38	0.59	0.82	---	---	0.40	---	---	---	---	0.55	0.52
P_2O_5.........	---	0.16	---	---	0.26	0.16	0.09	0.16	0.18	---	---	0.06	---	---	---	---	0.09	0.18
MnO..........	---	0.08	---	---	0.09	0.06	<0.02	0.04	0.09	---	---	0.04	---	---	---	---	0.05	0.09
CO_2..........	---	<0.01	---	---	0.03	0.02	0.02	0.01	0.03	---	---	<0.01	---	---	---	---	0.01	0.03
Minor elements (parts per million) by X-ray fluorescence analysis																		
Rb.............	---	153	---	162	112	105	176	178	137	93	202	92	---	---	---	107	143	150
Sr..............	---	300	---	265	488	458	297	423	438	573	457	549	---	---	---	381	390	343
Y...............	---	9	---	10	---	---	14	15	11	10	8	11	---	---	---	9	10	20
Zr..............	---	92	---	237	282	---	297	230	120	211	213	218	---	---	---	191	188	109
Nb............	---	24	---	17	---	---	24	33	47	22	22	18	---	---	---	18	26	10

Minor elements (parts per million) by instrumental neutron activation analysis																		
U	---	---	---	---	3.1	5	1.77	2.91	3.99	---	---	1.45	---	---	---	---	1.87	3.01
Th	---	---	---	---	10.3	22.6	38.1	21.7	9.79	---	---	9.96	---	---	---	---	11.2	10.3
Ta	---	---	---	---	3.95	2.95	---	2.57	3.78	---	---	2.30	---	---	---	---	---	0.60
Hf	---	---	---	---	7.6	8	7.93	6.46	6.75	---	---	7.30	---	---	---	---	---	3.40
Cr	---	---	---	---	25	18	12	23	7.8	---	---	15	---	---	---	---	---	74
Co	---	---	---	---	9.9	6.9	2.82	6.05	10.2	---	---	7.33	---	---	---	---	---	10.7
Sc	---	---	---	---	8	5.7	2.65	4.58	5.13	---	---	6.01	---	---	---	---	---	13.8
Zn	---	---	---	---	---	---	45	61	74	---	---	67	---	---	---	---	---	67
Ba	---	---	---	---	810	1,200	1,760	1,320	883	---	---	1,380	---	---	---	---	---	1,240
Cs	---	---	---	---	4.9	3.7	4.0	13.9	5.65	---	---	5.95	---	---	---	---	---	4.15
La	---	---	---	---	63	76	118	65.3	35.1	---	---	89.8	---	---	---	---	---	31.3
Ce	---	---	---	---	119	137	191	107	61.8	---	---	144	---	---	---	---	---	53.9
Nd	---	---	---	---	34	40	69	39	27	---	---	48	---	---	---	---	---	22
Sm	---	---	---	---	7.13	6.92	8.93	5.37	5.1	---	---	7.1	---	---	---	---	---	4.57
Eu	---	---	---	---	1.76	1.41	1.43	1.22	1.17	---	---	1.45	---	---	---	---	---	0.91
Gd	---	---	---	---	---	---	5.8	3.9	4.4	---	---	5.2	---	---	---	---	---	4.0
Tb	---	---	---	---	0.94	0.71	0.45	---	0.66	---	---	---	---	---	---	---	---	0.56
Yb	---	---	---	---	1.55	1.07	0.33	1.08	1.16	---	---	1.22	---	---	---	---	---	2.19
Lu	---	---	---	---	0.21	0.16	0.06	0.12	0.20	---	---	0.14	---	---	---	---	---	0.34

SAMPLE DESCRIPTIONS AND LOCALITIES

W43A NW¼SW¼ sec. 24, T. 37 N., R. 18 E. Fine-grained layer from banded gneiss.

W43C Same locality as W43A. Layer containing conspicuous potassium-feldspar megacrysts.

W144 NW¼SW¼ sec. 13, T. 37 N., R. 18 E. Inequigranular layer in banded gneiss.

W145 NW¼SE¼ sec. 27, T. 37 N., R. 18 E.

W143B NE¼SE¼ sec. 14, T. 37 N., R. 18 E. Equigranular gneiss.

W143A Same locality as W143B. Contains plagioclase megacrysts.

W679A SW¼SW¼ sec. 21, T. 37 N., R. 18 E. Medium-grained layer from banded gneiss.

W679B Same locality as W679A.

W140 SE¼NW¼ sec. 15, T. 37 N., R. 18 E. Layered biotite-hornblende gneiss.

W602 NW¼NW¼ sec. 14, T. 37 N., R. 18 E. Contains plagioclase megacrysts.

W587 NW¼NW¼ sec. 13, T. 37 N., R. 18 E. Contains plagioclase megacrysts.

W495A NW¼NE¼ sec. 15, T. 37 N., R. 18 E. Layered gneiss.

W42 SE¼NE¼ sec. 26, T. 37 N., R. 18 E. Layered gneiss.

W588 NW¼NW¼ sec. 13, T. 37 N., R. 18 E.

W613 NE¼NW¼ sec. 24, T. 37 N., R. 18 E. Contains plagioclase and microcline megacrysts.

W617 SE¼SW¼ sec. 26, T. 37 N., R. 18 E.

W491 SE¼NW¼ sec. 36, T. 38 N., R. 18 E. Augen gneiss.

W742 SE¼SE¼ sec. 19, T. 37 N., R. 18 E.

Al_2O_3, FeO, MnO, MgO, CaO, Na_2O, TiO_2, and P_2O_5, and increasing K_2O with increasing SiO_2 content. They have intermediate SiO_2 contents, ranging from 61.7 to 71.6 wt. percent, moderately high Al_2O_3 (14 to 17.5), and high K_2O (2.4 to 5.1 wt. percent). Most trace element abundances are relatively high in the Dunbar samples, particularly the abundances of Rb, Ba, Th, and the HFS (high field strength) elements. However, the trace elements show only a weak or no correlation with SiO_2 content. Steep negative slopes reflecting high light-REE contents ([La]=106 to 358 times chondrite) and relatively small negative Eu anomalies characterize rare earth element (REE) patterns (fig. 6). The sample with the highest SiO_2 content (W679A) has the highest light REE content and largest negative Eu anomaly but is strongly depleted in heavy REE (table 1, fig. 6); the depletion in heavy REE may reflect fractionation of a phase like zircon.

Sample W742 is included with the Dunbar Gneiss (table 1), but it is compositionally distinct from other samples in having lower Al_2O_3 and HFS element contents and relatively enriched heavy REE. These data suggest that this sample probably is not cogenetic with other Dunbar samples. The sample is similar in composition to the Twelve Foot Falls Quartz Diorite.

Compared to the Newingham Tonalite (see table 7), the Dunbar Gneiss has higher abundances of Ba, Rb, K_2O, high valence cations (U, Th, Ta, Nb, Zr, Hf), and REE. This difference in composition, particularly in the higher Ta and Nb concentrations, suggests that the two units were derived from different sources (Thompson and others, 1984; Pearce and others, 1984). In contrast to the Newingham Tonalite, the other igneous units of the Dunbar dome are also characterized by relatively high abundances of Ba, Rb, K_2O, and the high valence cations.

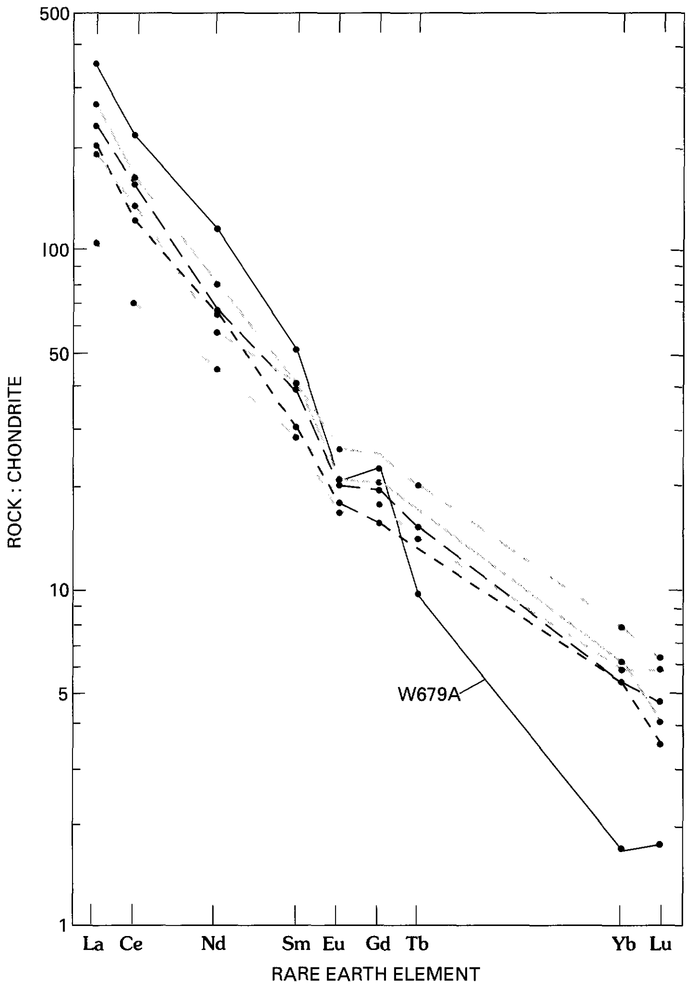


FIGURE 6.—Chondrite-normalized rare earth element plots for samples of Dunbar Gneiss. Chondrite-normalization values from Haskin and others (1968). Note that except for sample W679A, which is anomalously depleted in heavy REE, the samples have relatively constant heavy-REE abundances but variable light-REE contents and crossing patterns.

PETROGENESIS

The overall characteristics of the Dunbar Gneiss, including its medium grain size, presence of relict zoned plagioclase, and the existence of deformed pegmatite-aplite sheets and dikes, suggest that its protolith was plutonic. Widespread myrmekite adjacent to potassium feldspar grains is interpreted as a product of late-stage crystallization of hydrous magma (Hibbard, 1979). Accordingly, we interpret the gneissic fabric as resulting from deformation of an incompletely crystallized magma, as proposed elsewhere by Hibbard (1987). We conclude that the Dunbar Gneiss was deformed simultaneously with D_2, in a relatively high temperature environment wherein pressure was also relatively high and the rocks were subjected to slow-strain-rate penetrative deformation.

The approximately curvilinear compositional trends of decreasing Al_2O_3, CaO, MgO, and FeO with increasing SiO_2 content (64–72 wt. percent) could indicate that these samples of Dunbar Gneiss are related by fractional crystallization. However, such a relationship is not supported by trace element variations. For example, fractionation of the dominant minerals observed in the gneiss—plagioclase, potassium feldspar, quartz, and biotite with minor hornblende—cannot account for the observed enrichment trends in Rb and Ba (fig. 7A). Increasing the amount of hornblende fractionated improves the fit for the Rb trend but does not account for the Ba enrichment (fig. 7B). Also, samples W143A and W143B (table 1), which come from adjacent bands in the gneiss, have similar Rb and

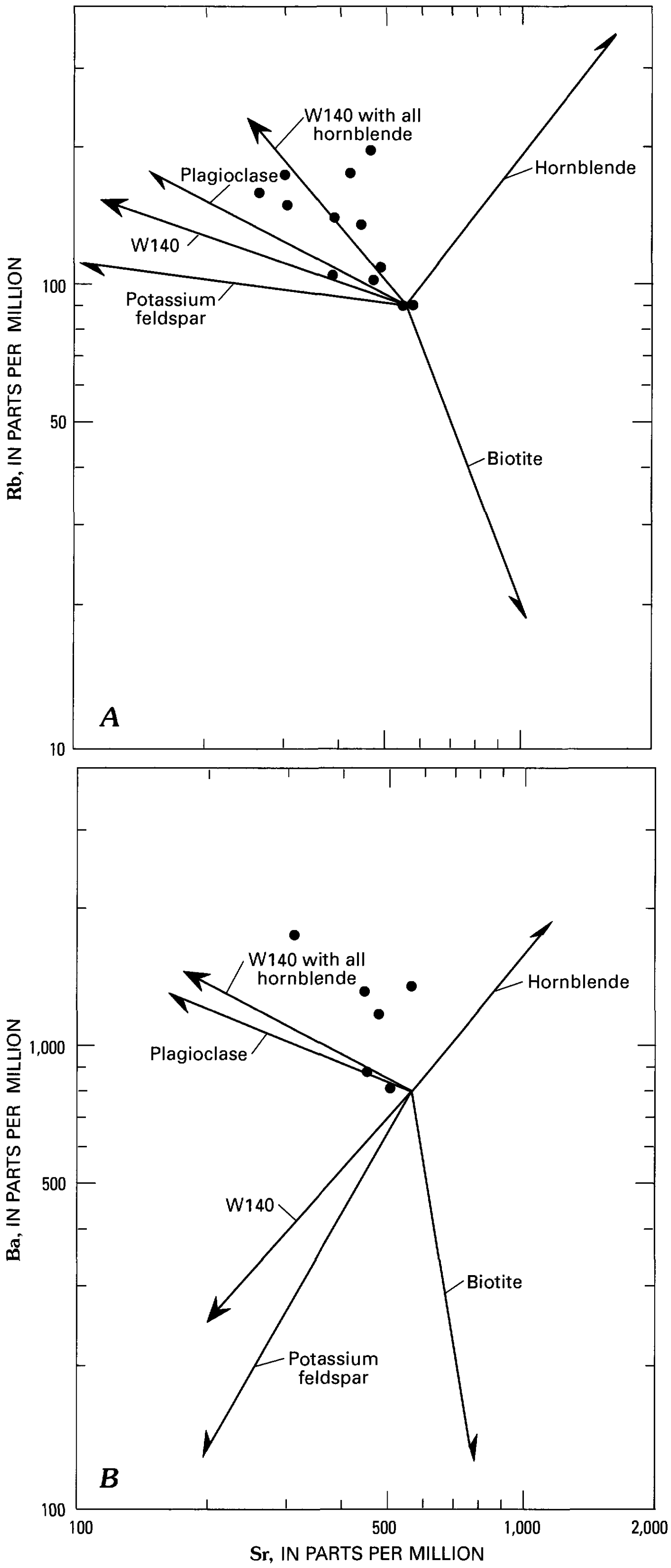
Sr contents but crossing REE patterns and Th contents that vary by a factor of two; this variation also cannot be accounted for by simple crystal-liquid fractionation.

Probably metamorphism affected the abundance of some elements in the Dunbar Gneiss, particularly the alkali elements, which are most prone to secondary mobilization. Alteration could also account for some of the variation in Rb, Ba, and K_2O, but it is unlikely to have significantly effected variations in Th, Ta, and REE, which generally remain relatively immobile even during granulite-facies metamorphism (Jahn and Zhang, 1984; Weaver and Tarney, 1981). Differences in mineral abundance possibly contributed to the variation of these elements. Recent studies have shown that granitoid rocks may not represent actual liquid compositions: instead, these rocks may consist of crystals with some proportion of trapped liquid (that is, are partly cumulate; McCarthy and Hasty, 1976; Zen, 1986). Further, studies have shown that the trace element budget of granitoid rocks in particular may be controlled largely by the abundance of minor and trace phases like sphene, zircon, and allanite (Gromet and Silver, 1983; Sawka, 1988). The higher Co, Sc, and heavy-REE content of sample W143B may reflect the higher abundance of hornblende relative to W143A, whereas the higher Th and light REE of sample W143A could reflect slightly greater amounts of allanite. Such variations could be the result of the original igneous protolith or could have occurred during later metamorphism and deformation (Hibbard, 1987).

In the simplest end member models, the igneous protolith of the Dunbar Gneiss could have been derived by fractional crystallization from a more primitive magma, or alternatively, the parental magma(s) could have been derived directly by partial melting of a more mafic source followed by some degree of crystal fractionation. For either case, the generally high abundance of LIL (large-ion-lithophile) elements in Dunbar Gneiss samples suggests derivation of the parental magma(s) from a relatively enriched source. Also, the high content of HFS elements in the gneiss, particularly Ta and Nb, suggests a "within-plate" source component, as defined by Brown and others (1984) and Pearce (1983).

FIGURE 7 (facing column).—Minor elements in Dunbar Gneiss. *A*, Rb versus Sr; *B*, Ba versus Sr, showing Dunbar Gneiss samples, fractionation vectors for hornblende, plagioclase, biotite, and potassium feldspar, and fractionation vectors for sample W140 with the observed mineral proportions and calculating all biotite as hornblende. Note that fractionation of the observed or recalculated mineral assemblage of a relatively unevolved sample like W140 cannot account for the observed compositional variation in Dunbar Gneiss. Mineral-melt distribution coefficients from Hanson (1978).

Derivation of the igneous protolith of the Dunbar Gneiss by crystal fractionation from a more primitive, probably mantle derived melt cannot be totally



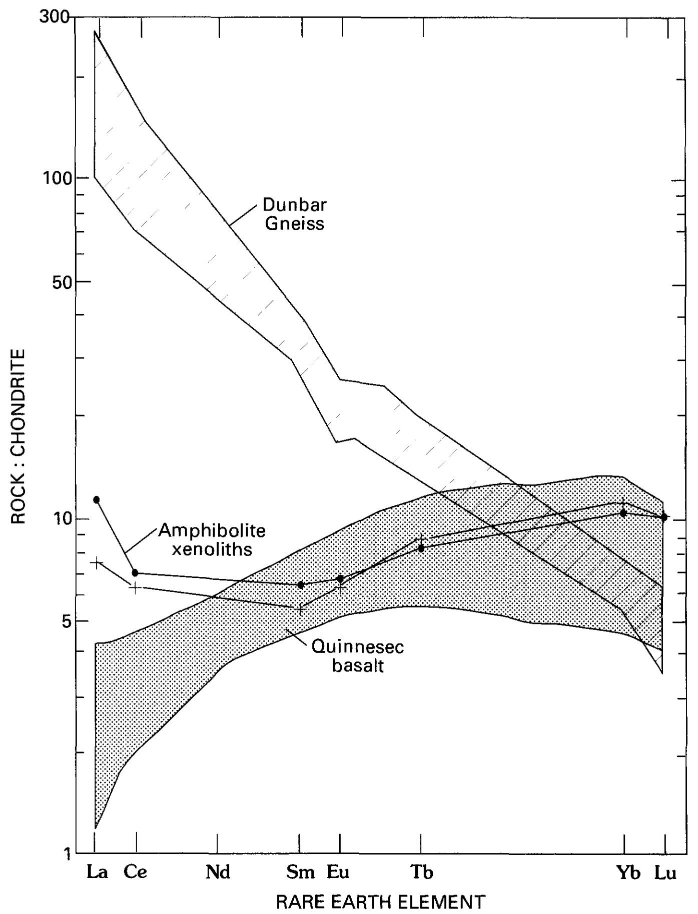


FIGURE 8.—Chondrite-normalized rare earth element plots for amphibolite xenoliths in Dunbar Gneiss. Also shown are fields for Quinnesec basalts and Dunbar Gneiss. Note that REE patterns for amphibolite xenoliths are generally similar to those of the Quinnesec basalts except for anomalously enriched La and Ce, probably resulting from contamination of amphibolite by Dunbar Gneiss.

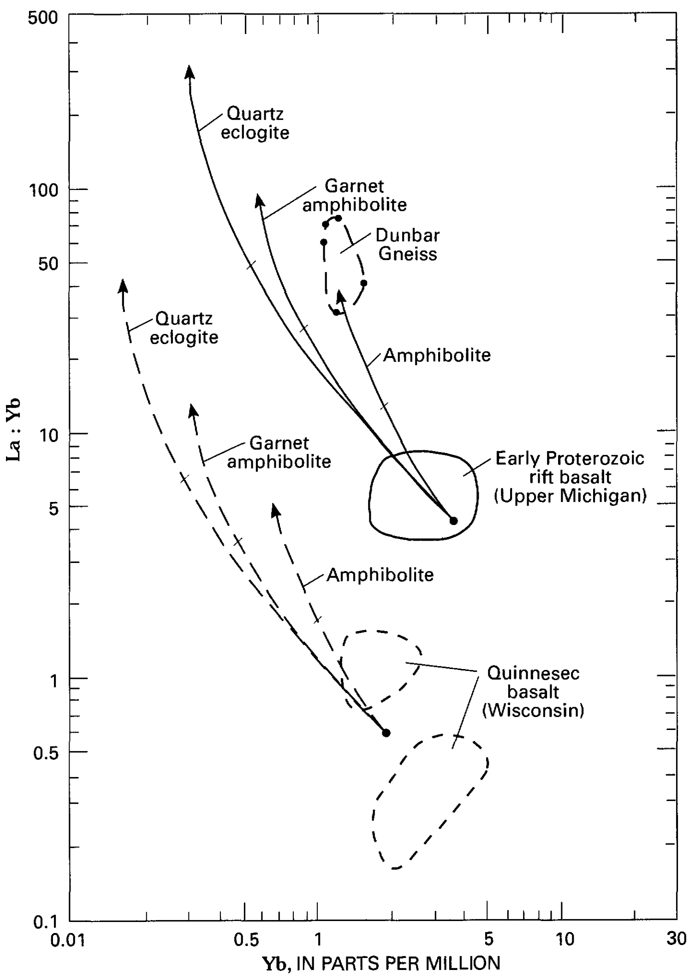


FIGURE 9.—La:Yb versus Yb diagram showing Dunbar Gneiss, Early Proterozoic rift basalts of Upper Michigan, Quinnesec basalt, and partial melting curves for quartz eclogite, garnet amphibolite, and amphibolite mineral assemblages. Partial melting curves calculated using mineral proportions of Jahn and Zhang (1984) and mineral-melt distribution coefficients of Hanson (1978). Origin of upper curves was arbitrarily taken near limit of compositional field of Early Proterozoic rift basalt; origin for lower curves near average Quinnesec basalt. Arrowhead on curve, theoretical limiting value for 0 percent melting; tick mark, 50 percent melting. Partial melting of amphibolite of average rift basalt composition gives the best fit to the Dunbar Gneiss data. Dots, individual samples of Dunbar Gneiss.

discounted, but no samples more primitive than about 62 percent SiO_2 (table 1) have been recognized as clearly cogenetic with the Dunbar Gneiss. Except for somewhat elevated light-REE contents, possibly resulting from contamination with the gneiss, amphibolite layers within the gneiss have compositions similar to the Quinnesec tholeiitic basalt (fig. 8) and could be inclusions. These amphibolites are too depleted in LIL and HFS elements to have been parental to the Dunbar compositions. If the protolith was derived from a more mafic parent melt by fractional crystallization, fractionation would have had to occur in a separate, deeper chamber.

Several experimental studies have shown that partial melting of mafic compositions under both relatively low pressure hydrous conditions (amphibolite) and high pressure anhydrous conditions (granulite to eclogite) can produce melts of intermediate (tonalitic-granodioritic) composition (Green and Ringwood, 1968; Stern and Wyllie, 1978). As shown in figure 9 (see also fig. 24*B*), partial melting of LIL element-enriched amphibolite could produce melts with the general REE characteristics of the Dunbar Gneiss. Such a model is supported by the weakly peraluminous composition of even the most mafic of the samples and their low abundances of Sc, Cr, Co, and Yb, which suggest equilibration with a mafic metaluminous phase like hornblende (Noyes and others, 1983; Zen, 1986).

A possible mafic source for the parent magma(s) of the Dunbar Gneiss may be the basalt in the continental-margin assemblage, exposed in Upper Michigan. This basalt, emplaced during the rifting stage of evolution of the passive margin (Schulz and others, in press), has relatively enriched LIL and HFS element compositions (Schulz, 1984a; Ueng and others, 1988); it probably underplated the continental crust, as has been suggested for basaltic magmatism related to Holocene continental margins (White and others, 1987). Partial melting of such underplated basalt could have occurred in response to thermal perturbations resulting from the collision and overthrusting of the Pembine-Wausau arc terrane over the Superior craton margin.

The Nd isotope data for samples of Dunbar Gneiss (this report, page 49, and Barovich and others, 1989) provide further constraints on the petrogenesis of the Dunbar protolith. The calculated Sm-Nd model ages of 2.28 to 2.06 Ga, which are significantly older than the determined crystallization age, indicate derivation from or interaction with a source relatively enriched in isotopes. Barovich and others (1989) suggested that this source was produced by mixing between Early Proterozoic depleted mantle and Archean crustal material, probably through subduction of Archean sediments. Although such a model can explain the range in Nd isotopes in samples from the northern part of the Wisconsin magmatic terranes, it would not directly account for the high abundance of HFS elements in units like the Dunbar Gneiss because both the depleted mantle (and the magmas derived from it) and typical Archean sediments are relatively depleted in HFS elements (Taylor and McLennan, 1986) and are unlikely to produce a "within-plate" signature (Thompson and others, 1984). Alternatively, if the Dunbar protolith was derived by partial melting of a mafic source similar in composition to the basalts found in the continental-margin assemblage, then both the enriched Nd-isotope and HFS element characteristics of the Dunbar Gneiss could directly reflect those characteristics of the source. With this scenario, the older Sm-Nd model ages could represent the approximate time of separation of the mafic source from a depleted mantle reservoir (McCulloch, 1987). Alternatively, the older model ages could also reflect some degree of contamination with Archean crust of the mafic source.

In summary, the Dunbar Gneiss is interpreted to have had an igneous, probably intrusive protolith. Although petrogenesis of that protolith remains somewhat equivocal, a model involving partial melting of a LIL-enriched mafic source such as the Early Proterozoic rift basalts of the continental margin appears most compatible with present data. Potential underplated equivalents to these basalts could have undergone partial melting during the overthrusting of the arc terrane onto the continental margin during the Penokean orogeny.

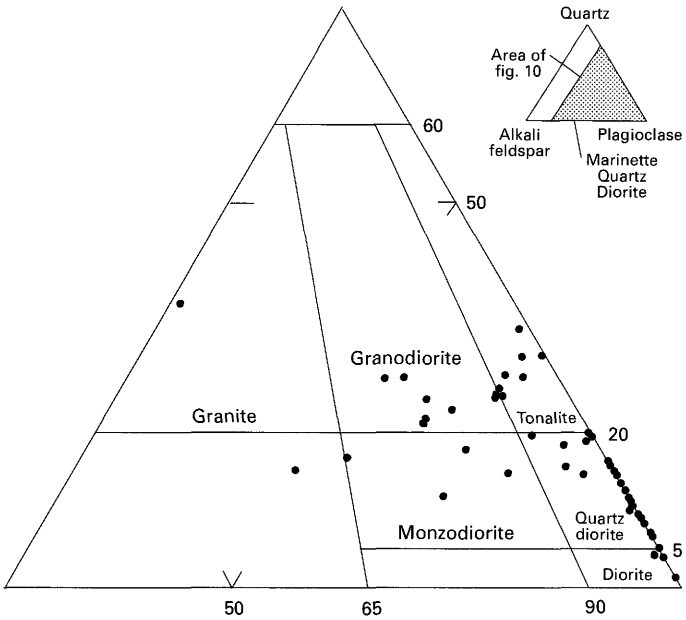


FIGURE 10.—Quartz–alkali feldspar–plagioclase diagram for Marinette Quartz Diorite.

MARINETTE QUARTZ DIORITE

GENERAL FEATURES

The Marinette Quartz Diorite is a large, layered, sill-like intrusive body that was emplaced in the contact zone between the Dunbar Gneiss and the Quinnesec Formation and associated Newingham Tonalite (fig. 2). The rocks are dominantly quartz diorite or diorite in composition (fig. 10), and can be distinguished from the Dunbar Gneiss and the Newingham Tonalite by their moderately low quartz content and moderately high biotite, hornblende, and sphene contents (tables 2 and 3). Because of abundant sphene, the Marinette Quartz Diorite can also be distinguished from the generally finer grained, sphene-poor volcanic rocks of the Quinnesec Formation.

In the north-central part of the Dunbar dome, the layered rocks contain variable amounts of potassium feldspar. These are interpreted as hybrid rocks: the primary quartz diorite was modified by the addition of post-crystallization potassium feldspar. At a few places in the east-central part of the dome, intrusive rocks of

TABLE 2.—*Approximate modes and chemical analyses*

[Leaders (---), not determined; Tr, trace; <, less than. For samples W522A, W523, W520, W527, and W406A, major oxide analyses by J.S. Wahlberg, A. Bartel, J. Taggart, and J. Baker; by instrumental neutron activation analysis by J.S. Mee. For samples SIM-2-1 through SPB-959, major oxide analyses by Z.A. Brown, F.W. Brown, and H. Smith; minor elements by X-ray fluorescence analysis by Z.E. Peterman (slabs)]

Sample No.	W659–1A	W522A	W524A	W519	W520	W521	W406A	W36A	W36B	W527	W527–1
Modal analyses											
Plagioclase	64.5	50	42.5	50	51	45	37	41	49	38	45
Quartz	11	18	13.5	19	6	23	8.6	11	5	14	9
Potassium feldspar	0	14	22.5	5	0	0.7	0.2	9	Tr	4	7
Biotite	22	10	9	19	24.5	29	17	28.5	32	25	21
Hornblende	0	7	12	4	14.5	0	27	4.7	3	10	14.5
Clinopyroxene	0	0	Tr	0	0	0	0	0	0	0	0
Sphene	Tr	1	Tr	2.5	4	1.5	4.3	5.8	6	4.3	3
Opaque oxides	Tr	Tr	Tr	Tr	Tr	Tr	Tr	Tr	Tr	0.2	Tr
Epidote-clinozoisite	Tr	Tr	Tr	Tr	Tr	Tr	0.2	Tr	5	3	Tr
Accessory minerals	0.5	Tr	0.5	0.5	Tr	0.8	5.7	Tr	Tr	1.5	0.5
Major oxides (weight percent)											
SiO_2	---	65.0	---	---	51.8	---	51.1	---	---	54.1	---
Al_2O_3	---	15.9	---	---	16.8	---	16.2	---	---	17.0	---
Fe_2O_3	---	0.64	---	---	1.97	---	1.42	---	---	1.45	---
FeO	---	3.01	---	---	7.77	---	7.36	---	---	6.84	---
MgO	---	1.56	---	---	3.50	---	5.48	---	---	3.09	---
CaO	---	3.20	---	---	6.18	---	7.67	---	---	5.55	---
Na_2O	---	4.30	---	---	3.88	---	3.16	---	---	3.54	---
K_2O	---	3.69	---	---	2.62	---	2.32	---	---	3.43	--
H_2O^+	---	0.44	---	---	1.23	---	1.59	---	---	1.13	---
H_2O^-	---	0.02	---	---	0.07	---	0.14	---	---	0.14	---
TiO_2	---	0.79	---	---	2.42	---	2.24	---	---	1.77	---
P_2O_5	---	0.18	---	---	0.64	---	0.63	---	---	0.59	---
MnO	---	0.07	---	---	0.15	---	0.15	---	---	0.13	---
CO_2	---	0.04	---	---	<0.01	---	<0.01	---	---	0.10	---
Minor elements (parts per million)											
Rb	---	118	---	---	84	---	117	---	---	85.9	---
Sr	---	459	---	---	727	---	727	---	---	696	---
Y	---	13	---	---	25	---	20	---	---	23	---
Zr	---	150	---	---	208	---	231	---	---	235	---
Nb	---	46	---	---	51	---	55	---	---	44	---

[1]Sample descriptions are on p. 16–17.

granodioritic composition (as for example sample W667B, table 3) contain inclusions of mafic rocks typical of the Marinette; these intrusive rocks are included in the Marinette as mapped (fig. 2), but their extent and nature are obscure because of their sparse exposure.

The northern part of the Marinette Quartz Diorite, mapped previously by W.C. Prinz (Bayley and others, 1966, pls. 1 and 2), is moderately well exposed in small hills and cuts along the Chicago, Milwaukee, St. Paul, and Pacific Railway. Except for an area of about 2 km^2 west of Fischer Lake (Sims and Schulz, 1988), exposures elsewhere in this part of the dome are rather sparse. Another moderately good outcrop area occurs in the southwestern part of the dome, south of Dunbar. In this area as well as a smaller area east of Dunbar (sec. 20, T. 37 N., R. 19 E.), white to pink granite pegmatite and aplite similar to that cutting the Dunbar Gneiss profusely intrude the Marinette Quartz Diorite.

The Marinette Quartz Diorite has a moderately well developed foliation expressed mainly by oriented biotite that generally is subparallel to, but locally transects, compositional layering. The rocks were deformed together with the Dunbar Gneiss and the Quinnesec volcanic rocks on northwest-oriented fold axes (F_2).

The quartz diorite in the elongate body north of the Niagara lobe (fig. 2), which is separated from the main body by the intrusive Spikehorn Creek Granite, shows sharp intrusive contacts against the Quinnesec Formation; a lit-par-lit intrusive zone about 50 m wide is exposed in the east half of sec. 22, T. 38 N., R. 20 E. In the central part of the dome, the Marinette intrudes *of selected samples of Marinette Quartz Diorite*[1]

FeO, H_2O, and CO_2 analyses by H. Neiman, J. Ryder, and G. Mason; minor elements by X-ray fluorescence analysis by J. Storey, S. Donahey, B. Vaughn, and M. Coughlin; minor elements by X-ray fluorescence analysis by R. Johnson, K. Dennen, and B. Scott; minor elements by instrumental neutron activation analysis by L.J. Schwarz. For sample W404, minor elements

W688A	W511A	W511B	W511	SIM–2–1	SIM–6–82	SDNE 100–82	SDNE 104–1–82	SDNE 104–2–82	SPB 959	W507	W704	W664B	W404	W465–1A
(volume percent)														
48	53	25	---	55.7	54	42.5	49.8	60.2	29.4	52	53	56	56	47
7	23	35.5	---	13.9	7.6	4.4	2.4	2.3	8.8	11	4	13	5.5	11.5
0	0	35.5	---	2.3	1.6	0	0	0	21	11.5	Tr	2.5	0	0
23	21	4	---	18	25.4	12.7	23.6	27.2	27.9	11.5	27	22.5	25	27
18	0	0	---	3	3.7	31.1	19.7	2.3	7.8	11	11	0	11	9
0	0	0	---	0	0	0	0	0	0	0	0	0	0	0
3	2.5	Tr	---	3.3	3.4	1.1	5	3.4	3	3	4.5	4	2.5	5
Tr	0	0	---	0.4	0.9	4.7	0.5	0.4	0.8	Tr	Tr	Tr	0	0
0	Tr	Tr	---	0.8	Tr	0.2	Tr	Tr	Tr	Tr	Tr	Tr	Tr	Tr
1	0.5	Tr	---	2.6	3.3	3.3	Tr	4.2	1.3	Tr	0.5	2	Tr	0.5
by X-ray fluorescence analysis														
---	63.0	74.2	75.0	58.7	55.3	49.5	49.7	55.9	57.8	---	---	---	---	---
---	16.7	13.9	13.5	16.0	19.3	16.4	18.1	20.3	17.6	---	---	---	---	---
---	0.81	0.14	0.08	2.0	1.6	2.3	2.4	1.7	1.7	---	---	---	---	---
---	4.34	0.80	1.22	5.6	5.2	9.6	8.2	3.8	4.7	---	---	---	---	---
---	1.81	0.29	0.22	2.7	2.6	4.2	3.6	2.0	2.5	---	---	---	---	---
---	3.13	1.11	1.25	3.9	5.0	7.3	6.5	4.9	4.6	---	---	---	---	---
---	4.17	2.94	3.17	3.7	4.8	3.7	3.9	5.7	4.9	---	---	---	---	---
---	2.79	5.16	4.51	2.7	3.1	1.8	2.6	2.0	3.8	---	---	---	---	---
---	0.76	0.25	0.24	2.5	1.2	1.5	1.5	1.3	1.1	---	---	---	---	---
---	0.02	<0.01	0.10	0.26	0.20	0.18	0.18	0.25	0.18	---	---	---	---	---
---	1.06	0.09	0.14	1.5	1.5	2.4	2.3	0.93	1.2	---	---	---	---	---
---	0.13	<0.05	<0.05	0.5	0.42	1.0	0.55	0.32	0.33	---	---	---	---	---
---	0.12	<0.02	<0.02	0.13	0.09	0.17	0.15	0.07	0.12	---	---	---	---	---
---	0.01	0.01	0.03	0.23	0.08	0.17	0.24	0.48	0.06	---	---	---	---	---
by X-ray fluorescence analysis														
---	---	213	237	51	109	65	91	85	159	---	---	---	60	---
---	---	---	236	497	703	692	655	715	502	---	---	---	1,153	---
---	---	---	7.8	26	21	28	26	14	21	---	---	---	16	---
---	---	---	68	104	579	176	301	302	277	---	---	---	247	---
---	---	---	14.5	50	60	42	48	22	70	---	---	---	41	---

the megacrystic facies of the Newingham Tonalite, but details of their relationship are not known. Also, dikes of the Marinette intrude both the Quinnesec Formation and the Dunbar Gneiss, but major contacts against the Dunbar Gneiss are not exposed.

The Hoskin Lake Granite has intruded the north margin of the Marinette Quartz Diorite and has produced a contact zone of mixed rocks as much as 4 km wide. As described by Bayley and others (1966), this broad contact zone contains many rock types gradational between the two end members, quartz diorite and granite, and broad areas composed predominantly of one or the other of the members. In this zone, fine-grained leucogranite dikes cut both the quartz diorite and the granite. The Spikehorn Creek Granite in the Niagara lobe (fig. 2), on the other hand, has moderately sharp contacts with the quartz diorite, and at places crosscuts previously foliated Marinette Quartz Diorite.

DESCRIPTION

The Marinette Quartz Diorite ranges in composition from diorite or gabbro to tonalite and granodiorite and rarely, granite (fig. 10). Quartz diorite is the predominant rock type, as noted by Prinz (1959; 1965). Subsequent to crystallization, the rocks were differentially metamorphosed and metasomatized, as well as deformed. Prograde metamorphism of the rocks in the north-central part of the core (fig. 2) was accompanied by the metasomatism.

In the relatively unmetamorphosed eastern part of the body, the Marinette Quartz Diorite is a dark-gray to black, medium-grained, generally massive rock that

TABLE 2.—*Approximate modes and chemical analyses*

Sample No.	W659–1A	W522A	W524A	W519	W520	W521	W406A	W36A	W36B	W527	W527–1	W688A
										Minor elements (parts per million) by		
U	---	2.8	---	---	3.1	---	1.79	---	---	2.4	---	---
Th	---	23	---	---	19.4	---	5.9	---	---	22.4	---	---
Ta	---	3.50	---	---	3.85	---	3.46	---	---	2.58	---	---
Hf	---	4.63	---	---	5.87	---	5.6	---	---	7.70	---	---
Cr	---	17	---	---	6.4	---	92	---	---	9.6	---	---
Co	---	9.67	---	---	28.5	---	29.7	---	---	22.7	---	---
Sc	---	6.35	---	---	13.1	---	20.2	---	---	13.6	---	---
Zn	---	58	---	---	94	---	107	---	---	97	---	---
Ba	---	1,050	---	---	701	---	964	---	---	1,205	---	---
Cs	---	3.42	---	---	4.36	---	14.9	---	---	2.56	---	---
La	---	54.3	---	---	57.2	---	67.7	---	---	114	---	---
Ce	---	83.1	---	---	118	---	124	---	---	195	---	---
Nd	---	28	---	---	60	---	56	---	---	75	---	---
Sm	---	4.21	---	---	10.2	---	8.92	---	---	11.5	---	---
Eu	---	1.12	---	---	2.50	---	2.33	---	---	2.59	---	---
Gd	---	3.3	---	---	8.6	---	6.9	---	---	9.3	---	---
Tb	---	0.34	---	---	1.01	---	0.80	---	---	0.94	---	---
Yb	---	1.08	---	---	2.20	---	1.73	---	---	2.16	---	---
Lu	---	0.15	---	---	0.32	---	0.26	---	---	0.29	---	---

SAMPLE DESCRIPTIONS

W659–1A	NE¼SE¼ sec. 22, T. 38 N., R. 20 E. Dike that cuts Quinnesec Formation.
W522A	NW¼ sec. 22, T. 38 N., R. 20 E. Shows retrogressive metamorphism.
W524A	SE¼NE¼ sec. 22, T. 38 N., R. 20 E. Contains relict grains of clinopyroxene.
W519	Railway cut, NW¼ sec. 10, T. 38 N., R. 20 E. Hypidiomorphic granular texture.
W520	Near locality W519. Granoblastic texture.
W521	SW¼SW¼ sec. 21, T. 38 N., R. 20 E. Hypidiomorphic granular texture.
W406A	NW¼SE¼ sec. 35, T. 37 N., R. 18 E. Mafic layer in banded rock.
W36A	Railway cut, SW¼NE¼ sec. 18, T. 38 N., R. 20 E. Granoblastic texture.
W36B	Same locality as W36A. Granoblastic texture.
W527	Railway cut 90 m south of locality W36A. Granoblastic texture.
W527–1	Railway cut, NE¼SE¼ sec. 18, T. 38 N., R. 20 E. Hypidiomorphic granular texture.
W688A	SW¼NW¼ sec. 34, T. 38 N., R. 19 E.
W511A	Small prospect pit, NE¼NW¼ sec. 33, T. 38 N., R. 19 E. Granoblastic texture.

is hypidiomorphic granular but porphyroblastic near the Hoskin Lake Granite. The rocks have a well- to ill-defined layering and generally lack a penetrative foliation. Plagioclase, the most abundant mineral in the rocks, occurs as tabular, normally zoned crystals as much as 5 mm long that display combined carlsbad and albite twins, and less commonly pericline twins. Oscillatory zoning was noted in a few thin sections. The plagioclase is dominantly calcic oligoclase, but has cores as calcic as An_{40}. The plagioclase tends to be weakly to moderately altered to epidote-clinozoisite and lesser sericite. Small patches of microcline occur in some grains. Microfractures (Hanmer, 1982; type IIP) occur locally in the plagioclase. Potassium feldspar occurs in variable amounts (tables 2 and 3) as fresh, twinned microcline or microperthite, and generally has inclusions of myrmekitic plagioclase, quartz, biotite, and hornblende. Quartz commonly is clear or shows strong undulatory extinction and generally has been recrystallized as aggregates that are finer grained than the original crystals. It tends to be a late, interstitial mineral. Clinopyroxene occurs as rare relict grains in the cores of some hornblende crystals. Hornblende is grayish olive or pale green and moderately pleochroic. Generally it is weakly altered along grain margins and cleavages to a medium-green or bluish-green hornblende, which pseudomorphically replaces the older generation. In some of the more mafic rocks, hornblende is absent and a green chlorite having dark-blue interference colors is the dominant mafic silicate mineral. This chlorite is associated with calcite, sphene, opaque oxides, and epidote-clinozoisite. Biotite is grayish red and commonly has frayed crystal outlines. In the elongate body north of the Niagara lobe, hornblende appears to be unstable where in contact with microcline and may be partly replaced by *of selected samples of Marinette Quartz Diorite*—Continued

511A	W511B	W511	SIM–2–1	SIM–6–82	SDNE 100–82	SDNE 104–1–82	SDNE 104–2–82	SPB 959	W507	W704	664B	W404	W465–1A
instrumental neutron activation analysis													
4.24	2.49	2.28	3.3	3.6	1.3	1.2	2.6	1.75	---	---	---	---	---
14.8	14.8	11.5	24.1	23.0	6.5	5.9	12.4	18.9	---	---	---	---	---
---	---	2.58	3.36	3.22	3.19	3.6	1.16	5.54	---	---	---	---	---
---	2.37	2.39	2.2	16.4	4.7	6.5	7.3	5.77	---	---	---	---	---
---	1.3	17	4.0	1.7	12.0	6.6	18.0	16	---	---	---	---	---
---	1.01	1.34	18.8	18.4	34.4	31.3	11.9	14.95	---	---	---	---	---
	0.88	1.05	8.46	8.45	13.7	12.5	5.78	10.33	---	---	---	---	---
---	24	26	86	76	109	121	71	---	---	---	---	---	---
---	1,680	1,620	855	902	666	1,020	884	860	---	---	---	---	---
---	8.11	10.9	0.7	3.7	2.4	2.8	3.2	---	---	---	---	---	---
---	23.5	26.1	57	88	71	67	66	71.9	---	---	---	---	---
---	39.2	42.1	107	137	126	123	96	108	---	---	---	---	---
---	16.0	15.9	---	---	---	---	28	---	---	---	---	---	---
---	2.28	2.47	8.8	7.9	9.8	9.7	3.8	6.42	---	---	---	---	---
---	0.66	0.67	1.97	2.08	2.52	2.40	2.06	1.54	---	---	---	---	---
---	2.2	1.4	---	---	---	---	---	---	---	---	---	---	---
---	---	---	0.83	0.71	0.85	0.84	0.27	0.65	---	---	---	---	---
---	0.32	0.52	1.3	1.7	2.1	2.4	0.5	1.47	---	---	---	---	---
---	0.05	0.06	0.25	0.30	0.29	0.33	0.15	0.22	---	---	---	---	---

AND LOCALITIES

W511B	Same locality as W511A.
W511	Same locality as W511B. Layered rock.
SIM–2–1	Railway cut, NW¼ sec. 20, T. 38 N., R. 20 E. Hypidiomorphic granular texture.
SIM–6–82	Railway cut, SW¼NE¼ sec. 18, T. 38 N., R. 20 E. Granoblastic texture.
SDNE–100–82	SE¼SW¼ sec. 33, T. 38 N., R. 20 E. Hypidiomorphic granular texture.
SDNE–104–1–82	NW¼SW¼ sec. 33, T. 38 N., R. 20 E. Granoblastic texture.
SDNE–104–2–82	Same locality as SDNE–104–1–82. Has plagioclase phenocrysts.
SPB–959	NE¼SE¼ sec. 21, T. 38 N., R. 20 E. Hypidiomorphic granular texture.
W507	NE¼NW¼ sec. 26, T. 38 N., R. 19 E. Contains radiating hornblende crystals.
W704	SE¼SE¼ sec. 23, T. 38 N., R. 19 E.
W664B	Railway cut, NW¼SE¼ sec. 33, T. 38 N., R. 20 E.
W404	SW¼NE¼ sec. 36, T. 37 N., R. 18 E.
W465–1A	NW¼SW¼ sec. 20, T. 37 N., R. 19 E. Granoblastic texture.

microcline. Sphene and opaque oxides, either separately or as grains having opaque oxide cores and sphene rims, constitute about 5 percent of the rocks. They are closely associated with hornblende and biotite, as is epidote-clinozoisite. Epidote in the more mafic rocks is yellow. The common accessory minerals, in order of decreasing abundance, are apatite, allanite, and zircon. Typically, allanite has narrow rims of epidote-clinozoisite. Much of the zircon is zoned.

Many of the rocks in the easternmost part of the dome have retrograde mineral assemblages indicative of the following partial reactions: clinopyroxene→light-green hornblende+quartz→medium-green hornblende→biotite+opaque oxides+sphene. At places, the light-green hornblende alters directly to epidote-clinozoisite.

The most southerly exposures of the Marinette Quartz Diorite, in the vicinity of Dunbar, are medium- to dark-gray mesocratic, layered quartz diorite and diorite (fig. 11) that are cut by abundant pegmatite and aplite; they are generally equigranular rocks but include porphyritic varieties containing plagioclase as much as 1 cm in diameter. Layering is mainly expressed by compositional differences; some layers are clinopyroxene-bearing gabbro. Plagioclase (An_{30-50}) occurs as generally equant, anhedral grains 1 mm or less in diameter in which albite and pericline twins are moderately well developed. The plagioclase grains locally have numerous microfractures—type IIP structures of Hanmer (1982). Pale-olive or light-olive, but locally bluish green hornblende forms subhedral crystals that tend to elongate in the direction of foliation. At places, hornblende is partly altered to biotite+opaque oxides+sphene. The biotite is reddish brown and generally associated with a few percent of sphene, some of which has opaque oxide cores. Epidote-clinozoisite is the main alteration mineral, of both

TABLE 3.—*Approximate modes and minor element analyses of selected samples of Marinette Quartz Diorite*

[Leaders (---), not determined; Tr, trace. Minor element analyses of rock slabs by Z.E. Peterman]

Sample No.	W671	W710B	W669A	W672B	W710A	W469	W667A	W667B	W623	W624	W626	W651	W408A	W409A	W357	W652	W707	W515	W471	W708C	W493C
Modal analyses (volume percent)																					
Plagioclase	59.5	53.3	49	52	56	52	47.5	56.5	57	51	60	57	51	38	43	54	54.5	53	50	56	49.7
Quartz	10.5	18	20	28	20	6.5	0.5	20.5	10	21	2.5	6	9	12	3	9	10	20	5	25	10
Potassium feldspar	2.5	15.5	13	10	6.5	0	0	12	0	4	0	0	15	18	0	Tr	0	6	0	2	2.8
Biotite	24.5	12.5	17.5	10	17	25.5	29.5	9.5	16	20	15.5	19	19	27	14	21.5	22	14.5	19	17	12
Hornblende	1	0	0	0	0	9.5	15.5	0	14	3.5	12.5	14	4	3	37.5	13.5	10	4.5	22	0	19.8
Sphene	2	0.5	0.5	Tr	0.5	6	5	1	1	0.5	5	2.5	2	2	1.5	1.5	3	1.5	3	Tr	2.8
Apatite	Tr	Tr	Tr	Tr	Tr	Tr	Tr	Tr	0.5	Tr	Tr	Tr	Tr	Tr	0.5	Tr	Tr	Tr	1	Tr	Tr
Opaque oxides	Tr	Tr	0	0	0	0	2	Tr	0.5	0	Tr	Tr	Tr	Tr	0.3	0.5	0	Tr	Tr	Tr	Tr
Allanite	Tr	Tr	Tr	0	Tr	Tr	Tr	Tr	Tr	Tr	0	Tr	Tr	Tr	0.2	Tr	Tr	0.5	Tr	Tr	Tr
Accessory minerals	Tr	0.2	Tr	Tr	Tr	0.5	Tr	0.5	Tr	Tr	4.5	1.5	Tr	Tr	Tr	Tr	Tr	Tr	Tr	Tr	1.9
Minor elements (parts per million) by X-ray fluorescence																					
Rb	112	66	86	76	---	468	68	83	47	110	---	141	---	---	---	---	---	---	---	---	---
Sr	1,109	594	562	551	---	871	841	436	1,159	777	---	722	---	---	---	---	---	---	---	---	---
Y	12	5	13	8	---	22	20	9	10	7	---	10	---	---	---	---	---	---	---	---	---
Zr	186	177	238	160	---	256	264	220	230	126	---	65	---	---	---	---	---	---	---	---	---
Nb	37	11	39	26	---	40	45	25	31	11	---	23	---	---	---	---	---	---	---	---	---

SAMPLE DESCRIPTIONS AND LOCALITIES

W671	NW¼NE¼ sec 9, T. 37 N., R. 19 E. Foliated megacrystic granitic rock. Granoblastic texture.
W710B	NE¼SW¼ sec. 34, T. 38 N., R. 19 E. Foliated megacrystic granitic rock. Protomylonite.
W669A	NE¼NE¼ sec. 1, T. 37 N., R. 19 E. Foliated granitic rock. Protomylonite.
W672B	SW¼SE¼ sec. 4, T. 37 N., R. 19 E. Prominently layered granitic rock.
W710A	Same locality as W710B. Foliated granitic rock. Protomylonite.
W469	NW¼SE¼ sec. 10, T. 37 N., R. 19 E. Granoblastic texture.
W667A	SW¼NW¼ sec. 32, T. 38 N., R. 20 E. Mafic phase that occurs as inclusions in W667B. Has weak retrogressive metamorphism.
W667B	Same locality as W667A. Foliated granitic rock that contains inclusions of more mafic phase of Marinette Quartz Diorite.
W623	SW¼SE¼ sec. 27, T. 37 N., R. 18 E. Foliated megacrystic rock.
W624	SW¼NE¼ sec. 34, T. 37 N., R. 18 E. Similar to W623.
W626	SE¼SW¼ sec. 25, T. 37 N., R. 18 E. Sheared and moderately altered.
W651	SW¼NW¼ sec. 20, T. 37 N., R. 19 E. Foliated megacrystic rock, locally migmatitic.
W408A	NE¼NE¼ sec. 34, T. 38 N., R. 19 E. Foliated megacrystic granitic rock.
W409A	SE¼SE¼ sec. 28, T. 38 N., R. 19 E. Foliated granitic rock that contains conspicuous potassium feldspar megacrysts. Granoblastic texture.
W357	SW¼SE¼ sec. 21, T. 37 N., R. 18 E. Strongly lineated mafic rock.
W652	SE¼NW¼ sec. 20, T. 37 N., R. 19 E. Foliated granitic rock that contains mafic inclusions.
W707	NW¼SW¼ sec. 34, T. 38 N., R. 19 E. Foliated rock in layered succession. Granoblastic texture.
W515	SW¼NE¼ sec. 33, T. 38 N., R. 19 E. Foliated megacrystic rock. Granoblastic texture.
W471	NW¼SW¼ sec. 36, T. 37 N., R. 18 E. Foliated megacrystic rock.
W708C	NE¼SE¼ sec. 33, T. 38 N., R. 19 E. Foliated megacrystic rock in layered succession.
W493C	SW¼NW¼ sec. 31, T. 38 N., R. 19 E. Dike that transects Dunbar Gneiss.

FIGURE 11.—Primary compositional layering in Marinette Quartz Diorite (sec. 36, T. 37 N., R. 18 E.) defined by variations in grain size and mineral proportions. Hammer (30 cm) for scale.

andesine and hornblende. Sericite is a rare alteration product of plagioclase. Minor minerals are sphene and opaque oxides, and accessory minerals are apatite, allanite, and zircon.

The northern and northwestern parts of the Marinette Quartz Diorite, adjacent to the Hoskin Lake Granite, have been metamorphosed to gneisses under amphibolite-facies conditions. These are dominantly dark mesocratic rocks that are layered and have a well-developed foliation and lineation. The layering is expressed by subtle differences in composition and texture, and individual layers range in thickness from less than a meter to a few tens of meters. The foliation is subparallel to the layering and is expressed mainly by oriented biotite plates and more locally by hornblende and plagioclase.

The metamorphosed quartz diorite is a nearly equigranular gneiss that has xenomorphic granular textures. Except for relict normally zoned plagioclase phenocrysts(?), probably all the mineral constituents have been recrystallized. The gneisses contain 40–60 percent plagioclase, 0–20 percent potassium feldspar, 10–45 percent combined hornblende and biotite, and as much as 6 percent sphene. Opaque oxides occur rarely as cores of grains rimmed by sphene. Epidote-clinozoisite is sparse, mainly forming narrow rims on altered allanite grains. Apatite and zircon are the other accessory minerals.

Plagioclase occurs in two forms: (1) as somewhat tabular, relict, zoned crystals (andesine) as much as 5 mm long having slightly more calcic cores, and (2) as anhedral unzoned, smaller recrystallized grains. In part, the latter occur as polycrystalline aggregates surrounding the older, zoned grains, forming a mortar structure or a core and mantle structure (White, 1976). Carlsbad combined with albite and pericline twins are prevalent. Adjacent to potassium feldspar, myrmekite commonly is present. The plagioclase is weakly altered to epidote-clinozoisite or sericite. The potassium feldspar is microcline, which forms interstitial grains and poikilitic porphyroblasts containing plagioclase, myrmekite, biotite, and quartz. The hornblende is fresh, dark green or dark blue green, and contains poikilitic quartz inclusions. Most hornblende has oriented exsolution (hornblende?) textures and, locally, narrow rims of blue-green hornblende. At places, biotite appears to be an alteration product of hornblende, but more commonly it is intergrown with hornblende and appears to have crystallized together with it. The biotite is fresh, olive gray or grayish red, and commonly has inclusions of quartz. It is intergrown with sphene and opaque oxides with sphene rims. Quartz occurs mainly as polycrystalline aggregates, in places having undulatory extinction. It also occurs as elongate aggregates (ribbon quartz). Sphene forms well-crystallized subhedral grains that are associated with biotite and hornblende.

GEOCHEMISTRY

The lithologic diversity of the Marinette Quartz Diorite is reflected in the chemistry presented in tables 2 and 3. Compositions range from mafic with high K_2O and REE contents to granite. Samples are characterized by high abundances of Al_2O_3, K_2O, total alkalies, TiO_2, P_2O_5, Ta, Nb, Zr, Rb, and REE and are alkali-calcic in character (fig. 5). Chondrite-normalized REE patterns (fig. 12) show steep negative slopes with enriched light REE ([La]=170–340) and generally small negative to no Eu anomalies; sample 104–2–82, which has a large positive Eu anomaly and very steep REE pattern, contains abundant plagioclase megacrysts. Major elements show a general inverse correlation with increasing SiO_2 content, except for K_2O which increases slightly. Trace elements generally show a poor correlation with SiO_2 content and are probably controlled by the abundance of minor and trace phases such as titanite, apatite, and zircon (see Noyes and others, 1983).

Field relations and petrography clearly show that the mafic to intermediate rocks are part of a layered, differentiated complex and that individual samples represent crystal cumulates to varying degrees.

Two samples (W511 and W511B, table 2) taken adjacent to the Hoskin Lake Granite are high-silica minimum-melt granites. They have low rare earth element abundances, and low Zr and Y, and they lack prominent Eu anomalies (fig. 13). The granites from

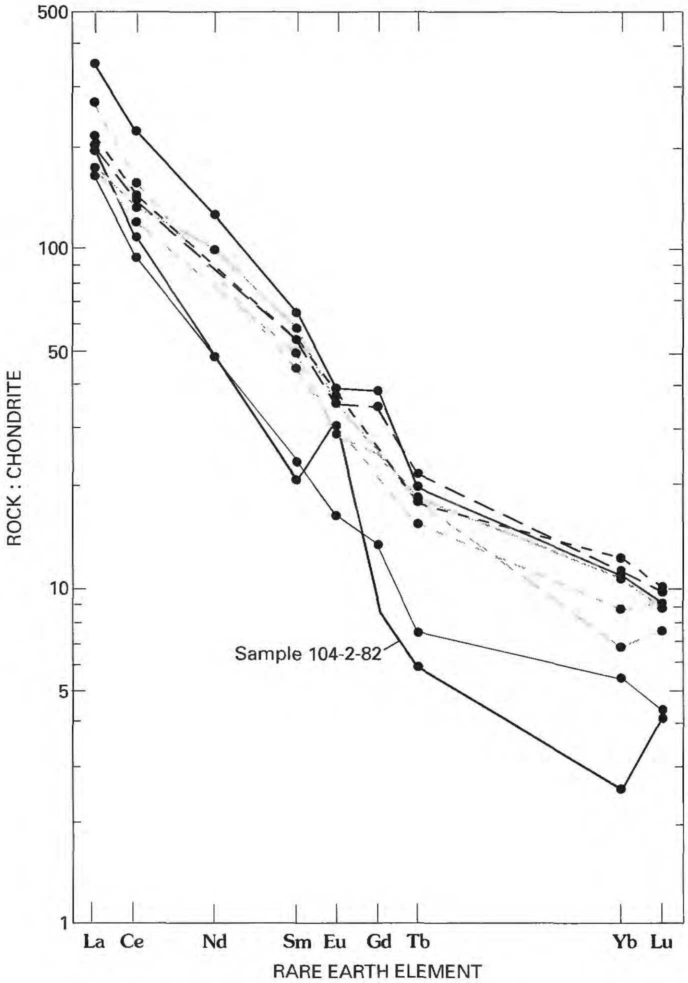


FIGURE 12.—Chondrite-normalized rare earth element plots for mafic samples of Marinette Quartz Diorite. Chondrite-normalization values from Haskin and others (1968).

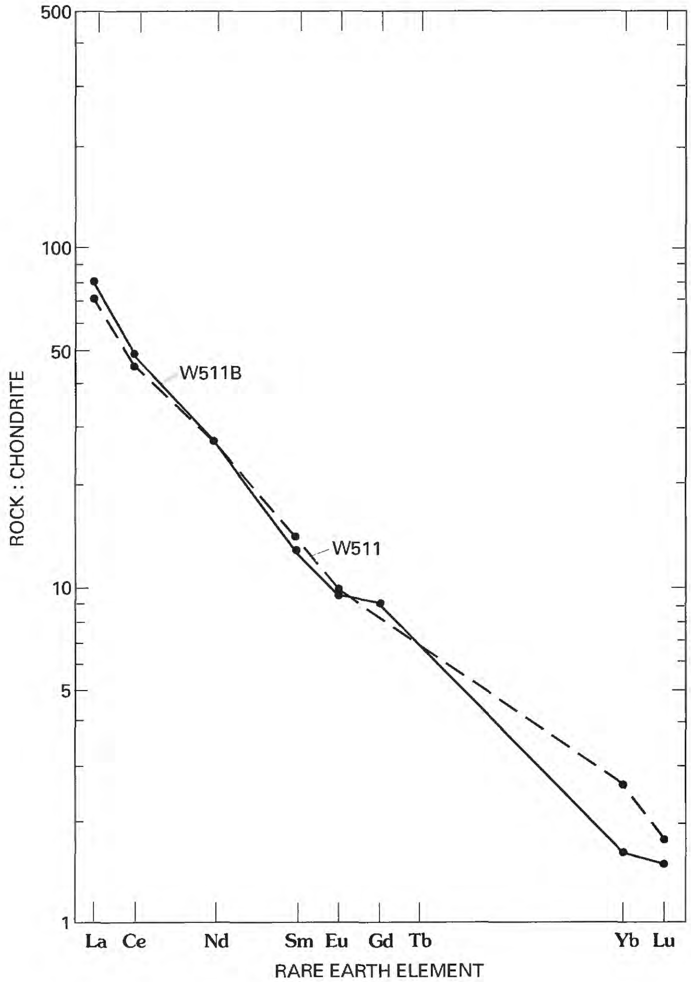


FIGURE 13.—Chondrite-normalized rare earth element plots for felsic samples (W511 and W511B) from Marinette Quartz Diorite. Chondrite-normalization values from Haskin and others (1968).

which these samples came are finely interlayered with tonalite (sample 511A, table 2). Whether these granites are related to the Marinette or Hoskin Lake intrusions is uncertain.

PETROGENESIS

MAFIC ROCKS (SiO_2<65 PERCENT)

The mafic rocks of the Marinette Quartz Diorite (SiO_2<65 percent) show field and petrographic evidence of crystal accumulation and probably do not represent liquid compositions. Figure 14 shows the Sc content of mafic rocks plotted against the combined modal abundance of clinopyroxene, hornblende, and sphene, phases in which Sc is a compatible element. The strong positive correlation demonstrates that the Sc abundance is a function of the modal mineralogy and that these samples are at least in part cumulate in nature. The cumulate nature is further demonstrated by the P_2O_5 content, which is highest in samples with high modal apatite. Also, the REE pattern of sample 104–2–82 (table 2), which has abundant plagioclase megacrysts, exhibits a strong positive Eu anomaly (fig. 12).

The cumulate nature of the mafic rocks may be further supported by their near-constant FeO_T:(FeO_T+MgO) ratio (fig. 15), which is unlikely to result from a magma crystallizing mafic phases over a compositional range of 50 to 65 percent SiO_2 (Barker, 1978). Instead, the near-constant FeO:MgO ratio would be compatible with mixing (or unmixing) of mafic phases of about constant composition and variable proportions of liquid. From the variation in Sc, Yb, Ti, and other elements, it appears that variable hornblende content is a major factor controlling the composition of these samples.

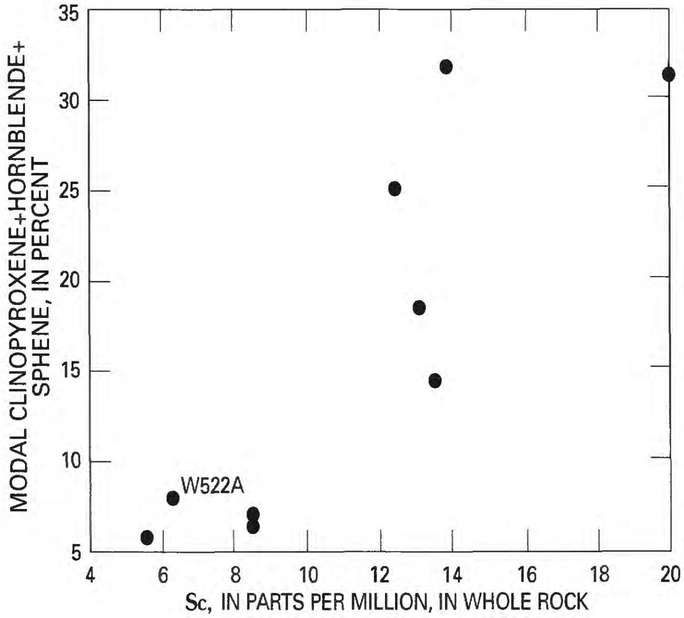


FIGURE 14.—Plot of modal clinopyroxene+hornblende+sphene versus whole rock Sc content from mafic samples of Marinette Quartz Diorite. General positive correlation between Sc content and modal mineralogy suggests that samples are at least in part cumulates.

Although the mafic rocks cannot be taken as directly reflecting liquid compositions, they do provide constraints on the nature of the magma from which they crystallized. The general compositional characteristics of high total alkalies and high TiO_2, P_2O_5, Zr, Nb, Ta, Hf, and light-REE contents, together with the hydrous mineral phases hornblende and biotite, suggest crystallization from a hydrous magma with alkaline affinities (Bender and others, 1984). The existence of such a magma is further supported by the similarity of the REE patterns of the Marinette Quartz Diorite mafic rocks to those of diorites and lamprophyres having alkaline affinities (fig. 16; also Thompson and others, 1984).

A chilled-margin for the Marinette Quartz Diorite has not been recognized, but lamprophyre dikes that cut the volcanic rocks outside the Dunbar dome have chondrite-normalized trace-element patterns similar to those of the Marinette but with lower elemental abundances (fig. 16). These dikes attest to the presence of late hydrous alkaline magmas in the area of the Dunbar dome.

SILICIC ROCKS (SiO_2>65 PERCENT)

The most silicic rocks in the Marinette Quartz Diorite, represented by samples W511 and W511B (table 2), have nearly equal modal amounts of quartz,

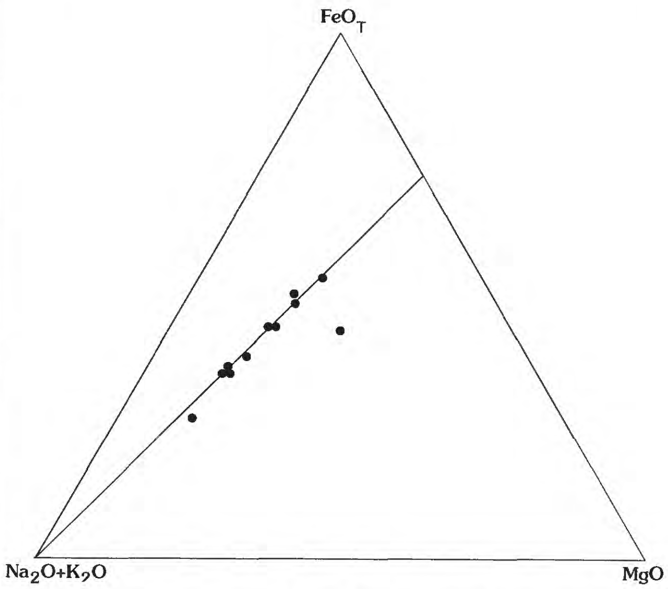


FIGURE 15.—FeO_T-MgO-(Na_2O+K_2O) diagram for mafic samples (SiO_2<65 percent) of Marinette Quartz Diorite. Note the constant FeO:MgO.

plagioclase, and potassium feldspar, high SiO_2 and K_2O contents, low HFS element contents (Zr, Nb, and Y), and variable Rb:Sr. They occur within 25 m of the contact with the Hoskin Lake Granite.

Samples W511 and W511B are compositionally distinct from the Hoskin Lake and Spikehorn Creek Granites and do not lie along the same compositional trends. Of particular importance is their low total REE content, lack of a Eu anomaly, low HFS element contents, and variable Rb:Sr ratios. They represent highly evolved, near-minimum-melt granite compositions.

Highly evolved granitic rocks like these samples can form from silicate melts or aqueous solutions (Hanson, 1980; Nabelek, 1986). In the case of formation from aqueous fluids derived from melts or from crystalline granitic rocks, the resulting product would have REE concentrations significantly lower than that of the parent (Hanson, 1980) and would tend to have relatively low abundances of HFS elements but high Rb, Ba, and Sr (Nabelek, 1986). We propose that these granites represent the crystallized products from a fluid phase exsolved from either the Marinette Quartz Diorite or the Hoskin Lake Granite late in its crystallization history. Exsolution of such an alkali-enriched fluid from the Hoskin Lake is supported by the occurrence of late potassium feldspar crystals that transect quartz veins in the granite. Such a fluid phase could account for the late molybdenum mineralization that is associated with these evolved granites.

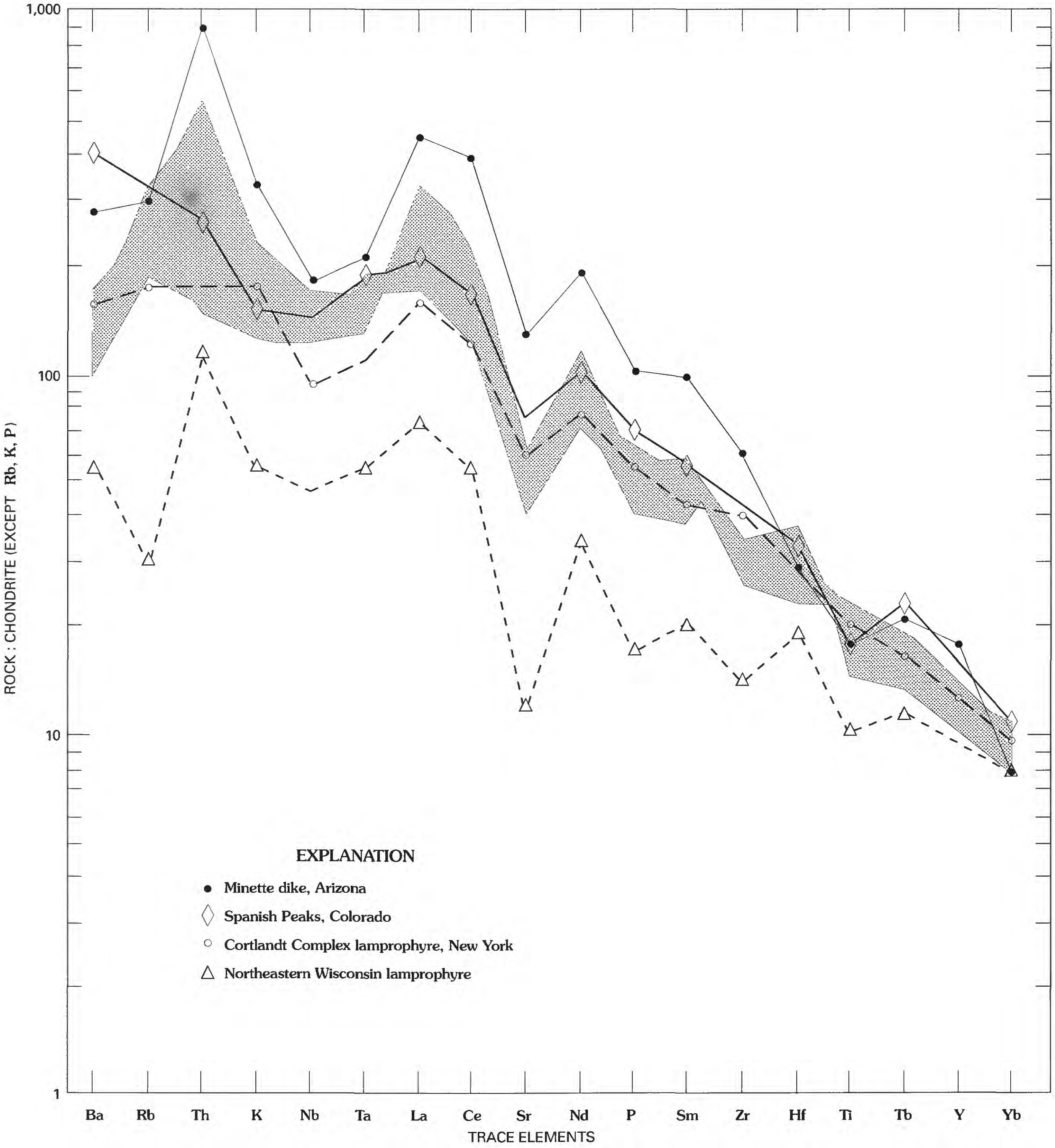


FIGURE 16.—Normalized trace-element diagram for mafic samples of Marinette Quartz Diorite (shaded field) and a lamprophyre dike from east of the Dunbar dome (K.J. Schulz, unpub. data, 1989). Diorite from the middle Tertiary intrusive complex at Spanish Peaks, Colo. (Lipman, 1987), lamprophyre from the Ordovician Cortlandt Complex, N.Y. (Bender and others, 1984), and a minette dike (Oligocene), Arizona (Thompson and others, 1984) are shown for comparison. Note that all samples have similar trace element patterns with relative enrichments of Th and light REE and depletion of Sr. Normalization values from Thompson and others (1984).

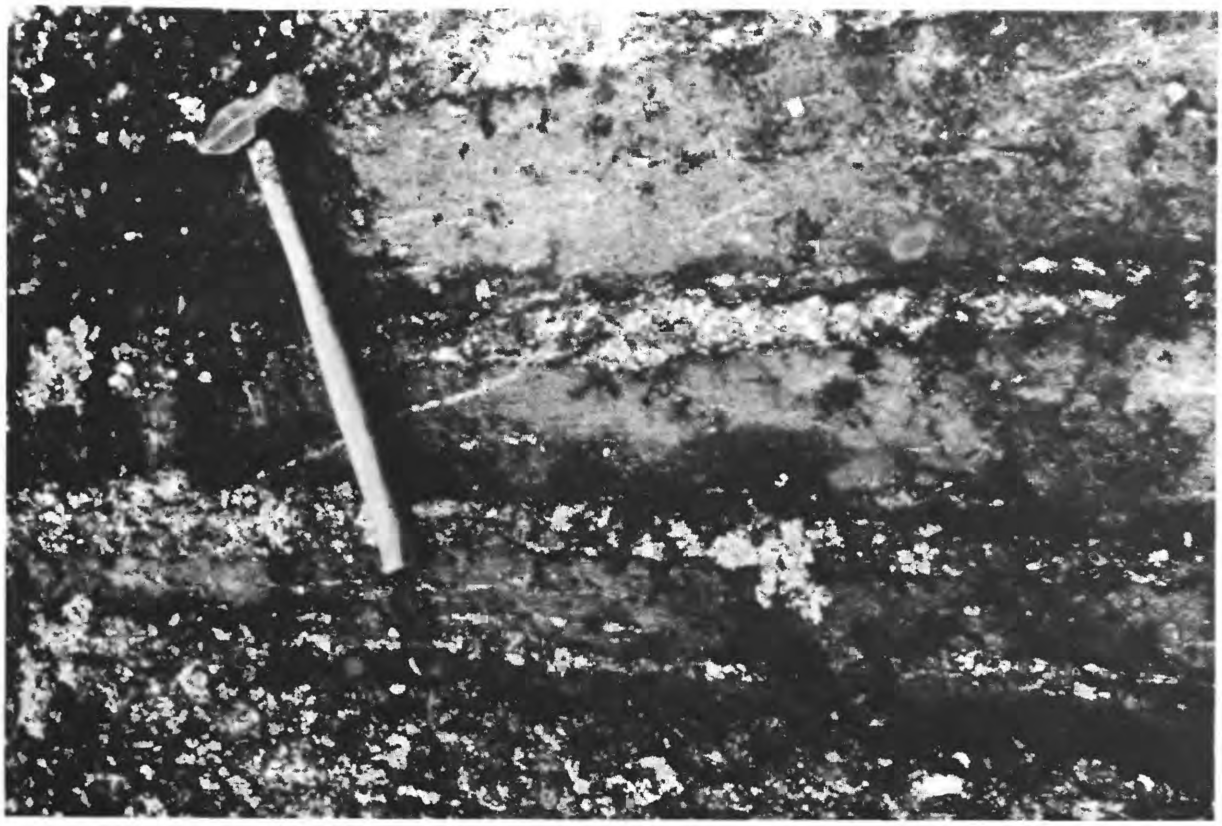


FIGURE 17.—Marinette Quartz Diorite (dark layers) intruded and metasomatized by Hoskin Lake Granite (light layers) (sec. 33, T. 38 N., R. 19 E). Note potassium feldspar megacrysts in diorite layers. Hammer (40.6 cm) for scale.

HOSKIN LAKE GRANITE

GENERAL FEATURES

The arcuate, convex-northward body of granite along the north margin of the dome and that in the Niagara lobe (fig. 2) were named the Hoskin Lake Granite by Prinz (1959, 1965), and the granite that composes the Bush Lake lobe was given the same name by Dutton (1971). However, because of structural and compositional differences, we have delineated the granite bodies separately and renamed the granites in the Niagara lobe (Spikehorn Creek Granite) and the Bush Lake lobe (Bush Lake Granite). The Spikehorn Creek Granite seems to grade into the Hoskin Lake Granite in the area south of Niagara (fig. 2). The Bush Lake Granite appears to be an isolated body.

The Hoskin Lake Granite is characterized by (1) large, oriented, tabular potassium feldspar crystals, (2) abundant inclusions of mafic-intermediate rocks, and (3) late, euhedral crystals of potassium feldspar that lie athwart an older foliation and, at least locally, transect centimeter-thin quartz veins. The rock has been deformed and metamorphosed. Excellent descriptions of the granite appear in Bayley and others (1966).

The Hoskin Lake Granite is well exposed in the Iron Mountain 7½-minute quadrangle (Bayley and others, 1966); it intrudes the Quinnesec Formation and the Marinette Quartz Diorite. The contact between the granite and the Quinnesec is sharp, and generally is a ductile shear zone. In a broad area near the contact, the granite contains small to large inclusions of Quinnesec volcanic rocks, the largest (in sec. 18, T. 38 N., R. 20 E.) being nearly 1 km long. A fault-bounded inclusion, 15 m wide, exposed along the Chicago, Milwaukee, St. Paul and Pacific Railway in SE¼SW¼ sec. 7, T. 38 N., R. 20 E., appears to be a tectonic wedge. Shear zones within the Hoskin Lake in this general area are common, and are oriented subparallel to the contact with the Quinnesec. The granite is reddened within the shear zones. A shear zone in the NE¼NE¼ sec. 18, same township, contains substantial purple fluorite and associated milky quartz lenses about 0.5 m wide. The shear zone has been prospected, apparently as a possible source of sulfide minerals. The contact with the Marinette Quartz Diorite, on the other hand,

TABLE 4.—*Approximate modes and chemical analyses of representative samples of Hoskin Lake Granite, Spikehorn Creek Granite, and Bush Lake Granite*

[Leaders (---), not determined; Tr, trace; <, less than. For samples W38, W506, W711, W155, W411, and W411F, major oxide analyses by J.S. Wahlberg, A. Bartel, J. Taggart, and J. Baker; FeO, H_2O, and CO_2 analyses by H. Neiman, G. Mason, and F. Newman; minor elements by X-ray fluorescence analysis by J. Storey, S. Donahey, B. Vaughn, and M. Coughlin; minor elements by instrumental neutron activation analysis by K.J. Schulz and J.S. Mee. For samples 3817–11–2 and SPB 120–10, major oxide analyses by Z.A. Hamlin; H_2O and CO_2 analyses by R. Somers; minor elements by X-ray fluorescence analysis by R. Johnson, K. Dennen, and B.Scott; minor elements by instrumental neutron activation analyses by K.J. Schulz]

	Hoskin Lake Granite							Spikehorn Creek Granite					Bush Lake Granite			
Sample No...	W38	W528	W508	W509	W506	W711	W523	W155	W158	W162	W407	SPB 120–10	W411	W411F	3817–11–2	W744
Modal analyses (volume percent)																
Plagioclase.......	23.5	54	50	33.7	28	37	49	29	38	33	48	---	30	24	---	---
Quartz.......	19	18	26	36	42	26	24	38	25.4	28	20	---	34	30	---	---
Potassium feldspar.	48	16.3	14.8	27	22	19	15	24	28	32	27	---	26	37	---	---
Biotite.......	8.5	10	9.2	3	5.5	8.9	12	4.5	8	2	4.5	---	9	7.7	---	---
Muscovite.	Tr	Tr	Tr	0	0.8	5.1	0	0.4	Tr	Tr	Tr	---	0.3	Tr	---	---
Sphene......	Tr	Tr	Tr	Tr	0.8	1	Tr	0.4	Tr	0	Tr	---	0.1	0.1	---	---
Opaque oxides....	Tr	Tr	Tr	Tr	0	Tr	Tr	0.4	0	0	0	---	0.1	0	---	---
Apatite......	Tr	Tr	Tr	Tr	0.1	0	Tr	0	0	0	0	---	0.3	Tr	---	---
Allanite.....	Tr	0	Tr	0	0.4	0	Tr	0	Tr	Tr	0	---	0.3	Tr	---	---
Epidote-clinozoisite....	Tr	Tr	Tr	Tr	0.4	0.6	Tr	1.6	Tr	3	Tr	---	Tr	Tr	---	---
Chlorite.....	0	0	0	0	0.7	0.1	0	1.7	Tr	Tr	Tr	---	0.1	0.1	---	---
Carbonate.	Tr	Tr	0	0	1.8	0	0	Tr	0	Tr	0	---	0	0	---	---
Accessory minerals.	1	1.7	Tr	0.3	Tr	Tr	Tr	Tr	Tr	2	0.5	---	Tr	Tr	---	---
Major oxides (weight percent) by X-ray fluorescence analysis																
SiO_2..........	70.8	---	---	---	69.1	70.1	69.6	73.0	---	---	---	71.4	74.6	72.7	74.1	---
Al_2O_3.........	14.2	---	---	---	15.0	14.6	15.2	13.7	---	---	---	15.2	13.7	13.6	14.4	---
Fe_2O_3........	0.48	---	---	---	0.56	0.46	0.63	0.31	---	---	---	0.79	0.23	0.27	0.34	---
FeO..........	1.87	---	---	---	2.67	2.22	1.80	1.78	---	---	---	1.8	0.96	2.07	1.4	---
MgO.........	0.72	---	---	---	0.91	0.83	1.02	0.41	---	---	---	0.92	0.15	0.40	0.38	---
CaO..........	1.55	---	---	---	1.91	1.64	1.71	1.06	---	---	---	1.9	0.65	1.19	1.2	---
Na_2O........	3.24	---	---	---	3.45	3.72	3.71	3.51	---	---	---	4.1	3.43	2.74	3.0	---
K_2O..........	5.15	---	---	---	4.69	4.60	4.59	4.63	---	---	---	4.0	5.06	5.48	5.5	---
H_2O^+........	0.60	---	---	---	0.63	0.58	0.41	LOI=	---	---	---	0.64	LOI=	0.39	0.56	---
H_2O^-.........	<0.01	---	---	---	0.13	0.10	<0.01	0.44	---	---	---	0.12	0.28	0.12	0.09	---
TiO_2..........	0.39	---	---	---	0.51	0.42	0.46	0.18	---	---	---	0.35	0.03	0.27	0.18	---
P_2O_5.........	0.08	---	---	---	0.11	0.10	0.10	<0.05	---	---	---	0.15	<0.05	0.09	0.12	---
MnO.........	0.04	---	---	---	0.05	0.04	0.04	0.05	---	---	---	0.06	0.03	0.04	0.04	---
CO_2..........	0.08	---	---	---	0.08	0.52	0.07	---	---	---	---	0.08	---	0.02	0.17	---
Total......	99.2	---	---	---	99.8	99.9	---	---	---	---	---	---	---	99.4	---	---
Minor elements (parts per million) by X-ray fluorescence analysis																
Rb.............	---	---	---	---	153	180	175	244	---	186	---	150	329	270	249	285
Sr..............	---	---	---	---	279	223	249	117	---	119	---	245	58	143	155	120
Y...............	---	---	---	---	15	14	13	---	---	12	---	18	---	31	39	13
Zr..............	---	---	---	---	163	135	173	137	---	115	---	182	57	154	167	144
Nb............	---	---	---	---	39	42	48	---	---	19	---	30	---	32	35	44

Minor elements (parts per million) by instrumental neutron activation analysis																
U	---	---	---	---	4.5	4.9	5.4	11	---	---	---	4.2	15	11	19	---
Th	---	---	---	---	33.7	33.2	41.5	25.4	---	---	---	23.1	24	46	40	---
Ta	---	---	---	---	4.61	---	4.86	5.50	---	---	---	4.17	5.78	3.90	4.52	---
Hf	---	---	---	---	5.91	---	5.76	4.9	---	---	---	5.58	2.80	5.0	6.18	---
Cr	---	---	---	---	22	---	14	23	---	---	---	10	22	18	3.3	---
Co	---	---	---	---	5.75	---	5.33	2.41	---	---	---	5.06	0.74	2.54	2.09	---
Sc	---	---	---	---	3.31	---	3.86	1.70	---	---	---	3.33	3.20	4.85	5.42	---
Zn	---	---	---	---	44	---	44	---	---	---	---	---	---	47	---	---
Ba	---	---	---	---	705	---	714	466	---	---	---	639	184	787	797	---
Cs	---	---	---	---	4.21	---	4.75	7.1	---	---	---	5.4	12.8	5.73	7.8	---
La	---	---	---	---	56.3	---	50.6	32.0	---	---	---	53.9	19.6	69.2	57.6	---
Ce	---	---	---	---	85.6	---	77.2	59	---	---	---	89.4	42	121	113	---
Nd	---	---	---	---	28	---	26	18	---	---	---	27.5	16	48	36	---
Sm	---	---	---	---	4.43	---	4.10	3.22	---	---	---	4.62	4.05	8.45	7.96	---
Eu	---	---	---	---	0.89	---	0.83	0.54	---	---	---	0.86	0.29	0.88	0.83	---
Gd	---	---	---	---	3.3	---	---	---	---	---	---	---	---	8.9	---	---
Tb	---	---	---	---	0.39	---	0.30	0.52	---	---	---	0.38	0.99	0.85	0.98	---
Yb	---	---	---	---	1.38	---	---	1.14	---	---	---	1.18	3.51	2.60	2.99	---
Lu	---	---	---	---	0.19	---	0.18	0.19	---	---	---	0.18	0.54	0.43	0.50	---

SAMPLE DESCRIPTIONS AND LOCALITIES

Hoskin Lake Granite:

W38 Railway cut in SW¼SE¼ sec. 7, T. 38 N., R. 20 E. Coarse-grained granite containing tabular plagioclase porphyroclasts and subrectangular potassium feldspar porphyroblasts.

W528 Same locality as W38. Sphene is partly converted to opaque oxides.

W508 SE¼SE¼ sec. 28. T. 38 N., R. 19 E. Same lithology as W38. Potassium feldspar porphyroblasts are unoriented and transect a cataclastic foliation. Contains sparse fluorite.

W509 SW¼SE¼ sec. 28, T. 38 N., R. 19 E. Fine-grained, slightly inequigranular granite dike that cuts Hoskin Lake Granite.

W506 SW¼NW¼ sec. 13, T. 38 N., R. 19 E. Contains tabular porphyroblasts of potassium feldspar; has mortar structure.

W711 SE¼NE¼ sec. 16, T. 39 N., R. 20 E. Coarse-grained granite containing bluish-gray quartz grains.

W523 Center sec. 22, T. 38 N., R. 20 E. Hypidiomorphic granular texture.

Spikehorn Creek Granite:

W155 Road cut, U.S. Highway 141–8 in NE¼NW¼ sec. 36, T. 38 N., R. 20 E.

W158 NE¼NW¼ sec. 1, T. 37 N., R. 20 E. Has cataclastic foliation.

W162 NE¼SW¼ sec. 1, T. 37 N., R. 20 E.

W407 NW¼NE¼ sec. 28, T. 38 N., R. 20 E. Rock contains narrow fractures coated by muscovite.

SPB–120–10 Road cut, old U.S. Highway 141–8 in NW¼NE¼ sec. 1, T. 37 N., R. 20 E.

Bush Lake Granite:

W411 NW¼NW¼ sec. 13, T. 38 N., R. 17 E.

W411F Same locality as W411. Rolationship to W411 is not known.

W3817–11–2 Dike cutting metasediments, drill core, NW¼NE¼ sec. 11, T. 38 N., R. 17 E.

W744 SE¼SE¼ sec. 18, T. 38 N., R. 18 E. Medium- to coarse-grained biotite granite.

is gradational, and is marked by a zone of mixed rock many meters wide, as discussed earlier. An excellent exposure showing the relations with the Marinette is in the small hill in SE¼SE¼ sec. 28, T. 38 N., R. 19 E. At this locality, Hoskin Lake lithology is finely intercalated stratigraphically with equigranular layered rocks, and appears to have resulted from partial replacement of certain layers in the gneiss by potassium-rich fluids (fig. 17).

The Hoskin Lake Granite is sill-like in form and convex northward, and it ranges in exposed width from about 0.6 km to 3.2 km. Toward the east, it grades transitionally into the little-deformed Spikehorn Creek Granite in the Niagara lobe. Toward the west, the granite is covered, and its western limit (fig. 2) is uncertain. The granite is cut by numerous small dike-like bodies of aplite and granite pegmatite and by minute ramifying veinlets of quartz and black tourmaline, some of which also contain pyrite and arsenopyrite. Aplite dikes that have been intensely deformed by brittle-ductile deformation are common in the adjacent Quinnesec Formation as much as 0.1 km away from the contact. At places in the vicinity of the contact the dikes are boudined.

Many inclusions of biotite and hornblende schist in the granite—apparently derived from the Marinette Quartz Diorite protolith—were folded prior to emplacement of the Hoskin Lake. In an excellent exposure in a rock knob in SE¼SE¼ sec. 13, T. 38 N., R. 19 E., adjacent to Florence County Highway N, the granite has a foliation expressed by biotite alignment that postdates the folds in the inclusions, and this foliation in turn is transected by relatively undeformed tabular potassium feldspar grains 2–4 cm long. Such tabular feldspars that lie athwart the foliation in the granite are common, and in SE¼SE¼ sec. 28, T. 38 N., R. 19 E., some transect a centimeter-wide quartz vein.

DESCRIPTION

The Hoskin Lake Granite is a pink or gray, coarse-grained inequigranular rock that contains large tabular feldspar megacrysts. At most places, it has a foliation and lineation given by the alignment of the feldspars and biotite plates. Large tabular to subrectangular megacrysts of potassium feldspar and plagioclase (oligoclase) as much as 7 cm long dominate the texture. These are set in a matrix of oligoclase, quartz, potassium feldspar, and biotite. The potassium feldspar, quartz, and oligoclase are present in about equal amounts (fig. 18; table 4), although the proportion of potassium feldspar to plagioclase varies considerably, even within one and the same outcrop. Biotite and its alteration product chlorite compose 5–10 percent of the

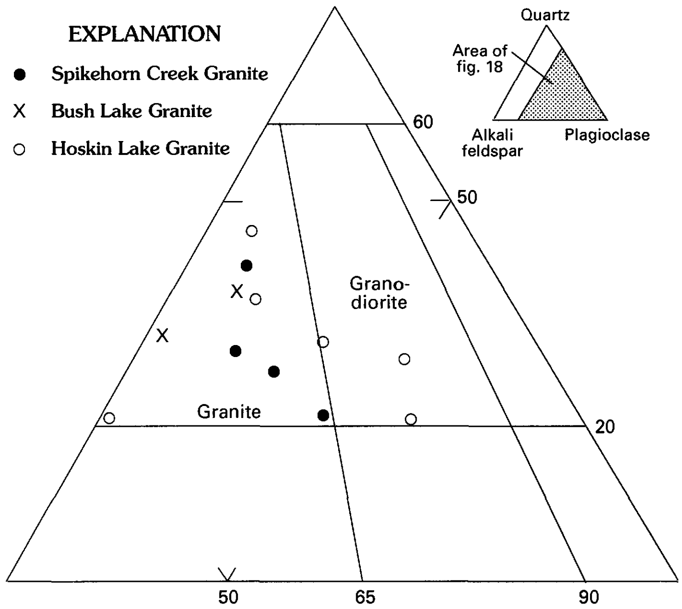


FIGURE 18.—Quartz–alkali feldspar–plagioclase diagram showing composition of representative samples of Hoskin Lake Granite, Spikehorn Creek Granite, and Bush Lake Granite.

rock. The accessory minerals are allanite, sphene, apatite, monazite, and zircon.

The granite is a mylonitic rock in which plagioclase, quartz, biotite and, at places, potassium feldspar have been partly or entirely recrystallized to a finer grain size, commonly resulting in a mortar structure. The deformation was dominantly ductile, but late shear zones indicative of brittle deformation are present locally, particularly along the north margin of the body. The ductile deformation was accompanied by the development of potassium feldspar porphyroblasts, and the brittle deformation by local retrogressive metamorphism.

Plagioclase occurs as (1) tabular or lath-shaped anhedral porphyroclasts that are twinned according to the combined carlsbad-albite law and, rarely, the pericline twin law, and as (2) polycrystalline aggregates of smaller anhedral grains derived by the recrystallization of the older, tabular grains (type IP structure). The older grains have a moderate normal zoning and, at places, oscillatory zoning. They are variably embayed and corroded by potassium feldspar and locally contain small patches of potassium feldspar. The recrystallized plagioclase generally is unaltered and untwinned and contains vermicules of quartz where in contact with potassium feldspar. All the plagioclase seems to be oligoclase, but some of the recrystallized plagioclase could be more sodic. The plagioclase in granite from the transitional zone with the Spikehorn Creek Granite in the Niagara lobe (locality W711, table 4) lacks both a distinct mortar structure and myrmekite, although potassium feldspar is present in the rock. Potassium feldspar has two modes of occurrence: (1) large porphyroblasts of microcline microperthite that includes poikilitic plagioclase, myrmekite, quartz, and biotite, and (2) smaller, anhedral grains interstitial to plagioclase that are associated with polycrystalline aggregates of plagioclase, quartz, and biotite. Some large porphyroblasts contain minute remnants of myrmekitic plagioclase and abundant tiny quartz globules. The potassium feldspar typically has undulatory extinction. Quartz is recrystallized as finer grained aggregates of sutured grains, and at places is drawn out into leaves or lenses. It has a ubiquitous undulatory extinction. The primary quartz grains seem to have been interstitial to plagioclase grains. Biotite is reddish brown where fresh, and generally occurs in clots or as trains of oriented plates associated with moderately fine grained recrystallized plagioclase and quartz. Accordingly, biotite tends to wrap around plagioclase grains and to occur at plagioclase-quartz grain boundaries. Associated coarse sphene locally is partly altered to opaque oxides, and less commonly to epidote-clinozoisite. Dark-yellowish-green hornblende is a minor constituent that occurs in the cores of biotite aggregates. Retrograde alteration is moderately weak except in the vicinity of the late, brittle fractures, where the granite is reddened because of the presence of hematite, which is associated with chlorite. Alteration products of plagioclase are epidote-clinozoisite, muscovite (sericite), and less commonly, calcite. Biotite is variably altered to muscovite and (or) chlorite. Locally, it has sheaths along its cleavage of a zeolite(?) mineral.

SPIKEHORN CREEK GRANITE

GENERAL FEATURES

The Spikehorn Creek Granite is a gray to pinkish-gray, fine- to medium-grained rock containing scattered anhedral potassium feldspar grains as much as 2 cm in diameter. It generally is massive, but locally (especially near the margins of the body), it bears a mylonitic foliation expressed mainly by recrystallized quartz leaves and oriented biotite. This foliation tends to be subparallel to the contacts, but at places is oriented northeastward, at a large angle to the trend of the contact.

The granite has sharp intrusive contacts against the Quinnesec volcanic rocks and the Marinette Quartz Diorite, and small ramifying dikes intrude these rocks for distances of as much as 400 m from the contact. The granite in the dikes mainly has an aplitic texture and is leucocratic. Where observed, as in sec. 1, T. 37 N., R. 20 E., along U.S. Highway 141–8, the contact against the Quinnesec is highly sheared and marked both by brittle and ductile deformation and by dikes of granite in the adjacent volcanic rocks. The granite in the contact zone is altered and contains abundant gray quartz pods and scattered almond-shaped inclusions of the volcanic rocks. A large xenolith of the metavolcanic rocks about 0.15 km wide and perhaps a kilometer long is present about 0.5 km north of this contact zone. The contact of the granite with this xenolith also is marked by shears and a cataclastic foliation and by thin ramifying aplite dikes. The contacts of the granite against the Marinette Quartz Diorite, where observed along the power line in NW¼ sec. 28, T. 38 N., R. 20 E., crosscut layering in the quartz diorite. Certain layers in the quartz diorite contain large potassium feldspar megacrysts, indicating that, at least here, they were formed before the Marinette was intruded by the Spikehorn Creek Granite.

DESCRIPTION

The Spikehorn Creek is a granite as defined by Streckeisen (1976) (fig. 18; table 4). All thin sections examined show some mylonitic or cataclastic textures, mainly expressed by recrystallized quartz but locally expressed by recrystallized biotite and by narrow shears filled mainly with sericite. Plagioclase (sodic oligoclase) has weak normal zoning, and is mainly twinned according to the albite and pericline twin laws. In addition to occurring as single zoned crystals, the plagioclase also forms glomeroporphyritic aggregates of small zoned crystals as much as 5 mm across. The grains locally contain small patches of potassium feldspar, and are weakly altered to epidote-clinozoisite and sericite. Potassium feldspar is microcline microperthite of both string and bleb types and, less commonly, grid-twinned microcline. The microcline embays plagioclase, and its principal form is as crystals as much as 6 mm in diameter that contain inclusions of plagioclase with albite rims, quartz, and biotite. Quartz has undulose extinction and is locally recrystallized as new grains that are aggregated and locally aligned, to give a mylonitic foliation. Biotite is dark reddish brown and typically has ragged outlines. Most sphene and opaque oxides occur at grain margins and at the junction between unaltered biotite and marginal zones that contain wormy intergrowths of quartz. Sheaths of a fibrous mineral, possibly a zeolite mineral, occur at places parallel to the cleavage, and epidote-clinozoisite is a common alteration mineral along grain margins. Chlorite is another common alteration mineral after biotite. In some sections, opaque oxides occur in late fractures in the rock.

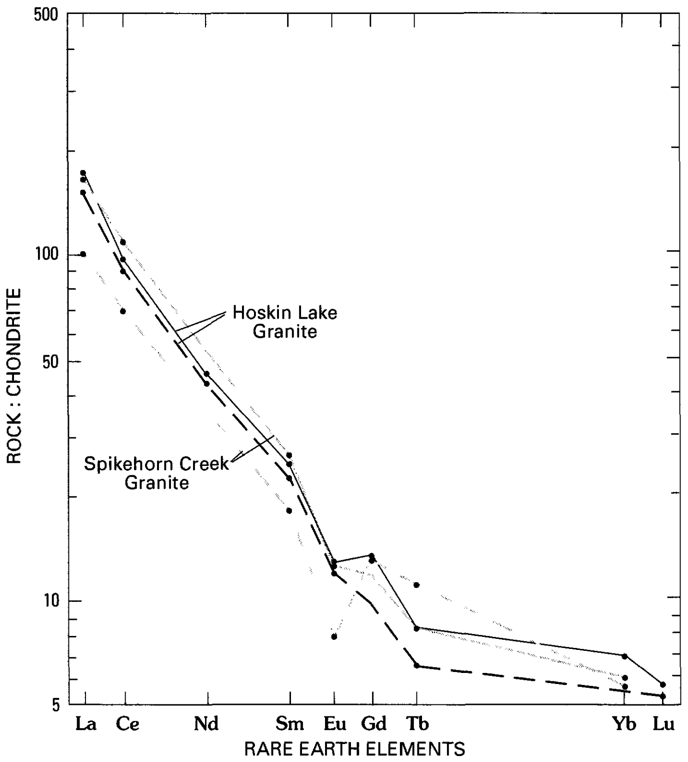


FIGURE 19.—Chondrite-normalized rare earth element plots for Hoskin Lake and Spikehorn Creek Granites. Chondrite-normalization values from Haskin and others (1968).

Accessory minerals are zircon, allanite, apatite (rare), and fluorite (rare).

GEOCHEMISTRY

The Spikehorn Creek Granite (of the Niagara lobe) and the Hoskin Lake Granite are compositionally similar except that the Hoskin Lake has a slightly higher K_2O content. Rb:Sr ratios range from about 0.55 to 2.0 and show a positive correlation with increasing SiO_2 content; K:Rb ratios range from about 250 to 155 and show a negative correlation with increasing SiO_2 content. The samples show light-REE enrichment, small to moderate negative Eu anomalies, and decreasing light-REE abundance with increasing SiO_2; they also show slightly fractionated heavy-REE (fig. 19).

PETROGENESIS OF HOSKIN LAKE AND SPIKEHORN CREEK GRANITES

The granitic rocks in the dome have had a complex crystallization and post-crystallization history. The Spikehorn Creek Granite in the Niagara lobe clearly was injected as a magma, as indicated by its internal homogeneity and the sharp-walled inclusions of the Quinnesec Formation. Although its contact with the Quinnesec is mainly tectonic, local crosscutting intrusive relations with the Quinnesec are indicative of magmatic injection. As noted later, this body is interpreted to have been emplaced diapirically. In accordance with earlier field observations by Prinz (1959) and Bayley and others (1966), no evidence exists that the granite was produced by granitization of Quinnesec basalt, as suggested by Lyons (Emmons and others, 1953, p. 108).

Both field and petrographic observations indicate that the Hoskin Lake Granite also formed as an igneous intrusion with local selective replacement of preexisting rocks of the Marinette Quartz Diorite. Subsequent to or late in the crystallization of the intrusion, the rock was modified by neocrystallization of potassium feldspar porphyroblasts. Apparently much of the porphyroblastic potassium feldspar formed during ductile deformation, as indicated by oriented tabular feldspar grains in many outcrops; but at least some of it formed after the deformation, as indicated by disoriented crystals that lie across the foliation and by rare crystals that transect thin quartz veins in the granite.

A possible mechanism for the origin of the potassium feldspar porphyroblasts is partial replacement of plagioclase via potassium-rich fluids released late in the crystallization history of the Hoskin Lake Granite and fluid movement through cataclastic and mylonitic zones in the host rocks. The replacement mechanism frees sodium that moves to adjacent, altered plagioclase grains, which are recrystallized as more sodic species. Myrmekite is formed where more silica is trapped ahead of advancing potassium replacement than can be incorporated in plagioclase relative to its anorthite content. Where silica is not excessive, it is incorporated in potassium feldspar, as was noted in several thin sections.

Judged from the distribution of amphibolite-facies mineral assemblages in rocks both within and outside the core of the Dunbar dome, the Hoskin Lake Granite generally coincides with the zone of highest temperature attained during the doming process. This zone also was the locus of maximum potassium metasomatism of preexisting rocks.

The general field, petrographic, and compositional characteristics of the Hoskin Lake and Spikehorn Creek Granites are similar and suggest that the two granites may be related or at least derived by comparable processes. Also, their spatial association and similarity in mineralogy and composition with the Marinette Quartz Diorite suggest that the granites could be directly related to the diorite. The lack of straight-line compositional trends between the granites and mafic Marinette Quartz Diorite samples precludes a simple mixing or unmixing relationship

TABLE 5.—*Mass balance calculation using samples W522A and W523 and mineral compositions from Bender and others (1984)*

The parent is W522A:		
Coefficient	Percent	
0.090	0.355	AMPH2
0.010	0.040	BIOT2
0.146	0.579	PLAG2
0.001	0.006	SPH2
0.005	0.032	APATITE
0.748		W523 is the daughter

	SiO_2	TiO_2	Al_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O	P_2O_5
W523	70.45	0.47	15.39	2.40	0.04	1.03	1.73	3.76	4.65	0.10
OBS	66.14	0.80	16.18	3.65	0.07	1.59	3.26	4.38	3.75	0.18
CALC	66.22	0.88	15.96	3.64	0.06	1.70	3.16	4.45	3.69	0.31
DIF	–0.03	–0.08	0.11	0.01	0.01	–0.11	0.10	–0.07	0.06	–0.13
Sum of squares of residuals =0.067										

between them. As shown in figure 20, the granite samples plot along trends offset from the mafic samples but extending back through sample W522A. Sample W522A is intermediate in composition between the mafic Marinette Quartz Diorite and the Hoskin Lake and Spikehorn Creek Granite samples but has a REE pattern similar to the granites except for the lack of a Eu anomaly. This suggests that crystal fractionation of the Marinette Quartz Diorite, or a compositionally similar magma, may have given rise to the Hoskin Lake and Spikehorn Creek Granites. Mass balance calculations using sample W522A as a parent and mineral compositions from the literature for compositionally similar rocks (Bender and others, 1984) suggest that the granites could be derived by about 25 percent fractional crystallization involving plagioclase+amphibole+biotite+apatite+sphene. (See table 5.) These phases are compatible with those observed in the Marinette rocks (table 2). The slightly peraluminous composition of the granites, suggesting equilibration with metaluminous phases like amphibole, is also compatible with this model.

We tentatively conclude that the Hoskin Lake and Spikehorn Creek Granites could have been derived by crystal fractionation from the Marinette Quartz Diorite or a similar magma at depth. It is likely that the actual fractionation process was more complex than simple fractional crystallization and may have involved segregation of phases by sidewall crystallization as described by Noyes and others (1983) and Sawka (1988) for plutons of the Sierra Nevada batholith.

It should be noted that sample W155 (table 4), the most evolved of the Spikehorn Creek granitic rocks, shows a relative depletion in light REE, Th, Ba, Sr, and Eu. Preferential depletion of light-REE content in evolving high SiO_2 systems has been shown to be a common feature (Miller and Mittlefehldt, 1984) variously attributed to (1) fractionation of minor phases such as allanite or monazite, which have partition coefficients much greater than one for the light REE (Mittlefehldt and Miller, 1983), (2) liquid-state diffusion processes in zoned magma chambers (Hildreth, 1979), and (3) transport in a fluid phase due to complexing of light REE (Hudson and Arth, 1983). In the case of the Spikehorn Creek Granite, the combined depletion of both the light REE and Th in sample W155 would be compatible with fractionation of a phase like monazite (Mittlefehldt and Miller, 1983) or allanite (Brooks and others, 1981). The depletion in Ba, Sr, and Eu suggests further that potassium feldspar (±biotite) was a major fractionating phase. This is in agreement with the observed presence of megacrystic potassium feldspar in the Hoskin Lake and Spikehorn Creek Granites.

BUSH LAKE GRANITE

GENERAL FEATURES

The granite in the Bush Lake lobe is poorly exposed, and the outline of the lobe as shown in figure 2 is based largely on aeromagnetic data and scattered exposures mapped earlier by Dutton (1971). Contacts with adjacent rocks have not been observed, but granite dikes presumably related to the main body cut the country

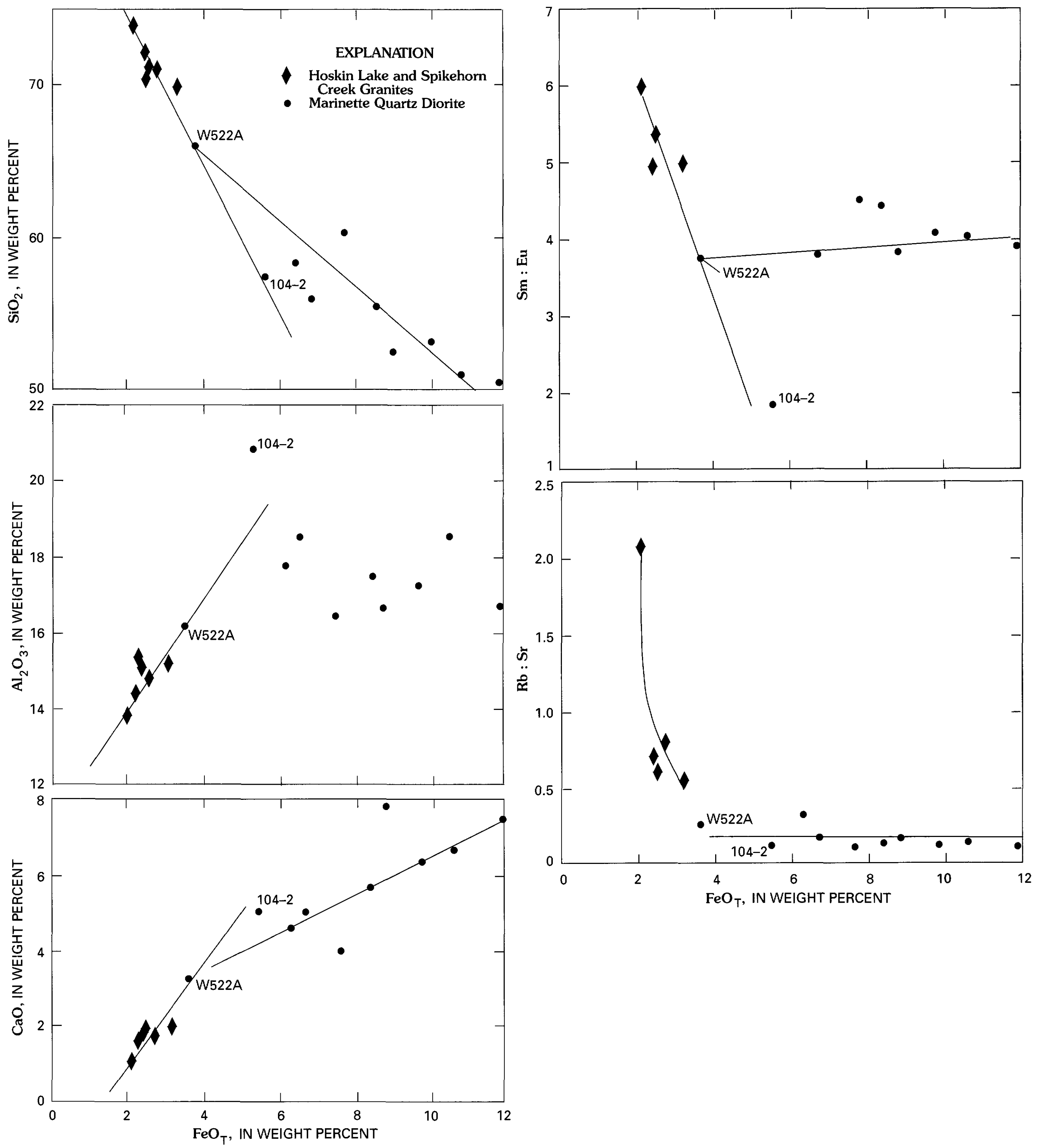


FIGURE 20.—Variation of CaO, Al_2O_3, SiO_2, Rb:Sr, and Sm:Eu versus FeO_T for Marinette Quartz Diorite and Hoskin Lake and Spikehorn Creek Granites. Note that the data define two trends that project through sample W522A, the most evolved diorite sample. Sample SDNE–104–2–82, the plagioclase cumulate sample, is also identified by reference (104–2).

rocks on the west and north sides of the lobe. The following description is partly taken from Dutton's (1971) report.

DESCRIPTION

The granite is a gray, buff, or orange-pink, coarse-grained, slightly porphyritic rock containing feldspar phenocrysts that commonly range from 1 to 4 cm in length. Where observed, it has been deformed and has a weak to moderate secondary foliation expressed mainly by oriented biotite and fine-grained aggregates of recrystallized quartz. Potassium feldspar generally is more abundant than plagioclase, but in some rocks the two are nearly equal in amount (table 4). Plagioclase is strongly inequigranular and is mainly calcic oligoclase or sodic andesine. Larger, primary crystals as much as 0.5 mm in size are anhedral and are twinned according to combined carlsbad-albite and pericline twin laws. Typically, the primary grains are partly surrounded by finer grained polycrystalline aggregates (type IP of Hanmer, 1982). These aggregates are intergrown with fine-grained recrystallized quartz, biotite, and potassium feldspar. Not uncommonly, the polycrystalline aggregates occupy shears. Potassium feldspar is dominantly microcline microperthite, which forms anhedral or subhedral porphyroblasts as much as 4 cm long; they are twinned according to the carlsbad law. The larger grains are strongly poikilitic and include plagioclase, quartz, and biotite. The potassium feldspar embays and replaces plagioclase. Smaller grains associated with new plagioclase aggregates tend to have grid twinning, and the adjacent plagioclase has extensive vermicular intergrowths of quartz. Quartz is gray and has wavy extinction. It embays plagioclase, and is recrystallized as aggregates of sutured grains, some of which are lensoid. Biotite is the only mafic mineral. It is dark reddish brown and typically occurs as ragged crystals. It has mainly been recrystallized and is associated with new grains of plagioclase and quartz; as a result, it commonly forms partial rims around primary plagioclase crystals. The accessory minerals, apatite, allanite (altered), and zircon, are associated closely with biotite. Zircon, a common mineral, forms zoned prismatic grains. The rocks show weak retrogressive alteration, mainly expressed by the alteration of biotite to sphene and muscovite or to chlorite.

GEOCHEMISTRY

The Bush Lake Granite is compositionally distinct from the Hoskin Lake and Spikehorn Creek Granites; it has slightly higher SiO_2 and K_2O contents and higher K_2O:Na_2O, U:Th, and Rb:Sr ratios. The REE patterns are also distinctive (fig. 21), exhibiting large negative Eu anomalies, relatively flat heavy-REE slope, and significant depletion in the light REE with increasing SiO_2 content.

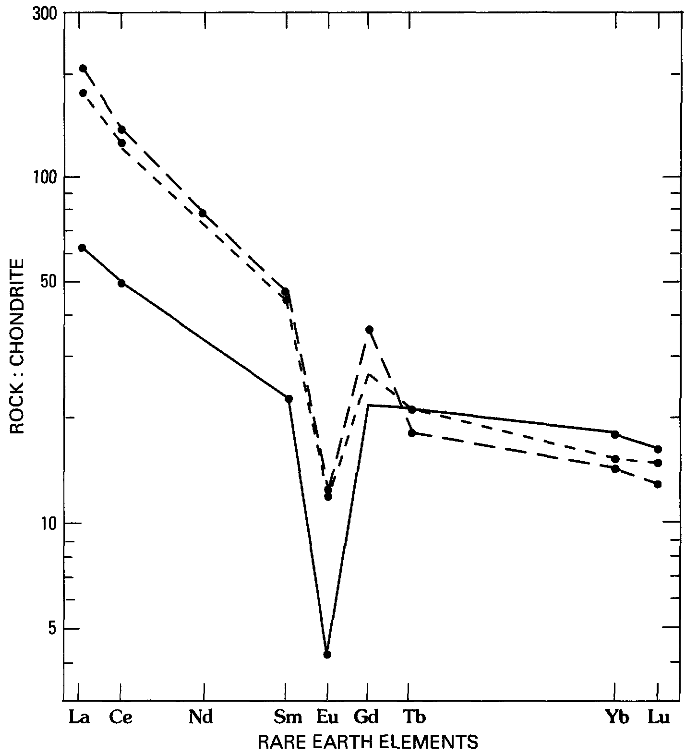


FIGURE 21.—Chondrite-normalized rare earth element plots for samples of Bush Lake Granite. Chondrite-normalization values from Haskin and others (1968).

GRANITOID DIKES, PEGMATITE, AND APLITE

GENERAL FEATURES AND DESCRIPTION

Small granitoid dikes, mainly granodiorite and tonalite, intrude all the rocks in the core of the Dunbar dome and a narrow fringe of metasedimentary and metavolcanic rocks. These dikes range from less than 1 m to several meters wide. Pegmatite forms sheetlike bodies that are as much as hundreds of meters thick, mainly in the Dunbar Gneiss in the southwestern part of the dome, and aplite-pegmatite composite dikes are common. Except for a late garnet-bearing aplite and associated pegmatite, all these rocks are foliated, indicating that they were deformed (D_2) together with their host rocks subsequent to emplacement.

TABLE 6.—*Approximate modes and chemical analyses of representative samples of aplite dikes in the Dunbar Gneiss*

[Leaders (---), not determined; Tr, trace; <, less than. Major oxide analyses (W143C) by J.S. Wahlberg, A. Bartel, J. Taggart, and J. Baker, and (W495, W679) by Z.A. Brown. Minor elements by X-ray fluorescence analysis (W143C) by J. Storey, S. Donahey, B. Vaughn, and M. Coughlin, and (W495, W679C) by R. Johnson, K. Dennen, and B. Scott; minor elements by instrumental neutron activation analysis by L.J. Schwarz]

Sample No.	W143C	W495	W679C
Modal analyses (volume percent)			
Plagioclase	46	36	24
Quartz	29	35	27
Potassium feldspar	23	27	46
Biotite	1.4	0.1	0.8
Muscovite	0	0.7	0.8
Garnet	Tr	0.2	1
Zircon	Tr	Tr	0
Apatite	0	0	0.1
Tourmaline	0	0	0.1
Allanite	Tr	Tr	0
Monazite	Tr	Tr	0
Chlorite	Tr	Tr	Tr
Sericite	Tr	0	Tr
Major oxides (weight percent) by X-ray fluorescence analysis			
SiO_2	74.2	76.3	72.9
Al_2O_3	14.5	13.3	14.4
Fe_2O_3	0	0.13	0.14
FeO	0.65	0.58	0.74
MgO	0.12	0.06	0.06
CaO	0.43	0.39	0.29
Na_2O	5.21	4.4	3.0
K_2O	3.90	4.6	7.8
H_2O^+	LOI=	0.86	0.67
H_2O^-	0.17	0.14	0.14
TiO_2	<0.02	0.02	0.03
P_2O_5	<0.05	0.04	0.05
MnO	<0.02	0.14	0.23
CO_2	---	0.02	0.02

TABLE 6.—*Approximate modes and chemical analyses of representative samples of aplite dikes in the Dunbar Gneiss*—Continued

Sample No.	W143C	W495	W679C
Minor elements (parts per million) by X-ray fluorescence analysis			
Rb	454	429	666
Sr	9.75	38	47
Y	98	64	36
Zr	35	39	43
Nb	60	55	66
Minor elements (parts per million) by instrumental neutron activation analysis			
U	7.6	4.9	4.6
Th	6.6	8.6	6.7
Ta	23.2	23.0	16.2
Hf	2.9	2.5	2.6
Cr	28	20	14
Co	0.6	0.4	0.6
Sc	18.10	3.84	2.79
Zn	<6	7	23
Ba	65	68	138
Cs	1.5	2.1	4.9
La	5	15	8
Ce	11	38	21
Nd	---	---	---
Sm	3.9	5.3	5.7
Eu	0.03	0.03	0.15
Gd	---	6.7	---
Tb	1.44	0.88	0.84
Yb	11.1	8.9	3.5
Lu	1.56	1.23	0.46

SAMPLE LOCALITIES

W143C	NE¼SE¼ sec. 14, T. 37 N., R. 18 E.
W495	NW¼NE¼ sec. 15, T. 37 N., R. 18 E.
W679C	SW¼SW¼ sec. 21, T. 37 N., R. 18 E.

Granitoid dikes are locally common in the Dunbar Gneiss. They are mainly light gray or pinkish-gray, fine-grained foliated rocks—dominantly tonalite or granodiorite and generally leucocratic. Plagioclase commonly has a weak concentric zoning and well-developed albite twinning, and is myrmekitic adjacent to potassium feldspar. Where present, potassium feldspar is generally interstitial to plagioclase and quartz. A brown biotite is the dominant mafic mineral; it is associated with sphene, apatite, and epidote-clinozoisite. Green hornblende is a local mafic mineral. Zircon, apatite, and allanite are common accessory minerals.

Pegmatite is abundant in much of the Dunbar Gneiss; its volume locally exceeds that of the country rock. The pegmatite is white and very coarse grained, and consists of bleb-type microcline microperthite, quartz, and plagioclase with distinct albite rims. Muscovite is an abundant alteration mineral, and tiny red garnet crystals are present in some finer grained parts. The pegmatite generally has an ill-defined foliation expressed by highly strained and sutured elongate aggregates of quartz. Brown or pink aplite of two generations also is common; an older generation is deformed, whereas a younger garnet-bearing generation, which is largely confined to the western part of the Dunbar Gneiss, is not deformed. Commonly, this latter aplite is associated with pegmatite in zoned dike-like bodies, but it also occurs as discrete dikes. Modes of three selected samples of garnet-bearing aplite are given in table 6. The plagioclase is albite, and in one section (W143C) the feldspar has a fine-scale rapakivi texture.

Granitoid dikes that cut the Newingham Tonalite are medium-gray, locally slightly porphyritic tonalite and granodiorite, and presumably are cogenetic with the Newingham Tonalite. They are distinctly finer grained than the host rock and are foliated. Plagioclase has a moderate concentric zoning, and biotite is the principal mafic mineral. Locally, sphene is associated with the biotite; other accessory minerals are zircon and allanite.

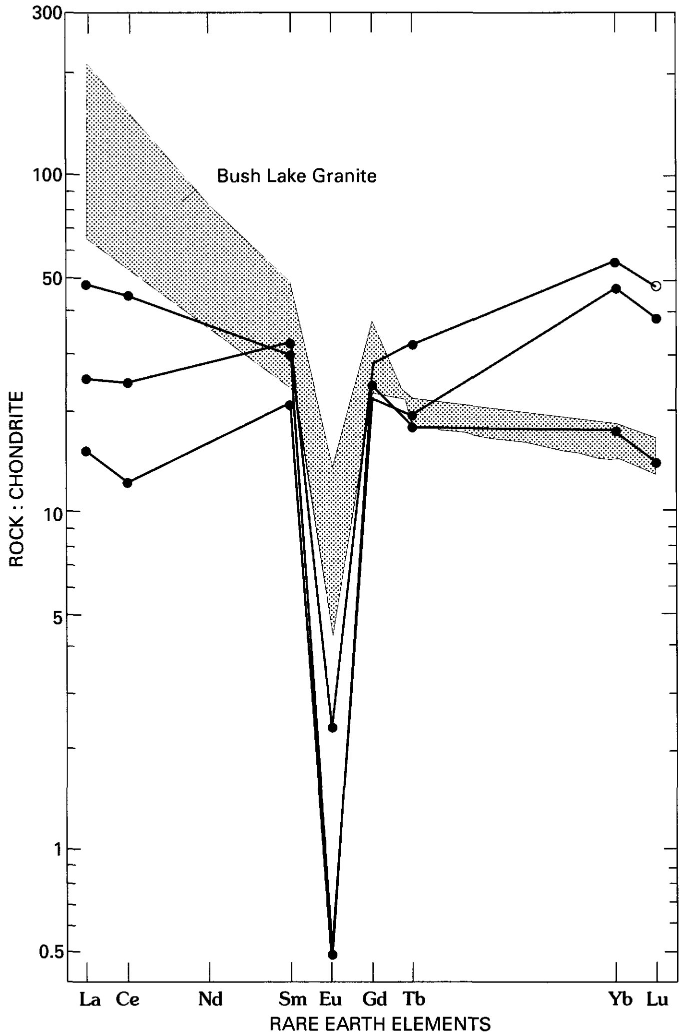


FIGURE 22.—Chondrite-normalized rare earth element plots for three garnet-bearing aplite dikes. Shaded field shows rare earth patterns for Bush Lake Granite. Chondrite-normalization values from Haskin and others (1968).

Granite dikes are also common in the Hoskin Lake Granite, at least near the south margin of the body. They are medium gray or pinkish gray, equigranular, and much finer grained than the Hoskin Lake Granite. A mode of a typical dike (W509) is given in table 4. The granite has elongate aggregates of quartz and the biotite is strongly oriented, indicating that it was deformed and metamorphosed subsequent to its emplacement. Plagioclase occurs as anhedral grains that have good albite twinning; myrmekite is developed adjacent to potassium feldspar. Potassium feldspar is slightly perthitic and includes poikilitic quartz and plagioclase. Biotite is reddish brown and subhedral, and contains tiny rutile needles. Zircon occurs as zoned crystals as much as 0.1 mm long.

GEOCHEMISTRY OF GARNET-BEARING APLITE DIKES

Garnet-bearing aplite (table 6) has highly evolved compositions with high SiO_2 and total alkali contents. Relative to the granite of the Dunbar dome, the aplite is strongly depleted in FeO_T, MgO, CaO, TiO_2, P_2O_5, MnO, Sr, Zr, Ba, Eu, and light REE, but is enriched in SiO_2, total alkalies, Y, Ta, Nb, Rb, and the heavy REE. The REE patterns are particularly distinctive, with flat to positive slopes and very large negative Eu anomalies (fig. 22). The aplites also have very low Zr:Hf (<17) and Nb:Ta (<5) ratios. The aplite probably represents an evolved differentiate from the Bush Lake Granite, as discussed following.

PETROGENESIS OF BUSH LAKE GRANITE AND LATE APLITE DIKES

The Bush Lake Granite is compositionally distinct from the Hoskin Lake and Spikehorn Creek Granites, as discussed previously. Further, examination of the compositional characteristics of the Bush Lake Granite reveals that it is probably not genetically related to the Marinette Quartz Diorite–Hoskin Lake–Spikehorn Creek suite. For example, for a similar SiO_2 content, the Bush Lake Granite has higher K_2O, Rb, Th, U, and REE contents than the Hoskin Lake and Spikehorn Creek Granites as well as a distinctive chondrite-normalized REE pattern. Such differing abundances are incompatible with the observed compositional trends shown by the Hoskin Lake and Spikehorn Creek Granites.

On compositional variation diagrams, the garnet-bearing aplite dikes plot on extensions of the trends defined by the Bush Lake Granite, suggesting that the aplites (and related pegmatites) may represent highly evolved differentiates of this granite. As shown in figure 23, relative to the least evolved Bush Lake Granite sample (W411F), the relatively evolved sample W411 is enriched in Na, Cr, Rb, Ta, and heavy REE (Tb-Lu) and depleted in all other components. Further, the figure shows that the aplites are depleted and enriched in the same components as W411, but the magnitudes of difference are greater than in sample W411, the only exception being Sc, which is depleted in sample W411 but enriched in the aplites.

Miller and Mittlefehldt (1984) have shown that a general similarity exists in compositional trends (that is, enrichments and depletions) among extremely fractionated felsic magmas. They have also demonstrated that these trends are at least qualitatively like those anticipated from fractional crystallization. Alternatively, Hildreth (1981) has argued for noncrystal-

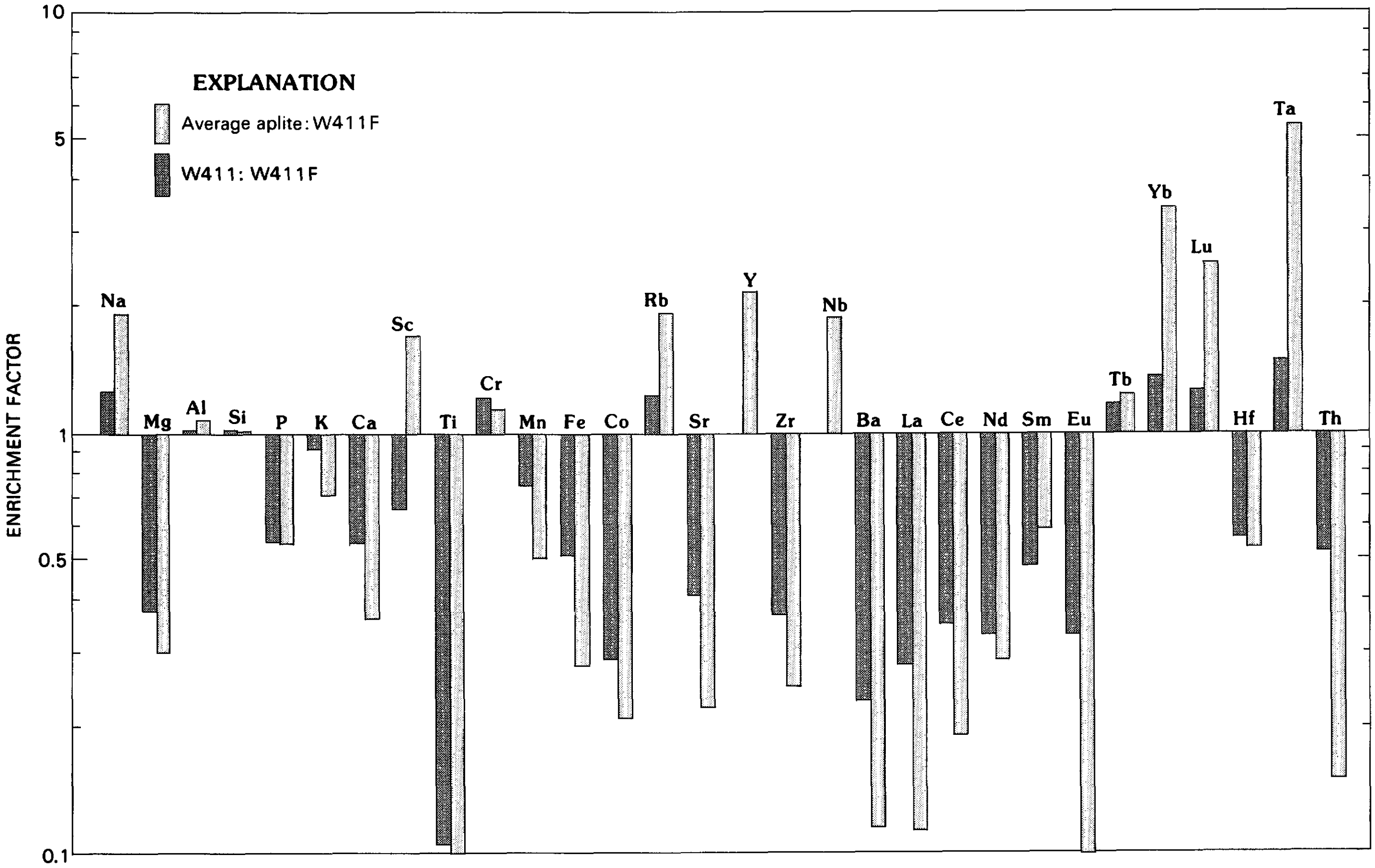


FIGURE 23.—Elemental enrichments and depletions in sample W411, the most evolved Bush Lake Granite sample, and in the average aplite composition relative to the least evolved Bush Lake Granite sample, W411F. Values plotted are concentration of element in sample W411 and average aplite divided by the concentration in sample W411F. Note that except for Sc, the average aplite composition has a similar enrichment and depletion pattern to that of sample W411.

liquid processes like thermogravitational diffusion to control evolution in high SiO_2 magmatic systems.

Comparison of the compositional trends between samples of the Bush Lake Granite and the garnet-bearing aplite dikes in the Dunbar Gneiss with those discussed by Miller and Mittlefehldt (1984) for other high SiO_2 suites confirms the remarkable similarity between these trends, including: (1) only slight variation of Si and Al, (2) slight increase of Na and depletion of K, (3) large depletion of Mg, Ti, Ba, Sr, Zr, and light REE, and (4) large enrichments of Rb, Y, and heavy REE. As discussed by Miller and Mittlefehldt (1984; see also Michael, 1984), this enrichment-and-depletion pattern is compatible with fractionation of biotite (Fe, Mg, Ti, Mn, and Ba depletion), potassium feldspar (K, Sr, Ba, and Eu depletion), plagioclase (Ca, Sr, and Eu depletion), and accessory phases like allanite and monazite (light REE and Th depletion).

An interesting compositional characteristic of the aplites is their anomalously low Nb:Ta (<5) and Zr:Hf (<17) ratios. These low ratios result from the preferential enrichment of Ta to Nb and depletion of Zr to Hf (fig. 23). In contrast, typical igneous suites show Zr:Hf ratios decreasing from 60 in gabbroic rocks to about 40–30 in diorite and granitic rocks, and near constant Nb:Ta ratios of about 17–11 (Cerný and others, 1985). However, low ratios similar to those of these aplites are observed in some other highly evolved, often rare-metal granitic pegmatites and aplites (Cerný and others, 1985).

The cause of the low Zr:Hf and Nb:Ta ratios remains uncertain. The absolute depletion of Zr and Hf in the aplites is compatible with fractionation of zircon in a F-dominated felsic system, in contrast to an enrichment of these elements observed in Cl-rich systems (Hildreth, 1981). It has been suggested that differences in the stabilities and mobilities of Zr- and Hf-bearing complexes may enhance the effects of crystal-melt fractionation in highly evolved granitic systems and result in decreasing Zr:Hf ratios (Cerný and others, 1985). The same relationship may apply to Nb and Ta. Thus, fluid complexing may also have contributed to the fractionation of some element ratios in the highly evolved Bush Lake granite-aplite system.

In conclusion, the Bush Lake Granite and garnet-bearing aplites from the Dunbar dome appear to represent the products of a fractionated high SiO_2 magmatic system. Compositional trends are generally compatible with fractionation of observed crystal phases, although late-stage fluid complexing may have contributed to enhanced fractionation of Zr:Hf, Nb:Ta, and LREE:HREE ratios.

The Bush Lake Granite is compositionally similar to many other high SiO_2 felsic systems. Hildreth (1981) and Thompson and others (1984) stated that granite-rhyolite systems having such characteristic extreme depletions of Ba, Sr, P, and Ti appear only in continental settings. Examples include the late orogenic granites from the Variscan and Caledonian belts of England (Watson and others, 1984), the Pikes Peak pluton (Colorado) and its associated Redskin cupola (Ludington, 1981), and the high- and low-temperature parts of the Western United States Bishop Tuff (Hildreth, 1979). Following Hildreth (1981) and Thompson and others (1984), we conclude that the Bush Lake Granite was probably derived largely from a crustally derived melt possibly generated and modified by basalt additions to the crust.

METASEDIMENTARY ROCKS

Metasedimentary rocks are exposed sparsely on the west side of the Dunbar dome and adjacent to and northwest of the Bush Lake lobe, in the extreme northwestern part of the map area (fig. 2). They were included by Dutton (1971) in the Quinnesec Formation, but subsequent mapping, geophysical data, and core drilling by a private company show that these strata underlie the Quinnesec (Schulz and Sims, 1982). The thickness of the succession is not known because of folding, but on the west side of the Bush Lake lobe proprietary aeromagnetic, electromagnetic, and geologic data suggest a thickness at places of a kilometer or more. Metasedimentary rocks seem to be absent or at least thinner on the east side of the Bush Lake lobe. The belt of metasedimentary rocks extending north-northwest from the end of the Bush Lake lobe has an estimated width of about 0.5 km. We interpret these rocks as lying above a north-northwest-trending buried extension of the granite in the Bush Lake lobe, because of their amphibolite grade and the scattered small bodies of granitoid rocks that cut them and the adjacent metavolcanic rocks (Dutton, 1971, pl. 1).

The exposed metasedimentary rocks are mainly quartz rich schist, impure marble, calc-silicate rocks, and biotite schist. The electromagnetic data and drilling indicate that a graphitic schist lies along the west side of the Bush Lake lobe, near the granite in the Bush Lake lobe. It is apparently stratigraphically below the exposed metasedimentary rocks.

A large exposure of metasedimentary rocks in the SE¼ sec. 7, T. 37 N., R. 18 E. consists of interbedded calc-silicate rocks and biotite schist that are cut by granite pegmatite and aplite, identical to that exposed within the western part of the Dunbar dome. On the northwest margin of the Bush Lake lobe, a succession at least 100 m thick of marble, calc-silicate rocks, and thin interbeds of biotite schist and ferruginous quartzite is exposed on a low hill in the NW¼SW¼ sec. 11, T. 38 N., R. 17 E. The marble has structures suggestive of stromatolites. The principal exposures northwest of the Bush Lake lobe are in and adjacent to the Popple River (Dutton, 1971, pl. 1). At Popple Rapids, in the NW¼ sec. 27, T. 39 N., R. 17 E., a succession at least 300 m thick of thin- to medium-bedded quartz-rich schist is exposed. It is pink and resembles quartzite, and has interbedded siliceous biotite schist and calc-silicate rocks. Similar schist occurs in the SW¼ of sec. 27 and in the SW¼ of sec. 34, same township. These rocks are cut by thin dikes of leucogranite. All the observed metasedimentary rocks have assemblages characteristic of amphibolite-facies metamorphism (Nielsen, 1984; 1985). The presence of stromatolitic marble suggests that the metasedimentary succession was deposited in shallow water.

The metasedimentary rocks on the west margin of the dome resemble in many respects those of the Chocolay Group of the Marquette Range Supergroup (Early Proterozoic), as exposed in the Menominee iron range to the north (Bayley and others, 1966). Probably, the rocks compose a tectonic slice of continental-margin rocks that is interleaved with granitoid rocks of the Dunbar dome and volcanic rocks of the Quinnesec Formation.

QUINNESEC FORMATION

GENERAL FEATURES

The principal supracrustal rocks in the area are assigned to the Quinnesec Formation, a predominantly volcanic succession that has thin interbeds of metasedimentary rocks—biotite schist, quartz-magnetite-hornblende iron-formation, and impure quartzite—and thick mafic layered sills (Bayley and others, 1966, p. 17). The Quinnesec Formation was redescribed by James (1958). Earlier the rocks were called "Quinnesec schists" (Van Hise and Bayley, 1900) and "Quinnesec greenstone" (Leith and others, 1935). Bayley and others (1966) considered the metagabbro sills as being younger intrusive rocks, but we consider them as coeval with the volcanic rocks of the Quinnesec Formation. Schulz (1984b) has subdivided the Quinnesec on the east side of the Dunbar dome into four separate stratigraphic units; these units are being delineated on map compilations of this area. Bayley and others (1966) have given a complete description of the Quinnesec rocks from the northeastern part of the map area (fig. 2), and Dutton (1971) has described those from the northwestern part.

GEOCHEMISTRY

The Quinnesec Formation consists of basalt, andesite, dacite, and rhyolite flows and pyroclastic rocks (Schulz, 1984b) that have been altered regionally to greenschist-facies mineral assemblages. An older succession consists of tholeiitic basalt and basaltic andesite. These rocks show limited FeO enrichment, low TiO_2, and are relatively depleted in HFS (high field strength) elements (Sims and others, 1989). Rare earth element patterns show moderate to extreme light-REE depletion (fig. 24*A*). Gabbro sills within the basalt have major and trace element characteristics similar to the basalt, and probably are consanguineous with it. Plagiorhyolite is locally interlayered with the basalt; it has low K_2O and REE abundances and a flat REE pattern (fig. 24*A*). The association of plagiorhyolite with serpentinite and sheeted dikes within the tholeiitic basalt (Schulz, 1987) suggests that these rocks are a dismembered ophiolite. The basalts are compositionally similar to some basalts of the Troodos ophiolite (Kay and Senechal, 1976) and the Oman ophiolite (Pearce and others, 1981), and suggest formation in a supra-subduction zone environment.

A calc-alkaline suite, ranging in composition from andesite to rhyolite, locally overlies the tholeiitic basalt succession. Major and trace element characteristics of the suite are similar to normal-K_2O calc-alkaline andesite suites. They have moderately enriched light REE with relatively flat heavy-REE patterns, and increasingly negative Eu anomalies with increasing SiO_2 content (fig. 24*B*). The calc-alkaline volcanic rocks plot in fields for calc-alkalic subduction-related magma suites (Wood, 1980) and are geochemically similar to calc-alkaline volcanic suites from oceanic island-arcs such as the New Hebrides.

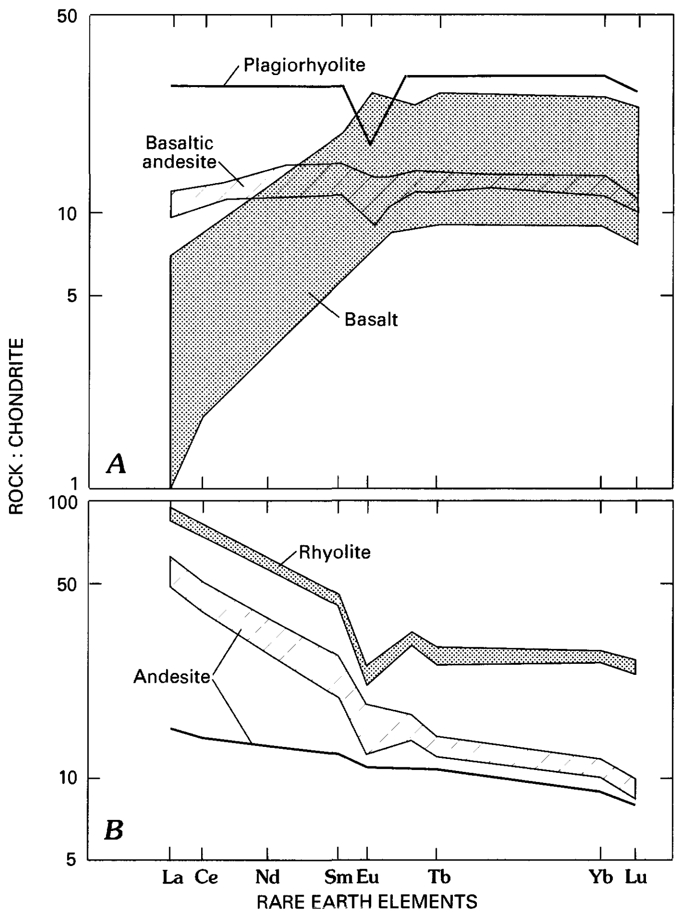


FIGURE 24.—Chondrite-normalized rare earth element plots for *A*, basalt and plagiorhyolite, and *B*, calc-alkaline andesite and rhyolite from the Quinnesec Formation. Chondrite-normalization values from Haskin and others (1968).

NEWINGHAM TONALITE

GENERAL FEATURES

The Newingham Tonalite is a homogeneous intrusive body that composes the Pembine lobe and extends into the core of the Dunbar dome. Within the core, the tonalite has been largely converted to granodiorite or granite by post-crystallization addition of potassium feldspar. This hybrid facies of the Newingham, informally called herein the megacrystic facies of the Newingham Tonalite, is discussed separately on page 41.

The Newingham Tonalite forms a large body about 75 km^2 in area that intrudes the volcanic rocks of the Quinnesec Formation (fig. 2) with sharp contacts. A contact is well exposed in NE¼ sec. 5, T. 36 N., R. 20 E., north of U.S. Highway 8. The contact zone is at least 100 m wide, and consists of an interlayering of tonalite tongues and amphibolite (metabasalt) that takes place within a few meters to a few tens of meters, and of small anastomosing tonalite dikes that transect the layering within the metabasalt. The larger tonalite bodies are essentially conformable with the southeast-dipping foliation in the metabasalt. Other good exposures of the contact zone are in secs. 3 and 4, T. 36 N., R. 19 E. In this area, the Newingham contains moderately abundant, generally angular inclusions of the metavolcanic rocks that retain the orientation of the foliation and lineation that was developed during D_2 deformation. (See section, "Structure.") Although the Newingham Tonalite was deformed together with the Quinnesec during D_2 deformation, in part it has a distinct, younger, northeast-trending mylonitic foliation. This foliation crosscuts the intrusive contacts.

The Newingham is well exposed in scattered knobs and low hills and in road cuts along U.S. Highway 8, west of Pembine (Sims and Schulz, 1988). The principal foliation in the western part of the intrusion is a mylonitic foliation expressed by oriented biotite plates and quartz leaves or ribbons. This is superposed on, and partly destroys, a weak older foliation (S_2) that trends N. 45°–55° E., dips steeply southeastward, and postdates porphyritic tonalite dikes that intrude the main intrusive phase. A weak east- or northwest-trending foliation is present in the eastern part of the body. An exception is in secs. 30 and 31, T. 37 N.,R. 20 E., near Belgian Lake, where a pronounced northwest-trending mylonitic foliation largely obscures the older foliation. The age of the mylonitic foliations is discussed in a later section.

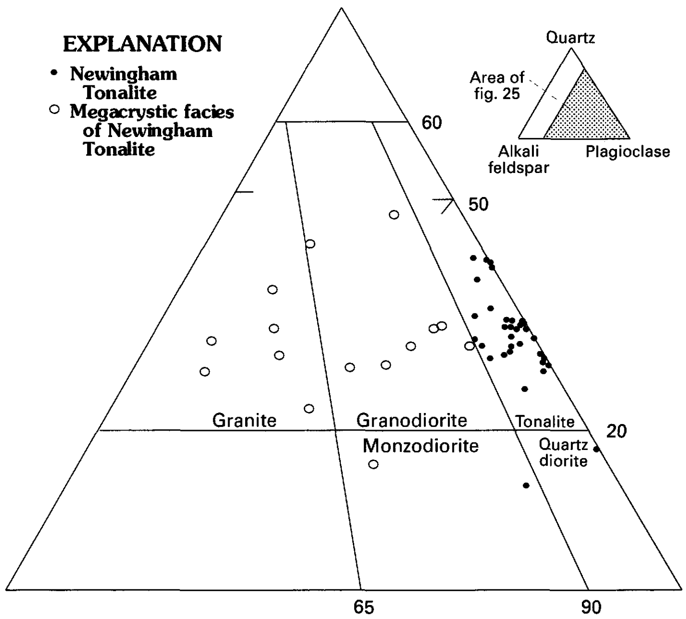


FIGURE 25.—Quartz–alkali feldspar–plagioclase diagram showing composition of representative rocks in Newingham Tonalite and megacrystic facies of Newingham Tonalite.

DESCRIPTION

The Newingham Tonalite is a uniform light-gray, medium-grained, slightly porphyritic rock that generally has a good secondary foliation expressed mainly by oriented biotite, but locally in the eastern portion lacks any penetrative structure. It is dominantly tonalite but is granodiorite at places in the eastern part of the body (fig. 25). Dikes of slightly porphyritic tonalite and, locally, granite pegmatite cut the tonalite. Throughout, it is weakly to moderately altered by retrogressive metamorphism.

The tonalite has a hypidiomorphic granular texture, but at many places this fabric has been modified by recrystallization of new biotite and quartz and, locally, of polycrystalline aggregates of new plagioclase, forming a core and mantle structure or type IP structure. The primary texture dominantly consists of plagioclase that occurs mainly as glomero-porphyritic aggregates of equant or lathlike normally zoned crystals with interstitial quartz.

The rock contains an average of about 55 percent plagioclase, 29 percent quartz, 2.5 percent microcline, 10 percent biotite, and a fraction of a percent of hornblende and sphene (table 7). Epidote-clinozoisite, muscovite and, locally, chlorite are minor minerals. The accessory minerals are apatite, allanite, and zircon.

Plagioclase mainly occurs as subhedral, normally zoned crystals of calcic oligocase–sodic andesine composition, but rarely it has oscillatory zoning. Typically, it is twinned according to the combined carlsbad-albite twin laws. Generally, the zoning was not destroyed by deformation and recrystallization. New plagioclase grain aggregates occur at places adjacent to the primary grain boundaries and within zones of high intergranular strain such as isolated shears. Quartz is clear or has strong undulatory extinction, and mainly has been recrystallized as grain aggregates as a result of high strain. Some aggregates are elongate and are oriented parallel to the foliation expressed by biotite. In least deformed rocks, the quartz clearly is interstitial to aggregates of plagioclase grains. Potassium feldspar (microcline) is mainly fresh and interstitial to plagioclase and quartz; it occurs at places as patches in plagioclase. Grid twinning is poorly developed; some grains are microperthite. Typically, the microcline embays and corrodes plagioclase, and adjacent plagioclase has myrmekitic intergrowths of quartz. Biotite occurs as grayish-red, typically ragged grains that always have associated sphene. The sphene commonly occurs as beads at biotite grain boundaries. Biotite is generally oriented, and in part forms sheaths around plagioclase grains, where it clearly has recrystallized. The ubiquitous presence of sphene with biotite suggests that it possibly recrystallized in part from alteration of original hornblende or an older generation of biotite. Epidote-clinozoisite is a common alteration mineral of biotite, being most abundant at biotite grain

TABLE 7.—*Approximate modes and chemical analyses of representative*

[Leaders (---), not determined; Tr, trace; <, less than. Major oxide analyses (W122A, W134, W132, W152A, W167A, W126B, W129B, W170A) by J.S. Wahlberg, A.J. Bartel, J. Taggart, and elements by X-ray fluorescence analysis by J. Storey, S. Donahey, B. Vaughn, and M. Coughlin; samples W122A, W134, and W757 through W723–4 analyzed by Z.E. Peterman. analyses (W132, W128, and SPB 357–66) by K. J. Schulz; (W152A, W167A) by J.S. Mee; (SDNE 39–82) by L.J. Schwarz]

Sample No.	W122A	W127	W134	W175	W497A	W497B	W131A	W131B	W132	W123	W729	W152A	W573	W167A	W126B	W129B
Modal analyses																
Plagioclase	57.3	51	54.2	62.3	53	53.2	63.3	56.3	45	46	55.5	57	57	57	60	59.5
Quartz	27.4	36.7	31.5	26.2	38.7	36	25	29.5	33	34	30	28	30.5	25.5	30	28.5
Potassium feldspar	1.3	0	2.6	0	0	2.4	0.6	5.8	0.1	1	0.5	2	1.5	3.5	0	Tr
Biotite	10.5	11.1	8.8	8.8	8.3	8.4	10	6.2	14	7.5	10.5	11	9	11	9	11
Muscovite	Tr	Tr	Tr	Tr	Tr	Tr	Tr	1.4	1	1	Tr	Tr	Tr	Tr	Tr	Tr
Epidote-clinozoisite	3.5	1.2	2.9	2.3	Tr	Tr	0.6	0.8	5	5.5	3	2	1.5	Tr	1	0.5
Hornblende	0	0	0	0	0	0	0.3	0	0	0	0	0	0	3	0	0
Other minerals	Tr	Tr	Tr	Tr	Tr	Tr	0.2	Tr	1.9	5	0.5	Tr	0.5	Tr	Tr	0.5
Major oxides (weight percent)																
SiO_2	67.1	---	67.8	---	---	---	---	---	68.0	---	---	69.1	---	66.8	68.8	67.3
Al_2O_3	16.6	---	16.4	---	---	---	---	---	15.7	---	---	16.2	---	16.9	16.4	16.9
Fe_2O_3	0.80	---	0.52	---	---	---	---	---	0.51	---	---	0.57	---	0.66	0.60	0.55
FeO	1.62	---	1.98	---	---	---	---	---	2.64	---	---	1.49	---	2.04	1.84	2.0
MgO	1.29	---	1.23	---	---	---	---	---	1.38	---	---	0.97	---	1.48	1.07	1.30
CaO	3.68	---	3.58	---	---	---	---	---	3.53	---	---	3.34	---	3.53	4.06	4.01
Na_2O	3.86	---	3.93	---	---	---	---	---	3.88	---	---	4.20	---	3.99	3.47	4.24
K_2O	1.93	---	2.25	---	---	---	---	---	2.15	---	---	1.97	---	2.37	1.58	1.54
H_2O^+	1.10	---	0.79	---	---	---	---	---	LOI=	---	---	0.72	---	0.69	1.07	0.53
H_2O^-	0.01	---	<0.01	---	---	---	---	---	0.51	---	---	0.03	---	0.05	0.04	0.07
TiO_2	0.27	---	0.34	---	---	---	---	---	0.37	---	---	0.24	---	0.33	0.22	0.32
P_2O_5	0.10	---	0.10	---	---	---	---	---	0.13	---	---	0.09	---	0.09	0.08	0.10
MnO	0.04	---	0.03	---	---	---	---	---	0.04	---	---	0.03	---	0.03	0.03	0.03
CO_2	0.12	---	0.08	---	---	---	---	---	---	---	---	0.03	---	<0.0	0.22	0.02
Minor elements (parts per million)																
Rb	29	---	44	---	35	---	---	---	66.9	39.3	51	48	---	52	---	---
Sr	732	---	806	---	513	---	---	---	656	744.2	584	607	---	---	---	---
Y	5	---	4	---	4.5	---	---	---	7	---	8	14	---	---	---	---
Zr	120	---	114	---	85	---	---	---	124	---	97	128	---	---	---	---
Nb	17	---	6	---	5	---	---	---	14	---	14	17	---	---	---	---
Minor elements (parts per million) by																
U	2.07	---	2.24	---	---	---	---	---	1.8	---	---	1.78	---	0.8	1.55	1.52
Th	9.36	---	7.67	---	---	---	---	---	6.0	---	---	9.81	---	6.41	5.01	6.19
Ta	---	---	---	---	---	---	---	---	1.75	---	---	0.71	---	1.12	---	---
Hf	---	---	---	---	---	---	---	---	3.4	---	---	2.94	---	3.44	---	---
Cr	---	---	---	---	---	---	---	---	42	---	---	9	---	16	---	---
Co	---	---	---	---	---	---	---	---	6.97	---	---	4.28	---	7.11	---	---
Sc	---	---	---	---	---	---	---	---	4.5	---	---	3.85	---	5.23	---	---
Zn	---	---	---	---	---	---	---	---	---	---	---	47	---	55	---	---
Ba	---	---	---	---	---	---	---	---	961	---	---	450	---	797	---	---
Cs	---	---	---	---	---	---	---	---	6.1	---	---	1.13	---	1.44	---	---
La	---	---	---	---	---	---	---	---	27.9	---	---	33.5	---	39.6	---	---
Ce	---	---	---	---	---	---	---	---	52	---	---	51	---	64.8	---	---
Nd	---	---	---	---	---	---	---	---	17	---	---	17	---	22	---	---
Sm	---	---	---	---	---	---	---	---	2.67	---	---	2.16	---	3.11	---	---
Eu	---	---	---	---	---	---	---	---	0.77	---	---	0.61	---	0.84	---	---
Gd	---	---	---	---	---	---	---	---	---	---	---	1.9	---	2.6	---	---
Tb	---	---	---	---	---	---	---	---	0.26	---	---	---	---	---	---	---
Yb	---	---	---	---	---	---	---	---	0.42	---	---	0.44	---	0.44	---	---
Lu	---	---	---	---	---	---	---	---	0.06	---	---	0.06	---	0.06	---	---

[1]Sample descriptions and localities appear on p. 40.

samples of Newingham Tonalite and associated dikes[1]

J. Baker, (W128, SPB 357–66) by Z.A. Hamlin, (SDNE 39–82) by Z.A. Brown and F.W. Brown. FeO, H_2O, and CO_2 by H. Neiman, J. Ryder, and R. Somers. Minor Sample W498A analyzed by R. Yeoman. Rb and Sr analyses for samples W132, W123, W170, and W739 by isotope dilution method by K. Futa. Neutron activation

W128	SPB 357–66	SDNE 39–82	W170A	W170	W730	W817	W757	W739	W765B	W680A	W498A	W474	W466	W742S	W717	W725B	W723–4
(volume percent)																	
---	---	---	58	50	58	54	57	58.5	58	54	57	58.5	67	59	53.5	62.5	55
---	---	---	27	26	29.5	32	30	27	29	28	23.5	28	18	30.5	32	23	27
---	---	---	5.5	6	1	4	2	2.5	Tr	0	0	0.5	0.5	1.5	4	3.5	5.5
---	---	---	7.5	9.5	9	9	9	10.5	11	14	18	13	13	9	10	10	12
---	---	---	Tr	3	Tr	Tr	Tr	Tr	0.5	0	Tr	Tr	0	0	Tr	Tr	Tr
---	---	---	Tr	4	1	0.5	1	1.5	0.5	Tr	1	Tr	Tr	0	Tr	Tr	Tr
---	---	---	0	1	0	0	0	0	0	0	0	0	0.5	0	0	0	---
---	---	---	Tr	0.5	1.5	0.5	1	Tr	1	4	0.5	Tr	1	Tr	0.5	1.0	0.5
by X-ray fluorescence analysis																	
66.6	71.7	70.3	67.8	---	---	---	---	---	---	---	---	---	63.9	---	---	---	---
18.0	16.0	16.1	16.4	---	---	---	---	---	---	---	---	---	19.7	---	---	---	---
1.1	0.65	0.85	0.66	---	---	---	---	---	---	---	---	---	0.19	---	---	---	---
2.4	1.4	1.3	1.88	---	---	---	---	---	---	---	---	---	2.27	---	---	---	---
1.6	0.96	0.93	1.24	---	---	---	---	---	---	---	---	---	1.01	---	---	---	---
4.5	2.6	3.6	3.50	---	---	---	---	---	---	---	---	---	4.82	---	---	---	---
4.1	4.8	4.0	4.02	---	---	---	---	---	---	---	---	---	4.89	---	---	---	---
1.7	1.4	1.6	2.16	---	---	---	---	---	---	---	---	---	1.50	---	---	---	---
0.82	0.87	1.1	0.91	---	---	---	---	---	---	---	---	---	0.52	---	---	---	---
0.10	0.11	0.14	0.04	---	---	---	---	---	---	---	---	---	0.02	---	---	---	---
0.37	0.23	0.23	0.32	---	---	---	---	---	---	---	---	---	0.32	---	---	---	---
0.18	0.14	0.09	0.11	---	---	---	---	---	---	---	---	---	0.06	---	---	---	---
0.03	0.02	0.04	0.03	---	---	---	---	---	---	---	---	---	0.03	---	---	---	---
0.05	0.47	0.02	0.01	---	---	---	---	---	---	---	---	---	0.02	---	---	---	---
by X-ray fluorescence analysis																	
54	38	42	43	42.2	---	---	29	75.3	64	26	75	60	57	32	41	45	63
724	604	630	620	662.8	---	---	743	634.5	623	884	604	772	928	658	338	702	606
9	10	10	7	---	---	---	5	5	3	5	6	4	10	2	2	6	7
100	130	103	114	---	---	---	97	107	96	121	148	135	159	56	92	171	112
8	7	12	10	---	---	---	12	13	10	11	14	8	13	8	7	19	23
instrumental neutron activation analysis																	
0.96	2.2	1.3	1.22	---	---	---	---	---	---	---	---	---	1.07	---	---	---	---
2.6	5.91	4.5	7.44	---	---	---	---	---	---	---	---	---	16.4	---	---	---	---
0.44	0.57	0.93	---	---	---	---	---	---	---	---	---	---					
2.0	4.0	2.4	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---
14	11	5.5	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---
8.77	4.71	4.4	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---
6.74	4.0	3.72	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---
---	54	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---
570	640	620	---	---	---	---	---	---	---	---	704	---	---	---	---	---	---
0.81	0.93	1.2	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---
12.5	22.9	20	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---
22.8	40.0	31	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---
7.2	13	11	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---
1.27	2.24	1.8	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---
0.44	0.53	0.49	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---
---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---
0.11	0.16	0.14	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---
0.18	0.29	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---
0.03	---	0.05	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---

SAMPLE DESCRIPTIONS AND LOCALITIES

W122A	Cut on U.S. Highway 8 in NW¼ sec. 5, T. 36 N., R. 20 E. Rock has two conspicuous foliations expressed by biotite laths.
W127	SW¼SW¼ sec. 35, T. 37 N., R. 19 E.
W134	South-center sec. 22, T. 37 N., R. 19 E. Plagioclase has microcline patches and quartz blebs.
W175	Rock knob in SW¼ sec. 20, T. 37 N., R. 20 E.
W497A	Low outcrop east of Walton road in NW¼ sec. 24, T. 37 N., R. 19 E.
W497B	Same locality as W497A. Dike that cuts rock sample W497A.
W131A	NE¼SE¼ sec. 34, T. 37 N., R. 19 E. Medium-grained tonalite; contains cataclastic zones.
W131B	Same locality as W131A. Fine-grained tonalite.
W132	Rock cut on Threemile road in SW¼SW¼ sec. 26, T. 37 N., R. 19 E.
W123	Rock knob on north side of U.S. Highway 8 in NE¼NE¼ sec. 6, T. 37 N., R. 20 E.
W729	Knob, 0.16 km north-northwest of W730.
W152A	Rock knob on south side of Marinette County Highway 00 in NW¼SW¼ sec. 33, T. 37 N., R. 20 E.
W573	Rock knob on east side of Chicago, Milwaukee, St. Paul, and Pacific Railway in SW¼ sec. 22, T. 37 N., R. 20 E.
W167A	SE¼SE¼ sec. 14, T. 38 N., R. 20 E. Pinkish-gray, weakly foliated tonalite.
W126B	Rock cut on U.S. Highway 8 in NW¼NW¼ sec. 2, T. 36 N., R. 19 E. Fine-grained, possible dike.
W129B	NE¼NW¼ sec. 3, T. 36 N., R. 19 E. Fine-grained tonalite dike that cuts tonalite.
W128	SW¼SE¼ sec. 34, T. 37 N., R. 19 E.
SPB 357–66	Railway cut, NW¼NE¼ sec. 27, T. 37 N., R. 20 E. Porphyritic tonalite.
SDNE 39–82	Road cut, U.S. Highway 8, NW¼NW¼ sec. 2, T. 36 N., R. 19 E. Porphyritic tonalite dike.
W170A	NE¼ sec. 26, T. 37 N., R. 20 E.
W170	Same as W170A.
W730	NW¼NE¼ sec. 19, T. 37 N., R. 20 E.
W817	Rock knob on southwest side of lake in SW¼SE¼ sec. 30, T. 37 N., R. 20 E.
W757	SW¼SE¼ sec. 16, T. 37 N., R. 20 E.
W739	SE¼SW¼ sec. 13, T. 37 N., R. 19 E. Intrudes Quinnesec Formation.
W765B	NE ¼NW¼ sec. 23, T. 37 N., R. 19 E. Intrudes Quinnesec Formation.
W680A	SE¼NE¼ sec. 16, T. 37 N., R. 20 E.
W498A	SE¼SW¼SW¼ sec. 13, T. 37 N., R. 19 E. Intrudes Quinnesec Formation.
W474	SW¼SE¼ sec. 34, T. 37 N., R. 18 E. Apparent isolated body.
W466	SE¼SE¼ sec. 9, T. 37 N., R. 19 E. Granoblastic texture.
W742S	SW¼NW¼ sec. 11, T. 37 N., R. 19 E.
W717	SE¼NW¼ sec. 7, T. 37 N., R. 20 E. Contains amphibolite inclusions.
W725B	SW¼NW¼ sec. 6, T. 37 N., R. 20 E.
W723–4	NW¼SW¼ sec. 12, T. 37 N., R. 19 E.

boundaries. Hornblende is a rare primary mineral in the body. Where enclosed by quartz, it is fresh, but in other associations it is partly altered to biotite and epidote-clinozoisite.

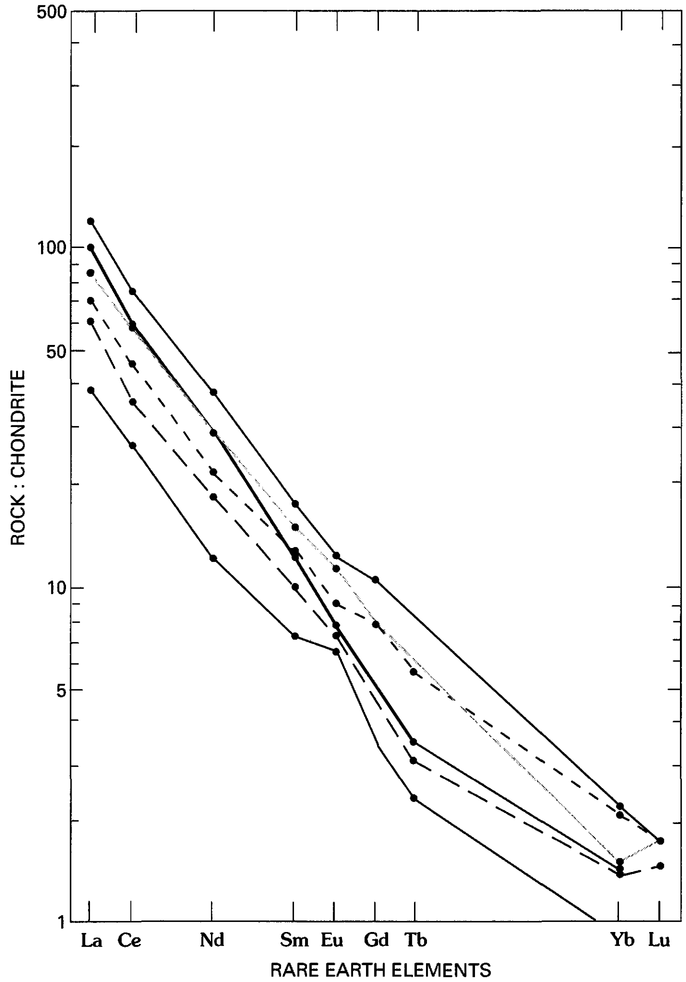


FIGURE 26.—Chondrite-normalized rare earth element plots for samples of Newingham Tonalite. Chondrite-normalization values from Haskin and others (1968).

Mild retrogressive metamorphism is ubiquitous. The cores of plagioclase grains are weakly to moderately altered to epidote-clinozoisite, muscovite, rare calcite, and possibly, zeolite minerals. Biotite is partly altered to epidote-clinozoisite, zeolite(?), and locally to green chlorite; at places it alters directly to muscovite plus quartz. The zeolite(?) occurs as small lenses along biotite cleavages. As noted above, the ubiquitous association of sphene with biotite is interpreted also to indicate retrogressive metamorphism. Allanite nearly always has epidote-clinozoisite rims.

GEOCHEMISTRY

The Newingham Tonalite is calcic in character (fig. 5) and relatively homogeneous (table 7). Two samples from a slightly porphyritic phase of the tonalite have higher SiO_2 contents and are trondhjemitic. All samples have steep, near-linear REE patterns with slight positive or no Eu anomaly (fig. 26).

Samples show a positive correlation between increasing Rb:Sr, increasing total REE abundance, and decreasing magnitude of the Eu anomaly that suggests fractionation of feldspar. Relative to the Dunbar Gneiss, the Newingham Tonalite has higher Al_2O_3, CaO, Na_2O, and Sr contents, lower Rb:Sr ratio, lower REE contents, and much lower abundances of Ta, Nb, Zr, Hf, U, Th, Rb, and K_2O.

PETROGENESIS

The Newingham Tonalite has the compositional characteristics of the high-Al_2O_3 type trondhjemitic-tonalitic suite as defined by Arth (1979) and Barker and Arth (1976). These characteristics include high Al_2O_3 (>15 percent) and Sr content, low Rb content, and moderately enriched light REE but depleted heavy-REE patterns with either no or positive Eu anomalies.

Barker and Arth (1976) have attributed petrogenesis of high-Al trondhjemitic-tonalitic magmas to fractional crystallization of wet basalt parent melts or to partial melting of a basaltic parent leaving a residue with little or no plagioclase but residual amphibole and (or) garnet. Gromet and Silver (1987) have also suggested melting of eclogite for the petrogenesis of tonalites from the Peninsular Range batholith of southern California and the Baja Peninsula. For the Newingham Tonalite, the partial melting model is preferred because of the relatively uniform composition of the tonalite and the absence of more mafic lithologies with suitable compositions to represent more primitive melts. The near-linear chondrite-normalized REE patterns with strongly depleted heavy REE and relatively enriched light REE (fig. 26) suggest that the source for the Newingham Tonalite magma contained garnet as a residual phase (for example, Hanson, 1981) and was not light REE depleted. However, it should be noted that the basaltic source required for the Newingham Tonalite is compositionally distinct from that required for the Dunbar Gneiss. Specifically, the Newingham Tonalite lacks strong "within-plate" source characteristics and must have been derived from a relatively HFS-element-depleted basaltic source. The calc-alkaline basalts of the Wisconsin magmatic terranes, with flat to slightly enriched light-REE and depleted HFS element compositions, could have been the source for the melt that formed the Newingham Tonalite.

MEGACRYSTIC FACIES OF NEWINGHAM TONALITE

GENERAL FEATURES AND DESCRIPTION

The megacrystic facies of the Newingham is a hybrid rock of granodiorite or granite composition (fig. 25) that has resulted from partial replacement of the primary tonalite by post-crystallization potassium feldspar. The facies is delineated separately from the unaltered Newingham Tonalite (fig. 2), but it contains relicts of virtually unaltered tonalite, as for example samples W717, W725B, and W723–4 (table 7). In addition, the megacrystic facies rocks contain inclusions of amphibolite and biotite schist, derived from the Quinnesec Formation, and tongues of Marinette Quartz Diorite, which presumably cut the megacrystic facies. The contact between the Newingham Tonalite and the megacrystic facies is gradational (fig. 2), as is to be expected from the nature of the replacement mineralogy in the megacrystic facies.

The megacrystic facies of the Newingham is a heterogeneous, medium- to dark-gray, generally medium grained igneous rock that commonly contains porphyroclasts of plagioclase as much as 1 cm in diameter, moderate amounts of biotite, and variable, but appreciable amounts of potassium feldspar. It has a pervasive, wavy, mylonitic foliation expressed mainly by oriented biotite but also by elongate and flattened quartz aggregates. It also forms the leucosome of migmatite in contact zones with biotite gneisses. A light-gray or pink, medium- to coarse-grained, equigranular leucogranite is associated with the unit locally.

The rock ranges in composition from tonalite to granite (fig. 25). It has a xenomorphic granular texture modified by later partial recrystallization to finer grained mineral aggregates. Except for the plagioclase phenocrysts(?), probably all the major minerals were recrystallized during the deformation that imposed the northeast foliation. The plagioclase phenocrysts have been partly modified by recrystallization and embayment and replacement by potassium feldspar.

Plagioclase is the dominant mineral, composing between 30 and 50 percent of the rock. It has a bimodal texture consisting of (1) relict, mainly anhedral phenocrysts(?) as much as 1 cm in diameter and having weak normal zoning, and (2) smaller (generally less than 0.5 mm in diameter) recrystallized anhedral plagioclase crystals that generally are unzoned. The larger grains typically have combined carlsbad-albite twins and, less commonly, pericline twins. The overall texture expressed by the plagioclase is a core-mantle structure or type IP structure. The plagioclase has a composition of calcic oligoclase, generally An_{25-30}. Patches of microcline are common in the plagioclase, and where in contact with potassium feldspar it has a fringe of myrmekite. The plagioclase typically is weakly altered to epidote-clinozoisite, sericite, and possibly, zeolite minerals. Quartz generally composes 20–30 percent of the rock; it is highly strained and generally recrystallized to aggregates of sutured grains. Biotite is

TABLE 8.—*Approximate modes and minor element analyses of selected samples of megacrystic facies of Newingham Tonalite*

[Leaders (---), not determined; Tr, trace. Minor element analyses of rock slabs by Z.E. Peterman]

Sample No.	W504A	W504	W726S	W677	W715	W719	W743S	W566	W724	W499B	W760A	W465A
Modal analyses (volume percent)												
Plagioclase	43	33	30.5	50	50.5	37.5	50	52.5	48	49	46	34.5
Quartz	21.5	40	25.5	30	30	28	28.5	25	31	26	26.5	30.5
Potassium feldspar	26.5	17	35	9.5	8	27.5	12.5	13	10.5	16	21	25.5
Biotite	9	7	7.5	10.5	10	7	9	9	10	9	6.5	9
Hornblende	0	0	0	0	0	0	0	0	0	0	0	0
Sphene	Tr	0.5	0.5	Tr	1	Tr	Tr	0.3	Tr	Tr	0	Tr
Apatite	Tr	Tr	Tr	Tr	Tr	Tr	Tr	Tr	Tr	Tr	Tr	Tr
Opaque oxides	Tr	0	0	Tr	Tr	0	0	0	Tr	0	0	0
Allanite	Tr	Tr	Tr	Tr	Tr	Tr	Tr	Tr	Tr	Tr	Tr	0
Accessory minerals	Tr	2.5	1.0	Tr	0.5	Tr	Tr	0.2	0.5	Tr	Tr	0.5
Minor elements (parts per million) by X-ray fluorescence analysis												
Rb	95	95	71	58	56	103	85	---	85	83	68	---
Sr	262	289	478	498	496	344	343	---	395	434	648	---
Y	2	9	7	8	5	5	9	---	5	7	2	---
Zr	81	101	104	146	104	71	145	---	88	138	93	---
Nb	13	21	11	21	13	16	22	---	20	25	8	---

SAMPLE DESCRIPTIONS AND LOCALITIES

W504A SE¼SE¼ sec. 2, T. 37 N., R. 19 E. Foliated granitic rock that contains mafic inclusions.

W504 Same locality as W504A. Foliated granitic rock.

W726S NE¼NE¼ sec. 13, T. 37 N., R. 19 E. Contains mylonitic shears.

W677 NW¼NE¼ sec. 10, T. 37 N., R. 19 E. Protomylonite.

W715 SW¼NE¼ sec. 7, T. 37 N., R. 20 E. Protomylonite.

W719 SE¼NW¼ sec. 7, T. 37 N., R. 20 E. Protomylonite.

W743S NE¼SE¼ sec. 10, T. 37 N., R. 19 E. Protomylonite.

W566 NW¼NW¼ sec. 22, T. 37 N., R. 19 E.

W724 SW¼SW¼ sec. 6, T. 37 N., R. 20 E.

W499B SE¼SE¼ sec. 12, T. 37 N., R. 19 E. Contains inclusions of amphibolite and biotite schist.

W760A SE¼SE¼ sec. 23, T. 37 N., R. 19 E.

W465A NW¼SW¼ sec. 20, T. 37 N., R. 19 E. Granoblastic texture.

moderate brown to reddish brown and fresh or weakly altered to epidote-clinozoisite, chlorite, and zeolite(?) minerals. The minerals believed to be zeolite occur as rare barrel-shaped, fibrous masses along cleavage planes in the biotite. Typically, the biotite has two intersecting orientations in a single thin section, one of which is expressed by biotite oriented along plagioclase grain margins or along plagioclase-quartz interfaces. It is associated with minor amounts of sphene, opaque oxides, apatite, allanite, and epidote-clinozoisite; some of the sphene occurs as small beads along biotite grain margins. Potassium feldspar ranges in amount from about 5 to 35 percent, and is mainly grid-twinned microcline and microcline microperthite. It is irregularly distributed, occurring mainly as interstitial grains intergrown with recrystallized plagioclase, quartz, and biotite but also as poikiloblasts as much as 5 mm in diameter containing inclusions of plagioclase, myrmekite, quartz, and biotite. Some inclusions of plagioclase have albitic rims. The larger grains embay and replace plagioclase. Accessory minerals are sphene, opaque oxides, apatite, allanite, and zircon. Zircon is a moderately abundant accessory, and it commonly forms acicular, zoned crystals.

GEOCHEMISTRY AND PETROGENESIS

The megacrystic facies differs chemically from the unaltered Newingham Tonalite in having higher SiO_2, K_2O, and Nb contents and Rb:Sr ratios, and lower Al_2O_3, FeO_T, MgO, CaO, and Na_2O (compare table 7 with tables 8 and 9), reflecting its more granitic modal composition (fig. 25).

The potassium metasomatism indicated by replacement textures of potassium feldspar in the megacrystic rocks apparently followed a late, post-crystallization ductile deformation, which produced type IP structures. These mylonitic textures could have provided relatively permeable channelways for circulation of potassic solutions. The restriction of the potassium metasomatism to the northern part of the Newingham Tonalite body suggests a northerly source for the circulating solutions, perhaps the Hoskin Lake Granite or its equivalent, which extensively hybridized the adjacent Marinette Quartz Diorite, as discussed earlier. That a source like the Hoskin Lake Granite probably caused the potassium metasomatism is suggested by a plot of Nb versus Rb for the Newingham Tonalite, the Hoskin Lake Granite and related dikes,

TABLE 9.—*Approximate modes and chemical analyses of selected samples of megacrystic facies of Newingham Tonalite*

[Leaders (---), not determined; Tr, trace; <, less than. Major oxide analyses by J.S. Wahlberg, A. Bartel, J. Taggart, and J. Baker; FeO, H_2O, and CO_2 analyses by H. Neiman, J. Ryder, and G. Mason; minor elements by X-ray fluorescence analysis by J. Storey, S. Donahey, B. Vaughn, and M. Coughlin]

Sample No.	W503A	W465–1B
Modal analyses (volume percent)		
Plagioclase	30	38
Quartz	30	36
Potassium feldspar	34	19.5
Biotite	5.5	6.5
Sphene	0.5	Tr
Opaque oxides	Tr	0
Accessory minerals	Tr	Tr
Major oxides (weight percent) by X-ray fluorescence analysis		
SiO_2	72.9	71.5
Al_2O_3	14.1	14.6
Fe_2O_3	---	0.34
FeO	1.42	1.41
MgO	0.62	0.81
CaO	1.55	1.47
Na_2O	3.16	3.52
K_2O	4.25	3.73
H_2O^+	0.34	0.71
H_2O^-	0.03	0.06
TiO_2	0.22	0.15
P_2O_5	<0.05	<0.05
MnO	0.02	0.04
CO_2	<0.01	<0.01
Minor elements (parts per million) by X-ray fluorescence analysis		
U	1.03	1.9
Th	6.98	3.8
Rb	99	113
Sr	350	312
Y	6	7
Zr	60	47
Nb	9	20

and the megacrystic facies of the Newingham (fig. 27). With respect to the Newingham Tonalite, the megacrystic facies is enriched in both Nb and Rb, and these values fall between the fields for the Newingham and the Hoskin Lake Granite and the dikes, strongly suggesting their introduction from the granite or a similar magmatic source. In the same way, a plot of Rb versus Sr for the Newingham and the megacrystic facies (fig. 28) shows a higher Rb:Sr ratio for the megacrystic facies that is intermediate to ratios in the Newingham and the Hoskin Lake Granite and granitic dikes. These data suggest that both Rb and Nb were introduced together with potassium from the Hoskin Lake Granite or a comparable magmatic source. This interpretation is consistent with the previous conclusion that potassic metasomatism from the Hoskin Lake Granite or a similar buried granite also modified the adjacent Marinette Quartz Diorite body.

TWELVE FOOT FALLS QUARTZ DIORITE

The Twelve Foot Falls Quartz Diorite, named by Cain (1964), is a poorly exposed body on the south side of the Dunbar dome that occupies an area of about 80 km^2. It intrudes and locally contains inclusions of metavolcanic rocks of the Quinnesec Formation. The quartz diorite is massive in the eastern part of the main body, but becomes foliated toward the west. The foliation strikes west-northwest and dips steeply. At the type locality at Twelvefoot Falls on the North Branch of the Pike River (NW¼ sec. 32, T. 36 N., R. 19 E.), the quartz diorite has been intensely sheared, has a pronounced foliation, and is mainly a mylonitic gneiss. This zone of mylonitic gneiss extends northward at least to Eighteenfoot Falls, a distance of 0.8 km. Foliation within the gneiss in the shear zone strikes N. 70° W. and dips steeply to either north or south; a steep (stretching) lineation occurs in the plane of the foliation. The gneiss within the zone is strongly altered to retrograde assemblages; hornblende is partly to completely altered to epidote and chlorite, and plagioclase is saussuritized. Quartz is extremely comminuted. The mylonitic rocks lack any primary mafic mineral and are fine-grained aggregates of plagioclase, quartz, epidote, sericite, and chlorite. Garnet and opaque oxides are local accessory minerals. The shear zone partly coincides with a northwest-trending positive magnetic anomaly, apparently because of a high content of magnetite developed in the quartz diorite in the shear zone.

Outside the shear zone, the quartz diorite is medium to coarse grained and is characterized by subhedral plagioclase (sodic andesine) crystals as much as 1 cm long, smaller subhedral hornblende crystals, in part pseudomorphic after pyroxene, and anhedral crystals of blue quartz as much as 1 cm in diameter. Microcline locally occurs as a late interstitial mineral. The primary texture is hypidiomorphic granular, but finer grained secondary textures are superposed on it at many places. Characteristically, the rock is considerably retrograded: plagioclase is partly to largely altered to epidote and albite, and hornblende is partly altered to biotite, epidote, and chlorite. Other alteration minerals are sphene, opaque oxides, and calcite. In some sections, quartz is a late interstitial mineral that has sharp crystal faces against plagioclase and hornblende, suggestive of a granophyric texture.

Its chemical composition, especially its low Ta and Nb concentrations (table 10) and REE pattern (fig. 29), suggests that the Twelve Foot Falls Quartz Diorite is a subvolcanic intrusion comagmatic with calc-alkaline rocks of the Quinnesec Formation.

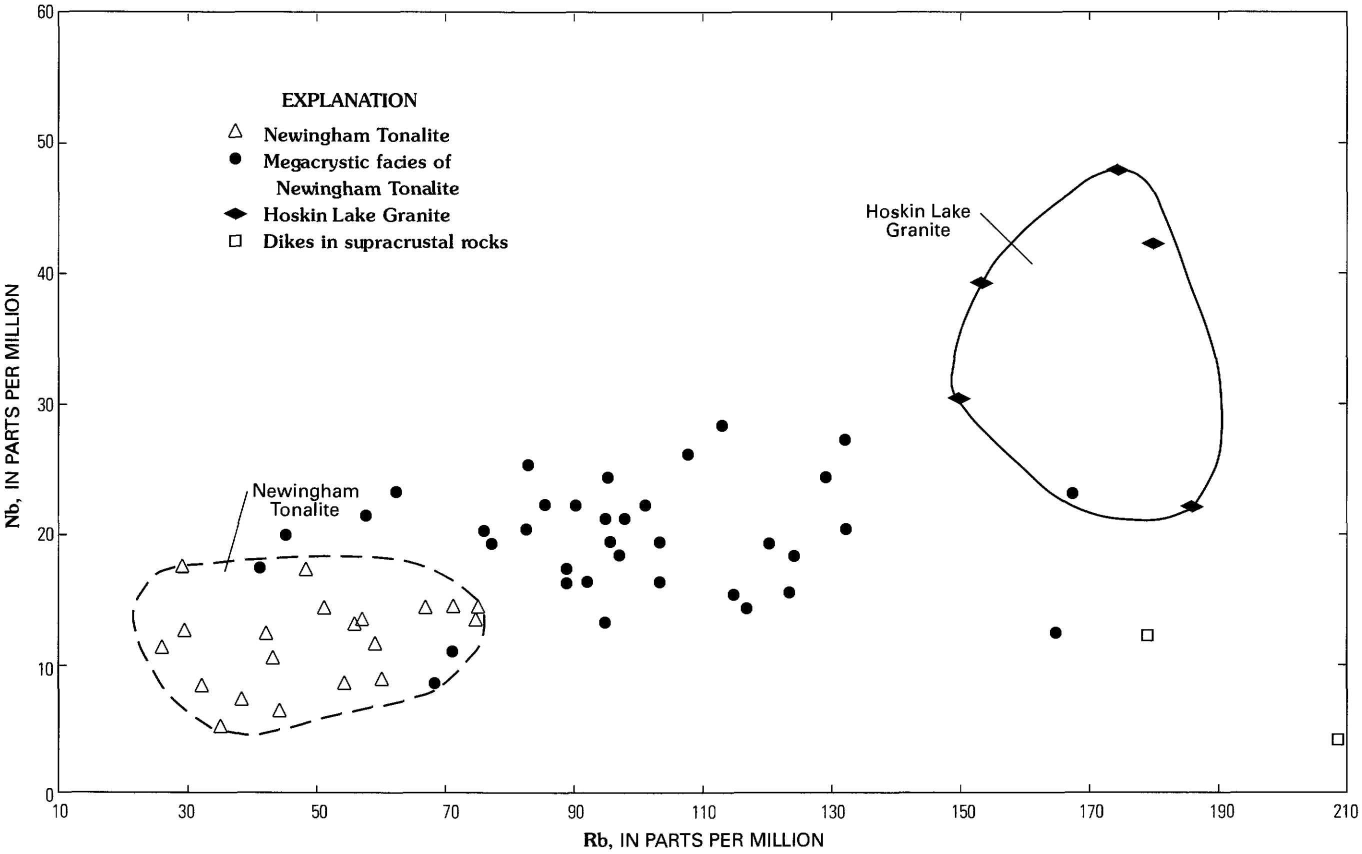


FIGURE 27.—Plot of Nb versus Rb contents for Newingham Tonalite, megacrystic facies of Newingham Tonalite, and Hoskin Lake Granite and related dikes.

GRANITOID DIKES IN SUPRACRUSTAL ROCKS

In addition to the Marinette Quartz Diorite, Newingham Tonalite, and Spikehorn Creek Granite, which are known to intrude the supracrustal rocks in their contact zones, granitic dikes of uncertain affinity are common in the volcanic and sedimentary rocks adjacent to the crystalline core. On the north side of the dome, southwest of Niagara (fig. 2), granitic dikes extend outward from the boundary for a distance of at least 0.3 km, and on the northwest margin of the Bush Lake lobe for a distance of about 0.8 km. Elsewhere, the dike-bearing zone in the supracrustal rocks is narrower. In general, the dikes increase in abundance as the boundary is approached. The dikes are as much as 30 m wide, and generally are discordant. Dutton (1971, pl. 1) has shown that granitic dikes occur sporadically in the Quinnesec Formation and metasedimentary rocks northwest of the northwest margin of the Bush Lake lobe; the dikes are known to be present as much as 5 km from the dome margin, and they probably extend even farther from it.

Pegmatite is rare in the supracrustal rocks, but a body about 85 m wide crosscuts metasedimentary rocks at a small angle adjacent to Macintire Creek (sec. 7, T. 38 N., R. 18 E.) on the west margin of the dome. The pegmatite and associated aplite are megascopically identical to that cutting the Dunbar Gneiss in the vicinity of Dunbar.

The dikes are pinkish-gray, fine- to medium-grained, partly inequigranular, leucocratic rocks that range in composition from granite to tonalite (table 11). They are mostly metamorphosed and deformed, and generally have a xenomorphic granular texture and a superposed mylonitic texture.

The granitic dikes on the north margin of the Dunbar dome are in part inequigranular; they contain poikilitic microcline grains as much as 1 cm long in a finer grained matrix of plagioclase, microcline, biotite, and rarely, green hornblende. Microcline is late paragenetically; it embays and replaces plagioclase and not uncommonly contains inclusions of plagioclase having albite rims and myrmekitic intergrowths of quartz. Where subjected to strong ductile deformation, patch antiperthite and bleb microperthite are developed. Biotite was recrystallized in deformation zones. Accessory minerals are rare; those that occur include sphene, opaque oxides, allanite, apatite, and zircon. A mild retrogressive metamorphism was superposed on

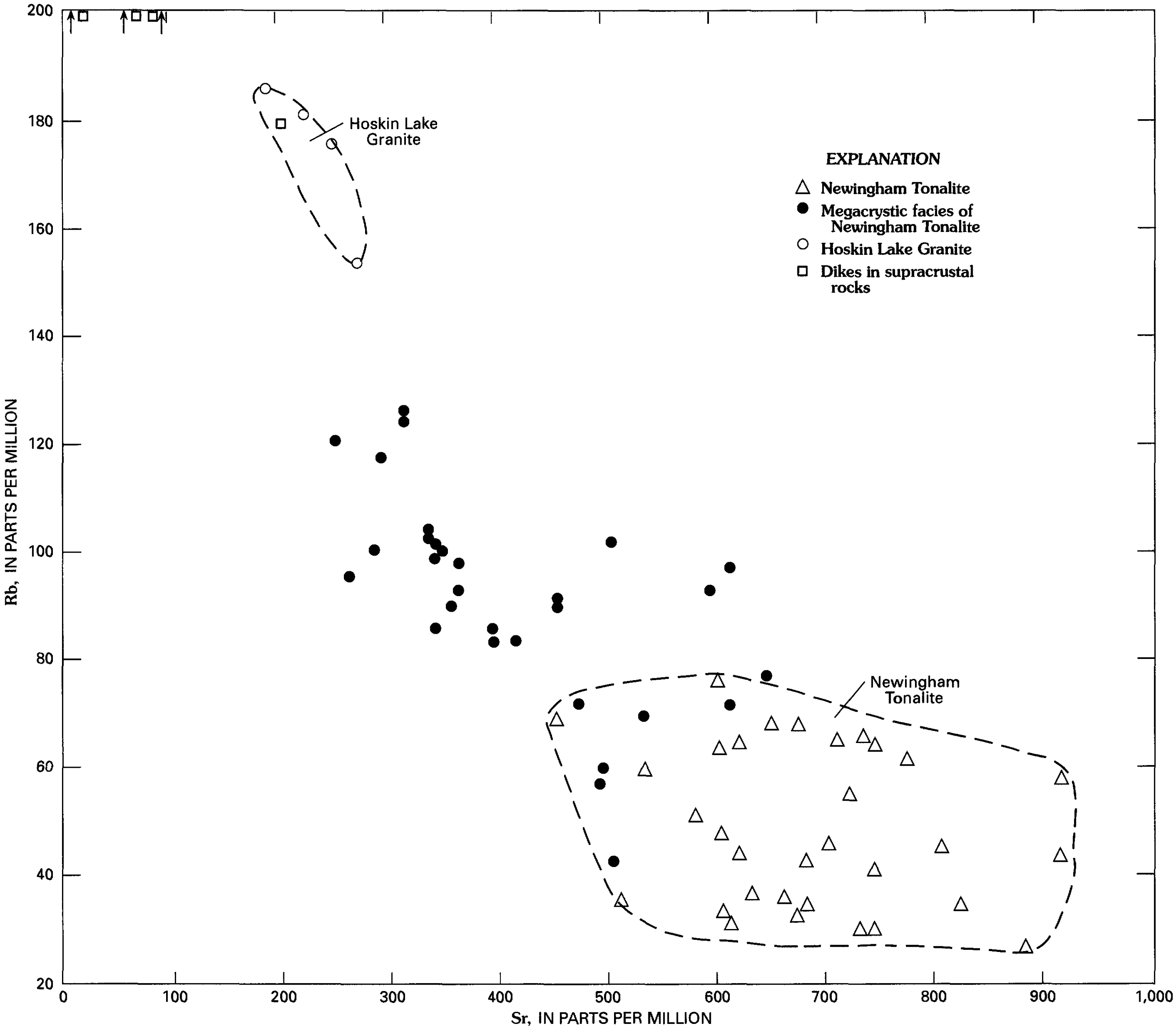


FIGURE 28.—Plot of Rb versus Sr contents for Newingham Tonalite, megacrystic facies of Newingham Tonalite, and Hoskin Lake Granite and related dikes. Note that as in figure 27, values for the megacrystic facies are intermediate between those for the Newingham Tonalite and the Hoskin Lake Granite and related dikes. Arrows associated with dike symbols show that values exceed scale given.

the primary textures. Rare hornblende is partly altered to epidote, and plagioclase is altered to unidentified low-birefringence minerals.

On the southwest margin of the Dunbar dome, garnet is a local component of the tonalitic (aplite) dikes, and muscovite and sericite are common alteration minerals. Albite twinning in plagioclase is wedgelike. Possibly the aplite is that described previously from the Dunbar area.

In the reentrant between the Niagara and Pembine lobes, dikes in the metavolcanic rocks are tonalite in composition and contain plagioclase with oscillatory zoning, indicating a definite magmatic origin. The dikes have been deformed and now are mainly protomylonite. The composition and porphyritic texture of the dikes make them most closely resemble the Newingham Tonalite.

The granitoid dikes occur in the inner margin of the supracrustal rocks, adjacent to the core, in rocks metamorphosed to amphibolite facies. The greater abundance of dikes on the north margin of the dome, and their dominant granite composition, support the

TABLE 10.—*Chemical analysis of Twelve Foot Falls Quartz Diorite*

[Major oxide analysis by rapid rock ICP by Z.A. Brown and F.W. Brown; minor elements by X-ray fluorescence analysis by R. Johnson, K. Dennen, and B. Scott; minor elements by instrumental neutron activation analysis by L.J. Schwarz]

Sample No.	STF-4-1
Major oxides (weight percent) by X-ray fluorescence analysis	
SiO_2	56.9
Al_2O_3	15.5
Fe_2O_3	4.2
FeO	5.1
MgO	4.1
CaO	7.7
Na_2O	2.3
K_2O	0.84
H_2O^+	1.8
H_2O^-	0.15
TiO_2	0.44
P_2O_5	0.16
MnO	0.20
CO_2	0.17
Minor elements (parts per million) by X-ray fluorescence analysis	
Rb	21
Sr	475
Y	15
Zr	57
Nb	6
Minor elements (parts per million) by instrumental neutron activation analysis	
U	0.3
Th	2.0
Ta	0.09
Hf	1.3
Cr	18
Co	22.2
Sc	33.6
Zn	83
Ba	258
Cs	0.8
La	18
Ce	36
Nd	19
Sm	3.8
Eu	0.94
Gd	2.8
Tb	0.37
Yb	1.6
Lu	0.28

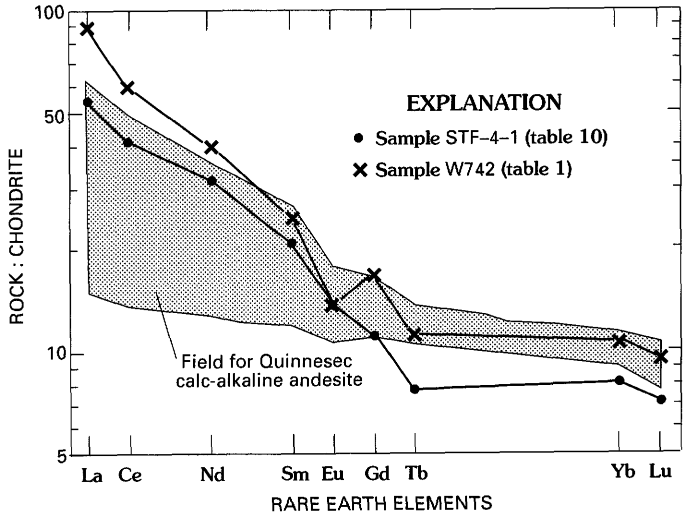


FIGURE 29.—Chondrite-normalized rare earth element plot for a sample of Twelve Foot Falls Quartz Diorite. Also shown is sample W742 from the Dunbar Gneiss and the field for Quinnesec andesite. Note that sample W742 has a rare earth element pattern very similar to that of the Twelve Foot Falls Quartz Diorite. Chondrite-normalization values from Haskin and others (1968).

conclusion based on the width of the amphibolite-facies zone and the intensity of metamorphism here that this margin was the hotter, perhaps deeper, part of the dome. Diopsidic pyroxene developed during culmination of the metamorphism in the andesine amphibolite facies zone, adjacent to the core.

As a group, the dikes are characterized by high Rb:Sr ratios, low Zr and Y contents, and variable Nb contents. Sample W412C is modally and compositionally-like other dikes around the Bush Lake Granite (table 6, sample W679C) and is probably related to them. The remaining dikes listed in table 11, all of which intrude the Quinnesec Formation, are similar both modally and compositionally to the Hoskin Lake Granite and are probably related to it.

DIABASE DIKES

Two diabase dikes are associated spatially with the dome (fig. 2), one of which is partly exposed. The diabase is presumed to be Middle Proterozoic in age (Sims and Schulz, 1988), correlative with 1,100 Ma Keweenawan magmatism in the Midcontinent rift system, but this is uncertain because of the diabase's retrograde alteration. A diabase dike that cuts the Newingham Tonalite has an ophitic texture and is moderately altered to greenschist-facies mineral assemblages. The mafic minerals are now actinolite, chlorite, and opaque oxides. The opaques compose as much as 15 percent of the altered rock, and account for the moderate positive magnetic anomaly over the dike.

The compositions of a chilled-margin and interior sample of the dike cutting the Newingham Tonalite are given in table 12. The dike is a high-iron, quartz-normative tholeiite with a light-REE enriched pattern and is similar compositionally to some Keweenawan dikes in the region (Green and others, 1988).

TABLE 11.—*Approximate modes and minor element analyses of selected samples of granitoid dikes in the supracrustal rocks*

[Leaders (---), not determined; Tr, trace. Minor elements by X-ray fluorescence analysis by Z.E. Peterman]

Sample No.	W477B	W569B	W569–1	W412C	W490A	W660A
Modal analyses (volume percent)						
Plagioclase	60.3	54.5	37	21.5	25	55
Quartz	34	33.5	31	21	32	31
Potassium feldspar	2.5	10.5	30	53.5	40	7
Biotite	0	1.5	2	3	2	7
Muscovite	3	Tr	Tr	Tr	0	Tr
Garnet	0.1	0	0	0	0	0
Hornblende	0	0	0	Tr	1	0
Other minerals	0.1	Tr	Tr	1	Tr	Tr
Minor elements (parts per million)						
Rb	246	27	209	313	179	---
Sr	24	28	67	89	202	---
(Rb/Sr	10.25	0.97	3.12	3.52	0.89	---)
Zr	4	40	31	21	71	---
Y	4	4	5	---	1	---
Nb	36	24	4	---	12	---

SAMPLE DESCRIPTIONS AND LOCALITIES

W477B	NE ¼NW¼ sec. 3, T. 36 N., R. 18 E. Pink aplite dike in Quinnesec Formation.
W569B	NW¼NE¼ sec. 16, T. 38 N., R. 20 E. Leucocratic granodiorite dike in Quinnesec Formation, 200 m north of contact with Hoskin Lake Granite.
W569–1	60 m south of locality W569B. Leucogranite dike in Quinnesec Formation.
W412C	NW¼SW¼ sec. 11, T. 38 N., R. 17 E. Leucogranite dike in marble.
W490A	SW¼SE¼ sec. 25, T. 38 N., R. 18 E. Leucogranite dike in Quinnesec Formation.
W660A	SE¼NE¼ sec. 4, T. 37 N., R. 20 E. Tonalite dike in Quinnesec Formation.

TECTONIC IMPLICATIONS OF GRANITOID ROCK GEOCHEMISTRY

Igneous suites with differing compositional characteristics, such as mid-ocean ridge basalts, calc-alkaline rocks, continental tholeiites, and kimberlites, appear to be related to particular tectonic settings. Trace-element discrimination diagrams have been used as a means of fingerprinting the tectonic setting of several igneous suites, using particularly the basaltic compositions (for example, Pearce and Cann, 1973; Floyd and Winchester, 1975; Shervais, 1982; Thompson and others, 1984), and more recently, felsic (granitic-rhyolitic) compositions (Pearce and others, 1984; Harris and others, 1986; Brown and others, 1984; Thompson and others, 1984; Whalen and others, 1987). Although the use of these diagrams as discriminants of tectonic setting can be inconclusive, the diagrams do provide useful constraints on possible tectonic environments.

As discussed previously (Sims and others, 1985), the lithologic units in the Dunbar area range from calcic (Newingham Tonalite) to calc-alkaline (Dunbar Gneiss) to alkali-calcic (Marinette Quartz Diorite) in terms of the alkali-lime (Peacock) index (fig. 5), and they are generally similar to rocks formed along convergent plate margins. However, further evaluation of the relevant data in light of recent reports on the tectonic characterization of granitoid rocks suggests some modification of earlier conclusions.

Except for the Newingham Tonalite, the granitoid rocks of the Dunbar area are characterized by relatively high K, Rb, Ba, Th, Nb, Ta, and light-REE abundances. Their high Nb and Ta concentrations are particularly distinctive, and contrast markedly with the lower values that characterize most other Penokean igneous rocks in the Wisconsin magmatic terranes (see fig. 33). This distinctive compositional characteristic of the granitoid rocks in the Dunbar area suggests that they were derived from a different source and probably formed in a tectonic setting different from that of the volcanic arc-related rocks in the Wisconsin magmatic terranes.

The tectonic characterization of the Dunbar area rocks is shown in figures 30, 31, and 32. Samples of the Newingham Tonalite plot well within the field of volcanic arc granites in figure 30, but overlap the fields of volcanic arc and syn-collisional granites in figures 31 and 32. The overlap into the syn-collisional field reflects the somewhat high Ta values in some samples

TABLE 12.—*Chemical analyses of Middle Proterozoic diabase dike*

[Major oxide analyses by rapid rock ICP by H. Smith and F.W. Brown; minor elements by X-ray fluorescence analysis by R. Johnson and K. Dennen; minor elements by instrumental neutron activation analysis by J.S. Mee]

Sample No.	SDNE–113–1–83	SDNE–115–83
Major oxides (weight percent) by X-ray fluorescence analysis		
SiO_2	49.5	50.2
Al_2O_3	13.4	13.2
Fe_2O_3	4.8	4.2
FeO	10.5	11.1
MgO	4.7	4.8
CaO	8.1	8.4
Na_2O	2.8	2.8
K_2O	1.2	1.1
H_2O^+	1.4	1.4
H_2O^-	0.35	0.35
TiO_2	2.0	2.0
P_2O_5	0.50	0.49
MnO	0.22	0.23
CO_2	0.15	0.17
Minor elements (parts per million) by X-ray fluorescence analysis		
Rb	45	39
Sr	502	538
Y	34	35
Zr	155	149
Nb	11	10
Minor elements (parts per million) by instrumental neutron activation analysis		
U	<2	1.1
Th	2.88	2.87
Ta	0.46	0.55
Hf	3.78	3.85
Cr	26.3	27.8
Co	43	44.5
Sc	40	40.7
Zn	150	146
Ba	645	617
Cs	2.88	2.28
La	27.1	26.8
Ce	54	49
Nd	33	35
Sm	8.1	6.97
Eu	1.81	1.84
Gd	6.0	5.2
Tb	1.0	1.0
Yb	3.6	3.6
Lu	0.55	0.52

SAMPLE DESCRIPTIONS AND LOCALITIES

SDNE–113–1–83	Chill margin
SDNE–115–83	Diabase

of the tonalite, possibly resulting from metasomatism by Hoskin Lake-type granite. The other granitoid rocks of the Dunbar dome are distinct from the Newingham Tonalite and plot with volcanic arc and syn-collisional granites in figure 30 and with syn-collisional and within-plate granites in figures 31 and 32. Although some ambiguity remains in their classification, the granitoid rocks of the Dunbar area are considered from their overall chemistry to be most similar to syn- to post-collisional granites of some recent orogens (Harris and others, 1986). This classification agrees with the structural relationships of the granitoids as syntectonic to posttectonic.

Harris and others (1986) and Thompson and others (1984) have attributed the compositional characteristics of some syn- to post-collision-related granitoids to derivation from parental mantle-derived basaltic magmas having within-plate compositional characteristics derived from continental lithosphere. These basaltic melts undergo differentiation and possible mixing with crustal-derived melts during their ascent through the thickened crust of the collisional orogen. The contribution of a within-plate component to the Dunbar dome granitoids is emphasized in figure 33 where, with the exception of the Newingham Tonalite, the granitoid rocks plot close to within-plate magma types (see inset). It should be noted that basalts from the Early Proterozoic epicratonic sequence in Michigan and Minnesota also plot with recent within-plate basalts, consistent with their rift-related origin (Schulz, 1984a).

We conclude that, except for the Newingham Tonalite, granitoid rocks of the Dunbar area show compositional affinities with syn- to post-collisional granites of recent orogenic belts. Further, their compositions suggest derivation from and (or) mixing with within-plate mafic magmas rather than derivation from the subduction-related volcanic-arc assemblages that characterize the Wisconsin magmatic terranes. These collision-related magmas were probably generated by melting of continental lithosphere during the Penokean orogeny. Overthrusting of the Wisconsin magmatic terranes onto the margin of the Superior craton and subsequent decompression resulting from uplift (Harris and others, 1986) could have contributed to the generation of the magmas.

GEOCHRONOLOGY

Uranium-lead zircon ages of the rocks in and adjacent to the Dunbar dome (fig. 34) are clustered in the range 1,865 Ma to 1,835 Ma, but the data are not sufficiently precise to establish the timing of specific intrusive and tectonic events within this range. The Dunbar Gneiss has a concordia upper intercept age of 1,862±5 Ma, which is interpreted as the age of the igneous protolith for the layered succession. The Marinette Quartz Diorite, which intrudes the Dunbar Gneiss, has a comparable age, 1,862±15 Ma. A dacite

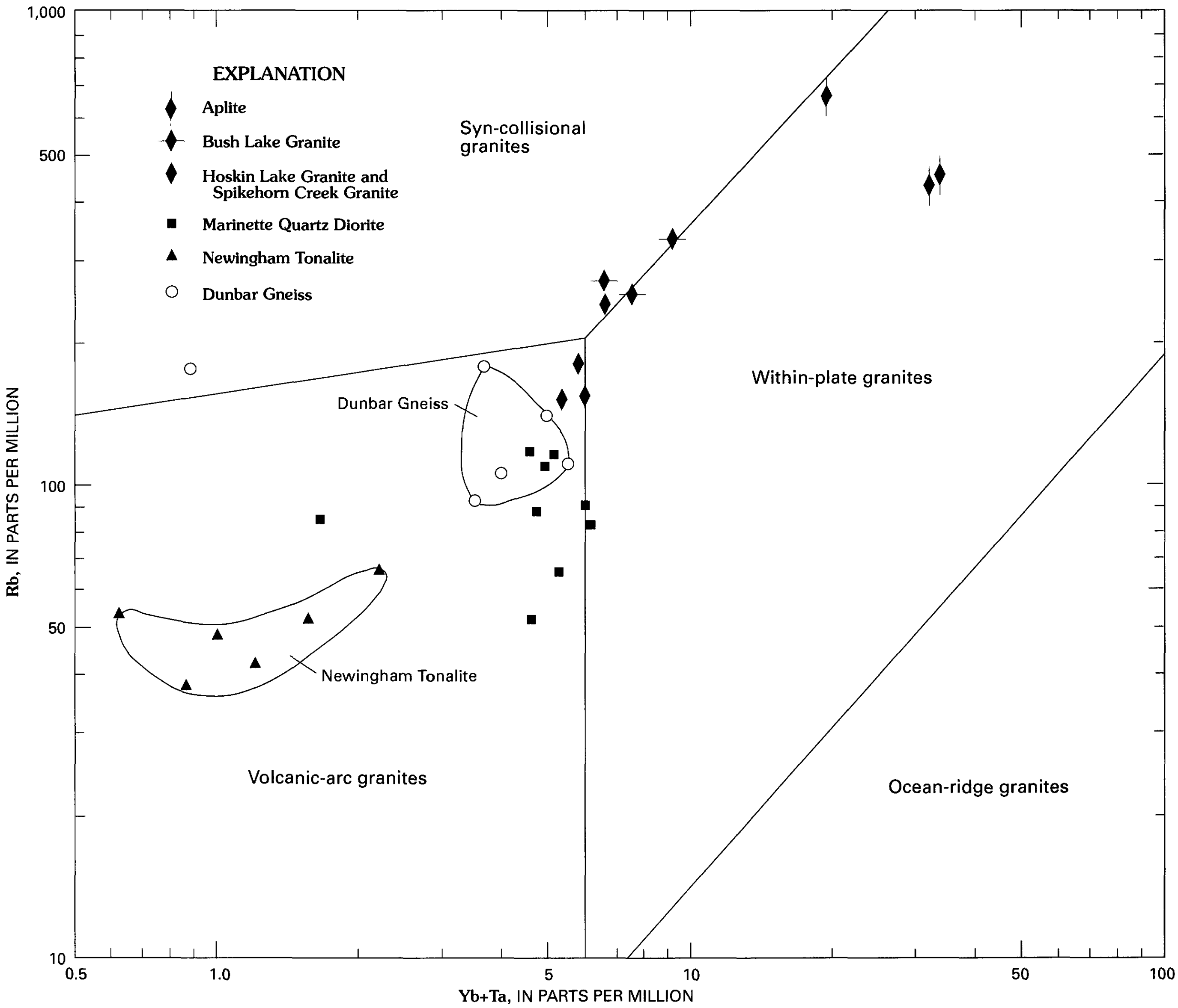


FIGURE 30.—Rb versus (Yb+Ta) diagram (modified from Pearce and others, 1984) for granitoid rocks of the Dunbar area with fields for volcanic-arc granites, syn-collisional granites, within-plate granites, and ocean-ridge granites.

from the Quinnesec Formation has an age of 1,866±39 Ma, and the volcanic rocks probably were deposited in the interval 1,870–1,860 Ma. The Newingham Tonalite, which intrudes the Quinnesec, has an age of 1,861±40 Ma; and the youngest rock unit, the Spikehorn Creek Granite, has an age of 1,836±6 Ma. These ages on the intrusive rocks are generally consistent with the geologic relationships. Rubidium-strontium whole-rock and biotite ages are consistently reset and are 100 m.y. or more younger than the zircon ages (Peterman and others, 1985).

Samarium-neodymium model ages of 2,130 and 2,280 Ma for two samples of Dunbar Gneiss (fig. 35) are substantially older than the crystallization ages defined by the zircon data. Other Early Proterozoic igneous rocks in northern Wisconsin have yielded similar "old" Sm-Nd ages (Nelson and DePaolo, 1985; Barovich and others, 1989). The Sm-Nd ages, together with Pb-isotope data (Afifi and others, 1984), indicate a major involvement of Archean crustal material in the genesis of some Early Proterozoic volcanic and plutonic rocks and syngenetic mineralization.

STRUCTURE

The principal structural feature in the Dunbar area, the Dunbar dome, is an irregular entity about 200 km^2 in area composed of a core of gneiss and foliated

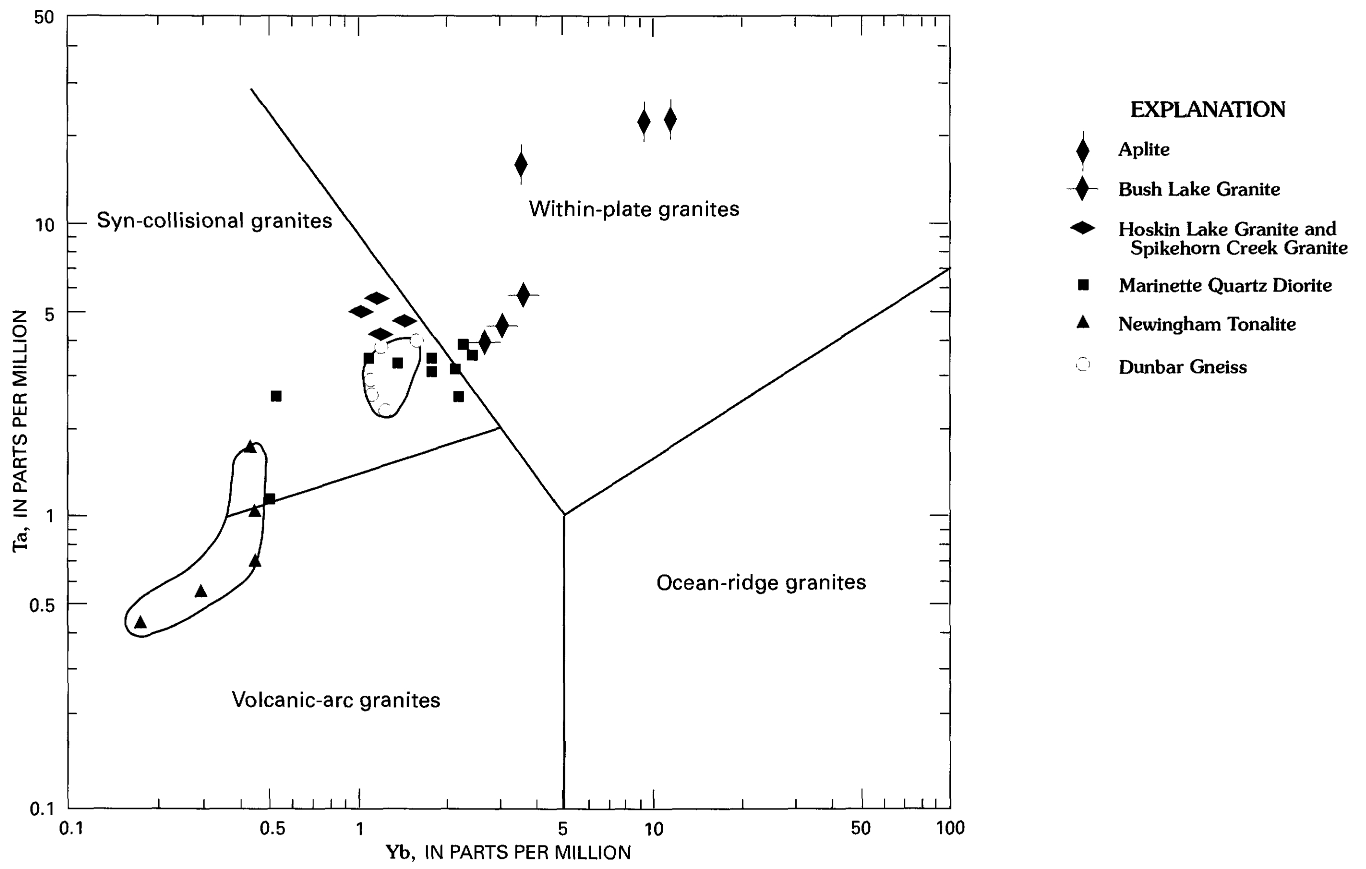


FIGURE 31.—Ta versus Yb diagram for granitoid rocks of the Dunbar area. Fields (which are the same as in fig. 30) from Pearce and others (1984). Note that with exception of the Newingham Tonalite, the granitoid rocks plot in the fields of recent syn-collisional and within-plate granites.

granitoid rocks mantled by steeply dipping metavolcanic rocks. Except for three conspicuous protuberances of the core (fig. 2), named the Bush Lake lobe, the Niagara lobe, and the Pembine lobe (Sims and others, 1985), the margins of the dome tend to be smooth. The outline of the dome is interpreted as resulting from polydeformation and emplacement of the granitoid rocks.

SMALL-SCALE STRUCTURES

Small-scale structures examined in the field indicate four principal deformational events in and adjacent to the Dunbar dome (table 13) and two younger episodes of shearing that produced local mylonitic foliations.

D_1 STRUCTURES

The oldest recognized structure is a pervasive foliation (S_1) that is parallel to compositional layering in the Dunbar Gneiss. It is defined mainly by a preferred orientation of biotite and hornblende and by quartz leaves. A lineation related to S_1 has not been recognized. The lack of an L_1 linear fabric suggests a strong component of flattening strain normal to S_1.

The other major rock type in the core of the dome, the Marinette Quartz Diorite, also has a foliation that is parallel to igneous layering and is expressed by a preferred orientation of biotite and hornblende. It is designated $S_{1'}$.

D_2 STRUCTURES

The major structures in the region are F_2 folds. They deform the Dunbar Gneiss, Quinnesec Formation, Newingham Tonalite, and Marinette Quartz Diorite, as well as an older generation of granite pegmatite and aplite associated with the Dunbar Gneiss. An antiform oriented N. 60° W. and plunging 35°–45° SE. in near-vertical axial surfaces has been delineated in the southwestern part of the Dunbar dome, and second-order folds are common on the limbs. The folds are upright, slightly asymmetrical, open to closed

TABLE 13.—*Structural sequence, Dunbar dome*

D_1	Foliation parallel to layering in Dunbar Gneiss (S_1). Foliation parallel to igneous layering in Marinette Quartz Diorite ($S_{1'}$).
D_2	Northwest-oriented folds in Dunbar Gneiss, Quinnesec Formation, Newingham Tonalite, and Marinette Quartz Diorite (F_2). Foliation (S_2) parallel to axial planes of folds. Local. Lineation (L_2) parallel to fold axes. Local.
D_3	Asymmetrical S- and Z-type folds along northwest and southeast margins of core. Lineation (L_3) parallel to fold axes (F_3).
D_4	Mylonitic foliation S_4 and stretching lineation (L_4) along north margin of dome and in adjacent Quinnesec Formation.

structures. Except locally, the folds do not transpose the older foliation (S_1) and layering (S_0). An axial-plane foliation (S_2) is best developed in relatively massive tonalitic Dunbar Gneiss, where it is defined by oriented tabular feldspars and biotite. A lineation (L_2) that is parallel to fold axes (F_2) is best developed in mica- and hornblende-rich schists of the Marinette Quartz Diorite and is expressed by elongate minerals and mineral aggregates in S_1 planar structures. The existence of a southeast-plunging synform in the covered area to the north is suggested by the orientation of layering in exposed Marinette Quartz Diorite (fig. 36).

The F_2 antiform in the vicinity of Dunbar is defined by outward-dipping S_1 and S_0 planar structures on opposite limbs of the fold (fig. 37). The girdle expressing folding of S_1 and S_0 surfaces has a beta fold axis plunging 63° S. 75° E. The axial orientations of F_2 fold hinges and L_2 lineations, plotted in figure 37, show considerable variation, being steeper on the southwest limb than on the northeast limb, but they generally plunge less steeply than the beta fold axis. The fold hinges that plunge gently N. 90° E. probably represent a slightly older set of upright, tight to isoclinal intrafolial folds in the SW¼ sec. 13, T. 37 N., R. 18 E. Probably, they were formed during an early stage of D_2 deformation.

D_3 STRUCTURES

Following deformation D_2, the rocks adjacent to the southeast and northwest margins of the core were refolded. Along the southeast margin (loc. B, fig. 36), Z-type asymmetrical folds (F_3) were locally superposed on the previously deformed Marinette Quartz Diorite and Quinnesec Formation; hinge lines plunge 30°–50° S. 5° W. and axial surfaces dip southeastward, subparallel to the core-cover boundary. Along the northwest margin (loc. C, fig. 36), a steep, close-spaced foliation that parallels the core-cover boundary and an associated southwest-plunging stretching lineation were

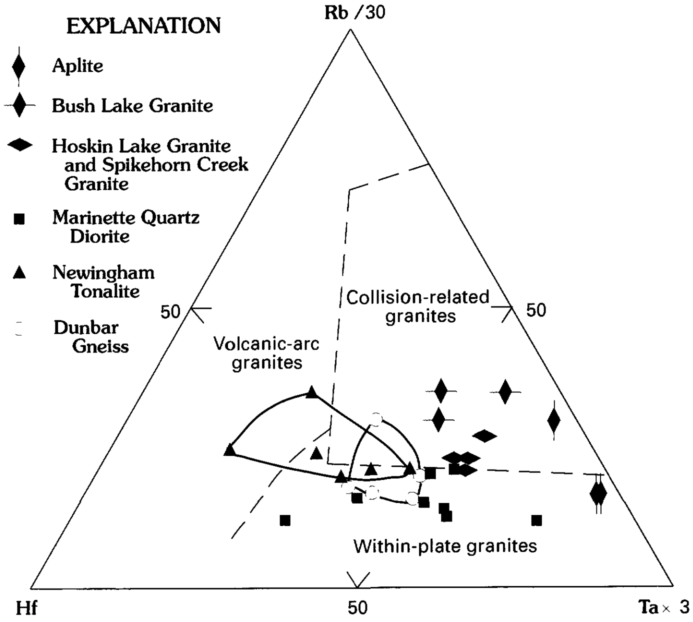


FIGURE 32.—Rb-Hf-Ta diagram (modified from Harris and others, 1986) showing granitoid rocks of the Dunbar area. Except for collision-related granites, fields are the same as in figure 30.

developed in the Dunbar Gneiss, and S-shaped asymmetrical folds that plunge at a moderate angle southwest were formed in the adjacent volcanic rocks of the Quinnesec Formation. Amphibolite-facies metamorphism accompanied or followed D_3 on the northwest margin. Possibly the gentle east-northeast-trending folds in the Marinette Quartz Diorite at locality E (fig. 36) and the south-southwest-plunging folds in it at locality A were formed at this time also, under amphibolite-facies metamorphic conditions.

D_4 STRUCTURES

D_4 structures, which are characterized by a stretching lineation (L_4) defined by elongate clasts, mullions, slickenside striae, and rare folds, and a mylonitic foliation, are predominant in the Quinnesec Formation and in the core-cover boundary on the north side of the dome. Similar structures extend to and beyond the Niagara fault zone (Sedlock and Larue, 1985). On the north margin of the dome (loc. D, fig. 36), the foliation dips 70°–80° S. and the lineation plunges uniformly 60°–70° SW. The zone of mylonitic foliation is as much as 500 m wide; the foliation at this locality increases in intensity toward the boundary, indicating that the deformation here is strongly controlled by the core-cover boundary.

Similar northwest-striking foliations (S_4) and southwest-plunging stretching lineations (L_4) are present in the Quinnesec Formation on the northeast side of the Bush Lake lobe. Lineations measured in this

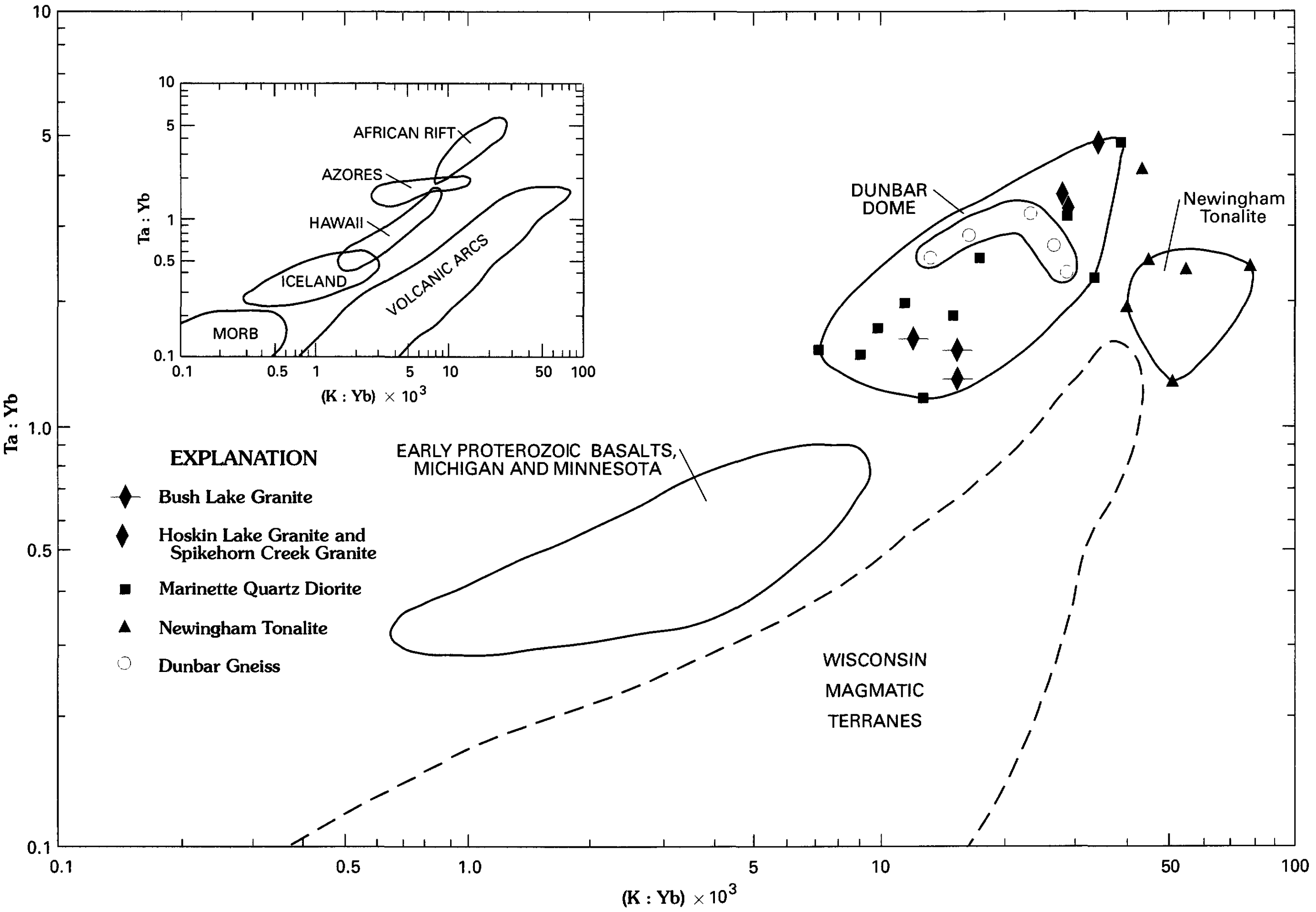


FIGURE 33.—Ta:Yb versus K:Yb diagram (modified from Rogers and others, 1987) showing rocks of the Dunbar area, and fields for Early Proterozoic rift basalts from the epicratonic continental-margin sequence in Upper Michigan and east-central Minnesota (K.J. Schulz, unpub. data, 1989), and rocks of the Wisconsin magmatic terranes (K.J. Schulz, unpub. data, 1989). Shown in inset are fields for recent volcanic and intrusive rocks: data for MORB (Mid-Oceanic Ridge basalt) from leg 33 DSDP (Jackson and others, 1976), Iceland from Wood (1978), Hawaii from Frey and Claque (1984) and Roden and others (1984), Azores from White and others (1979), African rift from DeMulder and others (1986), and volcanic arcs from Leo and others (1980), McBirney and others (1987), Lopez-Escobar and others (1977), Deruelle (1982), Innocenti and others (1981), Barker and others (1986), and Noyes and others (1983).

area mainly plunge 50°–60° S. 40°–70° W. (fig. 38). D_4 structures in this area are most intense near the margin of the Bush Lake lobe, where they obliterate older structures related to D_2 deformation. The structures are the predominant ones in this area and compose a unique belt of highly strained rocks about 10 km wide that straddles the Niagara fault (Sedlock and Larue, 1985).

LARGE-SCALE STRUCTURE

The Dunbar dome is interpreted as a fold-interference structure resulting from superposition of F_2 and F_3 folds modified by diapirism and the emplacement of granitoid intrusive rocks. Many of the criteria indicative of diapirism, as listed by Brun and others (1981), are observed in the dome: (1) cleavage parallel to dome boundaries, (2) steeply plunging lineation in dome boundaries, and (3) higher strain intensities located on dome boundaries.

The outline of the core of the dome (fig. 36) is mainly the result of superposed F_2 and F_3 folding. The southwest margin is the southwest limb of the major northwest-trending F_2 antiform cored by Dunbar Gneiss. Small-scale structures indicate that the antiform plunges southeast. Possibly, the antiform is doubly plunging, to account for the westward closure of the dome, but this was not confirmed because of the absence of exposures in the extreme western part of the dome. The steeply dipping cover rocks along the west margin and fabrics indicative of high strain indicate that diapirism was intense in this boundary zone.

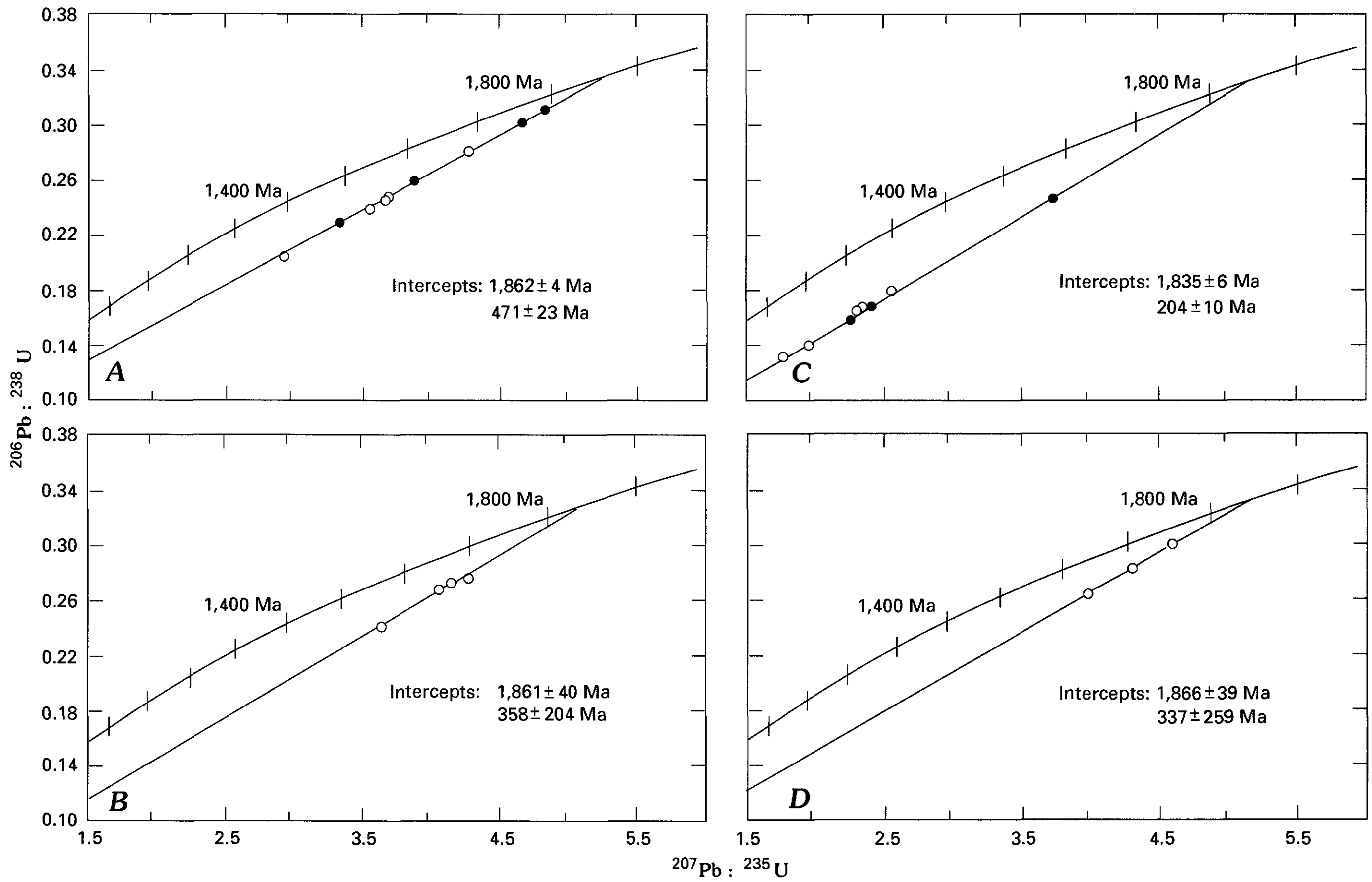


FIGURE 34.—U-Pb concordia plot of zircon data for rocks of the Dunbar area. Filled circles are unpublished U.S. Geological Survey data. Open circles are from Banks and Cain (1969), Banks and Rebello (1969), and Aldrich and others (1965). For *A*, Dunbar Gneiss, regression includes only the USGS data. Array for *B*, Newingham Tonalite, includes one data point for the Marinette Quartz Diorite. Regression for *C*, Spikehorn Creek Granite, is based on the three USGS data points from an undeformed phase of the granite. Inclusion of all the data shown in *C* does not appreciably change upper intercept (1,842±59 Ma). *D*, Quinnesec Formation.

The northwest and southeast margins of the core of the dome are subparallel to small-scale D_3 structures, and are interpreted as the limbs of a major northeast-oriented antiform. The asymmetry of the F_3 flattening folds on these margins suggests that this episode of deformation resulted from a shear couple oriented northeastward, with sinistral shear on the northwest margin and dextral shear on the southeast margin. The folds on both margins are slightly overturned to the northwest.

On the north, overturned margin of the core, the contact between the Hoskin Lake Granite and the Quinnesec Formation was the site of intense D_4 deformation, which almost completely obliterated older structures.

The Niagara lobe, which is composed of nearly undeformed Spikehorn Creek Granite, transects at nearly right angles the outer (east) margin of the central core, composed of Marinette Quartz Diorite, and it is interpreted as a second-order diapir in the first-order one, of the type described by Schwerdtner and others (1979). The granite in the Niagara lobe has a steep foliation near its walls, and the contact is sheared, indicating that in part at least it is a tectonic contact. The volcanic rocks of the Quinnesec Formation are molded around the margin of the dome. We conclude that the lobe of granite flowed diapirically upward and outward in a plastic state during a late stage of dome inflation. As a consequence of the second-order diapir, a "cleavage triple point" (Brun and others, 1981) developed in the supracrustal volcanic rocks at the intersection of the Niagara lobe and the east margin of the core (loc. F, fig. 36).

The geology and isotopic ages of rocks in the Dunbar dome indicate that the dome was developed during regional deformation within the Penokean orogen. The

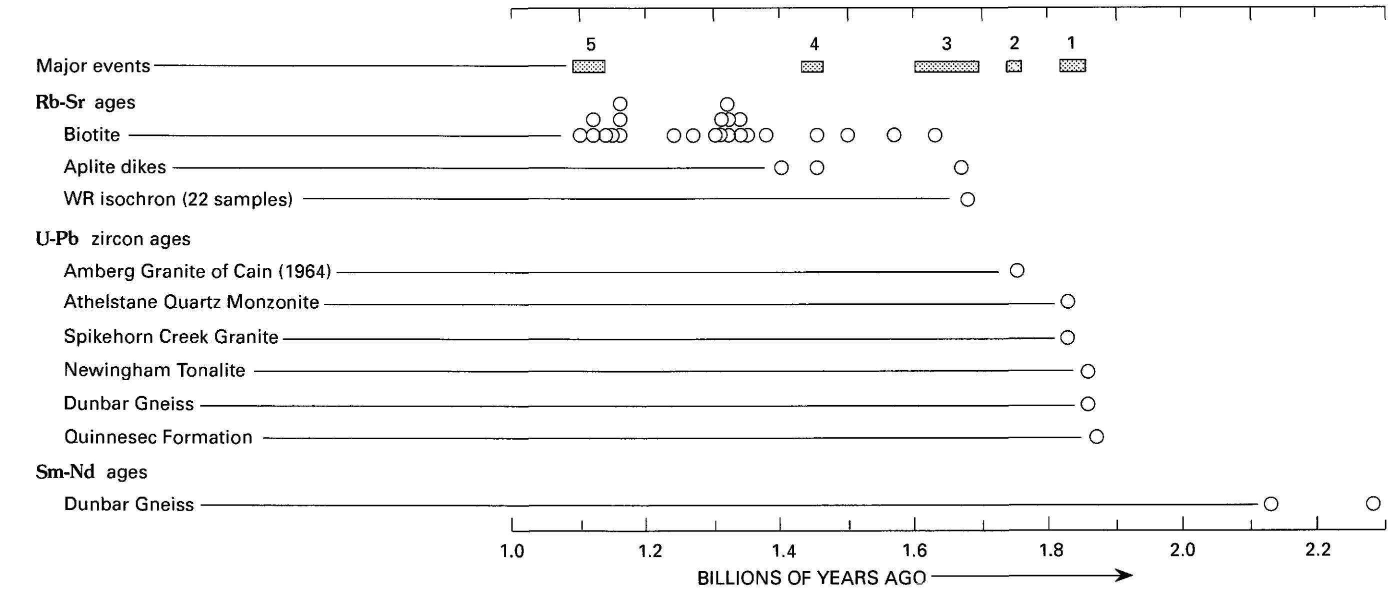


FIGURE 35.—Summary of isotopic ages for rocks of the Dunbar area and environs. Events shown are: 1, main interval of Penokean igneous activity; 2, post-Penokean 1.76 Ga igneous event; 3, 1.60±0.05 Ga event that disturbed isotopic systems throughout much of the Precambrian of Wisconsin; 4, emplacement of the Wolf River batholith at about 1.47 Ga; and 5, Keweenawan igneous activity at about 1.1 Ga. Isotopic ages shown are from Aldrich and others (1965), Banks and Cain (1969), Banks and Rebello (1969), Van Schmus (1980), and Z.E. Peterman (unpub. data, 1989, Rb-Sr, U-Pb, and Sm-Nd ages). Each circle represents an age.

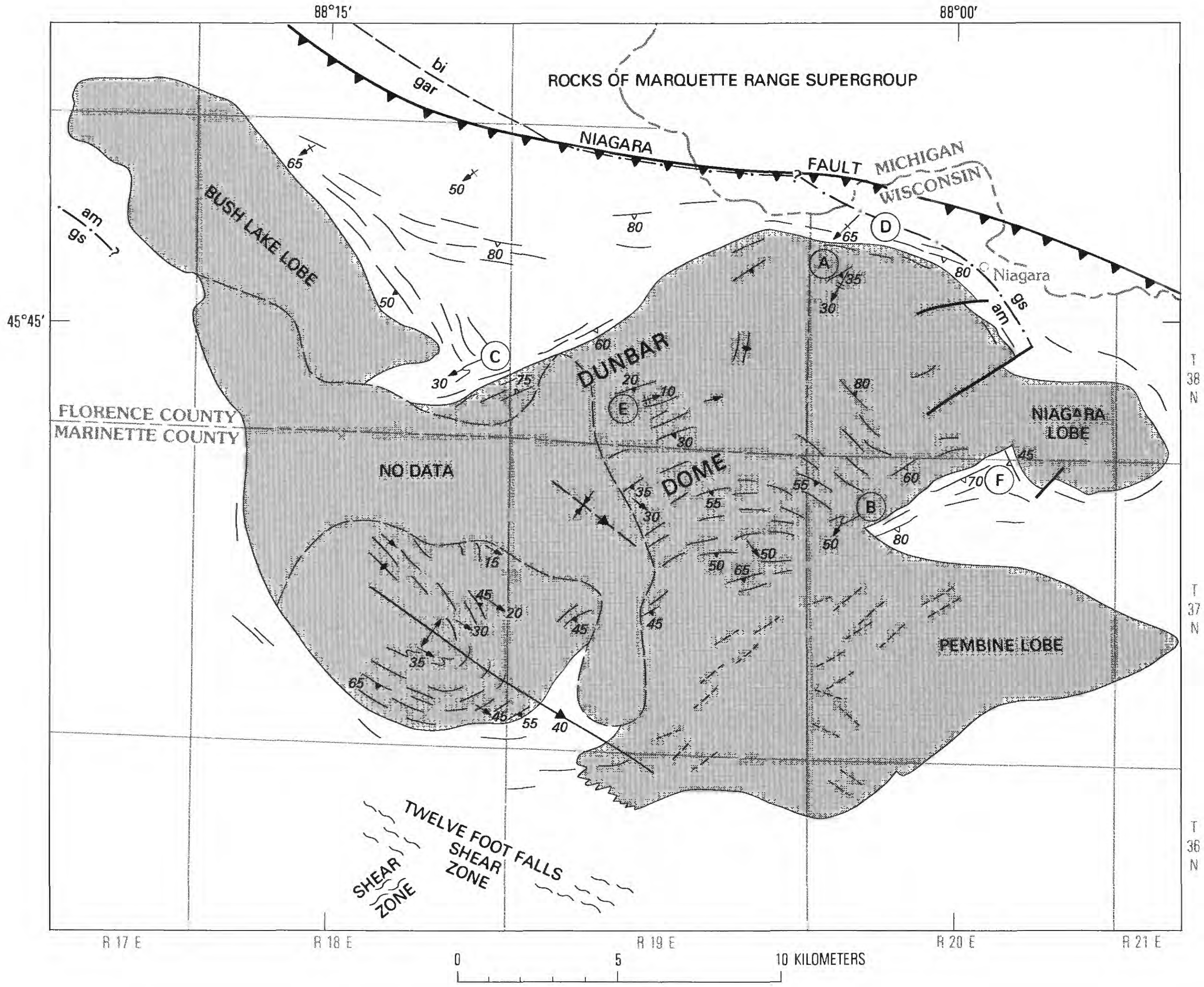
regional (northwest-trending) structural fabric and perhaps also the persistent southwest-plunging stretching lineation in the core-cover boundary of the dome (L_4) could have resulted from subhorizontal compression oriented north-northeastward during collision along the arc-continent margin (Niagara fault zone). The northeast elongation of the core appears to be related to more local forces, perhaps thermal perturbations within the core of the dome, because D_3 structures are virtually confined to the dome.

Sedlock and Larue (1985) concluded from a study of the highly strained rocks along the Niagara fault zone, including those on the north side of the Dunbar dome (D_4 structure of this report), that the zone of ductile shear is as much as 10 km wide and the axes of the finite strain ellipsoid during this deformation were oriented: X, subvertical; Y, subhorizontal, west-northwest–east-southeast; and Z, subhorizontal, north-northeast–south-southwest. Accordingly, the steep, southwest-plunging stretching lineation (L_4) records the X axis of the strain ellipsoid.

Whether the core-mantle boundary of the Dunbar dome is a stratigraphic or a tectonic contact was not determined during the mapping because (1) the contact itself is nowhere exposed and (2) high strain was imposed on the boundary during D_3 and D_4 deformations. Support for the core-cover boundary's being a tectonic surface is provided by the aeromagnetic map of the region (King and others, 1966) and by two-dimensional density modeling of the Dunbar dome by Klasner and Osterfeld (1984). The aeromagnetic map clearly shows that the strong anomaly associated with the iron-formation exposed on the Menominee range extends southward beneath the body of Spikehorn Creek Granite (fig. 2), suggesting that the rocks of the Dunbar area have overridden the continental-margin assemblage (see fig. 1). Klasner and Osterfeld concluded that the granitoid rocks of the Dunbar dome, including the Dunbar Gneiss, extend to a depth of only about 1.6 km. This depth is much shallower than many large granitic plutons (Bott and Smithson, 1967), and accordingly, Klasner and Osterfeld suggested that the Dunbar dome possibly is truncated at depth by a south-dipping, low-angle thrust fault, tectonically related to the Niagara fault. Therefore, the dome itself could be allochthonous with respect to at least a part of the surrounding rocks.

SHEAR ZONES

Two steep zones of shearing that postdate the structural development of the Dunbar dome were



EXPLANATION

- F_2 **antiform**—Showing plunge
- F_2 **synform**—Showing direction of plunge, inferred
- L_2 **lineation**—Showing plunge
- F_3 **asymmetrical fold**—Showing plunge
- F_3 **symmetrical fold**—Showing plunge
- S_3, S_4 **foliation**—Showing dip
- L_4 **stretch lineation**
- **Trend of** S_1, $S_{1'}$ **foliation**
- S_1, $S_{1'}$ **foliation**—Showing dip
- S_1, $S_{1'}$ **foliation, vertical**
- **Post-1,835 Ma mylonitic foliation**
- **Mylonite in shear zone**
- **Metamorphic isograd**—bi, biotite; gar, garnet. Modified from Dutton (1971)
- **Metamorphic isofacies**—Queried where extent uncertain; **gs**, greenschist facies; am, amphibolite facies. Modified from Bayley and others (1966)
- **Thrust fault**—Sawteeth on upthrust block
- **Fault**—Relative movement not known
- (A) **Location referred to in text**

FIGURE 36.—Structure map of the Dunbar dome. Dome (shaded) is surrounded by metavolcanic rocks of the Quinnesec Formation and associated subvolcanic rocks. Annular foliation around Niagara lobe is result of diapirism. Compiled by P.K. Sims, 1988.

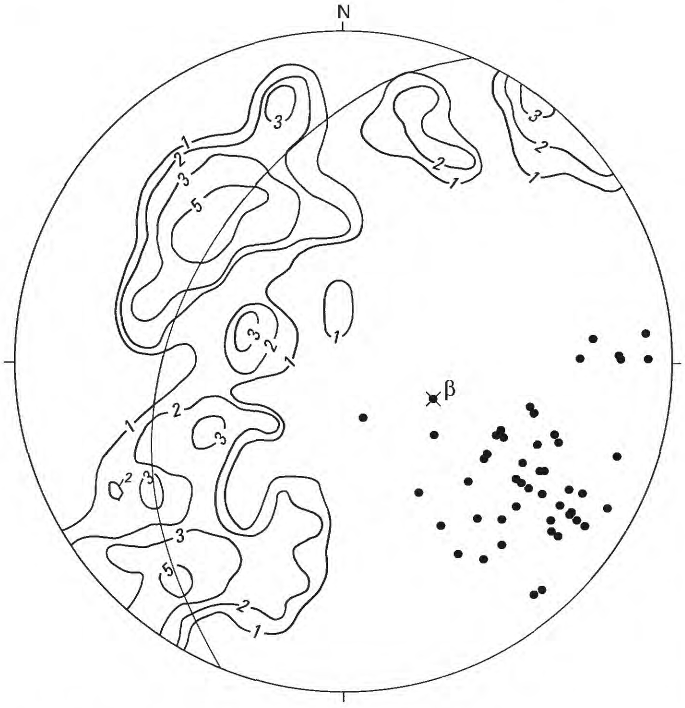


FIGURE 37.—Equal-area projection of structural elements in the Dunbar Gneiss and Marinette Quartz Diorite for southwestern part of the Dunbar dome. Poles to S_1 and $S_{1'}$ foliations are contoured in percent (1, 2, 3, and 5; 146 points). F_2 fold hinges and lineations (48 points) are represented by dots. X, Beta fold axis.

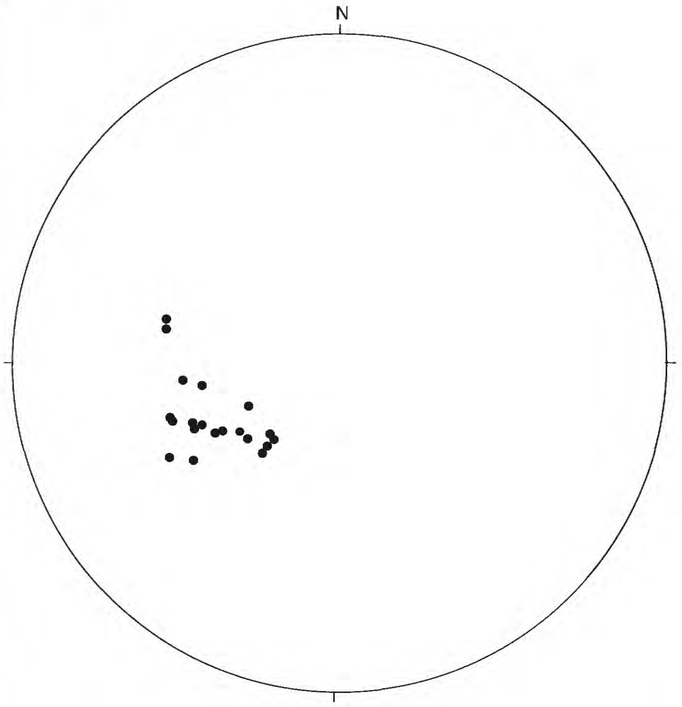


FIGURE 38.—Equal-area projection of L_4 stretching lineation (20 points) in core-cover boundary and in Quinnesec Formation, northeast of Bush Lake lobe.

recognized during our mapping: (1) the northwest-trending Twelvefoot Falls shear zone, which lies on the south side of the dome, and (2) an unnamed northeast-trending zone, also on the south side, that is interpreted to pass through the western part of the Newingham Tonalite pluton (fig. 36).

The Twelvefoot Falls shear zone is a broad, ill-defined ductile deformation zone that strikes about N. 70° W., dips steeply, and so far as known, ranges in width from about 1 km in the vicinity of Twelvefoot Falls on the North Pike River (sec. 21, T. 36 N., R. 19 E.) to about 15 km in the vicinity of Amberg, about 15 km to the southeast. We have traced it discontinuously along strike for a distance of about 35 km. The type area for the deformation is at Twelvefoot Falls. The shear zone is superposed on the volcanic rocks of the Quinnesec Formation (age ≈1,860 Ma) and the Twelve Foot Falls Quartz Diorite and, to the southeast of the map area (fig. 1), on the Athelstane Quartz Monzonite (age ≈1,835 Ma) and the Amberg Quartz Monzonite (age 1,760 Ma). Rocks within the shear zone have a weak to strong secondary foliation and, typically, a steep lineation. In the mafic rocks, the foliation is penetrative. In greenschist-facies rocks, it is expressed by finely comminuted plagioclase and quartz (mylonite) and oriented chlorite and epidote-clinozoisite aggregates; in amphibolite-facies rocks, it is expressed by recrystallization of the original minerals to medium or coarse grain size through penetrative ductile flow (terminology of Wise and others, 1984). In both metamorphic grades, the lineation generally is more conspicuous than the foliation. In the granitic rocks, the foliation is not penetrative; instead, it is a spaced fracture or flow cleavage commonly marked by crudely aligned micas and granulated quartz lenses.

The unnamed, northeast-trending shear zone was first recognized from small outcrops of the Quinnesec Formation in the northern part of sec. 22, T. 36 N., R. 18 E. (Peterman and others, 1985). At this locality it contains a pronounced vertical mineral lineation. This shear projects on a N. 60° E. bearing into the western part of the Newingham Tonalite body (fig. 36), where it is expressed by an ill-defined, broad zone of mylonitic foliation that strikes N. 60°–80° E. and dips steeply southeast. The shearing was not observed in other adjacent, more mafic and finer grained rocks. However, a cataclastic foliation striking N. 60°–80° E. and dipping about 60° SE. can be seen locally in the Spikehorn Creek Granite along U.S. Highway 141–8, in the northern part of sec. 1, T. 37 N., R. 20 E. (fig. 2); probably this structure has resulted from shearing within this northeast-trending zone. Thin sections of some samples of the granite show cataclastic textures.

Field relations indicate that the shearing in the Twelvefoot Falls shear zone postdates intrusion of the Amberg Quartz Monzonite, which has a U-Pb zircon age of 1,760 Ma (Van Schmus, 1980), but analysis of disparate Rb-Sr biotite ages (Peterman and others, 1985), discussed later, indicate still younger movement in this zone as well as on the northeast-trending shear zone.

METAMORPHISM

The north margin of the Dunbar dome has a narrow metamorphic aureole that increases in intensity inward, from low greenschist facies away from the dome to amphibolite facies adjacent to and within part of the core. The pattern is similar to but more asymmetrical than the nodal distribution pattern of metamorphism surrounding some domes cored by Archean rocks in northern Michigan (James, 1955), such as the Watersmeet dome (Sims and others, 1984).

Along the north margin of the dome, southeast of Niagara (fig. 36), the amphibolite-facies zone is about 1 km wide, and includes a narrow inner zone of andesine amphibolite grade, as defined by Bayley and others (1966). Westward, the amphibolite zone widens, diverging markedly from the core-cover boundary. It intersects the Niagara fault east of Aurora and follows the fault for about 5 km to the west (Bayley and others, 1966). Farther west, Dutton (1971, pl. 5) mapped a garnet isograd that crosses the fault and then extends northwestward subparallel to it. According to Dutton (1971, p. 41), the mafic rocks of the Quinnesec Formation and associated mafic sills on the northwest side of the dome are metamorphosed mainly to the andesine amphibolite facies, and thus equate with this facies mapped by Bayley and others (1966) to the east. Adjacent to the northwest margin of the core of the dome (loc. C, fig. 36), some mafic metavolcanic rocks are diopside bearing, and thus probably equivalent to the hornblende-hornfels facies of Fyfe and others (1958). This assemblage occurs adjacent to the Dunbar Gneiss as well as adjacent to the Hoskin Lake Granite, as was noted previously by Bayley and others (1966). A thin section from SW¼SE¼ sec. 25, T. 38 N., R. 18 E. in Florence County (fig. 2), in mafic metavolcanic rock adjacent to the gneiss, shows that diopside occurs as anhedral, poikilitic grains that formed later than mylonitization of the rock and recrystallization of plagioclase to a mosaic texture. Subsequently, the diopside was partly altered to minerals stable at lower temperatures: hornblende, biotite, and epidote.

As shown by Dutton (1971, pl. 5), all the rocks adjacent to the Bush Lake lobe of the dome are garnet grade, and equivalent to amphibolite facies. The metasediments adjacent to the west end of the lobe locally contain garnet, andalusite, and cordierite. Common mineral assemblages in calc-silicate rocks within the amphibolite zone are: diopside-oligoclase or andesine-(quartz); garnet-hornblende-andesine; oligoclase or andesine-biotite-hornblende-garnet-quartz; and hornblende-oligoclase or andesine-calcite-(microcline). Pelitic rocks contain the assemblage biotite-garnet-oligoclase-quartz±cordierite. This metamorphic zone extends northwestward, on the southwest side of the Niagara fault, at least to Nelma, Wis., a distance of 10 km from the northwest margin of the Bush Lake lobe. Nielsen (1984) has described a late-stage retrograde overprinting of the rocks in this area. The most prevalent alteration includes: cordierite→pinite+sericite; garnet→biotite+chlorite; and hornblende→actinolite+biotite±chlorite.

The mineral assemblage in the mafic rocks within the amphibolite zone is oligoclase-epidote-hornblende, andesine-hornblende (blue green or brownish green), and andesine-hornblende-diopside; the assemblage in the greenschist-facies zone is albite-chlorite-biotite-actinolitic amphibole-epidote (Bayley and others, 1966). The mineral assemblages are consistent with a low to moderate temperature-pressure environment. The diopside- and cordierite-bearing assemblages are indicative of a moderately low pressure environment.

Within the core, the northern parts of the Marinette Quartz Diorite and the Hoskin Lake Granite are metamorphosed to amphibolite facies. The width of this zone in core rocks along the north margin is as much as 4 km. Metamorphism of the northern part of the Marinette Quartz Diorite and the Hoskin Lake Granite followed deformation D_2 and probably continued until deformation D_4, when the Spikehorn Creek Granite flowed diapirically outward into the Niagara lobe. It postdated suturing of the magmatic terrane and the continental-margin assemblage. At least the latter part of this metamorphism was mainly thermal, as indicated by unoriented hornblende needles in the Marinette Quartz Diorite within the zone; these were observed both in sec. 26, T. 38 N., R. 19 E., and in sec. 22, T. 38 N., R. 20 E., near the Hoskin Lake Granite.

The locus of intense metamorphism along the north margin of the core is spatially associated with the Hoskin Lake Granite. Probably, emplacement of the magma was one phase of a continuum of thermal events during later stages of the doming that included diapirism of the granite in the Niagara lobe and mobility of potassium and other lithophile elements. The mobility of potassium is most evident in the

TABLE 14.—*Stratigraphic-tectonic evolution of Dunbar dome*

Age in Ma	Defor- mation	Event
		Quartz-tourmaline veinlets and fluorite in brittle fractures.
1,835		Spikehorn Creek Granite and, possibly, Bush Lake Granite emplaced as diapirs. Aplite and pegmatite intruded into Dunbar Gneiss and Marinette Quartz Diorite in western part.
	D_4	Intense deformation along and adjacent to north margin of dome.
		Metamorphism of rocks in northern part of core and adjacent cover rocks resulting from rise in isotherms (overlapped D_3 deformation); accompanied by potassium metasomatism.
	D_3	Northeast-trending folds on dome margins resulting from local shear couple.
	D_2	Deformation of Dunbar Gneiss, Marinette Quartz Diorite, Hoskin Lake Granite, Quinnesec Formation, and Newingham Tonalite on northwest-oriented fold axes (regional structure) resulting from continued arc-continent collision.
	D_1	Development of layering and foliation (S_1) in Dunbar Gneiss and Marinette Quartz Diorite and intrusion of granite pegmatite and aplite during early stages of regional deformation.
≈1,860		Formation of Quinnesec volcanic rocks and Newingham Tonalite in an oceanic arc environment followed by collision with continental margin and formation of collision-zone granitoid rocks that now compose the Dunbar dome.

northern part of the core, mainly in the Marinette Quartz Diorite and Hoskin Lake Granite; but as noted earlier, the primary composition of the Newingham Tonalite in the central core was also considerably altered by the addition of potassium feldspar porphyroblasts, as were certain layers in the Dunbar Gneiss. The potassium metasomatism was accompanied particularly by the addition of Rb and Nb.

The timing and cause of the retrogressive metamorphism that is so widespread, particularly in the Marinette Quartz Diorite in the eastern part of the dome, are not known; but most likely the metamorphism was associated with post-doming events, as recorded by discordant Rb-Sr whole-rock ages and Rb-Sr and K-Ar mineral ages that affected the region.

EVOLUTION OF DOME

The stratigraphic-tectonic evolution of the Dunbar dome spanned a relatively short time of about 30 m.y. during the Early Proterozoic, from about 1,862 Ma to 1,835 Ma (table 14).

The contrasting chemistry of the granitoid rocks within and outside the dome indicates that they formed from different sources, and presumably in different tectonic environments, and subsequently were tectonically superposed. The Newingham Tonalite and associated Quinnesec volcanic rocks were generated in a volcanic arc during subduction, and represent new mantle-derived crust. The intrusions within the dome have characteristics of syncollisional and post-collisional plutons, as defined by Pearce and others (1984), and presumably were generated from melting of continental lithosphere during collision of the arc with the continental margin.

The Dunbar Gneiss and the Marinette Quartz Diorite were intruded during regional deformation. We interpret the gneissic fabric (D_1) in the Dunbar Gneiss as resulting from deformation of an incompletely crystallized magma during the regional deformation (D_2). A compositional and textural layering as well as ductile flow textures resulted from this penetrative deformation.

The Marinette Quartz Diorite was emplaced approximately along the contact between the Dunbar Gneiss and the Quinnesec volcanic rocks. Somewhat later, the Hoskin Lake Granite was emplaced. This igneous event was followed by major regional deformation (D_2), which produced northwest-trending folds in both the core and cover rocks and produced regional greenschist-facies metamorphism. The fold axes are subparallel to the Niagara fault, implying that the folding in this area resulted from northeast-oriented compression during collision of the arc and continental margin. The regional folding was followed by local shear deformation (D_3) along the northwest and southeast margins of the dome, perhaps as a result of short-lived oblique collision. A rise of the isotherms in the northern part of the dome—in the area of the Hoskin Lake Granite—resulted in amphibolite-facies metamorphism of the Marinette in the northern part of the dome and the adjacent Quinnesec to the north. The culmination of deformation (D_4) occurred during peak metamorphic temperatures and was accompanied by inflation of the northern part of the dome. A steep lineation (L_4) and a mylonitic foliation (S_4) formed in rocks within the core-cover boundary and in the cover rocks to the north. This event is interpreted as the culmination of the collision of the Wisconsin magmatic terranes with the continental margin. Following suturing of the two terranes, isotherms within the northern part of the dome continued to rise, and the thermal aureole advanced northward across the Niagara fault (fig. 36), developing lower amphibolite mineral assemblages in the Michigamme Formation of the Marquette Range Supergroup. This increase in the thermal regime presumably led to emplacement of the Spikehorn Creek and Bush Lake Granites as diapirs. Later cooling of the rocks was accompanied by development of quartz-tourmaline veinlets and fluorite locally in brittle fractures both within and near the north margin of the dome.

POST-DOMING EVENTS

Post-doming events have severely perturbed Rb-Sr whole-rock and mineral ages. Twenty-two whole-rock samples (5–10 kg each), representing both massive and foliated units within the Dunbar dome, define a Rb-Sr isochron of 1,697±28 Ma, with an initial $^{87}Sr{:}^{86}Sr$ ratio of 0.7037±0.0014 (fig. 39). We attribute the disturbance of the Rb-Sr system at the whole-rock scale to open-system behavior related to cataclasis that variably affected all the units in the dome. Recrystallization of biotite (and microcline where present) and sericitization and epidotization of plagioclase increased the mobility of Rb and Sr. Fluids undoubtedly played a major role in the migration of Rb and Sr as well as other elements. A relation between rock composition and degree of resetting is suggested by an isochron age of 1,733±43 Ma obtained by regressing only those samples with $^{87}Rb{:}^{86}Sr$ ratios <3. This separation roughly divides the data according to rock type, with the granites (in the strict sense) having $^{87}Rb{:}^{86}Sr$ ratios >3 and the tonalites and granodiorites having ratios <3. This correlation between rock composition and degree of resetting of the Rb-Sr system is probably related to differences in physical properties of the rocks. The granites, being less biotitic and more quartz rich than the tonalites and granodiorites, probably deformed in a more brittle fashion, which led to a higher permeability and thus a greater opportunity for interaction with a fluid phase. Some of the units, although open systems on the sample-size scale (tens of centimeters), appear to have remained closed at larger scales. For example, average $^{87}Rb{:}^{86}Sr$ and $^{87}Sr{:}^{86}Sr$ values calculated for the Dunbar Gneiss (11 samples) by weighing each sample by its Sr content, are used to calculate a model age of 1,833±50 Ma, using an initial Sr ratio of 0.7020. Although the uncertainty is large, the model age is indistinguishable from the crystallization age given by the U-Pb zircon data.

Rubidium-strontium biotite ages of rocks within the Dunbar dome decrease from east to west (fig. 40). This variation is part of a regional pattern of Rb-Sr biotite ages that extends north to the Marquette trough (Peterman and others, 1985; Peterman and Sims, 1988). Within this area, 54 biotite ages define a tripartite distribution with well-defined modes at 1,580±70 Ma, 1,320±50 Ma, and 1,140±30 Ma. The older group is a composite that contains the tightly clustered 1,630±30 Ma ages for Archean rocks of the southern complex in northern Michigan (Van Schmus and Woolsey, 1975) and slightly younger ages from areas to the south. Van Schmus and Woolsey (1975) correlated the 1.63 Ga ages with a cryptic event that has affected Precambrian rocks over much of Wisconsin.

The younger group of ages, 1,140±30 Ma, in the western part of the dome, inspired us to undertake a comprehensive regional study of biotite ages in Wisconsin and northern Michigan (Peterman and Sims, 1988). This study delineated a large elliptical area approximately 100–150 km long and 50–100 km wide in northern Wisconsin where biotite from Early Proterozoic crystalline rocks yields Keweenawan ages (mean of 1,128±20 Ma). The biotite ages were interpreted to record uplift and cooling of this area, named the Goodman swell, during Keweenawan rifting and igneous activity. The west end of the Dunbar dome lies at the east end of the Goodman swell. The uplift is interpreted as a crustal forebulge that formed in response to depression of the crust along the axis of the rift by loading with heavy basaltic rocks. Tensional stresses developed in the upper part of the elastic plate along the crest of the forebulge, and these stresses facilitated the vertical reactivation of faults in and around the Dunbar dome.

The intermediate group of ages, 1,320±50 Ma, does not correlate with known events in the region (fig. 35). Aldrich and others (1965) suggested a thermal event at this time but did not elaborate on a cause. Possibly, the surface now characterized by the 1,320-Ma age group was uplifted and cooled during the Keweenawan from a depth at which the biotite systems were only partially reset.

MINERAL RESOURCES

The Early Proterozoic volcanic rocks in northern Wisconsin, which presumably are roughly equivalent to the Quinnesec Formation (Sims and others, 1989), contain scattered volcanic-hosted Cu-Zn-type massive sulfide deposits of future economic significance. At least four significant deposits are known (Afifi and others, 1984; Sims, 1987), the largest of which is the Crandon deposit (May and Schmidt, 1982), near Rhinelander, about 100 km west of Dunbar. The Crandon deposit contains estimated reserves of 83 million tons grading at 5 percent Zn, 1.1 percent Cu, 0.4 percent Pb, and 0.03 oz Au per ton (Mudrey, 1979). The other known deposits are smaller.

A small pyrrhotite-rich, massive sulfide lens that contains only trace amounts of base and precious metals has been drilled by the Duval Corporation in

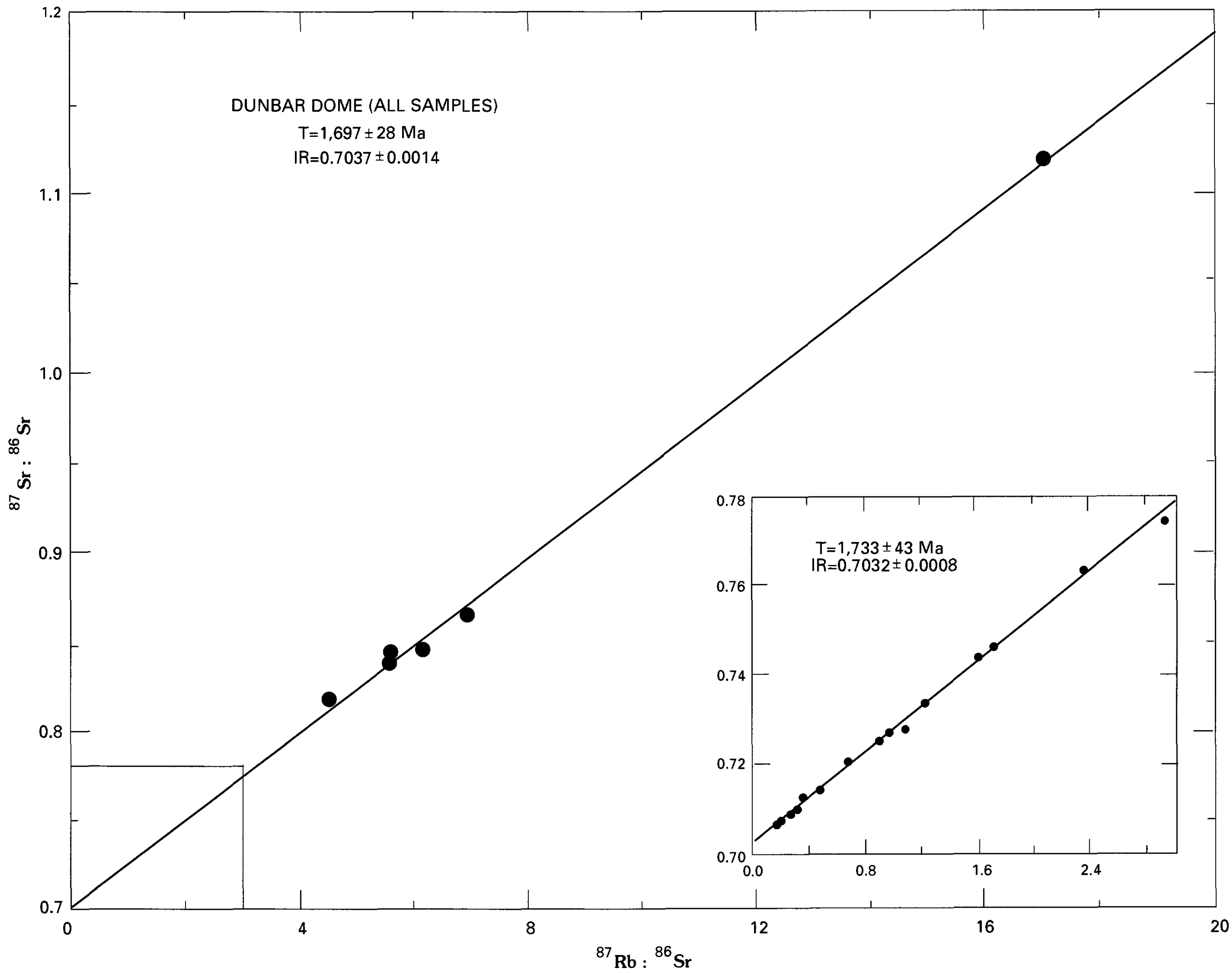


FIGURE 39.—Whole-rock Rb-Sr isochron for samples of all units within Dunbar dome. The isochron of 1,697±28 Ma is based on all the samples (22). Inset shows age and initial ratios (IR) for samples with ^{87}Rb:^{86}Sr ratios <3 (mainly tonalites and granodiorites).

secs. 2 and 3, T. 35 N., R. 18 E. and sec. 28, T. 36 N., R. 18 E. (Hollister and Cummings, 1982). The deposit is associated with an iron-formation unit and occurs in the interval between an upper marine clastic sedimentary succession and a lower submarine volcanic and iron-formation succession.

Presumably, other parts of the volcanic complex in northern Wisconsin are favorable for the occurrence of additional massive sulfide deposits. Also, the volcanic rocks merit consideration as exploration targets for gold and silver deposits.

A minor occurrence of molybdenite has been observed in the southwestern part of the Hoskin Lake Granite exposure area, in sec. 33, T. 38 N., R. 19 E. (fig. 2). The molybdenite occurs with pegmatite and aplite dikes and associated quartz veins. Several test pits were dug in this area in the 1950's, but commercial concentrations of molybdenite were not found (Fisher, 1957; Greenberg, 1983).

DISCUSSION

Prior to this study, it was generally accepted that the Niagara fault separated the Wisconsin magmatic terranes from the epicratonic continental-margin assemblage to the north. (See Sims and others, 1989, for discussion.) This conclusion was based mainly on the marked contrast in the geology, the volume of magmatism, and the geochemistry of the Early Proterozoic igneous rocks in the two areas (Sims and others, 1985). The volcanic rocks in the epicratonic assemblage are largely bimodal with abundant tholeiitic basalt and minor high-K_2O rhyolite. The basalts show strong iron

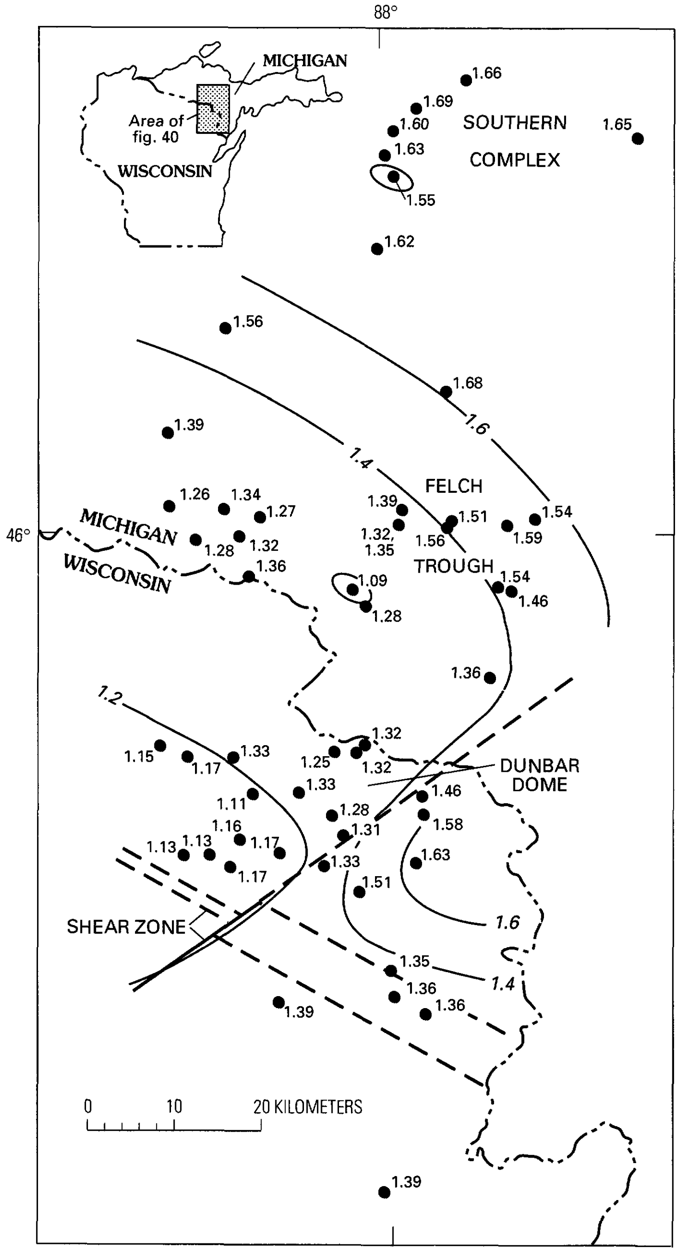


FIGURE 40.—Rb-Sr biotite ages (Ga) for Archean and Early Proterozoic rocks in northeastern Wisconsin and adjacent northern Michigan. Data are from Van Schmus and Woolsey (1975) for the southern complex, Aldrich and others (1965) for the Felch trough area, and Peterman and others (1985) for the Dunbar dome and vicinity.

enrichment and high TiO_2 and incompatible element contents (Fox, 1983); they are compositionally similar to continental rift basalts, such as those of the Keweenawan in Minnesota (Green, 1983). In contrast, the voluminous volcanic rocks in Wisconsin range from basalt through andesite to rhyolite, lack strong iron enrichment, and have volcanic-arc-derived characteristics (Greenberg and Brown, 1983; Schulz, 1984a, b). Van Schmus (1976), Cambray (1978), LaBerge and others (1984), Larue and Sloss (1980), and Schulz and others (1984) have proposed various plate tectonic models to account for the differences in the Early Proterozoic geology in Wisconsin and Michigan.

This study has shown some of the complexities of the boundary zone between the Wisconsin magmatic terranes and the continental-margin assemblage to the north. The boundary is not a simple suture that sharply separates the two contrasting terranes; instead, it is a complex tectonic zone of intricately mixed rocks that were formed in different tectonic environments and subsequently were juxtaposed, mainly by thrusting. The granitoid rocks are a key to the paleoenvironments. As discussed earlier, the Newingham Tonalite is typical of volcanic-arc granitoid bodies (Pearce and others, 1984), generated by subduction during closure of the ocean that intervened between arc complexes of the Wisconsin magmatic terranes and the continental margin. The remainder of the granitoid rocks in the Dunbar area show affinities with syn- to post-collisional granites of recent orogenic belts. Apparently these were generated by melting of continental lithosphere and crust during collision—an event unique to the collision zone.

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