TOURMALINITES, NELSONITES, AND RELATED ROCKS (EARLY PROTEROZOIC) NEAR PHILBROOK, TODD COUNTY, MINNESOTA

A THESIS
SUBMITTED TO THE FACULTY OF THE GRADUATE SCHOOL
OF THE UNIVERSITY OF MINNESOTA

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IN PARTIAL FULFILLMENT OF THE REQUIREMENT
FOR THE DEGREE OF
MASTER OF SCIENCE

FEBRUARY, 1987
ABSTRACT

Lower Proterozoic rocks near Philbrook, Minnesota consist of tourmaline-rich metasedimentary rocks, ultramafic to mafic igneous rocks, and metaconglomerates. The metasedimentary rocks have been intersected by drilling only; the igneous rocks have been intersected by drilling and also occur in outcrop; and the metaconglomerates occur in outcrop only.

The metasedimentary rocks intersected by drill holes west of Philbrook are a minimum of 230 m thick, strike N-NE, dip steeply east, and are believed to be part of the Mille Lacs Group. The metasedimentary rocks have been regionally metamorphosed to the upper greenschist facies. Regional geologic relations and mineralogy indicate that these metasedimentary rocks are Lower Proterozoic in age and in close proximity to Archean crustal rocks.

The general stratigraphy of the metasedimentary rocks, from bottom to top, consists of a lower arkose or feldspathic arenite, laminated quartz-magnetite or hematite iron-formation, upper arkose, and 'metagraywacke' (tourmaline- and magnetite-bearing micaceous schist). Neither the top nor bottom of the stratigraphic section has been penetrated by drilling.

The high feldspar content of the arkosic rocks indicates a predominance of chemical weathering and a
high relief source area. A submarine-fan depositional environment is indicated by the presence of very thin silt and sand beds intercalated with laminated iron-formation, with periodic slumping of clean sand from a shallow-water staging area upslope in the basin.

Tourmaline occurs in the arkosic rocks as disseminated crystals and as laminated tourmalinite, and in the micaceous schist as disseminated crystals. The chemical composition of the tourmaline (schorl-dravite) is similar to that of tourmaline which is associated with several syngentic, stratiform massive sulfide deposits elsewhere in the world, indicating a potential for sulfide mineralization at Philbrook.

The metaconglomerates occur in two poorly exposed clusters of outcrops along the edge of the Long Prairie River. Although less than 300 m apart, the two groups of conglomerate outcrops are quite different in mineralogy, suggesting a basal deposit of local provenance. One of the conglomerates is comprised mostly of quartz-rich pelitic clasts, and the other contains predominantly coarse-grained plutonic clasts. The clasts in the latter are mineralogically similar to a nearby hypabyssal, porphyritic tonalite which is believed to be Archean in age, providing evidence that the conglomerates are the basal part of an Early Proterozoic sequence which unconformably overlies Archean crustal rocks.

The ultramafic to mafic igneous rocks, based on
limited outcrop and drill hole data and aeromagnetic signature, are part of an intrusion which is approximately 2.4 km in diameter, circular in shape with steep sides, and concentrically zoned. Rock types in the intrusion range from ultramafic oxide-apatite rock (nelsonite), pyroxenite, and hornblendite (penetrated by drill holes) to apatite-rich melanocratic diorite and mesocratic diorite (in outcrop). Pegmatitic and anorthositic zones occur in the latter rock type. Chemically, this entire suite of igneous rocks is high in P₂O₅, TiO₂, and Fe₂O₃, and low in SiO₂, MgO, K₂O, and Na₂O. Abundant primary and deuteric hornblende (after pyroxene), net veins, pegmatitic patches, and local saussuritization of plagioclase supply evidence for a high H₂O content of the intrusion.

An immiscible split model is employed as the origin for the ultramafic parts of the intrusion. The melanocratic diorite may be a possible parent material which has split into an Fe-Ti-P-rich fraction and a Si-Al-Na-K-rich fraction. Some cumulus mineralogical layering may have further modified the splits, producing the zonal arrangement of the intrusion.

Late brittle shear zones and some weakly developed cleavages are the only observable structures in the intrusion, indicating that it has not been severely affected by the Penokean Orogeny. It is believed that the Philbrook intrusion is similar in age, and probably related to, Penokean granites of the St. Cloud type.
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INTRODUCTION

Drilling History

The bedrock geology of east-central Minnesota consists of a variety of Archean to Proterozoic meta-sedimentary, metavolcanic, and igneous rocks which bear evidence of a complex deformational history. Thick Pleistocene glacial cover obscures most of the bedrock geology in this area, and outcrops are sparse and small. Although the northern part of the area was extensively drilled for iron-ore exploration from the early 1900’s to the late 1950’s, detailed exploration was limited to the Cuyuna range and is marginally useful for geological interpretations of outlying areas.

Between December 18, 1909 and April 15, 1910, Cuyler Adams drilled several E-W sets of drill holes over a N-S trending, linear magnetic high located slightly more than 1 km west of the town of Philbrook (40-50 km southwest of the Cuyuna range) in Sec.5, T.132 N., R.32 W. and Sec.32, T.133 N., R.32 W., in Todd County, Minnesota (Fig. 1). Adams also drilled 12 holes, presumably for iron-ore, in the NW 1/4, NW 1/4, Sec. 3, T.132 N., R.32 W. which penetrate oxide-rich, strongly magnetic mafic intrusive rocks. The iron-formation and oxide rock proved to be magnetite-rich, but the iron content was not of ore-grade. Representative pieces of this core are now on file at the Minnesota Geological Survey, but only limited petrographic or stratigraphic work had been attempted on
Figure 1. Location map showing drill hole (secs. 3, 5, and 32) and outcrop locations (X).
these rocks. The construction of stratigraphic columns and logging of core is made difficult by the fact that only 3"-12" representative pieces of core per actual 5' of drilled core are available.

**Purpose**

The original purpose of this thesis was to provide a petrographic and stratigraphic report on these drill cores, and to integrate the drill cores with scattered exposures of schistose and gabbroic rocks just NE of Philbrook in Sec.34, T.133 N., R.32 W., Todd County. The outcrops were mapped in detail as a part of this project. The correlation of rock units between drill holes is based on petrographic and lithologic features, and is augmented by eight whole-rock analyses and several microprobe analyses.

Another aim of this thesis is to develop evidence for assigning a Proterozoic or Archean age to the metasedimentary rocks near Philbrook. They are situated near the western edge of the Animikie Basin, and, because of their geologic and geographic isolation, could be part of the Proterozoic basin fill or part of the Archean basement.

An unexpected early result of this work was the discovery of tourmaline in association with iron-formation and feldspar-rich rocks, near the respective top and bottom of the drilled stratigraphic sequence in sections 5 and 32 (App. A). The recently recognized
Figure 2. Regional geologic map (after Morey, 1986).
associations of tourmalinites with synsedimentary massive-sulfide deposits (Ethier and Campbell, 1977, 1983; Fleischer and Routhier, 1973; Taylor and Slack, 1984; Slack et al., 1984) prompted detailed study of the tourmalinite-bearing rocks at Philbrook, with emphasis on mineralization potential.

REGIONAL GEOLOGY

General

Early Proterozoic rocks (2,600 - 1,800 M.Y.) in east-central Minnesota define a broad, eastward-plunging synclinorium which is surrounded by Archean rocks on all sides except the east (Fig. 2). The Early Proterozoic rocks are deformed to various degrees, apparently depending on the competency of the underlying basement (Morey, 1983). Deposition of the Early Proterozoic strata occurred in an intracratonic basin which is inferred to have formed over a rift between the Archean gneissic and greenstone terranes; sea-level and basin shape fluctuations produced variations in the sedimentary sequences. Early deposition was in a shallow water, high energy environment, and is represented by conglomerates and quartzites in the lower Mille Lacs Group. Mafic volcanic and hypabyssal rocks were emplaced in the sedimentary sequences, also a result of the rifting environment. In the more advanced stages of rifting, the basin was deeper and quieter, and deposition of the overlying
Animikie strata occurred. Closing of the basin deformed the sediments, and was accompanied by the emplacement of mafic to felsic plutonic igneous rocks.

Regional Stratigraphy and Structure

The Early Proterozoic rocks of east-central Minnesota have been divided into two groups separated by a possible unconformity (Fig. 2). The Mille Lacs Group is the older of the two and contains the Denham, Glen Township, Randall, Little Falls, and Trout Lake Formations (Morey, 1978; Marsden, 1972). This group of rocks unconformably overlies Archean rocks. The Animikie Group is the younger of the two groups, and in the Cuyuna range includes (from oldest to youngest) the Mahnomen, Trommald, and Rabbit Lake Formations. These are believed to correlate with the Pokegama, Biwabik, and Virginia Formations of the Mesabi range and the Kakabeka, Gunflint, and Rove Formations of the Gunflint range (Morey, 1973). The rocks of the Animikie Group are gently dipping and relatively undisturbed on the Mesabi Range, where they overlie Archean granite-greenstone terrane. Further south, the Animikie strata unconformably overlie both Early Proterozoic schists and Archean gneisses, and are deformed and complexly folded. Deformation of the Proterozoic strata increases to the south, and structures range from open folds with near vertical axial planes to tight, overturned isoclinal folds with axial planes dipping up to 60° in the southern area (Morey, 1978).
The metamorphic grade also increases to the south, correlating in a general way to the degree of deformation. The highest metamorphic grade attained is the staurolite/garnet facies, near Little Falls.

The geology of the Early Proterozoic sequence in east-central Minnesota is largely inferred from geophysical data, augmented by test drilling, rare outcrops, and mine exposures which are now flooded. The distribution of non-magnetic units may be more complex than indicated in published literature (Morey, 1978).

Since the results of this thesis suggest that there is no reason to believe that any correlation exists between the rocks of the Philbrook area and the Animikie Group, only the Mille Lacs Group will be described in more detail.

**Mille_Lacs_Group**

**Denham_Formation**

The Denham Formation is exposed approximately 100 km east of the Philbrook area, where it overlies the Archean McGrath Gneiss and is composed mainly of metamorphosed feldspathic, arenitic, quartz-rich conglomerate and siltstone, with small amounts of dolomite, amphibolitized pillowed basalt, and mafic to intermediate meta-agglomerate (Morey, 1978). The conglomeratic units are confined to the base of the formation and contain clasts derived from the McGrath Gneiss in a matrix of mixed sand
silt-sized grains, with some carbonate cement. The matrix is extensively recrystallized and composed of quartz, muscovite, and lesser amounts of biotite and chloritoid.

On an outcrop scale, thin to thick vitreous quartzite beds are intercalated with thick argillite or argillaceous siltstone and thinner beds of dolomite. The quartzite varies from a white to dark reddish-gray color due to varying proportions of hematite and calcite. The argillaceous layers are generally light-colored and phyllitic due to strong orientation of muscovite and quartz parallel to bedding laminae.

Thin layers of pillowated metabasalt and thicker (<20m) agglomeratic units can be recognized, and several of the quartz-rich layers and amphibolite layers may be highly recrystallized felsic and mafic igneous rocks.

The Denham Formation passes laterally into the Randall and Glen Township Formations (Morey, 1978).

**Randall Formation**

The Randall Formation is named for exposures in Randall, Minnesota, about 30 km SSE of Philbrook. It consists dominantly of mafic to intermediate metavolcanic and hypabyssal rocks and interbedded clastic rocks which resemble those in the Denham and Glen Township Formations, and includes iron-formation, chert, quartzite, argillite, siltstone, and quartz-pebble conglomerate. Metavolcanic rocks include pillowed metabasalt, which is
extensively altered to sericite, actinolite, quartz, and chlorite. Hypabyssal rocks consist of fine- to medium-grained diabase which is associated with the flows and similiarly altered. Discontinuous units of banded iron-formation are present throughout the formation and are composed of finely laminated, dark, magnetite-bearing cherty rocks interlayered with thin to thick chert beds. Some mappable units of iron-formation have been located by drilling along strike (Morey, 1978).

The stratigraphic position of the Randall Formation is somewhat uncertain due to the small amount of outcrop, but it is believed to lie in close proximity to and to be infolded with the surrounding Archean rocks. The Randall Formation contains basal conglomerate and quartzite similar to the Denham Formation. It appears to be occupy the same stratigraphic position as the Glen Township Formation in the Animikie basin, but because the two units are separated by approximately 80 km, the rocks at Randall are given formational status (Morey, 1978).

**Glen Township Formation**

The Glen Township Formation consists of interbedded pyrite- and pyrrhotite-bearing carbonaceous slates, cherty iron-formation, metabasalt, diabase, and minor quartzite, argillite, and limestone. It is named after subsurface occurrences in Glen Township, southwestern Aitkin county. The Glen Township Formation may grade laterally into the Denham Formation and vertically into
the Mille Lacs group, undivided. Thin conglomeratic beds occur near the base of the formation, and several basaltic and diabasic rocks are present which may represent flows and hypabyssal intrusions, but the degree of deformation and metamorphism has made the relationships obscure (Morey, 1978).

**Little Falls Formation**

The Little Falls Formation is named for exposures on Mill Island in the Mississippi River Valley in Little Falls, Minnesota, 70 km south of Philbrook. The formation consists mainly of argillaceous rocks intercalated with thin quartz-rich metasiltstone beds and thin to thick beds of quartzose metagraywacke. The argillaceous units are metamorphosed to slate, phyllite, or finely crystalline schist. Carbonate concretions are found in the graywacke units.

Dark to light gray metagraywacke comprises 1/2 to 1/3 of the exposed thickness. It has framework grains of quartz with minor plagioclase set in a fine-grained quartz-muscovite matrix. The beds range from 3 cm to 3 m in thickness and are mostly massive in the more argillaceous units. Garnet and staurolite are present in the southern part of the Little Falls Formation where the metamorphic grade is the highest. Garnet and hornblende are locally found in the metamorphosed calcareous concretions. Drill hole and geophysical data indicate that the Little Falls strata are transitional northwards into
the Randall Formation (Morey, 1978).

**Trout Lake Formation**

The Trout Lake Formation was named by Marsden (1972) for a dolomitic unit which is completely restricted to the subsurface. Marsden (1972) suggested that the Trout Lake Formation unconformably underlies the Mahnomen Formation and overlies strata that Morey (1978) assigned to the Mille Lacs Group.

'Basalt Near Mora'

In addition to these formations, there are meta-basalt outcrops near Mora, Minnesota. Geophysical data indicate that the basalt forms a 3 x 17 km, east-west trending unit, which shows premetamorphic textural and mineralogical features similar to the volcanic rocks in the Denham and Randall Formations. It is therefore assumed by Morey (1978) that it is Early Proterozoic in age.

**Great Lakes Tectonic Zone**

The thesis area is located near the trace of the Great Lakes Tectonic Zone (GLTZ, Fig. 1), a boundary thrust fault between Archean gneisses of at least 3,000 Ma in southern Minnesota and the 2,700 Ma Archean greenstone-granite of northern Minnesota. The Early Proterozoic strata are located over this boundary, a center of complex deformation, sedimentation, and igneous activity.
In the Minnesota River Valley, Archean gneisses are exposed which record rock-forming events with ages of 3,600, 3,300, 3,000, 2,600, 1,850, and 1,760 Ma (Goldich et al., 1970). Rocks of the greenstone-granite terrane in northern Minnesota formed in the shorter time span of 2,750-2,450 Ma. In addition to the differences in age, the two terranes show very different metamorphic grades and structural features. The southern gneisses are warped into broad domes and basins. In contrast, most of the rocks of northern Minnesota are metamorphosed only to the greenschist facies and are isoclinaly folded into linear belts.

Near the end of the Penokean orogeny, granitic bodies intruded both terranes on a large scale, apparently welding the two terranes together (Sims and Morey, 1986).

The area of the break between the thick, highly deformed southern Animikie rocks and the thinner, undeformed northern sequence coincides with a good break in aeromagnetic and gravity geophysical anomalies. This geophysical break can be traced in a southwest direction to South Dakota (Lidiak, 1971), and to the northeast through Wisconsin, Michigan, and to the North Shore of Lake Huron to the Grenville front (Sims and others, 1981; Sims and Morey, 1986). Several historic earthquakes have occurred along this geophysical break (Mooney, 1979). This break, named the Great Lakes Tectonic Zone (GLTZ) by Sims and others (1981), is the locus of Penokean
deformation and Early Proterozoic sedimentation.

As mentioned previously, the large variation in the degree of deformation of the Animikie strata is thought to depend on the character of the underlying basement, with the highest deformation occurring in strata overlying the more mobile gneissic terrane; and on whether the rocks have been affected by one or two deformations (Morey, 1986; Morey and Sims, 1976; Holst, 1984).

The lack of outcrop and drill hole data have led to various interpretations of the Early Proterozoic tectonic regime and the GLTZ. In Wisconsin, some workers envision a southward-dipping subduction zone with Phanerozoic attributes (LaRue, 1983); others see it as a 'modified' subduction zone with some Phanerozoic attributes but with different processes at work due to the higher heat flow of the thinner Archean crust (Greenberg and Brown, 1983). In Minnesota the most highly regarded theory involves intracratonic rifting between the greenstone-granite terrane and the ancient gneiss terrane, along which down-dropped grabens formed a basin for Early Proterozoic sedimentation (eg. Morey, 1983). Subsequent closing of the basin during the Penokean Orogeny (below) deformed these sediments to various degrees, producing locally intense deformation, and metamorphism (Morey, 1983; Sims et al, 1981).

Possibly similar zones are known in Canada, between the Churchill and Superior Provinces, and further study
of the Minnesota GLTZ may provide valuable insight into comparisons with that Canadian Shield suture. Base metal deposits are also known in the Canadian suture, and this possibility could exist in Minnesota (Southwick, 1980).

Penokean Orogeny

The Penokean orogeny refers to a period of compressional deformation which culminated approximately 1850 Ma and followed the deposition of the Lower Proterozoic epiclastic sediments. During the orogeny, compressional forces from the southeast deformed the Lower Proterozoic sediments to various degrees. The greenstone-granite terrane of the Superior Province is thought to have acted as a stable foreland, protecting the overlying Animikie sediments (Mesabi range) from deformation. However, the ancient gneisses to the south behaved in a more ductile fashion, and the Proterozoic cover was more highly deformed and metamorphosed (Sims and Morey, 1986).

Thrust faulting probably occurred in the southern part of the Animikie basin, as evidenced by nappe structures, and two periods of deformation have affected the rocks in this basin. The tectonic front which separates the regions of one and two deformations is recognized to the north by the presence of one cleavage (upright, E-NE trending, later), or to the south by two cleavages (older, subhorizontal cleavage with later subvertical cleavage superimposed upon it (Holst, 1984).
Igneous intrusions involved in the late stages of the Penokean orogeny consist of syntectonic dikes, sills, and irregular bodies of gabbro and syenite, which have been subjected to greenschist-facies metamorphism. Subsequent to these were a series of large syn- to post-tectonic intrusions of tonalite to granodiorite. These rocks, commonly referred to as the St. Cloud-type granites and named the 'Stearns magma series' by Woyski (1949), are typically porphyritic, weakly foliated rocks containing inclusions of biotite or hornblende schist.

**Summary of Early Proterozoic Geologic History**

Much of the interpretation of the geology of East-Central Minnesota rests on magnetic and gravimetric surveys coupled with numerous test holes and local areas of exposures in the mines.

The Mille Lacs Group represents a fining-upward sequence which formed in the Animikie basin as rifting progressed. The interlayered volcanic rocks are the result of volcanism along the down-dropped grabens of the rift system. The lower conglomerates represent the initial stages of rifting and a shallow water, high-energy depositional environment. As the basin developed, sandstones were deposited along the basin margins and a basin-ward deep water and/or reducing environment resulted in the deposition of stratiform sulfidic, carbonaceous pelitic rocks. The younger Animikie Group was unconformably deposited on the Mille Lacs Group, and is
made up of a southward-thickening wedge of sedimentary rocks, the lowest of which have shallow-water attributes. The deep-water accumulation of the upper slaty sediments of the Animikie Group probably resulted from a drastic change in basin configuration at the end of iron-formation deposition, including rifting and associated foundering of the basin (Morey, 1978).

In the Proterozoic strata, five depositional phases can be recognized (Morey, 1978). The first two phases are miogeoclinal and include the Denham, Glen Township, Little Falls, and Trout Lake Formations, unconformably overlain by the Mahnomen Formation (second phase). The third phase is the main iron-bearing member of the iron ranges, which shows shallow-water attributes such as algal stromatolites and cross-bedding (Ojakangas, 1983). The fourth phase, which is quite carbonaceous, is transitional to a deeper water depositional environment, and includes the lower and Emily members of the Rabbit Lake Formation. The fifth phase, the upper member of the Rabbit Lake Formation, is eugeoclinal, southward thickening, and flysch-like. The proportion of mud increases both upward and outward (southward) in the Animikie strata.

The Early Proterozoic strata were intruded by mafic to intermediate dikes, sills, and plutons, which were emplaced in association with the Penokean orogeny (ca. 1,800 Ma). Evidence for the time of emplacement of these
intrusions relative to the Penokean orogeny includes minor deformation (shear zones, minor cataclasis) of the igneous bodies, probably produced during the late stages of the orogeny.

Acknowledgements

I wish to thank Dr. R. W. Ojakangas for serving as principal advisor and for his help and guidance. I also thank Dr. J. C. Green for his thorough and critical readings of this manuscript, and for serving as committee member. Special thanks to committee members Dr. D. L. Southwick (MGS), for suggesting the thesis, for his patient advice, and for serving as primary manuscript editor; and Dr. G. B. Morey (MGS), for his critical reading of the manuscript and thoughtful discussions. Also special thanks to the Minnesota Geological Survey for financial backing and for use of facilities, which were greatly appreciated; and to various staff members at MGS for supplying mental support and stimuli.
Figure 3. Geologic Map of the Philbrook area.
Figure 4. A.) Aeromagnetic map of Philbrook area. B.) Dip-needle contour map of anomaly west of Philbrook, over iron-formation. Boxed in area in (A.) corresponds to that in Figure 1.
INTRODUCTION TO SUBSURFACE METASEDIMENTARY ROCKS NEAR PHILBROOK

Thirty-three drill holes located in the SE. 1/4, Sec. 32, and down the middle of the N. 1/2, Sec. 5, just west of Philbrook, Mn. were drilled over a very pronounced N-NW trending magnetic high that was first recognized from dip-needle surveys (Fig. 4). This oval- to linear-shaped magnetic high, 9000 gammas over the surrounding background values on the modern aeromagnetic survey (Minnesota Geological Survey, 1983) is an effect of the under-lying magnetite-rich strata, which strike N to NW and dip steeply to the east (App. A).

A contour map of the early dip-needle work, (Fig. 4), shows a pock-marked pattern to the anomaly, the probable result of local leaching and oxidation of magnetite to hematite and goethite.

Petrographic Procedures

The drill core was macroscopically logged with as much lithologic detail as the incompletely represented samples would permit. Because the core was skeletonized in 5’ (1.5 m) intervals, the minimum thickness of a lithologic unit is restricted to 5’ (1.5 m). Macroscopic mineral identification is limited by the fine grain size, but mica/silicate/Fe-oxide ratios can be estimated from color and magnetic susceptibility. Quartz/feldspar ratios, however, are hard to determine in hand specimen, and in some cases, even microscopically. Other obser-
viable macroscopic features include types and orientations of bedding, foliations, fractures, and veins.

All heels from thin sections were stained for K-feldspar and plagioclase using Na-cobaltinitrate and the Laniz staining method. Select thin sections were also stained only for K-feldspar. A total of 125 thin sections were prepared to petrographically analyze all varieties of rock types, as well as to closely inspect unusual textures or mineralogies. Twenty-one of these thin sections, representing all of the basic lithological units, were chosen for point-counting. A minimum number of 1000 points per thin section were counted, grain size extremes were measured, and average grain size estimated. Two polished thin sections of metasedimentary rocks were prepared to determine the kinds and textures of the opaque oxide phases.

Classification of Metasedimentary Rocks

Due to recrystallization and metamorphism, the use of conventional sedimentary rock names and textures is not always possible, especially for the mica-rich rocks. Bedding is well preserved, however, and some grain size variations exist in the arenitic rocks which probably reflect size variations of the primary sediments. To a degree, the preservation of primary mineralogy is a function of grain size, as the fine-grained sediments are the most altered mineralogically. The micaceous schists are completely recrystallized, and the only clues to
Figure 5—Modal diagrams used to infer protoliths of metasedimentary rocks. OPQ = opaque oxides, Q = quartz, F = feldspar, PHY = phyllosilicates. I = quartz arenite, II = arkose, III = feldspathic arenite, IV = feldspathic graywacke, V = graywacke, VI = Mudstone
protoliths are laminations, relict igneous textures, rare recognizable rock fragments, and subtle mineralogical variations. Those rocks which are micaceous, foliated, and typically green are referred to as schist, and those rocks which are non-foliated, quartzofeldspathic, and whitish are referred to as arkose or feldspathic arenite.

As an aid to determining protoliths and in naming the rock members of the suite, a two-step classification scheme has been developed. The first step involves the use of ternary diagrams (Fig. 5), with varying combinations of quartz, feldspar, phyllosilicates, and opaques at the apices. Recognizable rock fragments are very rare, and are omitted from the classification. The triangle with opaque-phyllosilicates-(quartz + feldspar) apices (a.) is used as the basis to separate iron-formation (>20% Fe-oxides) from other rock types. Those rocks which contain <20% opaque Fe-oxides are replotted on a quartz-feldspar-phyllosilicate diagram (b.), and named in accordance with the classification scheme in Pettijohn and others (1973). Another way of expressing the Q-F-Ps diagram (Fig. 5c), helps to more clearly define the fields of variations in these rock types. In this classification scheme the phyllosilicates (muscovite, sericite, biotite, chlorite) are assumed to represent recrystallized matrix. Although this assumption certainly does not always hold true (e.g. in the case of sericitic alteration of feldspar), it has validity in most cases.
Other than iron-formation, most of the sedimentary rocks of this study range from feldspathic arenite to feldspathic graywacke, with lesser amounts of the more quartz-rich arkose and graywacke.

**Stratigraphy**

The metasedimentary rocks intersected by drilling near Philbrook consist of feldspathic arenite to plagioclase arkose, iron-formation, ferruginous graywacke (micaceous schist), and minor volcanogenic graywacke (?) (now carbonate-bearing chlorite-, biotite-, or amphibole-schist), and mafic volcanic rocks which may represent flows or hypabyssal intrusions (now epidote-chlorite schist). The minor units of volcanic rocks are less than 7 m thick, and are most commonly interbedded within iron-formation and micaceous schist (ferruginous metagraywacke). The lateral continuity of these thin units is somewhat uncertain, but they seem to be stratiform.

**Arkose and Feldspathic Arenite**

Arkose and feldspathic arenite occur in a separate 'lower' unit and an 'upper' unit (Fig. 3; App. A). These well indurated, variably metamorphosed and altered rocks constitute most of the lower part of the drilled section, and account for about half of the footage drilled. Individual sedimentary units range from 7 to 35 m in thickness, and some minor 1.5 m-thick lenses are present within other rock units. The arkose has a fairly continuous lateral extent over the length of the drill
pattern. However toward the northern end of the upper unit, the arkose apparently grades into chlorite schist and iron-formation (App. A).

The arkose ranges from a coarse-grained, pinkish-gray metasandstone to a fine-grained, greenish-gray or pinkish-gray metasiltstone. Laminations, created by thin magnetite- or tourmaline-rich layers are present but are not a prominent feature. A weak sheen may be imparted to the rock by the preferred orientation of foliated, fine-grained micas.

Iron-Formation

The term 'iron-formation' as used in this study, denotes those rocks containing greater than 20% modal volume of opaque oxide minerals. Iron-formation represents approximately 30% of the total stratigraphy intersected by drilling, ranging from 1.5 to 110 m in stratigraphic thickness. The iron-formation is grayish-green to dark gray, strongly to moderately magnetic, and thinlly laminated. Individual laminations range from less than 1 mm to 1 cm in thickness, and are defined by varying proportions of oxides to quartz, or oxide-quartz (cherty) laminae with intercalated quartz- and feldspar-rich clastic laminae (Fig. 12). A sub-unit of iron-formation in the upper arkose grades into the overlying tourmaline- and magnetite-bearing mica schist. This upper iron-formation differs from that just described in that the clastic laminae contain abundant magnetite and
are finer-grained than the clastic intercalations in the main iron-formation unit. Thin cherty (quartz-magnetite) laminae are still present in this upper iron-formation.

Magnetite and specular hematite are essentially the only iron-bearing phases of the iron-formation. Well-foliated specular schist accounts for approximately 50% of the iron-formation, and is most abundant in the southern part of the drilled area.

Thin carbonate-quartz veinlets are common, as are chloritized, lineated slickenside surfaces; both are normally at a low angle to the core axis and bedding.

Minor metamorphic actinolite is present at the contacts between the cherty (quartz-magnetite) and clastic (quartz-feldspar) laminae; these amphiboles are foliated parallel to bedding. In some cases the rock has been sheared; these shear zones contain semi-massive sulfides (pyrite), along with chlorite, quartz, and carbonate, in poorly defined veins that are <1 cm in thickness.

Effects of surface weathering are generally quite shallow and consist of goethite and limonite after magnetite, hematite, and iron-silicate minerals.

The lack of granular texture, the mineralogy, and the fine cherty laminae suggest this is an Algoma-type iron-formation.
Tourmaline and Magnetite-bearing Micaceous Schist

Dark green, fine- to medium-grained, laminated, slightly magnetic chlorite schist comprises approximately 10% of the intersected thickness. This unit has a maximum drilled thickness of 15 m, but could be thicker as the top is not intersected by drilling. The schist directly overlies the upper arkosic unit, except in the northern part of the drill area where an iron-formation intervenes (App. A). In this area the basal contact of the schist with iron-formation is gradational, and a gradation to iron-formation also exists in the upper part of the unit, where abundant thin (<1 cm), internally laminated cherty quartz-magnetite laminae occur.

Porphyroblasts of chlorite and muscovite occur in the coarser-grained part of this schist, as do tiny tourmaline needles (Fig. 12). Also present, and especially obvious in the north end of the unit, are soft white sericitized 'clasts' up to 3 cm long (Fig. 11). These are folded and elongate parallel to foliation, and set in a dark green, fine-grained chloritic matrix. They are interpreted as felsic rock fragments or sericitized feldspar crystals which have been flattened.

The lowermost 3 m of Hole #108 (App. A) is in a lighter-colored schist which has a fine-grained, relict sandy texture with relatively large magnetite porphyroblasts and visible tourmaline; this is apparently a thin unit within the iron-formation.
Other Schist Varieties

In addition to the tourmaline- and magnetite-bearing mica schists just described, several other varieties of chlorite-, biotite-, amphibole-, epidote-, or muscovite-rich schists are present within the stratigraphic column. These schists contain more carbonate and sulfide, and less magnetite than the tourmaline-bearing mica schists. They occur in thin units less than 10 m thick and are most commonly interbedded with iron-formation, but some are also found within the arkosic rocks. The protoliths of these schists are questionable, although some sedimentary and igneous textures can be positively identified (Fig 14). These schists are grouped and described according to texture and mineralogy. The lateral continuity and thickness of these schists is hard to determine because of core skeletonization.

Quartz-Chlorite-Epidote Schist

Epidote-rich rocks occur throughout the stratigraphic section as thin, continuous to discontinuous units less than 10 M thick. In one case the rock shows a definite igneous texture, but in the others primary textures are vague or non-existent. The difference in mineralogy between those schists with relict igneous texture and those without is that the volume of epidote is much higher and feldspar is completely obliterated in the latter. Considering the similarities in unit thickness,
metamorphic texture, and mineralogy, all of these epidote- and chlorite-rich schists are believed to have had a similar protolith.

The typical epidote-chlorite schist is a well-foliated, epidote-green, calcareous schist with thin, discontinuous calcite-epidote veinlets parallel to foliation. Carbonate is abundant in the matrix, and the rock effervesces readily when HCL is applied. Overall texture and grain size of the least altered samples, combined with the stratigraphic occurrences (see cross-sections, Appendix A), suggest a flow or shallow semi-concordant, sill-like origin for this rock type.

Calcite-Bearing (Chlorite-, Biotite-, or Amphibole-) Schist

These schists, although of varied mineral assemblages, show textural and compositional similarities and are probably derived from a similar protolith. Streaky, green-gray to green-black and white, moderately well-foliated, fine- to medium-grained carbonate-bearing chlorite-, biotite-, or amphibole-schist comprise less than 5% of the total thickness. Grain size increases and color darkens with increasing amphibole content. A rip-up clast of amphibolite is present in a pinkish-colored feldspathic bed which is adjacent to an amphibolitic bed, providing evidence of a sedimentary origin (Fig. 6). The more chlorite-rich schists contain abundant intergranular carbonate, and thin carbonate veinlets also cross-cut the foliation.
Possible volcanic quartz grains, strongly zoned blocky feldspar grains, and rock fragments with both of these minerals are present in the least metamorphosed example of this rock type. This implies that the rock originally had an intermediate to mafic volcanogenic affinity. Geometrically arranged masses of chlorite, sphene, leucoxene, and opaque oxides are suggestive of pseudomorphs after a mafic mineral such as pyroxene or hornblende. These pseudomorphic aggregates are slightly
smeared into the foliation in a wavy fashion. Much of the chlorite, biotite, and amphibole is the probable result of the metamorphism of pelitic components in the original rock, such as clay and silt-sized matrix or fine-grained shale and volcanic rock fragments.
The original sedimentary texture of the arkosic rocks is still evident, but grain boundaries are typically vague due to strain and recrystallization. Interstitial areas between grains, when not composed of phyllosilicates, are composed of silt-sized, indistinct granoblastic aggregates of quartz and/or feldspar. In most cases this material is the result of recrystallization along grain boundaries and therefore not representative of matrix material. Thus, material counted as 'matrix' in the point count (Table 1) is partly an intimate mixture of strain-growth quartz and feldspar, in which positive identification is impossible. No framework quartz or feldspar finer than 0.03 mm (silt) or clay material was observed, and the percent of primary matrix is assumed to be represented by the proportion of phyllosilicates (Fig. 5). Using this premise, the arkoses are composed of 50 to 90% framework quartz and feldspar grains.

A weak cleavage is typically present, and a second oblique cleavage may be present. These latter cleavage fractures are lined with chlorite.

No systematic variation in modal percent feldspar and grain size was noted. Feldspar:quartz ratios vary from 1.5:1 to 9:1, and average approximately 3:1 in the studied sections.
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**Table 1. Continued**
Middle Precambrian Graywackes

- Lower Precambrian Graywackes
- Chlorite-sericite schist (103-250)
- Feldspathic arenite and arkose (103-350; 106-240; 107-270)

(After Morey and Schulz, unpublished paper).

Figure 7. Chemical fields of Lower vs. Middle Precambrian Graywackes.

Feldspar ranges in shape from strongly zoned, blocky and broken grains to subrounded grains with wormy quartz inclusions. Feldspar alteration to sericite is nearly ubiquitous, ranging from incipient crystal-controlled sericitization to heavily altered grains which are barely recognizable as feldspar. A microprobe check of feldspar in a tourmaline-rich rock (Table C-2, App. C, Anal. PL-19), gives a composition of An05; and refractive indices are consistently lower than quartz, also indicative of an albitic composition. The presence of wormy quartz inclusions, relict twinning and zonation, and
Figure 8. Photograph of tourmalinite laminae in matrix. Hole #105, 433-438' depth.

Figure 9. Photomicrograph of 0.5 mm-thick tourmalinite laminae, in center of photo. Tourmaline crystals are oriented diagonally; matrix consists of quartz and sericitized feldspar, with abundant epidote at left. Field = 3x2 mm; hole #105, 433-438' depth, one polar.
vague grain boundaries suggests albitization of a more calcic feldspar. This albitization may be the result of greenschist metamorphism, but the presence of sericite and local albite-chlorite assemblages indicate that hydrothermal alteration may be responsible for the high Na content of the rock (Table 1, App. D; Fig. 7). Five suspect thin sections were stained for K-feldspar, and at most only trace amounts exist. No K-feldspar is suspected in any other of the unstained thin sections.

Quartz is typically granoblastic, and larger grains are typically flattened and recrystallized parallel to foliation. The volume of quartz does not appear to be dependent upon grain size. In rare cases, quartz grains are weakly- or non-undulose, subequant, and recrystallized only along grain boundaries. Although border recrystallization has obliterated grain shapes, this non-undulose quartz is commonly associated with bi-modal fine-grained detritus which contain grains of strongly zoned oligoclase (?). This association indicates a possible volcanic source for the sediments.

Chlorite, biotite, and sericite are found throughout the arkosic rocks, primarily as recrystallized matrix which is interstitial to and penetrative into framework grains. Sericite is also common as a feldspar alteration product, and is the dominant phyllosilicate wherever tourmaline-rich laminae are present, a possible hydrothermal alteration product.
Epidote is present in over half of the counted thin sections. It commonly occurs with chlorite or biotite in granular masses of irregular, anhedral grains, and the volume of epidote increases as biotite and chlorite increase. Epidote rarely occurs in veins with quartz and carbonate, where it forms radiating, tabular crystals up to 1 cm in length.

Tourmaline is common as disseminated prismatic grains and as thin tourmalinite laminae. One 5 mm-wide tourmalinite lamina contains over 50% tourmaline (Figs. 8 & 9). The tourmaline carries significant implications and will be dealt with in a later section.

Magnetite and pyrite comprise 0 - 2% of the volume in these rocks. Thin magnetitic laminae are present, but rare. Other accessory minerals include sphene, calcite, apatite, and zircon.

Iron-Formation

Microscopically, the finely laminated nature of the iron-formation is readily visible. These laminae consist of 1 to 7 mm-thick layers of quartz-magnetite and intercalated clastic beds of similar thickness. The clastic layers represent dilution, by silt- to sand-sized terrigenous material, of steady chemical precipitation of silica and iron. The quartz-magnetite laminae are internally laminated, this being defined by variations in magnetite and quartz which are individually as thin as 0.05 mm (Fig. 10).
Figure 10. Photomicrograph of iron-formation showing quartz-magnetite laminae (left) and intercalated clastic laminae. Field = 3x2 mm, hole #103, 410' depth. One polar.

The intercalated clastic, quartzo-feldspathic laminae generally contain only a few percent opaque oxides (Table 1), but locally contain as much as 50% oxides. These clastic quartzofeldspathic layers are similar in appearance to other quartzofeldspathic, carbonate-chlorite schists found within the section (pp. 28 to 29).

Blocky magnetite and platy, well-foliated specular hematite are the dominant iron-bearing phases of the iron-formation. A separate point count in the quartz-magnetite laminae and the magnetite-bearing clastic
laminae gives respective volumes of 38 and 15% opaques (Table 1).

In the quartz-magnetite laminae, quartz grain size decreases as magnetite content increases; this quartz is presumably recrystallized chert. Small inclusions of chlorite, biotite, and magnetite are common within this quartz, and micaceous minerals commonly line the mortar-textured quartz grain boundaries. Secondary quartz also occurs in thin veins, typically with calcite; it is commonly strongly undulose and mortar-textured or granoblastic.

Albitic feldspar comprises 0 to 20% of the overall modal rock volume, but accounts for nearly 43% of the volume in some clastic laminae. The grains range from clean and colorless to turbid, the latter having undergone ragged alteration to chlorite and biotite. A general inverse relation exists whereby feldspar abundance increases as magnetite abundance decreases.

Epidote is most common in veins with chlorite or biotite, and along contacts between quartz-magnetite laminae and clastic laminae. It forms clean and golden-yellow, 6-sided euhedral crystals. Small amounts of fibrous green actinolite grains are also concentrated along clastic and quartz-magnetite grain boundaries. Other accessory minerals include granular sphene, apatite, and sericite. Some bedding-related apatite concentrations are present, but their origin remains
Figure 11. Photomicrograph of flattened and sericitized fragments in chloritic groundmass. Field = 3x2 mm, partially crossed polars. Hole #103, 250’ depth.

Figure 12. Photomicrograph of broken muscovite porphyroblasts. Note curvilinear trail of tourmaline and magnetite. Field = 3x2 mm. Hole #108, 350’ depth; one polar.
unclear. Only one sample contains tourmaline, in the form of dark grayish, semi-opaque skeletal grains. The darker color indicates a more iron-rich (schorl) composition than the tourmaline in other lithological units.

_Tourmaline- and Magnetite-bearing Micaeous Schist_

The upper unit of chlorite-sericite schist is well-foliated; this foliation is defined by phyllosilicates, tourmaline, and elongate quartz and feldspar grains. All sections also show a second foliation, defined by chlorite pods, which are at a right angle to the main foliation. Large muscovite porphyroblasts are abundant, quartz is recrystallized and granoblastic, and feldspars are moderately to heavily altered to sericite. Relict clasts, now composed of sericite plus quartz and epidote, are relatively abundant in this schist. These range from 2-30 mm in length, and are presumably altered felsic volcanic rock fragments which have been flattened into the foliation (Fig. 11).

Muscovite porphyroblasts, up to 6 mm, are mostly foliation-parallel, and those that are not have large quartz and chlorite-filled pressure shadows behind them. Some of the muscovite porphyroblasts are broken or bent, or extended parallel to foliation, and judging from the textures and curvilinear trails of included minerals (Fig. 12), the porphyroblasts are early syntectonic. Sericite is intimately intergrown with chlorite, and is the dominant mineral phase in the altered rock fragments.
Chlorite occurs in irregular strings and veins foliated parallel to cleavage, and also in sigmoidal-shaped pods in which the cleavage is roughly normal to foliation; the latter is often radially-oriented and suggestive of pressure shadow fillings. The presence of intergrown green biotite, and associated sphene wedges and magnetite scraps indicate that the chlorite is a retrograde metamorphic product of biotite (Heinrich, 1965; p. 302). Optical properties (e.g. anomalous violet interference colors) indicate that the chlorite is rich in Fe. Chlorite is also concentrated along late oblique fractures.

Quartz is patchily distributed, due to recrystallization, and a large proportion of it is confined to muscovite pressure shadows. Phyllosilicate minerals commonly occur between the composite grains, and control the shape of the quartz into elongate, foliation-parallel grains.

Albitic feldspar shows well-preserved to vague clastic grain shapes as alteration increases. Chlorite, biotite, and sericite grow interstitial to and penetrate into feldspar grains, the result of recrystallization of clay-sized matrix material.

Epidote is most common in chlorite-epidote veinlets or fracture healings and in the sericitized fragments. The epidote crystals are cored by plechoic green to brown allanite, or a clay pseudomorph thereof. Up to 6
Figure 13. Photomicrograph of tourmaline- and magnetite-bearing chlorite schist. Tourmaline is blue-green. Field = 3x2 mm; one polar. Hole #108, 365-370' depth.

Fig. 14. Photomicrograph of metabasalt showing well-preserved relict igneous texture. Field = 3x2 mm, crossed polars. Hole #108, 1147' depth.
overgrowth rims of epidote, which all show the same optical properties, are visible; these indicate that repetitive conditions favorable for crystal growth occurred, such as recurring pulses of hydrothermal fluids.

Magnetite is present as scattered octahedra and is also concentrated in thin laminae, in places associated with concentrations of tourmaline. Trace amounts of pyrite are common as euhedral crystals in the matrix or as anhedral aggregates in veins with quartz. Trace amounts of apatite may be present, and carbonate is lacking.

Trace amounts of tourmaline are ubiquitous in the schist. The crystals are foliated, pinkish-orange to deep bluish-green in color (pleochroic), and up to 6 mm in length. The crystals contain a dark core and thin, lighter green overgrowths (Fig. 13).

Quartz-Chlorite-Epidote_Schist

The least-altered schist of this type shows a good remnant hypidiomorphic-granular igneous texture, which is defined by fresh, tabular feldspar (Fig. 14). Judging by the grain size of the feldspar, the original rock was holocrystalline with a diabasic texture. The primary ferromagnesian minerals, presumably pyroxene and Fe-Ti oxides, are altered to a fine-grained aggregate of calcite, chlorite, epidote, and sphene. Calcite is present in irregular zones throughout the rock, and
within the feldspar grains as an alteration product. The irregular, veined habit of calcite suggests that it originated via low-grade hydrothermal metamorphism, but an overall low Ca value in a whole-rock analysis (Table D-1, App. D, Anal. 4;) suggests replacement of Ca by K.

Granoblastic aggregates of quartz are an abundant interstitial alteration by-product. It is dispersed throughout the rock, in a mosaic with chlorite, epidote, and sphene. Given the abundant plagioclase, Fe-Ti oxides, and ferromagnesian silicates, it seems unlikely for appreciable quantities of igneous quartz to be present. The mineral assemblage, coupled with the texture, suggests that quartz is a by-product of Fe-Mg silicate breakdown or that SiO₂ was introduced during alteration.

Plagioclase occurs as fresh, slightly fractured, tabular, zoned grains, for which optical properties (refractive index combined with Michel-Levy extinction angles) indicate a composition of An₉₈ (Albite). This observation is rather inconsistent with the whole rock analysis (App. D, Table 1, Anal. 4), which shows low Na values. This discrepancy may lie in interpretation of composition by optical properties, which could alternatively be interpreted to mean that the rims are albitized, but most of the crystal is of andesine composition.
Chlorite occurs in aggregates made up of smaller, randomly oriented bundles and sheaves, as a retrograde metamorphic product of Fe-Mg silicates. Opaque Fe-Ti oxides occur as irregular grains which texturally suggest a primary interstitial occurrence. The opaque grains are embayed or skeletal and ladder-shaped, indicative of primary igneous texture rather than the characteristic trellis pattern of ilmenite-magnetite exsolution (Heinrich, 1965, p. 55). Epidote occurs as tiny euhedral or granular crystals within feldspar grains and as larger euhedral crystals within the chlorite-rich areas; and tiny wedges of sphene occur with chlorite, clustered near opaque minerals as Fe-Ti oxide alteration products.

In the more heavily altered schist of this type, no recognizable igneous texture is present. Epidote is much more abundant, and commonly forms rectangular granular aggregates which are apparently pseudomorphs of feldspar crystals. Feldspar comprises 0 to 4% of the volume, and where present is sericitized and irregular in shape. Rare relict twinning is present in the feldspars, and chlorite-filled pressure shadows are common behind the feldspar grains. For modal analyses of these schistose metavolcanic rocks, see Table 1.

Calcite-bearing (Chlorite-, Biotite-, or Amphibole-) Schist

These chloritic to amphibolitic schists are well-foliated rocks which are of similar, but slightly
varying, bulk composition. They all contain rounded feldspar grains which have indistinct borders due to infringing micas and amphiboles. The feldspars contain altered, biotite- or carbonate-bearing cores and clean, colorless perimeters, and where inclusion-free are difficult to distinguish from quartz. Some grains are blocky in shape and show a perthitic-type of texture, presumably from incomplete albitization.

The modal content of chlorite decreases as amphibole content increases. The chlorite folia bend around feldspar grains, resulting in a wavy appearance. Pods of chlorite with perpendicular foliation are present, found in pressure shadows next to sulfide cubes. Biotite is texturally similar to chlorite, and contains euhedral epidote, minor sphene, and opaque inclusions. It is also concentrated into a series of oblique fracture cleavages. Amphibole (actinolite) is strongly pleochroic (yellow-green to deep green) and well-foliated. Incipient alteration to biotite, epidote, and sphene is common in the actinolite, especially along crystal fractures.

Carbonate (calcite) is most abundant in the chlorite-rich schists, and decreases in volume as metamorphic grade increases. Calcite occurs in irregular masses and in thin veins, and shows a foliation-parallel elongate habit. The veined and irregular habit suggests that carbonate was introduced to the rock.
Quartz is mostly recrystallized, but in one case some quartz grains up to 1 mm in diameter are present which are non-undulose and are possibly volcanic in origin. Long needle-like inclusions, probably rutile, are locally present within the quartz.

Opaque minerals consist of oxide by-products of alteration and also of blocky pyrite, which contains inclusions of epidote and biotite. Granular- and wedge-shaped sphene is associated with chlorite, and is weakly to heavily altered to leucoxene. Epidote is concentrated in biotite and chlorite. It is mostly euhedral, and forms overgrowths over allanite, which is metamict and has destroyed the neighboring epidote to an altered and punky material.
The following general metamorphic mineral assemblages are present in the bedded metasedimentary-metavolcanic rock sequence near Philbrook:

Meta-arkose:
- quartz-albite-chlorite-biotite-epidote-actinolite

'Cherty' iron-formation:
- quartz-magnetite-hematite-chlorite-biotite-epidote-actinolite

Upper schist:
- quartz-albite-chlorite-biotite-muscovite-magnetite-tourmaline-
  epidote-sphene

Meta-silicic volcanic:
- quartz-chlorite-ralsite-epidote-magnetite-jennite-sphene-
  (albitic feldspar)

Conglomerates:
- quartz-albite-biotite-actinolite-chlorite-sphene-epidote

These assemblages are indicative of metamorphism under conditions of upper greenschist or lower amphibolite facies (Turner, 1968).

Chemical Analyses of Metasedimentary Rocks

Chemical data on the arkosic rocks (Table D-2, App. D, Anal. 1-3) show high \( \text{Na}_2\text{O}/(\text{CaO} + \text{K}_2\text{O}) \) ratios, reflective of the high content of albite in the rock. The low Fe and Mg values are a consequence of the lack of Fe-Ti oxides and silicates. It is unlikely for any potential source rocks to contribute such an albitic feldspar, and if the metamorphic system were closed, one would expect to see Ca-bearing mineral phases such as zoisite or calcite in the assemblage. An open system
reaction, such as sea-floor alteration, is suggested whereby Na\textsuperscript{+} was exchanged for Ca\textsuperscript{+} in a circulating fluid.

The high K\textsubscript{2}O content of PB-103-250 (Table D-2, App. D, Anal. 5; upper magnetite-chlorite schist with sericitized rock fragments) is tied up in the muscovite and biotite in the rock. This enrichment in K is consistent with the observation that sea-floor metamorphism commonly enriches basalt in K (Drever, 1983).

An unpublished paper by Morey and Schulz (pers. comm., 1986) shows distinctive groupings of Archean vs. Proterozoic graywackes on a SiO\textsubscript{2}-K\textsubscript{2}O-Na\textsubscript{2}O ternary diagram (Fig. 7). The arkoses from this study most closely match the Archean graywackes because of the high Na content, but the chloritic schist, which is probably the closest of the analyzed samples to a graywacke protolith, best matches the Proterozoic graywackes.

Using the fields of Irvine and Baragar (1971), sample PB-108-1147 (metabasalt) falls into the tholeiitic field on an AFM diagram, and into the sub-alkaline field on a SiO\textsubscript{2} vs (K\textsubscript{2}O + Na\textsubscript{2}O) plot.
SIGNIFICANCE OF TOURMALINE-BEARING ROCKS

History of Tourmalinites

Tourmaline is a common mineral which occurs in a variety of geological environments and may be of sedimentary, metamorphic, hydrothermal, or igneous origin. However, the significance of bedded, stratiform, tourmaline-rich rocks (tourmalinites) has been recognized only within the past ten years. The association of tourmalinites with stratiform sulfide deposits raise some interesting questions regarding the tourmaline-rich metasedimentary rocks of this study and therefore a short summary of tourmalinites is pertinent.

Tourmalinites have been defined as "...stratabound lithological units concordant with their enclosing host rocks and containing 15 to 20% or more tourmaline by volume." (Slack et al., 1984).

Tourmalinites typically may be internally laminated, and well-preserved sedimentary features in tourmalinites include fine-scale (<1mm) tourmaline-rich laminae, graded bedding, slump and flame structures, cross-bedding, and rip-up clasts of tourmalinite in a tourmaline-free matrix (Slack et al., 1984; Appel, 1984; Ethier and Campbell, 1977).

Tourmalinites are commonly associated with strata-bound massive-sulfide deposits containing tungsten, tin, gold, cobalt, copper, lead, and zinc (Fleischer and Routhier, 1973; Plimer, 1980; Ethier and Campbell, 1977.)
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<td>Woodcutters Prospect (Aus.)</td>
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(Above references taken from Taylor and Slack, 1984; Fleischer and Routhier, 1977; Ethier and Campbell, 1977).

South Carolina | Iron-formation, pelitic schist, quartzite, felsic volcaniclastic sandstone | Mittwede, S.K., 1984 |

New York | Feldspatic quartzite, metapelite, tremolitic schist | Brown and Ayuso, 1985 |

Table 2. Selected references to tourmalinite occurrences.
Several authors (Ethier and Campbell, 1977; Slack et al., 1984) have reported that tourmalinite has been misidentified as chert, amphibolite, or as a light-colored, fine-grained, argillaceous metasedimentary rock in which the tourmaline was not recognized at all due to its fine grain size.

Tourmaline Composition and Association as Described in the Literature

Tourmalinites can be associated with a variety of sedimentary and volcanic rocks, but they most commonly occur in Proterozoic or early Paleozoic clastic argillite or sandstone sequences (Table 2). Tourmalinite may also be locally present with mafic metavolcanic rocks, iron- or manganese-rich sediments (oxide-facies iron-formation), silicic volcanic rocks, and calc-silicate schists (Slack et al., 1984). The tourmaline typically occurs with quartz, and locally with chlorite, muscovite, biotite, albite, magnetite, spessartine garnet, and pyrite (Ethier and Campbell, 1977; Slack et al., 1984; Fleischer and Routhier, 1973; Appel, 1984). The abundance of boron appears to be the limiting factor on the genesis of tourmaline in any sedimentary environment. Because tourmaline typically is intercalated with laminated iron-formation, manganese-rich coticules, stratiform sulfides, and volcanogenic detritus it seems likely that the boron concentration is related to submarine volcanogenic-exhalative processes. Boron also can
accumulate in hypersaline, sabkha-type depositional environments (Appel, 1984; Slack et al., 1984; Ethier and Campbell, 1977; Taylor and Slack, 1984.) In either case, anomalously high boron values are present during sedimentation.

The lack of any associated heavy minerals such as zircon and rutile, and the fact that tourmaline-rich laminae sometimes occupy the normal position of the clay fraction within bedding features, precludes a placer-type heavy mineral deposit origin for tourmalinites.

Tourmaline Crystallization in Tourmalinites

The exact timing and process of tourmaline crystallization is still somewhat unclear. Based on textural relationships, the presence of chemical zones within tourmaline, and mineralogical associations, Taylor and Slack (1984) have suggested a premetamorphic origin for the tourmaline. Tourmaline growth is suggested to be initially syndepositional or diagenetic, from a boron-rich fluid, with later metamorphic modification (Ethier and Campbell, 1977; Taylor and Slack, 1984). Authigenic tourmaline (dravite) has been documented in an unmetamorphosed feldspathic lithic arenite in Montana, where it occurs as discrete euhedral crystals growing into pore spaces (Gautier, 1979).

An oxygen isotope study of tourmalinite from the Sullivan Mine, British Columbia, has demonstrated that boron was derived from connate fluids in sediments which
were probably deposited in a mildly restricted basin (Nesbitt et al., 1984). During sedimentation, loading and dewatering forced low temperature (<110°C) boron-bearing fluids up along at least 12 different conduits; these brines were vented and deposited at the sea floor, and also concentrated in permeable beds adjacent to the pipes. Subsequent to this, hotter (150°C) Pb-Zn-Fe bearing fluids, derived from a deeper source, rose through the same zones of weakness and deposited stratiform sulfides. A final hydrothermal event altered the tourmalinite, ore, and host sedimentary sequences of the hanging wall to an albite-chlorite assemblage; the fluids for this event were around 250°C. This low- to high-temperature alteration is similar to other successions of syngenetic sediment-hosted ore deposits. A deep fracture system is necessary to supply the heat to circulate the fluids and dissolved elements for the alteration and ore-forming processes; at Sullivan these are thought to be synsedimentary faults. Nesbitt et al., (1984) have pointed out the use of oxygen isotope studies as an exploration tool for locating hydrothermal deposits of this type.
Tourmaline accounts for as much as 2% of the volume in these schists, which comprise the upper 10% of the drilled stratigraphy (see page 27). It is typically present only in trace amounts, but is ubiquitous (see Table 1). It is strongly pleochroic, brownish-orange to deep blue-green, and occurs as euhedral to subhedral, pencil-shaped hexagonal prisms. Darker bluish-green cores can be seen in cross-sections and length-sections (Fig. 13). Small inclusions of opaque material, presumably magnetite, and colorless quartz or liquid-filled vacuoles (?) are common in both cores and rims.

Textural evidence, such as the relatively overall coarse grain size of the associated minerals and muscovite poikiloblasts indicate that recrystallization and metamorphism have significantly altered the textures of the rock. The relatively large grain size, stout and blocky habit, and magnetite and quartz inclusions within tourmaline provides evidence that it may be of metamorphic origin. Further evidence of this is that the tourmaline which is included within muscovite poikiloblasts is of a distinctly smaller grain size than independent tourmaline crystals. The darker-colored cores may be the vestige of premetamorphic tourmaline, or may simply be the result of slight compositional variations in metamorphic fluids during crystal growth.
Tourmaline is relatively homogeneously distributed throughout the schist, but some apparent concentrations of magnetite and tourmaline exist within the same laminae (Fig. 13). This implies that chemical deposition of iron and boron occurred synchronously and an exhalative source is a likely progenitor for both minerals.

Tourmaline in Feldspathic Arenite

A second type of tourmaline, distinct from that just described, is also present near Philbrook. The host for this variety is a highly feldspathic sericite-quartz schist which is magnetite-free (Figs. 8 & 9). The tourmaline, which occurs in amounts as great as 15%, ranges in color from a blue-green variety, similar to that in the micaceous schist just described, to a distinctly more brown variety which is pleochroic from orange-brown to a very dark green, semi-opaque color. The brown variety is more magnesian (closer to the dravite end-member) than the blue-green variety, and commonly forms a radial growth habit. The crystals lack the distinctive dark green core such as those in the magnetite-bearing schist. The blue-green variety lacks any noticeable overgrowths, but the brown variety has prominent, lighter colored overgrowths which preferentially form caps on crystal termini. Small colorless inclusions of quartz or fluid (?) are common in the blue-green variety and abundant in the brown type. The chemical attributes of both color types will be discussed in a later section.
The tourmaline from both schist and arenite form euhedral to subhedral hexagonal prisms which are a maximum of 1 mm in length and average approximately 0.3 mm in diameter. The tourmaline crystals are commonly foliated and slightly cross-fractured, but not significantly bent or deformed. In one case long, stringy, bent tourmaline crystals are present within a quartz vein; forming suns of acicular crystals which nucleate from brown tourmaline in contact with the quartz vein. This vein tourmaline is apparently hydrothermal in origin.

**Tourmalinite Bedding Features in the Arenite Rock**

0.5 to 1.5 mm thick tourmalinite laminae, some with sharp contacts (Fig. 9) and others with gradational, diffuse contacts, indicate a series of fine-scale repetitions of bedding, but no topping directions or rhythmic size variations are observable. In one case, a tourmaline-rich lamination is within a fine- to medium-grained quartzofeldspathic bed which is abruptly overlain (?) by a thick, coarse-grained quartzofeldspathic bed. This internal structure suggests that the distribution of tourmaline was influenced by primary sedimentological parameters, regardless of the timing of boron concentration and tourmaline crystal growth. However, non-laminated occurrences also are present in which tourmaline prisms are diffusely scattered throughout the rock; these occurrences suggest a hydrothermal origin for the tourmaline.
Alteration of the Arenitic Host Rock

The feldspar in the meta-arkose (as in the tourmaline-bearing mica schists described earlier) is variably sericitized, ranging from grains that are relatively fresh to those completely replaced by sericite. Where the grains are sufficiently unaltered and can be identified, the volume of feldspar in the arkoses is the same as or greater than that of quartz. Other minerals present include minor to abundant yellow-green epidote; and chlorite and green biotite are generally present in small amounts.

Chemical Analyses of Tourmaline

A concentrated sample of the blue-green tourmaline from the tourmaline-magnetite-mica schist was obtained for chemical analysis by: 1) crushing to a size small enough to go through a #100 sieve; 2) heavy liquid separation; 3) further concentration via use of a Frantz magnetic separator (limited success); and 4) hand picking under a binocular microscope. The final sample was quite pure, as the crystals were small enough and transparent enough to visually screen those with impurities. Quantitative chemical analysis (App. C, Table 1, analysis # TM-1-M) was done by R. L. Knoche, University of Minnesota geochemical laboratories. Due to time limitations, only one analysis of this kind was performed.
Additional major element data were obtained at the University of Wisconsin, Madison by electron microprobe techniques, using fifteen points; 6 on the blue-green type of tourmaline previously analysed chemically, and 9 on the brown-green type of tourmaline from the meta-arkose. Although oxide weight percents vary between samples, several rechecks of points in the tourmaline and in the standards showed consistent results. The variation in $\text{Al}_2\text{O}_3$ and $\text{SiO}_2$ (App. C, Table 1) is ascribed to chemical variations between samples, particularly in the case of Tour-1-M, for which the wet chemical analysis represent an average composition and the probe data represents local variations in composition.

The tourmaline from the arkosic rock shows markedly lower Al, Na, and Si content than tourmaline from the chloritic schist, and the Mg, Fe, and Ca values show corresponding increases. Although some authors (Ethier and Campbell, 1977) have reported problems with Al and Si values from the electron microprobe due to burnoff of volatiles, the values reported here are believed to be relatively accurate. To offset problems such as those discussed by Ethier and Campbell (1977), ratios of weight percents of oxides, such as $\text{Na}_2\text{O}/(\text{Na}_2\text{O} + \text{K}_2\text{O})$ and $\text{FeO}/(\text{FeO} + \text{MgO})$ are useful (Ethier and Campbell, 1977; Taylor and Slack, 1984).

Variation between core and rim pairs show consistent trends between samples. The blue-green tourmaline from
A.) Ca-Fe-Mo diagram for tourmaline rim and core compositions from Philbrook. The fields correspond to those in (B>). Fe is total iron, reported as FeO. Modified from Henry and Guidotti, 1985.

B.) Al-Fe-Mg diagram for tourmaline compositions from Philbrook. The fields represent: 1) Li-rich granitic pegmatites and aplites; 2) Li-poor granitic pegmatites and aplites; 3) Fe$^{3+}$-rich quartz-tourmaline rocks (hydrothermally altered granites); 4) Metapelites and metapsammites coexisting with an Al-saturating phase; 5) metapelites and metapsammites not coexisting with an Al-saturating phase; 6) Fe$^{3+}$-rich quartz-tourmaline rocks, calc-silicate rocks, and metapelites; 7) low-Ca metaultramafics and Cr, V-rich metasediments; and 8) metacarbonates and metabasaltic rocks. Modified from Henry and Guidotti, 1985.
the magnetite-bearing rocks have different absolute values than those in the brownish tourmaline from the quartz-albite rock, but the relative trends are similar. Most notable are the consistent increases in Fe and Mg and decrease in Al from core to rim. When classified according to the scheme of Henry and Guidotti (1985, as shown in Fig. 15), it is shown that both cores and rims are intermediate between dravite and schorl, with an additional uvite component in the brownish tourmalines. The rims show the strongest tendency towards the uvite end-member.

Chemical Signatures of Tourmalinites Associated with Massive-Sulfides

Chemical and isotopic $^{18}$O and $^{13}$C studies (p. 56) show tourmaline associated with massive-sulfide deposits to be distinct from tourmaline of igneous derivation, especially with regard to Mg/Fe ratios (Taylor and Slack, 1984; Ethier and Campbell, 1977; Appel, 1984). Stratiform tourmalines are typically Mg-rich (schorl-dravite to dravite) in contrast to more iron-rich igneous schorl. In addition, several other characteristic major and minor element signatures are present, and are summarized in Table 3.

The Mg/Fe ratio from the average Philbrook tourmaline is slightly lower than the average massive-sulfide tourmaline, but still within the range of values, as shown in Fig. 16 (Taylor and Slack, 1984). Alkali ratios are
Massive Sulfide Associated

<table>
<thead>
<tr>
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<th>Philbrook</th>
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<tbody>
<tr>
<td>FeO/(FeO + MgO)</td>
<td>0.06 - 0.4</td>
</tr>
<tr>
<td>ave 0.21</td>
<td>0.51 (mica schist, probe)</td>
</tr>
<tr>
<td>Mg &gt; Fe</td>
<td>0.59 (arkose, probe)</td>
</tr>
<tr>
<td>Na₂O/(Na₂O + K₂O)</td>
<td>0.31 - 0.91</td>
</tr>
<tr>
<td>ave 0.66</td>
<td>0.72 (mica schist, probe)</td>
</tr>
<tr>
<td>Na &gt; Ca</td>
<td>0.58 (arkose, probe)</td>
</tr>
<tr>
<td></td>
<td>0.66 (mica schist, chem)</td>
</tr>
<tr>
<td>High in: Cr</td>
<td>778 ppm</td>
</tr>
<tr>
<td></td>
<td>193 ppm</td>
</tr>
<tr>
<td>Cu</td>
<td>77 ppm</td>
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<tr>
<td></td>
<td>302 ppm</td>
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<tr>
<td>V</td>
<td>195 ppm</td>
</tr>
<tr>
<td></td>
<td>182 ppm</td>
</tr>
<tr>
<td>Low in: Mn</td>
<td>382 ppm</td>
</tr>
<tr>
<td></td>
<td>179 ppm</td>
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</table>

Table 3. Chemical characteristics of tourmaline associated with massive sulfide deposits.

also comparable, and plot close to the cluster of analyses from the Sullivan Mine massive sulfide deposit (Fig. 16). Several authors have stressed the possibility of using tourmalinite as an exploration tool for exhalative massive-sulfide deposits, because the distinct geochemical and textural characteristics are easily recognized, but have often been overlooked.

**Exploration Potential Near Philbrook**

Numerous occurrences of tourmalinite exist in the metasedimentary rocks near Philbrook. The chemical signatures of this tourmaline are similar to tourmaline associated with massive-sulfide deposits, and indicate economic potential. Hydrothermal alteration is indicated by sericitized albite and quartz-epidote-carbonate veins; and some of the tourmaline (e.g. the radial-growth
Il-Greenish-brown tourmaline from Sullivan massive-sulfide host rocks.
Ill-Amber tourmaline associated with Sullivan massive-sulfide ore.

Probe chemistry of tourmaline from trem-bamite schist
Probe chemistry of tourmaline from arkose rocks.
Oxidation chemical analysis of tourmaline from trem-bamite schist

![Diagram](image)

**Figure 16.** (Top) Compositions of tourmaline in terms of oxide ratios. Modified from Ethier and Campbell (1977).

(Bottom) Variation diagram of Philbrook tourmaline vs. massive-sulfide-related tourmaline.
tourmaline) may be of hydrothermal or metamorphic-growth origin.

Whether the tourmaline is authigenic, diagenetic, or metamorphic in origin, its presence results from concentrated boron within a given depositional interval. The thinly laminated occurrence of tourmalinite within the arenitic rocks suggests either sea-floor deposition distal to an active boron exhalative vent, or secondary boron concentration within permeable beds. Periodic pulses of boron-rich fluid may have emanated from a source, and spread across the seafloor as density flows, to be included within a thin sedimentary interval and later incorporated into tourmaline. Possible mechanisms of boron concentration have been discussed (p. 54-56), and if a variation of the exhalative model is valid, one would expect sulfide occurrences to be located either across or along strike in relation to the stratigraphic section of this thesis. The presence of sulfides in the skeletonized core is scant, but a complete drill core obtained in a different stratigraphic position within the section might yield more sulfides and good tourmalinite examples. Due to its high conductivity and magnetic susceptibility, the presence of iron-formation may act as a shield to a ground geophysical survey, but a likely target for a sulfide deposit would be a non-magnetic conductor. Isotopic, geochemical, and mineralogical
studies, based on proven examples such as the isotope study of Nesbitt et al. (1984) could provide a basis from which to evaluate exploration progress.
CONGLOMERATE AND TONALITE

Several small, elongate outcrops of conglomerate and hypabyssal igneous rocks are located east of Philbrook in Sec. 34, T. 133 N., R.32 W., along the Long Prairie River (Fig. 5).

One type of conglomerate is a dark green, well-foliated biotite schist with 2 to 15 cm long stretched clasts of tonalite and fine-grained, green metavolcanic rock fragments (Fig. 17). A few meters from this is a porphyritic, hypabyssal, fine-grained tonalite which is mineralogically similar to the tonalitic clasts in the schist. Approximately 300 meters to the NW of these outcrops, across the Long Prairie River, are a series of quartz-pebble conglomerate outcrops which form low ridges in the river and a 1 m vertical cliff along the north river bank. The maximum exposed thickness is less than 5 meters.

Quartz-rich metaconglomerates

These outcrops are rounded and coated with a red river varnish on the outer surface. On a fresh surface 0.5 to 15 cm angular to rounded clasts of white quartzite, light green metagraywacke, pelite, and volcanic fragments are visible in a light green, vitreous matrix; the conglomerate is clast-supported. The outcrops are cherty in appearance and hardness and have a hackly fracture cleavage which trends N-NW. Exposure is too poor to ascertain any systematic bedding variations, but from
FiConglomerate outcrop in NW1/4, NE1/4, Sec. 34. Clasts are dominantly tonalite, with minor volcanic fragments. Note stretched clasts above and to right of field book.

Figure 17.

Photomicrograph of quartz-rich conglomerate from north of Long Prairie River; NE 1/4, NW 1/4, Sec. 34. Both poly- and monocrystalline quartz grains are visible, in a fine-grained matrix. Field = 3x2 mm; partially crossed polars.

Figure 18.
what can be seen, no size grading or other features are evident.

Two thin sections from this outcrop were examined. One is largely within a single clast of intermediate vesicular metavolcanic rock; the other is cut so as to cover a range of clast size and lithology.

The metavolcanic clasts are recrystallized, partly amygdaloidal, and flow-banded, with a groundmass of fine-grained granular sphene (4%), epidote (9%), blocky magnetite (3%), decussate-textured fibrous actinolite (34%), and quartz plus feldspar (50%). Flow banding in the clasts is defined by volumetric variations of magnetite, epidote, and sphene. The even grain size and decussate texture of actinolite indicate that the rock was initially glassy or fine-grained and non-porphyritic.

The rock near the amygdules shows a narrow (<1mm) chill which supplies evidence that it was not glassy, as there is a variation in color and mineralogy between the rim and the groundmass. Amygdules are filled with ragged masses of actinolite. Quartz and feldspar are present as very fine-grained matrix around the ferromagnesian minerals. Staining for potassium reveals that K-feldspar is more abundant than plagioclase, leading to an andesitic composition.

Minor amounts of coarse-grained granitic rock fragments are also present, and are composed of microperthite, quartz, and biotite. The boundaries of
these fragments are vague due to penetration of actinolite from the surrounding groundmass. These granitic fragments are dissimilar from the tonalite fragments in the conglomerate to the southeast, as they do not contain plagioclase. Small amounts of pyrite occur between the broken vesicular fragments.

The other thin section from this outcrop (Fig. 18) contains 4 or 5 different types of clasts. The larger clasts (up to 2 cm in thin section) of quartzite are subangular and contain well to moderately sorted grains (clast type 1). The cleanest of these consists of weakly undulose quartz grains, 0.3 to 1 mm diameter, with sutured grain boundaries. A small amount of silt and clay (stains yellow) is present in the interstitial areas. Thin needle-like actinolite crystals emanate from the interstitial silty areas into the quartz grains. A minor amount of perthitic feldspar is present in the quartzite.

Although the primary grain shapes have been somewhat modified by pressure solution, the coarse size, high degree of sorting, and apparent roundness of the individual grains within the clasts indicates the quartzite clasts were derived from a clean and well-sorted sandstone. Other quartz-rich clasts contain approximately 10% silt and clay matrix, and gradations into recrystallized mudstone with scattered quartz grains are present; these are probably metagraywacke (second clast type).
Some of the quartz grains are rutileated, and rare zircons are scattered throughout the quartzose rocks.

A third type of clast consists of an Al-rich pelitic metasedimentary rock now composed of chlorite, chloritoid, and epidote in a very fine-grained quartz and feldspar matrix. The chloritoid forms decussate-textured, tabular porphyroblasts which are ≤1 mm in length and have low first order birefringence. Chlorite occurs pseudomorphically after chloritoid, and forms light green patches with inclusions of fine needly actinolite. The chlorite all shows Berlin-blue to gray interference colors.

A fourth type of clast is a fine-grained metamorphosed siltstone/mudstone. It consists of a very fine-grained granoblastic-recrystallized aggregate of quartz plus feldspar with abundant fine, granular epidote, chlorite, and minor actinolite. It is of an even grain size and relatively low in ferromagnesian mineral content.

The fifth type of clast is composed of actinolite similar to that in the vesicle infillings of the flow-banded andesitic rock. The coarser-grained actinolitic cores of the clasts are rimmed by fine-grained, fibrous, felted masses of actinolite and granular quartz-feldspar.
Schistose Tonalite-Bearing Conglomerate and Tonalite

The rocks south of the Long Prairie River (Fig. 5) are biotite schists, two of which are conglomerates rich in clasts of coarse-grained, intermediate plutonic rocks (exposed thickness less than 3 m); the other is a metamorphosed porphyritic, tonalite dike rock (exposed thickness less than 1 meter). The conglomerates contain abundant (possibly clast-supported) clasts of coarse-grained tourmaline-bearing tonalite, with graphic quartz/feldspar intergrowths. All of the plutonic fragments contain coarse-grained, red-brown biotite which is similar in color to the matrix biotite, but is coarser-grained and contains abundant exsolved rods of rutile.

Figure 19. Photomicrograph of conglomerate in Fig. 17. Note zoned plagioclase at left of photo and graphic quartz/feldspar intergrowth at lower right. Field = 3x2 mm; crossed nicols.
Sphene in the tonalite clasts is of secondary origin and forms worm-like rounded masses which have small, irregular oxide cores, apparently ilmenite. Skeletal blue-green schorl is also present within the clasts.

The matrix of the conglomerates is composed of semigranoblastic aggregates of quartz with various sizes and types of feldspar, plus sphene, chlorite, epidote, and biotite. The matrix biotite is red-brown and comprised of aggregates or small books concentrated into fracture-cleavage planes which anastomose between clasts (Fig. 19). Pleochroic haloes are common in both the igneous and metamorphic biotite, and most of the feldspar fragments are zoned (Fig. 19). Rare green hornblende is present in the matrix. Only one fine-grained metavolcanic fragment was seen in thin section, but hand samples indicate a higher percentage. This fragment is apparently a mafic metavolcanic fragment, composed of sphene, opaque iron-oxides, epidote, chlorite, and quartz plus feldspar.

A thin section of the hypabyssal tonalite intrusive rock shows both perthite and twinned plagioclase feldspar, and staining of the slide reveals no K-feldspar. The plagioclase (oligoclase to andesine) forms blocky, tabular, well-twinned and normally zoned phenocrysts up to 1 cm in length; the cores are preferentially altered to epidote, biotite, and chlorite. The perthitic feldspar phenocrysts are equant, blocky to rounded, and less
altered than the plagioclase; alteration is in the form of a fine-grained reddish product, presumably kaolinite with hematite. The groundmass is composed mostly of plagioclase feldspar, recrystallized granoblastic quartz, abundant flakes of red-brown biotite, granular sphene rimming opaque oxide scraps, epidote concentrated into fractures and with biotite, and trace amounts of chlorite.

A small inclusion is present, comprised of 20-30% lath-shaped plagioclase in a biotite-rich, low quartz matrix, and is probably a piece of wall rock incorporated into the magma.

Outcrop relations and the microscopic characteristics of the tonalitic porphyry indicates that it is of intrusive origin, perhaps a chilled margin of a larger intrusive body, or a synvolcanic dike or sill. The mineralogy is quite similar to the clasts in the conglomerate located a few feet away, giving evidence that the conglomerate is very local in source and basal in nature.
PROVENANCE AND SEDIMENTATION

Conglomerates:

The marked variation in lithology between the quartz-rich conglomerates and the tonalitic conglomerates, and their relations to the tonalite suggests the conglomerates are of local source and basal in nature. The texture and mineralogy of the tonalite porphyry and conglomerate clasts are similar to that in an exposure of tonalite near Staples (Southwick, pers. comm., 1986; Morey, 1978; Winchell et al., 1984), and to tonalitic gneisses intersected by drilling 10 km to the west of Philbrook (Southwick, pers. comm., 1986). The Staples outcrop has been dated at 2,600 ma (Z.E. Peterman, pers. comm., 1986), and is Archean in age. Thus, evidence points to the conglomerates as being the basal part of a Proterozoic sequence, developed directly on Archean rocks. This conglomerate could represent the base of either the Mille Lacs Group or the Animikie Group. The exposures are lithologically similar to the lower conglomerates in the Denham Formation, which developed in a fluvial to shallow marine environment directly on the Archean crustal rocks, in the initial stages of rifting of the Animikie basin (Morey, 1986). The Denham Formation is stratigraphically equivalent to the Randall Formation, and outcrops of mafic metavolcanics and conglomerates in the Randall Formation are located approximately 30 km to the SW of Philbrook.
It is doubtful that the conglomerates near Philbrook are directly correlative with those of the Denham Formation, which crop out approximately 100 km to the east, but the Philbrook conglomerates occupy a similar position in the stratigraphy, and are believed to be similarly developed on the Archean crust. Therefore, the Philbrook conglomerates may be correlative with the Randall, Glen Township, and Denham Formations. However, regional correlation (Southwick, pers. comm., 1986) indicates that the conglomerates might represent the westernmost basal extension of the overlying Animikie Group, developed on Archean basement rocks.

Metasedimentary Rocks West of Philbrook

The iron-formation package west of Philbrook is also believed to be part of the Mille Lacs Group rather than the stratigraphically higher Animikie Group. The best Animikie correlation would be with the Rabbit Lake-Thomson Formations, thick sequences of gray and black argillite, graywacke, iron-formation, and ferruginous slates. The Rabbit Lake is divided into three members, the lower which is carbonaceous slate and volcanic rock, the middle which contains iron-formation, and the upper which consists of slates, argillites, and graywackes. The rocks of this study are highly feldspathic, in close proximity to Archean crustal rocks, lack appreciable slaty components, do not have granular-textured iron-formation or stromatolites, and are not carbonaceous; and
therefore are believed to be part of the Proterozoic Mille Lacs Group.

Because of the limited data, only a general interpretation of the sedimentary environment can be put forth. The stratigraphy suggests that two distinct cycles of sedimentation took place. One cycle consists of the lower arkose and iron-formation; the other consists of the upper arkose and overlying metagraywacke/iron-formation. It cannot be assumed that the lower arkose represents the base of a depositional cycle, nor is it implied. It is merely stated that the variations in rock type represent at least two periods of changing sedimentation styles in the transitions from arkose to iron-formation and from arkose to metagraywacke (mica schist).

A depositional model similar to that of Morey (1983) for the Mille Lacs Group is useful (Fig. 20). Attempts to correlate the iron-formation package with the conglomerates are necessarily conjectural, and regional lithological associations indicate that the conglomerates at Philbrook may be younger than the iron-formation package at Philbrook (Southwick, pers. comm., 1986). In Fig. 20, it is assumed that the conglomerates are correlative to the Denham Formation, and therefore underlie the iron-formation package.
Lower Arkose and Feldspathic Arenite

High relief of the source area and rapid mechanical weathering are indicated by the high plagioclase content. The lack of orthoclase feldspar indicates a tonalitic source from the greenstone-granite terrane to the north rather than from the presumably gneissic terrane of the southern crust. However, the predominance of albite and the high Na₂O / (K₂O + CaO) ratio (Table D-2, App. D) of the rock indicates that it has been altered, and that the Na has replaced Ca and possibly K. A source area consisting entirely of albite is unlikely, and the rocks have undergone hydrothermal and/or greenschist-facies metasomatism. Hydrothermal alteration is indicated by non-stratiform occurrences of radial-growth tourmaline in conjunction with highly sericitized feldspar.

Thin stratiform tourmalinite laminae and rare magnetite laminae indicate one or all of three possibilities: 1) that these arkosic sediments were either deposited in a deep-water environment by turbidity-type currents, with occasional periods of quiescence and chemical deposition; 2) that periodic boron-rich density flows (see tourmalinite chapter) spread across the sea-floor and became incorporated into the accumulating sediments; or 3) that variations in permeability between adjacent beds allowed boron to percolate into the arkose, possibly in a soft-sediment state. In the first two cases, the environment at the time of boron deposition
must have been below wave base, to preserve the delicate concentrations of boron and iron.

**Iron-Formation**

The main thickness of iron-formation represents a period of reduced clastic input, presumably during a period of crustal stability. The iron-formation contains variable amounts of detrital material, grading from clastic-contaminated iron-formation to a clastic-dominated rock with magnetite-bearing laminae. Two separate accounts of Archean iron-formation/arkose/graywacke/shale sequences show many analogies to the Philbrook iron-formation sequence and provide supplementary clues to sedimentation. In these reports (Meyn and Palonen, 1980; Barrett and Fralick, 1985) well-documented sequences of graded sandstone-graywacke/iron-formation/shale beds are believed to be of submarine fan-type turbidite origin. In both cases the iron-formation is intimately associated with feldspar-rich clastic sequences, and magnetite concentrations are upward graded within individual Bouma beds. Analogies between Philbrook and these sequences are made for the purpose of sedimentary interpretation and not for age establishment. These papers (Barrett & Fralick, 1985; Meyn and Palonen, 1980) show that a turbidity current mechanism may have operated to produce the contamination of chemical sediments by clastic input. This reinforces the idea that sedimentation was occurring in a mildly
subsiding basin, even during the time of iron deposition, but also brings up the point that deposition of the Philbrook strata may have involved a shifting submarine-fan environment. In this highly speculative scenario, the arkoses might represent midfan facies, and the iron-formation a distal facies of a submarine fan. The predominance of oxide-facies iron-formation may indicate deposition in relatively shallow, oxygenated seas (James, 1954); but Walker (1979), has shown that this isn’t the case in a turbidite environment.

**Upper Arkose and Metagraywacke**

The upper arkose and micaceous schist (metagraywacke) represent a similar sequence to that just described. A higher input of fine-grained, presumably clay-sized material into the basin resulted in a ferruginous chloritic schist rather than a dominantly chemical sediment such as the main iron-formation member. The relatively abundant tourmaline within this upper schist indicates that boron concentration processes were at work in some proximity to the depositional site.

**Mafic Volcanic Rocks**

The thin beds of mafic metavolcanic rocks are probably the result of the upwelling of volcanic material along rift faults. The correlation of the metavolcanic rocks between drill holes is somewhat uncertain, as a flow of only 5 m in thickness would not be expected to cover the several tens of meters of horizontal distance.
However, the thicknesses shown in the cross-sections are minimal due to core preservation techniques, and they may actually be up to twice as thick as indicated. Also, some of the units are likely tuffaceous and could easily fit these dimensions.

**Interpretive Rift Model**

Referring to Fig. 20, the early rift phase (a), resulted in a fluvio-deltaic to shallow-marine, high energy depositional environment in which basal conglomerates may have been deposited. With further rifting, presumed high relief and rapid mechanical weathering of the source area and a shallow water, high energy depositional environment resulted in arkosic sandstone deposition, with synchronous deposition of graywacke and mudstone in the deeper parts of the basin. A period of quiescence and reduced clastic input at the end of this cycle allowed a dominantly chemical phase (iron-formation) to accumulate. Renewed rifting (b) created another sequence of deposition similar to that just described, with iron-formation diluted by terrigenous input of shaly material near the top of the sequence. Mafic volcanic rocks originated along extension fractures between down-dropped crustal blocks. Submarine sea-floor vents and/or hydrothermal convection systems would probably have developed along these fracture zones. Evidence of such hydrothermal processes exists in the stratiform occurrences of tourmalinite; the
Figure 20. Schematic north-south cross-section of depositional basin near Philbrook, before deformation. See text for interpretation.
iron-formation may also have originated from submarine exhalative vents.

If these sediments were deposited in a rifting basin, one would expect the presence of 'sub-basins', formed between down-dropped crustal blocks (Fig. 20). These sub-basins could have a profound effect on local depositional variations, producing relatively rapid and possibly recurring facies changes over short lateral distances.

**Summary of Sedimentation Model**

The sequence of metasedimentary rocks near Philbrook was deposited along the western edge of the rift-formed Animikie basin. The conglomerates represent fluvio-deltaic to shallow marine deposition of sediments directly on the Archean crust; this conglomerate may be younger or older than the iron-bearing rocks west of Philbrook. The metasedimentary rocks west of Philbrook either indicate deposition in relatively shallow, oxygenated water, as evidenced by the clean arkosic sandstones and oxide-facies iron-formation or shallow sorting followed by slumping and deeper-water submarine-fan deposition. In Archean analogs, the arkoses were deposited in a submarine-fan environment, being slumped down-slope after rapid mechanical breakdown and sorting in a temporary basin upslope Meyn and Palonen, 1980; Barrett and Fralick, 1984). The arkosic sediments may have been derived from an area of high relief, and are probably the

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result of little chemical weathering and rapid mechanical transport. Quiescence of tectonism slowed down the terrigenous input and allowed the main iron-formation member to form. Renewed rifting or tectonism may have created a recurrence of clastic sedimentation, resulting in the upper arkosic unit. The final depositional phase involved reduced fine-grained clastic input with synchronous deposition of cherty magnetite iron-formation and possibly boron.

Rift faults within the basin would provide deep channelways for the heat necessary to drive a hydrothermal alteration system. Evidence that such a system existed lies in the presence of mafic volcanic rocks, which should be synkinematic with hydrothermal alteration, the presence of tourmaline (for which the boron source was presumably a brine of hydrothermal origin), and the abundant sericitized albite.

Subsequent closing of the Animikie basin during the Penokean orogeny isoclinally folded the sediments and imposed a regional upper greenschist-facies metamorphism on the rocks. Neighboring gabbroic to granitic bodies were intruded in the late stages of the Penokean orogeny and possibly imposed a regional metamorphic overprint on the sedimentary strata.
MAFIC INTRUSIVE ROCKS NEAR PHILBROOK

Occurrence

Several small outcrops of medium- to coarse-grained mafic diorite and monzodiorite are located in sections 34, 35, and 3 (Figs. 1 & 3). Additionally, twelve drill holes in the NW 1/4 of section 3 penetrate ultramafic, Fe- Ti- P-rich gabbroic rocks (Fig. 5). These drill holes were targeted over magnetic highs and were drilled in the same time period as the metasedimentary rocks to the west of Philbrook.

The sharp break in aeromagnetic and gravity signatures, and the small diameter (~2 km) of the intrusion, as inferred from steep geophysical anomalies and rock distribution, implies a steeply-walled, cylindrical or funnel-shaped intrusion. The horseshoe-shaped magnetic high is probably a result of internal concentrations (layers or dikes) of oxide-rich rock (Figs. 3 & 4; App. B).

As with the metasedimentary rock drill core, the gabbroic cores have been skeletonized, making detailed layering studies impossible. However, general stratigraphic mineralogical variations are discernable.

Sample Preparation

Five polished plugs were made from the oxide rocks and pyroxenite/hornblendites, and 26 thin sections were made from drill core and outcrop samples. Several of these were polished for reflective optical work and for
Figure 21. Igneous rock classification diagrams. A) and B) modified from Streckeisen, 1973.
A=alkali feldspar, P=plagioclase, Q=quartz, Px=clinopyroxene, Ol=olivine, Hb=hornblende.
electron microprobe analyses. Representative samples from each major rock type were point-counted, with a minimum of 1000 points per slide. The varieties which were too coarse-grained to point count accurately were visually estimated from a slabbed hand sample.

**Classification**

The classification scheme of Streckeisen (1973) is used in Figure 21 as a basis of classification for the pyroxene-, hornblende-, or plagioclase-rich rocks, but is not satisfactory for the classification of the oxide-rich rocks. Those rocks containing more than 80% opaque oxide minerals are simply termed 'oxide rock' or 'nelsonite' (Philpotts, 1967).

**General Description**

Rock types range from medium-grained ultramafic nelsonite, pyroxenite, and hornblendite to gabbroic-looking melanocratic hornblende diorite and coarse-grained monzodiorite. The rocks are highly enriched in Fe, Ti, and P, are low in Si, Al, and K, and lack orthopyroxene and olivine. Whole rock chemistry and microprobe analyses of minerals show high Fe/Mg ratios, and values of 30 to 50% Fe are common from the oxide-rock (App. B & App. D, Table D-2).

Textural and mineralogical evidence indicates a high content of volatile fluids in the magma. Deuteric alteration and possible metamorphic effects are widespread and have significantly altered the primary
mineral assemblage, resulting in heavy uralitization of pyroxene, replacement of Fe-oxides by biotite or amphibole, and saussuritization of feldspar. Net veining and pegmatitic texture is present in the quartz-bearing diorites.

Although the color of the more mafic plagioclase-bearing rocks from outcrop is that of a typical gabbro, the plagioclase composition (An$_{45-50}$) and the abundant hornblende are more typical of a diorite. In consistency with Streckeisen (1973), and on the basis of modal mineralogy, this rock type will be called a mafic diorite. Other rocks from outcrop are pinkish-white and light-colored, and more easily classified as mesocratic diorite or monzodiorite.

Age of Intrusion

Several constraints can be placed on the age relations. With the exception of late, brittle shear zones in the diorite outcrops and the hornblendite/pyroxenites of drill core; and a weak foliation in the diorites near the edge of the intrusive body, the intrusive body appears to be relatively undeformed. Metamorphism within the intrusion is limited to epidotization, which is a common phenomenon in Archean to Proterozoic age rocks of all types in central Minnesota (Southwick, pers. comm., 1986). Alteration within the intrusive is deuteric, a result of the high water content of the magma.
The relatively fresh and unmetamorphosed nature and the fact that only late, brittle deformation is present suggests that the Philbrook intrusion is of Proterozoic or younger age. It may have been emplaced during the waning stages of the Penokean Orogeny.

The age of a neighboring pink, potassic granite is believed to be similar to that of the St. Cloud type granites, approximately 1700-1800 m.y. (Southwick, pers. comm., 1976; Morey, 1978). The Philbrook intrusion shows similar structural features to the St. Cloud-type granites, evidence for a similar age. Also, presumed basal conglomerates located a few feet from the NE contact of the intrusion do not contain clasts of this body in them (p. 68), but rather clasts of Archean (?) granodioritic rocks and quartz-rich sediments. The conglomerates are inferred to be equivalent to or younger than the Early Proterozoic Denham Formation, and since the Philbrook mafic body appears to intrude these, an upper time constraint of ~1800 Ma. (Morey, 1973) can be placed on the time of intrusion. Thus, a Proterozoic or Penokean age is inferred for the Philbrook intrusion from its cross-cutting relationships to stratified rocks of inferred Early Proterozoic age (Fig. 3.)
Figure 22. Photograph of nelsonite (left), pyroxenite (upper right), and hornblendite (lower right). Brownish-yellow apatite prisms visible in nelsonite at upper left.

Figure 23. Photomicrograph of nelsonite. Trellis ilmenite lamellae in equant magnetite grains; light-colored interstitial composite ilmenite visible in left of photo. Reflected light, crossed polars, field of view = 3 x 2 mm.
Table 4. Modal analyses of Intrusive Igneous Rocks.

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HBl 1- Magnetic hornblende, brown, blue-green, or near colorless; coarse prismatic aggregates or roughly tabular-shaped and zoned.

HBl 2- Fibrous aggregates of pale green to colorless hornblende, replacement in origin, associated with opaque scraps of Fe-Ti Oxides.

HBl 3- Uralitic-type hornblende, as patchy replacements or overgrowth rims on pyroxenes.
The term nelsonite was first used for rutile-apatite rocks from Nelson and Amherst Counties, Virginia, by Watson and Taber (1913, from Kolker, 1980), but the term has been expanded by others to include ilmenite- or magnetite-apatite rocks (Philpotts, 1967). At the type locality, the nelsonite is associated with alkalic anor-thosite, ferrodiorite, and charnockite (Herz and Force, 1964). Philpotts has shown that magnetite-apatite rocks may form by the immiscible separation of a 2:1 eutectic magnetite:apatite mixture from an intermediate, Fe-rich, alkaline magma.

Philbrook_Nelsonites

In the drill holes from section 3, oxide-apatite rocks (>80% oxide + apatite; Table 4; Figs. 22 & 23) account for approximately 10% of the total core interval, but no nelsonites were found in outcrop. Impurities in the ideal oxide-apatite composition consist of chlorite, amphibole, or pyroxene, and a continuous gradation from nelsonite to pyroxenite is present.

The oxide-apatite rocks are strongly magnetic, metal-lie black, medium-grained rocks with an equigranular texture. Apatite prisms are typically <1mm in length and yellowish-white. On a sawn surface 1-3 mm-long equant oxide grains are visible due to varying cleavages. Traces of pyrite and chalcopyrite are visible as blebs in the
rock, and as fracture linings with carbonate.

Both transmitted and reflected light were used to examine textural and mineralogical relations in the nelsonites. In transmitted light only apatite and chlorite are visible; in reflected light the abundant exsolved lamellar ilmenite stands out.

Euhedral to subhedral apatite grains generally contain solid cores of oxide or ring-shaped cores of chlorite, and apatite/oxide grain contacts are commonly rounded and smooth. In rare cases apatite forms optically continuous, sub-poikilitic masses around oxides. The apatite is fresh and relatively unfractured.

Except for a few percent of irregularly-shaped discrete ilmenite grains, all of the oxide grains are composed of subhedral, granoblastic magnetite with exsolved (111) lamellar ilmenite which forms triangular-octahedral outlines (Fig. 23). The exsolved (111) ilmenite lamellae are straight-edged along their entire length, but taper when in contact with other lamellae or with the edges of the grain; these lamellae are a product of titanomagnetite oxidation-exsolution. Oxide grains in contact with chloritic material are commonly 'shaggy' due to partial replacement of magnetite by chlorite, leaving the ilmenite lamellae protruding into the chlorite. Trace amounts of hematite are present within magnetite along the edges of exsolved ilmenite lamellae, also a product of titanomagnetite oxidation. Locally, abundant
tiny exsolved grains of a darker gray mineral are present in the magnetite portion; these impart a woven-textured appearance to the magnetite under crossed nicols (Fig. 23). Probe analyses show this to be an Al-bearing phase, probably hercynite (FeAl$_2$O$_4$).

Chemically, there is little variation between the composite ilmenite and the exsolved lamellar ilmenite. Magnetite is quite pure except in the phase with the exsolved hercynitic spinel, in which case substantial aluminum and increased titanium amounts are present (App. C, Table C-4; also see Fig. 24).

The degree of solid solution in hematite-ilmenite and ulvospinel-magnetite was calculated using the method
Mineral | Ox. Rock | Hblite-Phenite | Melodiorite | Meso- diorite
--- | --- | --- | --- | ---
Ilmenite | Early | Early to late | Late | Late
Composite Ilmenite | Late | Late | Late | Late
Apatite | Early | Early to late | Mid to late | Mid to late
Cumulus Hornblende | - | Early to middle | - | -
Pyroxene | - | Early | Mid | Mid
Andesine | - | - | Early | Early
Albite | - | - | Late | Late
Quartz | - | - | - | Late

Table 5. Relative timing of crystallization of igneous minerals.

of Stormer (1983; see p. 144).

Chlorite, the third most common component of the nelsonites, occurs in lobate masses, and the lack of opaque scraps, sphene, epidote, etc. within the aggregates suggest a metasomatic reaction between late-stage hydrous fluids and primary silicate minerals in which Fe was transferred elsewhere in the system. Minor amounts of sphene are present as a late interstitial product. (see App. C, Table C-3).

Trace amounts of pyrite, marcasite and chalcopyrite are present, typically in thin fractures with chlorite and carbonate. The sulfides locally replace magnetite in
exsolved Fe-Ti oxide grains, and some discrete blebs of
sulfides occur between oxide grains.

The oxide-apatite rocks appear to be cumulates, with
possible continued accretion growth resulting in the
present allotriomorphic equilibrium texture. Apatite is
largely of cumulus origin. Sphene is a late intercumulus
mineral, and chlorite is apparently of secondary,
deuteric origin. Table 5 shows the relative timing of
crystallization for the major minerals of the entire
igneous suite.

**Pyroxenite and Hornblende**

Magnetite pyroxenite and magnetite hornblende
comprise approximately 80% of the total drill core
interval, and plagioclase-bearing hornblende pyroxenite
occurs in one outcrop. Oxide contents are variable,
ranging from 10% to >80%, the latter grading into impure

In hand sample, these rocks are dark green and
black. The pyroxenite is medium-grained, equigranular,
and massive in structure, whereas the hornblende is
either massive and laminated. The lamination is defined
by prismatic hornblende crystals.

Massive varieties of hornblende are typically
coarse-grained (7-8 mm), lighter green, and more actin-
olitic than the laminated variety. Apatite is abundant
(4-20%) throughout all of these rock types. Some small 1
cm zones of magnetite-apatite concentrations are present
within the hornblende pyroxenite, suggestive of local concentrations of late-stage residual liquid or of trapped magnetite-apatite liquid within the pyroxenite magma.

Approximately 40% of the hornblendites from drill core are sheared, resulting in a strong foliation and subvertical lineation of the constituent minerals. In these rocks, alteration of amphibole/pyroxene to chlorite has produced a soft, splintery schist.

These ultramafic rocks show striking deep green and red-brown colors in thin section, due to various shades of pleochroic hornblende. Modes are given in Table 4.

Clinopyroxene is tabular to equant, subhedral to euhedral, and appears to be of cumulus origin (Fig. 25). It is clean and does not contain the numerous rods of exsolved oxides which are common in pyroxenes in the diorite. The original pale brownish pyroxene grains are rimmed and patchily replaced by green, uralitic amphibole; and commonly completely replaced by a nearly colorless, optically continuous actinolite. This colorless actinolite contains patches of the dark green uralitic actinolite. Microprobe analyses (App. C, Table C-5; Fig. 26) show that the pyroxene is a highly calcic variety similar to that in various Alaskan ultramafic complexes (Fig. 26; Wyllie, 1967), except for a higher Fe/Mg ratio at Phlbrook. Substitution of Al, Na, and Ti is minor, and the recalculated analyses show near ideal stoichio-
metric amounts of Ca, (Mg+Fe), and Si in the unit formula.

Three types of amphibole, either of magmatic or deuteric origin, are present in the ultramafic rocks. These types are: 1) red-brown to pale green, zoned magmatic amphibole; 2) patchy dark or light green uralitic actinolite, replacing and in optical continuity with pyroxene; and 3) fibrous, felted masses of actinolite and opaque scraps which are the latest pyroxene and amphibole replacement product.

Amphibole type (1) occurs as deep red-brown prismatic...
Pyroxene from diorite outcrop
Pyroxene from hornblende pyroxenite

Range of pyroxenes from SE Alaska-type ultramafic hornblende pyroxenite.

Range of pyroxenes from alkali basalt intrusives; Black Jack Intrusion (upper) and Garbh Eileen (lower).

Approximate trend of CPX from high-Al and tholeiitic Skaergaard, Stillwater, and Bushveld-type intrusions.

Pyroxene end-member compositions:

| PX-2-0G | Ca$_{43}$Mg$_{37}$Fs$_{20}$ | (Wo$_{43}$En$_{37}$Fs$_{20}$) (Di$_{65}$Hd$_{35}$) |
| PX-3-0G | Ca$_{47}$Mg$_{36}$Fs$_{20}$ | (Di$_{64}$Hd$_{36}$) |
| PX-3-0G | Ca$_{39}$Mg$_{38}$Fs$_{23}$ | (Di$_{62}$Hd$_{38}$) |

| PX-3-0G | Ca$_{48}$Mg$_{27}$Fs$_{25}$ | (Di$_{52}$Hd$_{38}$) |
| PX-3-0G | Ca$_{48}$Mg$_{29}$Fs$_{23}$ | (Di$_{56}$Hd$_{48}$) |
| PX-3-0G | Ca$_{50}$Mg$_{27}$Fs$_{23}$ | (Di$_{54}$Hd$_{46}$) |
| PX-3-0G | Ca$_{49}$Mg$_{31}$Fs$_{20}$ | (Di$_{61}$Hd$_{39}$) |

Figure 26. Pyroxene quadrilateral.
Figure 27. (Top) Photomicrograph of zoned cumulus amphibole consisting of red-green core with cyclic green and colorless overgrowth rims. Field = 3x2 mm, one polar; DDH #6, 50' depth, Sec 3.

(Bottom) Chemical variation diagrams across the zoned amphibole shown above. Sketch shows microprobe points on zoned amphibole.

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Figure 20. Amphibole compositional diagrams. Top diagram modified from Deer, Howie, and Zussman, 1966; lower two modified from Ernst, 1968.
grains, with euhedral cores composed of ferropargasitic to ferroedenitic hornblende. Dark green and nearly colorless ferroactinolitic hornblende (Fig. 27), forms rhythmic overgrowths on the ferropargasitic hornblende. Preferred orientations of prismatic hornblende impart a cumulus layering in some core samples. This hornblende contains small apatite or Fe-Ti oxide inclusions, and the rims are commonly complexly intergrown with oxides. Exsolved rutile (?) needles are present within some grains.

Chemically, the red-brown ferropargasitic cores are substantially higher in Al, Fe, Na, K, and Ti than the green actinolitic rims. Various chemical compositional plots show the distinctive groupings of these end members (see Fig. 28). The cause of the oscillatory zoning is somewhat unclear, but could be caused by sudden variations in Si/Al, Fe/Mg, or (Na+K)/Ca ratios in the late-stage residual liquids due to crystallization of other mineral phases, by changes in oxygen fugacity due to crystallization of oxide minerals, or a sudden change in confining pressure on the residual melt. The oscillatory nature of the overgrowths suggests that a swinging back and forth of differing crystallization conditions must have occurred.

Actinolite (type 2) occurs as olive-green patches within and as thin rims around pyroxene, or as very pale green pyroxene pseudomorphs which are in optical
continuity with relict scraps of the replaced grain.

Actinolite type 3 forms pale green to colorless, nonfoliated, fibrous mats replacing pyroxene and possibly hornblende. This type is compositionally the same as type 2, but is later in occurrence. It is of deuteritic origin and is associated with opaque oxide scraps. The degree of replacement by this type of amphibole varies widely from nil to nearly 100%, in which case the entire rock is composed of fibrous amphibole plus opaque scraps. Occurrences of the latter are indicative of the pervasive deuteritic alteration which has altered the mineral assemblage of the entire intrusive body.

Fe-Ti oxides occur as interstitial composite ilmenite and as blocky to rounded, cumulus magnetite which show oxidation-exsolution of ilmenite similar to that in the oxide-rock. The magnetite portion of these grains is incipiently to completely replaced by amphibole, biotite, chlorite, and minor sulfides, resulting in abundant skeletal ilmenite pseudomorphs. Small euhedral apatite inclusions are common within all of the oxide grains, and rarer small, rounded inclusions of amphibole or biotite and chlorite are also present. Trace amounts of hematite are seen as oxidation products rimming ilmenite lamellae within exsolved grains.

In some cases of actinolitic replacement, secondary oxides are concentrated into zones which mimic cracks in the replaced grains. The lack of any talc or iddingsite
suggests replacement of amphibole or clinopyroxene rather than olivine or orthopyroxene.

The titano-magnetite grains or pseudomorphs thereof are interpreted to be of cumulus origin, whereas the irregularly-shaped composite ilmenite forms the later, interstitial grains. This is suggestive of Ti-enrichment in the residual liquid as crystallization proceeded, and the concentration of apatite with the late interstitial ilmenite suggests that P was also enriched in the fluid residuum.

Apatite is quite abundant, but variable in amount (tr-15%). It is included within pyroxene, amphibole, and oxides, and in some cases the inclusions are seen to increase with size nearer the rim of the host pyroxene. The smaller grains are euhedral, whereas the larger ones (2-3 mm) are commonly anhedral and have irregularly rounded, partially embayed contacts with the oxides. As in the nelsonites, cores of amphibole, chlorite, or biotite are common.

Sulfides, consisting of marcasite, pyrite, and minor chalcopyrite, comprise a small percentage of the total volume. They occur as magnetite replacement products and as small rounded inclusions within magnetite-ilmenite grains. Textures indicate both primary and replacement origins for the sulfide minerals.

Biotite and chlorite occur as late, replacement-origin oxide pseudomorphs which form rims over or scaly
masses between skeletal ilmenite lamellae. Chlorite is of similar habit and occurrence to biotite, and is a retrograde product of biotite, as indicated by green biotite intergrown with chlorite.

Feldspar is conspicuously lacking in the pyroxenite-hornblendite rock suite. The only noted occurrence in thin section was a small, rounded feldspar inclusion within a pyroxene grain.

Calcite and sphene (App. C, Table C-3) are present only in trace amounts but are significant because of their magmatic origin (Fig. 25). They form interstitial fillings between cumulus pyroxene, oxide, and magmatic amphibole. Pyroxene alteration has also produced minor amounts of calcite.

Mineralogically and texturally, these hornblende-magnetite pyroxenites and magnetite hornblendites are very similar to the descriptions of the medial and outer zones of magnetite-hornblende pyroxenites of southeastern Alaska (see p. 126). The Fe/(Fe + Mg) ratio, however, is much higher in the Philbrook intrusion, suggestive of a higher degree of fractionation.
Figure 29. Photograph showing layered melanocratic diorite near Philbrok. NE 1/4 SW 1/4, Sec. 34, T. 133 N., R. 32 W., along Cty. Hwy. 23.
Diorite

Apatite-rich mafic diorite and diorite/monzodiorite are negligible in drill core, but occur in several outcrops near Philbrook (Fig. 3). There is a gradation from greenish-black, medium to coarse-grained, high P-Ti mafic diorite (see discussion under classification) to light green and white or pinkish-white, coarse-grained, quartz-bearing diorite, monzodiorite, and anorthosite. The lighter-colored diorites apparently occur in a circular zone within the mafic diorites (Fig. 3).

Melanocratic Diorite

High P-Ti meladiorite, the most mafic rock type in the diorite suite, is a hard, dark green to black, medium- to coarse-grained rock. It is quite fresh overall, but the southern-most outcrops are slightly oxidized and reddish in color. Feldspars are dark and platy, with visible albite twinning. Euhedral white to pale yellow apatite is coarsest and most abundant in the southern-most of the mafic diorite outcrops. Oxide minerals are abundant and evenly distributed throughout the rock. Cumulus layering is defined by 5 to 8 cm feldspathic bands and 18 to 28 cm darker green, more mafic bands. Its attitude is measurable in only one outcrop, but several large boulders in the area clearly show the cumulus layering (Fig. 29).

Several narrow shear zones ranging in width from 5 cm to at least 1 m are present within the mafic diorite.
Table 6. Plagioclase end-member compositions.

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</tbody>
</table>

Figure 30. Compositional plot (mol. %) of or-ab-an from diorite and anorthosite outcrops, based on probe analyses. (PL-19-Tm, solid circle, is from tourmaline-rich arkosic rocks).
This schistose sheared rock has a fine-grained protomylonitic texture, and is strongly foliated and lineated. The lineation results from a combination of crenulated foliation and amphibole alignment within the foliation plane. A brittle cleavage is present in some outcrops which is parallel to fractures in the rock. Three major sets of late, brittle fractures cut the outcrops; and mineralized veinlets, filled with a green, fibrous amphibole, are present.

The mafic diorites are medium- to coarse-grained, hypidiomorphic-granular, cumulate-textured, and rich in apatite and deuteric hornblende.

Cumulus plagioclase crystals are tabular, well-twinned, and moderately to strongly zoned. The composition of plagioclase was optically determined (Michel-Levy method) to be in the An$_{32-45}$ (Andesine) range, confirmed by microprobe analyses (Table 5; Fig. 30). Most plagioclase grains are stained by an unidentifiable reddish dust, probably hematite. Some of the plagioclase from the drill core samples is nearly opaque due to iron staining, and is pink in hand sample. Fine-grained, granular epidote and sericite are present as alteration products. The epidote is evenly distributed, but sericite selectively replaces plagioclase cores. Biotite, chlorite, and actinolite are present in small amounts along grain fractures, and the grain boundaries are typically rounded and irregular due to strong peripheral penetration by prismatic amphibole (see Fig. 32). A few samples contain broken feldspar grains with mortar-textured
feldspar and quartz (?) developed in the fractured zones. Small amounts of perthitic albite are present in interstitial areas, and contain wispy actinolitic hornblende. Rarely, a myrmekitic quartz/feldspar rim separates cumulus andesine from interstitial albite. Microprobe analyses (Fig. 30; App. C, Table C-2) show very little potassium in the feldspar, indicating that the last residual liquids may have been highly depleted in Ca and K and enriched in Na. This lack of K reflects the K deficiency of the entire intrusive system.

Subophitic to cumulus clinopyroxene (augite) has strong herringbone-type exsolution features and is replaced by actinolitic amphibole. The pseudomorphs are anhedral-interstitial to euhedral, and inclusions of small rounded plagioclase and euhedral apatite are common. The amphibole replacement has proceeded to various stages, with the most well preserved pseudomorphs consisting of fine herringbone structured actinolite normal to the central twin plane in the relict pyroxene. In the most advanced stages of replacement, fibrous mats of light green actinolite plus scraps of opaques have replaced the herringbone actinolite and/or pyroxene. Some remnants of brownish pyroxene remain inside the amphiboles (Fig. 31). These contain rod-shaped opaque oxide exsolution blebs, apparently of magnetite or ilmenite. Other alteration products of pyroxene include chlorite, leucoxene (?), and calcite. The parallel habit
of the amphibole replacing pyroxene and the herringbone-oriented oxide blebs give evidence that the pyroxene may have originally consisted of a Ca-rich clinopyroxene with exsolved Ca-poor orthopyroxene lamellae, but no petrographic evidence exists as proof. Probe analyses (App. C, Table C-5, Fig. 26) of the pyroxene show it to be a fairly iron-rich augite, consistent with the overall iron-enriched quality of the entire intrusive system.

A variety of amphibole types occurs within the mafic diorites. The three main types of amphibole (App. C, Table C-6) are: 1) fibrous light green actinolite which has replaced pyroxene grains and has abundant associated opaque scraps; 2) deep blue-green ferropargasitic hornblende which forms prismatic rims around relict pyroxene grains, typically penetrating plagioclase, and is associated with biotite; and 3) deep blue green interstitial magmatic (?) ferropargasitic hornblende with microscopic rows and fracture linings of oxides.

The fibrous pseuomorphous amphibole (type 1) has been discussed under pyroxene.

The second type of amphibole is a strongly pleochroic variety which forms coarse prismatic rims over pyroxene, penetrating into and replacing feldspar (Fig. 32). Chemically, this amphibole is quite similar to the red-brown cumulus hornblende in the hornblendite/pyroxenite, except that it is closer to the tschermakite end member due to higher aluminum content. Scraps of
Figure 31. Photomicrograph of relict pyroxene (light brown) from diorite. Note amphibole reaction rim between pyroxene and plagioclase. Field = 3x2 mm, one polar; NE1/4 SW1/4 NW1/4, Sec. 34, T.133 N., R.32 W.

Figure 32. Andesine crystal, in same sample as Figure 31. Note prismatic ferropargasitic hornblende penetrating the grain perimeter.
brown biotite and opaque oxides are commonly associated with this amphibole.

Using probe compositions, when available, and ideal formulas for unprobed minerals, a deuteric reaction such as pyroxene + andesine + H₂O → amphibole + albite-forming liquid is suggested, based on textural and compositional evidence.

\[
.3\text{Fe}_2\text{O}_3 + 1.7\text{FeO} + .5\text{MgO} + \text{Na}_0.08\text{Ca}_0.88\text{K}_{0.15}\text{Mg}_0.7\text{Fe}_4\text{Al}_{1.13}\text{Si}_{1.9}\text{O}_6 + 2(\text{Na}_0.52\text{Ca}_0.48\text{Al}_{1.47}\text{Si}_{2.5}\text{O}_8 + \text{H}_2\text{O} = \text{Na}_0.5\text{K}_{0.15}\text{Ca}_1.6\text{Fe}_2.7\text{Mg}_1.7\text{Al}_3.4\text{Si}_6\text{O}_{22}(\text{OH})_2 + .3\text{Na}_2\text{O} + .96\text{Si}_2\text{O}_5.
\]

Evidence also exists for complete replacement of feldspar by amphibole, i.e. relict feldspar twinning within zones of felted amphiboles and continuous plagioclase crystals separated by clusters of fibrous amphibole.

The third occurrence of amphibole consists of a magmatic(?) hornblende which is choked with dark brown blades of rutile(?) and cleavage-fracture controlled secondary opaque oxides. A deep brown biotite with similiar inclusions is also present, in growth-continuity with this hornblende. Chemically, this hornblende is similiar to the prismatic hornblende described above.

Irregularly-shaped, intercumulus Fe-Ti oxides and skeletal pseudomorphs of oxides contain abundant inclusions of, and are strongly associated with, apatite. Polished sections show that most of the oxides are composite ilmenite or trellis-shaped ilmenite pseudomorphs.
of exsolved magnetite-ilmenite grains, in which the magnetite has been selectively replaced by chlorite, biotite, or amphibole. Opaque scraps associated with degraded pyroxene and amphibole are in part ilmenite, but are too small for positive identification.

Apatite is an abundant late-crystallization product which is highly concentrated in interstitial areas with opaque oxides. Some small apatite inclusions are found in centers of pyroxene, feldspar, and opaque grains, but inclusion size and abundance increases near the rim of the host grain. Textural evidence indicates that apatite began forming early in the crystallization sequence in small amounts and bloomed in the late stages, with oxides. In rare cases large (6mm across) irregular apatite grains enclose feldspar and hornblende, and have rounded boundaries against oxide grains.

In one thin section from core, a 2 mm-wide epidote and zeolite (prehnite?) vein cross-cuts the slide; the feldspars in this rock are pink in hand sample, and in thin section are nearly opaque with a red dust and heavily altered to epidote-sericite.

_Mylonitized Gabbro from Shear Zones_

One thin section was examined from the mylonitized gabbro in the N-NW trending shear zones. A well-developed shear foliation is defined by aligned hornblende and foliated secondary actinolite. Broken feldspar porphyroclasts, ≤ 2mm in diameter, are still
well-twinned and fresh. The foliation wraps around the feldspars, which have quartz, chlorite, and recrystallized amphibole pressure shadows. Apatite grains are broken in a fashion similar to feldspar, and some amphibole porphyroclasts (pyroxene pseudomorphs) are also present. Trains of ilmenite rods, formed from the breakage of skeletal ilmenite, are strung out along foliation, and in some cases lead to a partially preserved pseudomorph. The matrix hornblende, a deep blue-green aluminous variety, is well lineated as a result of mechanical rotation and recrystallization. Thin carbonate/quartz veinlets (<0.5 mm) parallel the foliation. An abundant dusty red oxidation product with the carbonate suggests an earlier iron-rich carbonate.

**Mesocratic Diorite and Monzodiorite**

Coarse-grained mesocratic diorite, monzodiorite, and small zones of anorthosite occur in the north-central part of the outcrop area (Fig. 3). The dioritic rocks are green and white to pinkish-orange on the weathered surface, and are green-gray and light orange-white on a fresh surface. The degree of feldspar alteration controls the color to a large extent, i.e. highly sericitized plagioclase is white and fresh plagioclase is dark grayish-blue. These alteration zones are in places marked by a thin zone of high epidote content.

The green, mafic component of the diorite is composed of hornblende and chlorite. A moderate schistosity is
present in much of the diorite, confined to the mafic portions and defined by the preferred orientation of chlorite, giving the rock a phyllitic sheen. Jointing is present in most of the outcrops, with the same major directions as the cleavage.

Textures and grain sizes vary from porphyritic and laminated in the border phase to very coarse-grained and massive nearer the center of the intrusive body. The border phase contains blocky feldspar phenocrysts up to 1 cm in length which show lamination or flow alignment (Fig. 33). Interstitial antiperthitic albite is visible in the matrix, with dark green amphibole.

There are local pegmatitic patches in which grain sizes vary greatly, from 1 cm to 5 cm or more. These pegmatoid outcrops show net veining, probably a result of late-stage volatile streaming (Fig. 34). The veins are 1-20 cm wide, subvertical, and mineralogically similar to the host rock. Some small zones of anorthosite, less than 2 m across, are present in the pegmatitic areas of the outcrops.

None of the shear zones which are common in the gabbro/meladiorite outcrops were noted in the diorite outcrops.

Microscopically, the mesocratic diorite differs from the meladiorite just described in that it contains interstitial quartz and more interstitial antiperthitic feldspar and biotite (Table 4). It also contains less
Figure 33. Photograph of porphyritic, flow-laminated border phase of diorite. NE1/4 NW1/4, Sec. 34, T. 133N., R. 32W.

Figure 34. Photograph of diorite outcrop showing pegmatitic veins, visible from just left of hammer handle to upper left of photo. Plagioclase crystals are saussuritized, and green matrix consists of a chlorite-actinolite alteration assemblage. NE1/4 SW1/4 NW1/4, Sec. 34, T. 1233N., R132W.
apatite, oxides, and pyroxene or pyroxene pseudomorphs. Late-stage alteration is more evident in these rocks. The feldspars are commonly heavily saussuritized, and abundant late epidote/zeolite/carbonate veining or patchy alteration zones are common. The overall grain size is coarser than the meladiorite by a factor of five or more (up to 3 cm).

Plagioclase feldspar (An$_{45-53}$; App. C; Table C-2) grains are blocky, subhedral and well-twinned and range from fresh, clean crystals to heavily sericitized or epidotized (saussuritized) grains which no longer exhibit twinning. The alteration zones cross-cut several grains and are typically quite abrupt, separated by a transition zone only a few mm wide. The feldspars are compositionally zoned, and cores are commonly selectively saussuritized. Bent and strained grains are common, with recrystallized granoblastic albite and quartz (?) along strained areas.

Interstitial, antiperthitic albite (An$_{03-10}$; App. C, Table C-2) is present as a late crystallization product. The antiperthitic zones lack the reddish dust which is prevalent in the cumulus andesine grains, and are generally in sharp contact with the andesine. In rare cases a myrmekitic quartz/feldspar rim separates the cumulus and interstitial feldspars. As in the mafic diorite, this interstitial feldspar has an antiperthitic
texture, even though probe analyses show that they lack potassium. This lack of potassium cannot be readily explained, except that the texture may be the result of quartz-albite intergrowths.

The blue-green prismatic amphibole (type 2 under meladiorite), and fibrous actinolite (type 1 under meladiorite) are the most common amphibole types in the diorite outcrops. The blue-green variety (Hb-1 to 4-06; Hb-35 & 36-1; App. C, Table C-6) is ferropargasitic in composition. It is of similar origin to that in the meladiorite, i.e. via a reaction involving plagioclase + pyroxene, and penetrates the feldspar grain boundaries. Another amphibole variety (Hb-5 & 6-10 of Table C-6), army-green in color, appears to be of primary magmatic origin but chemically is closer to actinolite. It is apparently a deuteric alteration product of pyroxene. Interstitial hornblende of magmatic origin exhibits a lamellar type of twinning, and is concentrated with oxides and apatite, also late interstitial products (Fig. 35). Fibrous actinolitic pyroxene pseudomorphs, with the characteristic herringbone pattern, are texturally similar to those described in the mafic diorites.

Very little pyroxene remains, having been largely replaced by amphibole. Where present, it is of similar color and form (with exsolution rods) as the pyroxene in the mafic diorites described previously. The overall
abundance of pyroxene or recognizable pyroxene pseudomorphs decreases as the rocks become more felsic, either because less pyroxene was crystallizing or because it was more thoroughly replaced by amphibole by a reaction with late liquids.

Brown biotite is common, occurring most commonly as small scraps interstitial to randomly oriented blue-green hornblende scraps; and it is commonly choked with deep-brown, bladed rutile (?) inclusions.

Light green chlorite is present as an alteration
product of feldspar and ferromagnesian minerals. One outcrop of coarse-grained and net-veined diorite has rounded masses of chlorite which have radially oriented cracks emanating from them, suggesting that the chlorite is of replacement origin, with a volume increase.

Traces of sericite occur as feldspar alteration products.

Epidote (± clinozoisite) is an abundant alteration product of feldspar and of ferromagnesian minerals, where it is associated with chlorite. In some cases masses of epidote completely replace feldspar and ferromagnesian minerals; it is also common as fracture fillings. There is some textural evidence for magmatic epidote, i.e. interstitial grains between cumulate feldspars, but this is not without ambiguity (Zen, 1984). A possible reaction for the alteration of andesine to epidote is:

\[
10\text{Na}_2\text{O} \cdot 6\text{Ca}_0.4\text{Al}_1.5\text{Si}_2.5\text{O}_8 + 2\text{Fe}_2\text{O}_3 + 6\text{H}_2\text{O} + 5\text{SiO}_2 + 0.75\text{O}_2 \rightarrow \\
4\text{CaFe}^{3+}\text{Al}_2\text{Si}_2\text{O}_8(\text{OH})_2 + 6\text{NaAlSi}_3\text{O}_8 + 0.5\text{Al}_2\text{O}_3
\]

or \(\text{andesine + hematite + quartz + water + O}_2 \rightarrow \text{epidote + albite}\).

Apatite occurs as euhedral prismatic grains with oxides and as anhedral grains interstitial to other minerals. In either case, it is a late-stage crystalization product, concentrated with oxides in interstices between feldspar.

The oxides are anhedral and consist of composite ilmenite and trellis-shaped ilmenite after exsolved
magnetite-ilmenite grains. Sphene and leucoxene are commonly associated with oxides, and magnetite has been heavily replaced by chlorite, biotite or amphibole.

Clean, non-undulose, granoblastic, recrystallized aggregates of interstitial quartz are present. Quartz is not present in the mafic diorites, consistent with its being a product of more advanced differentiation in the rock suite.

**Crystallization Sequence**

Based of increasing SiO₂ and Al₂O₃ content, the crystallization order would seem to be in the order: oxide rock-hornblendite/pyroxenite-mafic diorite-mesocratic diorite. Minor and trace element data, however (Fig. 36), indicate that the hornblendite/pyroxenite had crystallized before the oxide rock. High V, Ni, Cr, Co–Cu, and Y and low Ba, Sr, Rb, and Zr values in the hornblendite/pyroxenite, relative to the other rock types (including nelsonite), are consistent with enrichment of the late stage liquid with the incompatible elements and concentration of the more compatible elements in the early-formed cumulus phases, such as pyroxene and hornblende.

Textures in the hornblendite/pyroxenite indicate an interstitial origin for the Fe-Ti oxides. This is consistent with the trace element data, and the crystallization sequence is thus one of early cumulus pyroxene + apatite joined by cumulus ferropargasitic-ferroedenitic
Figure 36. A.) Major-element variation diagram. Samples 11-6 & OG calculated from probe analyses and point counts.
B.) Minor-element variation diagram.
hornblende and some late cumulus Fe-Ti oxides. The later stages of pyroxenite formation involved the crystallization of interstitial Fe-Ti oxides plus an increase in apatite concentration, the rhythmic dark green and pale green actinolitic amphibole overgrowths, and interstitial carbonate and sphene. The last modification involved deuteric alteration of pyroxene to actinolite + Fe-Ti oxide scraps + secondary carbonate and sphene, via a schematic reaction such as pyroxene + H₂O → actinolite + magnetite + carbonate + sphene.
Although the exposures and drill cores of the Philbrook intrusion are limited, several statements can be made with regard to its petrogenesis. First, the mineralogy, small size, and evidence of high magma water content throughout the intrusion distinguish this body from the tholeiitic Skaergaard-Bushveld-Duluth Complex type of layered intrusion. It lacks several of the pertinent and ubiquitous features of these layered intrusions, including: a.) Clinopyroxene-orthopyroxene pairs and low-Ca augite; b.) typical mafic layered intrusion cumulus minerals, such as olivine, chromite, and spinel; c.) the typical large size of the layered intrusions; and d.) the rhythmic chemical and mineralogical layering which is developed in the stratiform layered intrusions.

Also, the extreme iron-enrichment of the Philbrook intrusive is higher than in a normal differentiation trend of a tholeiitic magma. Alternatively, the concentric magnetic and mineralogical patterns show analogies with a host of small zoned intrusions of varied lithologies, among them the ultramafic occurrences of SE Alaska (Wyllie, 1967) and carbonatite-alkalic occurrences in Ontario (Erdosh, 1979; Sage, 1985). Mineralogically, the magnetite-apatite rocks are similar to nelsonite occurrences in Virginia, and near Camsell River, N.W.T., Canada. Descriptions of these and certain aspects of the
Alaskan-type of occurrences will be discussed in the following sections.

**Duke Island (SE Alaska) Ultramafic Intrusions**

Concentrically zoned ultramafic complexes occur within a 48 x 500 km, SE-trending linear belt which includes the area of Duke Island, Alaska (Irvine, 1967). They were emplaced after deformation of Paleozoic sedimentary rocks, are usually associated with or intrusive into gabbroic to dioritic rocks, and are intruded by a later series of granitic batholithic intrusions.

All of these ultramafic intrusions are less than 2.5 km in diameter and show a concentric mineralogical zonation, which in complete form consists of a dunite core surrounded by successive shells of peridotite, olivine pyroxenite, magnetite pyroxenite, and hornblende pyroxenite. Some of these zones may be lacking in individual intrusive complexes, but all of the complexes commonly share some distinctive features, including: 1) lack of orthopyroxene and feldspar; 2) extensive uraltitization of clinopyroxene; 3) abundant primary hornblende and magnetite in pyroxenites; 4) consistently Ca-rich clinopyroxene (diopsidic augite); 5) small outcrop area (typically < 0.75 sq. km); 6) concentric zonal arrangement; and 7) a high content of volatiles, especially in the late stages of crystallization. All of the features just listed are present in the ultramafic rocks at Philbrook, but a few other features of SE Alaska.
type occurrences, such as late-stage pegmatitic dikes of highly calcic plagioclase (An<90–100) and hornblende, the dunite cores, and the complete lack of plagioclase as a cumulus crystallization product are not features shared by the Philbrook intrusion.

Additionally, the Philbrook intrusion has a higher content of apatite and a much higher overall Fe/Mg ratio than the Alaskan type of zoned ultramafic complexes. Despite these differences, the Alaskan occurrences provide a useful model for some features at Philbrook.

Ontario Carbonatites

In northwestern Ontario several small concentric intrusions, mostly between 1,000 and 1,800 m.y., are present (Erdosh, 1979; Sage, 1985). These intrusions are of alkalic-carbonatitic affinity, and are largely confined to linear belts whose geometries are thought to be controlled by deep first-order faults. The carbonatites have a small diameter (<2.5 km) and show pronounced circular gravity and magnetic highs (Erdosh, 1979; Sage, 1985; Mitchell and Platt, 1978). These rock bodies typically contain pyroxenite and apatite-magnetite rock (magnetite nelsonite), and P<sub>2</sub>O<sub>5</sub> values are usually high (0.5-1.5%). These complexes commonly contain a variety of rock types ranging from pyroxenite to diorite and syenodiorite, with varying amounts of carbonatite.
Oxide-Apatite Rocks (Nelsonites)

Most nelsonite occurrences form dike-like bodies, presumably the result of fracture infillings of a gravitationally-settled immiscible split of nelsonite magma from a dioritic parent magma (Herz and Force, 1984; Kolker, 1982).

At the type locality in the Roseland district, Virginia, the layers of impure nelsonite occurring near the base of ferrodioritic-charnockitic plutons have been interpreted as the product of crystal settling. Textural, chemical, and isotopic evidence indicates that the nelsonites probably formed as an immiscible split from the charnockites, and extreme K-depletion is postulated to have led to the formation of the nelsonite (Herz and Force, 1984). Chemical triangular plots show distinctive fields between nelsonite, diorite, and anorthosite (Fig. 37), affirming an origin by immiscibility processes.

Magnetite nelsonite from Camsell Bay, N.W.T. (Badham and Morton, 1976), consists of small veins and plugs which are partly of replacement origin and are surrounded by metasomatic aureoles of a rock rich in magnetite and apatite. The authors of this report have also invoked an immiscibility hypothesis for the origin of the oxide-rich rock. In their model the nelsonite intruded as a crystal mush, as evidenced by large broken magnetite and apatite grains which are fractured and rehealed by the same.
Oxide rock (nelsonite)
- Magnetite-hornblende pyroxenite
- Mafic diorite
- Mesocratic diorite, calculated %
- Neighboring pink granite porphyry
- Mafic diorite minus plagioclase
- Plagioclase from PB-1
- Anorthosite, based on probe data of plagioclase

Figure 37. A.) AFM diagram. B.) SiO₂-(Na₂O+K₂O+Al₂O₃)-(CaO+MnO+MgO+FeO+TiO₂+P₂O₅) diagram. C.) P₂O₅-TiO₂-K₂O diagram. Fields: I = field of Roseland anorthosite; II = Roses Mill diorite and charnockite field; III = Roses Mill and Turkey Mountain nelsonite (Virginia). From Herz and Force, 1984
These occurrences are thought to be similar to the 'Andean Type' of nelsonites as described by Park (1972), in which the nelsonites are interpreted to be of both intrusive and replacement origin. Flows of magnetite-apatite rock have been documented at El Laco, Chile (Park, 1961).

These occurrences of oxide-apatite rocks, although each slightly different, all evoke an immiscible liquid hypothesis for their origin. Various processes, such as crystal settling, forceful intrusion, etc. act on the liquids and give each occurrence a different quality. Concentrations of oxides are also reported in smaller SE-Alaskan ultramafic intrusions, but these lack appreciable apatite and are more likely the result of normal differentiation rather than immiscibility.

**Experimental work on Immiscible Liquids**

Philpotts (1967) has demonstrated that nelsonites may form by the immiscible separation of high P-Ti-Fe liquid from an intermediate parent magma. Kolker (1980, 1982) has summarized and chemically analysed several occurrences of oxide-apatite rock, and affirms that a general ratio of 2:1 oxide:apatite exists, as suggested by Philpotts (1967). Oxide phases may range from magnetite to ilmenite to rutile.

More recent studies on the effects of oxygen fugacity (Naslund, 1983), effects of P2O5 and TiO2 (Visser and Koster Van Groos, 1979), and effect of magma
oxidation state on a naturally split tholeiitic basalt magma (Philpotts and Doyle, 1983) show that immiscible splits could occur in a variety of natural magma systems.

Naslund (1983) has shown that in the system \( (\text{KAlSi}_3\text{O}_8 \text{ or NaAlSi}_3\text{O}_8) - \text{FeO-Fe}_2\text{O}_3-\text{SiO}_2 \), the degree of liquid immiscibility (more divergent compositions) and the upper temperature limit of the field of separation increase with increasing oxygen fugacity. His results show that the immiscible splits are most likely to occur in systems high in total Fe, TiO\(_2\), P\(_2\)O\(_5\), and Fe\(^{+3}\)/Fe\(^{+2}\), and low in MgO, Al\(_2\)O\(_3\)/
(\(\text{Na}_2\text{O + K}_2\text{O}\)), and CaO. Al, K, and Na are more likely to be concentrated into the silica-rich split as oxygen fugacity increases. With the addition of Ca, the immiscibility gap decreases, Al becomes even more preferentially concentrated into the silica-rich fraction, and K becomes less strongly enriched in the Si-rich liquid. MgO, which is relatively depleted in the Philbrook intrusive body, also causes a marked decrease in the degree of liquid-liquid separation. Under oxidizing conditions, the order of preference of fractionation into the Fe-rich melt is: Ti>Mn>P>Mg>Ca; under reducing conditions: P>Ti>Mn>Ca>Mg (Naslund, 1983). Visser and Koster Van Groos (1979) have experimentally shown that the addition of P\(_2\)O\(_5\) and TiO\(_2\) expands the two-liquid field in the system K\(_2\)O- FeO-Al\(_2\)O\(_3\)-SiO\(_2\), and that the widening effects are 2 to 3 times as strong for P than for Ti.
Philpotts (1967), has suggested that dioritic rocks which are rich in Na (plagioclase compositions An₄₀-₅₀ are more likely to form immiscible splits than the more calcic stratiform-type layered tholeiitic intrusions. He proposes a crystallization sequence in which the early crystallization of anorthosite enriches the melt in incompatible elements. Oxide-apatite rich diorite becomes a later crystallizing phase, and when the level of P and Ti reaches a critical level the diorite magma splits into iron-rich and silica-rich fractions. The more dense nelsonite liquid sinks to the bottom of the chamber and forms dike-like bodies in underlying anorthosite. A model in agreement with this has been proposed by Herz and Force (1983) for nelsonite-rich layers and dikes which occur as layers at the bottom of a ferrodiorite body and as dikes into underlying anorthosites.
Evidence for Immiscibility at Philbrook

The following discussions will focus on two alternative possible fractionation histories for the Philbrook intrusion. The first model invokes magma immiscibility; the second involves non-immiscible magmatic differentiation.

The presence of magnetite nelsonites in the Philbrook intrusion in itself is strong evidence that immiscibility has taken place. Liquid immiscibility is a widely accepted means of deriving nelsonites. The associated dicroites are high in Ti, P, and Fe, and are therefore related to the nelsonites. The lack of contact relations and outcrop seriously hinders detailed interpretation of the petrogenesis, but some inferences can be made.

When plotted on a ternary Greig diagram and an AFM diagram (Fig. 37), the presence of distinct fields for the various rock types is shown. Furthermore, if the feldspar (An45) component is subtracted from sample PB-1 (a melanocratic, Fe-P-Ti-rich diorite), the residual composition falls near the nelsonite-pyroxenite field. The modal composition would approximate that of a hornblende-pyroxene-magnetite-apatite rock. Indeed, it shows a close similarity to the composition of Pb-3-1-74, a magnetite hornblendite.

The nelsonites of this study plot slightly out of the fields of Herz and Force (1983), a result of the lower apatite contents in the analyzed samples. However,
because the core samples used in this study have been skeletonized, probably with the objective of saving the most Fe-rich portions, there is some probability that apatite-rich material has been discarded and that the overall composition of the nelsonites may approach the 2:1 oxide:apatite ratio of Philpotts' eutectic.

The preceding discussion shows that the nelsonites may have originated as an immiscible split. The question remains, however, as to what the compositional and spatial relations are to the Si-rich immiscible fraction. One possibility is that the lighter-colored, quartz- and perthite-bearing diorites from outcrop represent this Si-rich split and that the melanocratic, oxide-apatite-Fe-Mg silicate rich diorites, which show cumulus layering features, represent an unsplit 'parental' composition. This would be consistent with the process of Philpotts (1967) that a magma may undergo fractionation to a more Fe-P-Ti rich melt via early plagioclase settling to produce a diorite rich in these incompatible elements, which upon reaching a 'critical' saturation point will split to two immiscible liquids of nelsonite and normal diorite composition. If this is the case, the intrusion should be underlain by or be in some proximity to an anorthositic body. The possibility that such a split has occurred is also feasible according to the criteria set forth by Naslund (1983) in that the following are satisfied: a.) both 'splits' have similar
FeO/MgO and MnO/FeO values; b.) higher Na/K, Al₂O₃/(Na₂O + K₂O) values and P₂O₅, TiO₂, MgO, and MnO contents are found in the Fe-rich fraction; and c.) higher K₂O, Na₂O, and Al₂O₃ contents are present in the silica-rich fraction. Furthermore, the abundant magnetite in the oxide rock and pyroxenite, and the presence of trace amounts of magmatic sphene and calcite imply that oxygen fugacity was quite high during the bulk of the crystallization of the nelsonites. This inferred high oxygen fugacity (between experimental quartz-fayalite-magnetite and magnetite-hematite buffers) would widen the immiscibility gap (Naslund, 1983). The copious amounts of (now exsolved) ilmenomagnetite indicate that the system was buffered with respect to oxygen for most of the crystallization period of the nelsonites, except perhaps near the end of magnetite crystallization when interstitial composite ilmenite grains crystallized.

As noted by Kolker (1980, 1982) most nelsonites form dike-like bodies from the settling of a nelsonite magma or crystal mush through less dense dioritic mush into fractures in the country rock beneath the magma chamber (Herz and Force, 1983). Given the ring-like shape of the oxide layers as inferred from the shape of the magnetic high, and the absence of observable contact relations, one might conclude that the oxide-rich portion of this intrusion is itself a later intrusion into the dioritic rock mass. This could have originated via crystal
settling of oxide-rich layers in the pyroxenite-hornblendite rocks; this could account for the oxide-apatite clots within the pyroxenite (p. 97).

One a typical aspect of all of the whole rock and microprobe analyses (except biotite) is the distinct paucity of K₂O in the system (<1.0%, App. D, Table D-2). This anomalously low value leads one to conclude that perhaps an immiscible split had occurred before emplacement of the mafic intrusion, and that the mafic fraction has separately intruded and cooled in a normal fractionation pattern producing the concentric shape via normal layering and settling processes (p. 138). In this case, a possible analog to the Si-rich portion of the split lies in a pink, highly potassic, porphyritic granite which has been intersected by drilling 3 km to the east of the Philbrook intrusion (Southwick and others, 1986). Whole rock chemistry shows this rock to be very much like the average granite of LeMaitre (1976), except that it has unusually high K and low Ca values. The distinguishing patterns of element distribution between the two splits, as described above (Naslund, 1983), hold true for this granite to be an immiscible split. Similar granitic splits, however, have not been described at other localities, which instead have diorite, anorthosite, or quartz porphyry in association with the nelsonites.
Based on aeromagnetic and gravity data, the granitic body is assumed to consist of two separate bodies, one of which 'wraps around' the Philbrook intrusion (MGS preliminary map; Southwick, pers. comm, 1986). Thus, if a relationship does exist between the two rock types, it would be fairly easy to invoke separate but coeval intrusions of the granite and ultramafic rocks. Upon emplacement, the Ti-Fe-P rich magma may have crystallized in a normal fractional crystallization scheme of pyroxenite to quartz-bearing diorites. The overall iron-enrichment of the plug, which is reflected in both pyroxene and amphibole solid solutions and modal mineralogy, may be the result of fractionation at depth of Mg-rich phases such as olivine and Mg-pyroxene. The mineralogy at Philbrook is fairly similar to SE-Alaskan occurrences, except for the higher Fe, P, and Ti values.

Layered Intrusion Model

The distribution (Fig. 3; App. B) of the hornblendite /pyroxenite/oxide-apatite rocks suggests that some sort of cumulus action took place within the magma chamber, and differentiation of a mafic immiscible split may have proceeded to the point at which the diorite was produced as late-stage products of normal fractional crystallization.

Textural evidence for changes of pyroxene, oxides and apatite from early cumulus to late interstitial support this mechanism. However, the 'earlier' cumulus pyroxenes
(from pyroxenite) are more enriched in Ca and Fe/Fe + Mg relative to the 'later' intersitial pyroxene in diorite (Fig. 26). This could be the result of exsolution of Fe-Ti rods in the latter, which would lower the Fe content of the pyroxene. The apparent lack of Fe-enrichment of pyroxene is possibly due to the fact that the oxygen fugacity was buffered, crystallizing magnetite and preventing Fe-enrichment.

Further evidence for a layered intrusion exists in the form of oriented plagioclase and meager compositional layering in the diorite outcrops, and an apparent compositional continuum from pyroxenite to diorite. However, timing of feldspar crystallization is vague. It is assumed that the onset of feldspar crystallization occurred after the accumulation of the ultramafic pyroxene-oxide-hornblende apatite rocks, but this is not known for certain. The complete lack of plagioclase in the ultramafic rocks could be the result of an immiscible split, as some feldspar would otherwise be expected in the ultramafic suite. Not enough data are available to conduct a detailed study of the crystallization history. A continuously represented core sample through the ultramafic zones may show intrusive relations of the pyroxenite-nelsonite suite, layering relationships, or some other revealing mineralogy or texture.

The mineralogy and the irregular dike-like shape (?) of the oxide rock promote an immiscible split origin for
Fig. 3. T-fO₂ grid for coexisting Mt, ilm. pairs, based on the solution model. The stippled fields are our estimates of the limits of the model. The Mt-ilm miscibility gap is that calculated by the model for the three-phase assemblage Mt₂ = Usph = Ilmn. The ilm-pla miscibility gap (labeled "Approx. ILM, HEM") is our best guess from experimental data—it is not calculated.

The Philbrook intrusion. A layered intrusion would most likely be more uniform in rock distribution pattern.
Figure 39. A.) Isobaric (2 kb fluid pressure) log $f_{O_2}$-T diagram of ferropargasite stability field.

B.) Isobaric (3 kb fluid pressure) log $f_{O_2}$-T diagram of ferrotremolite stability field.

From Ernst, 1968
Figure 40. Cation vs. temperature plots for experimentally produced hornblende with a magnetite-wustite buffer. Dashed line represents cation contents of cumulus amphibole from Philbrook. (I) is olivine tholeiite; (II) quartz tholeiite; (III) and (IV) alkali basalt starting compositions. Number of cations all based on 23 (O). Modified from Helz, 1981.
Temperatures of Crystallization

Coexisting Fe-Ti oxides were analyzed to determine temperature and oxygen fugacity (Buddington and Lindsley, 1964; Spencer and Lindsley, 1981). This did not yield meaningful results for the Philbrook intrusion. The results (Fig. 38) only indicate that very low temperature (subsolidous) conditions prevailed at the time of ilmenite-magnetite oxidation-exsolution.

The presence of ferropargasitic hornblende implies temperature and fugacity ranges between 850°C and $10^{-16} f_0^2$ to 590°C and $10^{-27} f_0^2$ at 2 kb fluid pressure (Ernst, 1968). Ferrotremolite is stable between 525°C - $10^{-14} f_0^2$ and 400°C - $10^{-30} f_0^2$ at 2 kb pressure (Ernst, 1968; Fig. 39). The implied temperature variations between ferropargasite and ferrotremolite are consistent with observed core to rim compositions, and possibly imply a decrease in oxygen fugacity as cooling proceeded.

A synthesis of experimental studies on igneous hornblende from the literature shows consistent, but somewhat vague compositional variations with changes in temperature of crystallization (Helz, 1982). Using these composition-temperature diagrams, results of 950 - 1000°C are obtained as a crystallization temperature for ferropargasitic hornblende (Fig. 40). This figure is beyond the stability field of Ernst (1968) for end-member ferropargasite, but well within the field for end-member
pargasite; the ferropargasitic hornblende of this study contains significant Mg in the structure and therefore this is considered a reasonable upper temperature of crystallization.

**Evidence for H₂O-Rich Magma**

Evidence that the magmatic intrusion was highly enriched with water includes: a.) the ubiquitous deuteric alteration of pyroxene to hornblende; b.) reaction rims of ferropargasitic hornblende over pyroxene and into feldspar (possibly a liquid reaction); c.) the presence of cumulus magmatic hornblende; d.) alteration of magnetite to hydrous biotite, chlorite, and amphibole assemblages; and d.) late-stage pegmatite and net veining features in the diorites, evidence for volatile streaming (Fig. 34). A wet magma is consistent with the high degree of fractionation and enrichment of other incompatible elements. A high pH₂O would cause increased fO₂, which would help immiscibility processes (p. 134).
Conclusions

From this study several conclusive statements can be made. Regarding the metasedimentary rocks west of Philbrook and the conglomerates in the outcrops east of Philbrook:

1.) About half of the stratigraphic section at Philbrook consists of arkose and feldspathic arenite. These terrigenous rocks consist dominantly of albitic feldspar and quartz, with minor chlorite, epidote, biotite, carbonate, magnetite, tourmaline, and pyrite.

2.) Tourmaline and magnetite-bearing micaceous schist comprise approximately 10% of the total stratigraphic section. Chlorite is a retrograde mineral after biotite; and commonly the albitic feldspar is heavily sericitized. Tourmaline and magnetite commonly occur together in this schist, possibly indicative of a similar source for the necessary boron and iron. The protolith for this rock type is a ferruginous 'graywacke'.

3.) Iron-formation (>20% opaque oxides) represents about 30% of the total stratigraphic section. The iron-formation also contains a significant clastic component, and locally grades into ferruginous chlorite schist. Essential iron-bearing minerals
include magnetite and specular hematite, consequently the rock is highly magnetic.

4. Epidote-rich schist and biotite- or amphibole-bearing schist comprise the remaining 10% of the stratigraphic section. The epidote-rich schist was originally metabasaltic rock, and the biotite- or amphibole-bearing schist also was derived from a mafic rock, probably a mafic tuff or volcanogenic sediment. Both rock types occur as thin, stratiform units which are most common in the upper 3/4 of the stratigraphy.

5. The metasedimentary rocks intersected by drilling west of Philbrook are most likely correlative with the Denham or Randall Formations of the Mille Lacs Group. The intersected stratigraphic sequences reflect at least two cycles of deposition, represented by arkose-iron-formation and arkose-shale sequences. The high albite content and lack of orthoclase indicate rapid weathering of a tonalitic hinterland. Na-enrichment (albitization) may also have been enhanced by hydrothermal and/or regional greenschist-facies metamorphism, and the feldspars were sericitized by later hydrothermal alteration.
6. The metasedimentary rocks west of Philbrook were metamorphosed to the upper biotite—lower amphibolite isograd of the regional greenschist facies.

7. The conglomerates in the outcrops just east of Philbrook are basal in nature, of very local provenance, and probably fluvial to shallow marine in origin. Although they are possibly stratigraphically below the metasedimentary package west of Philbrook, regional geologic relations (Southwick 1986, pers. comm.) indicate that the conglomerate may overlie these metasedimentary rocks.

8. The occurrence of disseminated tourmaline in the epiclastic rocks, as well as tourmalinite itself, is very significant. Tourmalinite is known to be associated with several massive-sulfide deposits, and the tourmaline at Philbrook is chemically similar to tourmaline associated with stratiform massive sulfide deposits at other locations.
With regard to the igneous rocks east of Philbrook which occur in outcrop and are intersected by drilling:

1.) The four general rock types of the intrusion consist of nelsonite, pyroxenite to hornblende pyroxenite, melanocratic diorite, and mesocratic diorite with local anorthositic zones. Apatite and Fe-Ti oxides are abundant in all rock types except the mesocratic diorite. The intrusion is concentrically zoned, is of approximately 2.4 sq. km. in subsurface area, and is probably steeply-walled and funnel-shaped.

2.) The igneous intrusion crystallized from a water-rich ultramafic to mafic magma which may have formed as an immiscible split from a dioritic parent magma. This mafic portion of the split either underwent further immiscible splitting, crystallized in a layered manner, or some combination of these to produce the mineralogical layering of the intrusion.

3.) The oxide-apatite rocks (nelsonites) are mineralogically similar to those at the type locality in Virginia; but the intrusion size, morphology, and hornblende-pyroxenite rocks are
similar to occurrences in southeastern Alaska (Duke Island). The intrusion as a whole does not show similarities to the Duluth Complex or Bushveld-type of tholeiitic layered intrusion.

5. The unmetamorphosed nature, brittle shear zones, and round shape of the intrusion provide evidence that it was largely unaffected by, and emplaced towards the waning stages of, the Penokean orogeny.
Appendix A

OF METASEDIMENTARY ROCK DRILL HOLES
CROSS-SECTIONS AND CORE LOGS

Key

1. Tourmaline-bearing, quartz and feldspathic arenite, white to pinkish-gray sandstone

2. Clayey-rich, tuffaceous beds

3. Interbedded claystone, carbonaceous shale, and tuffaceous sandstone

4. Mudstone, siltstone, and fine-grained sandstone

5. Quartz-feldspar matrix and feldspar-foid mixture

6. Chlorite-muscovite matrix

7. Carbonaceous-bearing chlorite-muscovite and green, well-foliated tourmaline

8. Tourmaline, chlorite, and biotite

9. Magnetite-bearing chlorite-muscovite and green, well-foliated tourmaline

10. Tourmaline-bearing quartz arenite, white to pinkish-gray sandstone

11. Interbedded claystone, carbonaceous shale, and tuffaceous sandstone

12. Clayey-rich, tuffaceous beds

13. Tourmaline-bearing, quartz and feldspathic arenite, white to pinkish-gray sandstone
Figure A-1. Subcrop map of metasedimentary rocks west of Philbrook.
Figure A-2. Cross-section of drill group 32A, facing north.
Figure A-3. Cross-section of drill group 32B, facing north.
Figure A-4. Cross-sections of groups 32-104, 5A, 5B, and 5C, facing north. See subcrop map (Fig. A-1) for location of drill groups.
<table>
<thead>
<tr>
<th>Depth (ft)</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>0 - 50</td>
<td>Glacial overburden.</td>
</tr>
<tr>
<td>50 - 180</td>
<td>Schist; gray, shiny; specular hematite-quartz.</td>
</tr>
<tr>
<td>180 - 200</td>
<td>Iron-formation; gray, finely laminated; chl-qtz-mag laminae intercalated with green-gray fsp-qtz-mag-chl laminae.</td>
</tr>
<tr>
<td>200 - 250</td>
<td>Iron-formation; gray, thinly laminated; hem-qtz-mag.</td>
</tr>
<tr>
<td>250 - 300</td>
<td>Iron-formation; gray, thinly laminated; qtz-mag-spec hem.</td>
</tr>
<tr>
<td>300 - 350</td>
<td>Less specularite, more magnetite.</td>
</tr>
<tr>
<td>350 - 400</td>
<td>Missing interval.</td>
</tr>
<tr>
<td>400 - 500</td>
<td>Schist; green, fine-grained; epi-chl-carb; metavolcanic.</td>
</tr>
<tr>
<td>500 - 600</td>
<td>Missing interval.</td>
</tr>
<tr>
<td>600 - 700</td>
<td>Iron-formation, gray, laminated; qtz-mag laminae interbedded with greenish fsp-chl-fsp-qtz laminae.</td>
</tr>
<tr>
<td>700 - 800</td>
<td>Iron-formation; gray, schistose; qtz-mag-spec hem.</td>
</tr>
<tr>
<td>800 - 900</td>
<td>Greenish feldspathic laminae interbedded with iron ftn.</td>
</tr>
<tr>
<td>900 - 1000</td>
<td>Arkose; greenish-gray, fine-grained; qtz-fsp.</td>
</tr>
<tr>
<td>1000 - 1100</td>
<td>Arkose; greenish-gray; ep-qtz-fsp; as above.</td>
</tr>
</tbody>
</table>
Glacial overburden.

Schist; green, slightly weathered; chl-bio.
Arkose; pinkish-gray, fine-grained; trm-qtz-mag.
Meta-siltstone; gray, fine-grained; pytymatically-folded qtz vein; chl-musc-fsp-qtz.
Arkose/graywacke; gray-green, fine-grained; fsp-chl-qtz.
Meta-siltstone; gray, fine-grained; fsp-chl-musc-qtz; musc
Meta-siltstone; grayish-pink, slightly micaceous; mica-fsp-qtz.
Metasiltstone or fine-grained sandstone; gray; mica-fsp-qtz.

Feldspathic metasandstone; gray-white, coarser-grained than above; mica-qtz-fsp.

Schist; dark green, well foliated, fine-grained; garnet-fsp-chl-bio-mag.
Metasiltstone; gray, fine-grained; chl-fsp-qtz; similar to above.

Iron-formation; gray, finely laminated; trm-py-fsp-bio-chl-qtz-mag.

Missing interval.
Schist; mottled grn & wht; amp-qtz-fsp-chl; probable metagraywacke.
Iron-formation; gray, laminated; act-qtz-mag.
Schist; dark green, soft, sheared and lineated; chlorite.
Iron-formation; gray, laminated; quartz-magnetite with minor qtz-fsp-chl-mag laminae.

Actinolite porphyroblasts.
Hole 32-102, p.2

Quartz-carbonate vein.

Iron-formation; gray, laminated; specular hematite-qtz-mag.

Schist; mottled green & white as above; fsp-qtz-chl; metagraywacke.

Iron-formation; gray, laminated; hematite-qtz-mag.

Schistose fsp-chl laminae in vuggy iron-formation.

Iron-formation; specular.

Iron-formation; black & white, laminated; specular hematite schist.

Iron-formation; gray, act-qtz-mag.

Missing interval.

Iron-formation; green-gray; chl-qtz-mag.
Hole 32-102, p.3

Arkose; light gray; tra-chl-qtz-fsp.

Schist; dark green sheared, foliated and well lineated; qtz-fsp-carb-amph-chl.

Schist; green, laminated (?); carb-chl-amph laminae intercalated with fsp-epi laminae; possible pseudolaminae due to shear.

Schist; green, weakly crenulated; fsp-ep-chl.

Iron-formation; dark gray, laminated; spec hem-qtz-mag.
Hole 32-103 P.1

Glacial overburden.

Schist; brown-green, slightly oxidized; fsp-qtz-musc-epi-chl; granular weathered appearance; meta-graywacke or mafic tuffaceous rock.

Schist; green, well foliated; with white, elliptical, altered and flattened, fine-grained felsic rock fragments or feldspar crystals. Matrix of tma-mag-fsp-qtz-bio-epi-chl; clasts composed of epi-qtz-sericite.

Crystal tuff? Similar to above, but sericite fragments are blocky-shaped and suggestive of highly sericitized feldspar crystals.

Schist; greenish-gray, slightly oxidized, sandy-appearance; fsp-qtz-mica; meta-siltstone (?)..

Schist; Iron-formation, Arkose; interbedded on a fine scale; mag-qtz-chl; qtz-mag; qtz-fsp

Arkose; gray-white, weakly schistose; musc-qtz-fsp.

Iron-formation; gray-green, laminated; qtz-mag laminae intercalated with fsp-qtz-chl laminae.

Schist; light gray, weakly foliated; qtz-fsp-chl-musc

Arkose; light pink-gray, massive; qtz-fsp-mica; schistose.

Schist; dk green, lineated; sheared chl-carb, carb veins.

Arkose; pinkish-gray, brecciated; qtz-fsp, healed w/quartz veins.

Iron-formation; gray, laminated; qtz-mag w/green fsp-qtz-chl laminae.

Schist; green; tma-mag-fsp-qtz-chl-musc; possible volcanic rock fragments.
Hole 103, p.2

**Iron-formation; dark gray-green,qtz-chl-mag**

<table>
<thead>
<tr>
<th>Depth</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>0</td>
<td>Schist; dark green, well foliated; fsp-carb-bio-chl; metavolcanic.</td>
</tr>
<tr>
<td>116</td>
<td>Schist; mottled dark green and white; fsp-carb-qtz-bio-chl; blocky white sericitic clots (volcanic rock fragments).</td>
</tr>
<tr>
<td>52</td>
<td>Schist; soft, green, sheared and greasy-feeling; chlorite. Entire interval represented by quartz vein.</td>
</tr>
<tr>
<td>116</td>
<td>Iron-formation; gray, laminated; quartz-magnetite.</td>
</tr>
<tr>
<td>40</td>
<td>Iron-formation; gray, contorted laminae; qtz-mag-act, with interbedded green fsp-qtz-chl schist laminae.</td>
</tr>
<tr>
<td>52</td>
<td>Iron-formation; pink and gray, laminated; qtz-carb-mag.</td>
</tr>
<tr>
<td>40</td>
<td>Missing interval.</td>
</tr>
<tr>
<td>116</td>
<td>Iron-formation; gray, laminated; qtz-mag.</td>
</tr>
<tr>
<td>52</td>
<td>Iron-formation; gray, laminated; qtz-mag.</td>
</tr>
<tr>
<td>52</td>
<td>Iron-formation; gray, laminated; qtz-mag.</td>
</tr>
<tr>
<td>55</td>
<td>Iron-formation; gray, laminated; qtz-mag.</td>
</tr>
<tr>
<td>55</td>
<td>Iron-formation; gray, laminated; qtz-mag.</td>
</tr>
<tr>
<td>55</td>
<td>Iron-formation; gray, laminated; qtz-mag.</td>
</tr>
<tr>
<td>557</td>
<td>Iron-formation; gray, laminated; qtz-mag.</td>
</tr>
<tr>
<td>555</td>
<td>Iron-formation; gray, laminated; qtz-mag.</td>
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<tr>
<td>555</td>
<td>Iron-formation; gray, laminated; qtz-mag.</td>
</tr>
<tr>
<td>557</td>
<td>Iron-formation; gray, laminated; qtz-mag.</td>
</tr>
<tr>
<td>555</td>
<td>Iron-formation; gray, laminated; qtz-mag.</td>
</tr>
</tbody>
</table>

Arkose; as above, but gray.

Arkose; pink, coarse-grained; trq-qtz-fsp.

Arkose; as above, gray-white.
### Hole 32-104, p.1

**Glacial overburden.**

Schist; soft, weathered, yellow-green; qtz-fsp-mica.

Schist; dark brown-green, laminated, slightly weathered; bio-chl-hem-mag-qtz

Iron-formation; dark gray, thinly laminated; quartz-magnetite, with green, foliated actinolite porphyroblasts.

Schist; light green; fsp-qtz-chl.

Chlorite porphyroblasts.

Schist; green, laminated; chlorite; quartz vein.

Schist; dark gray-green; epi-carb-bio-amph. (metagraywacke?).

Schist; gray, thinly laminated; hematitic; minor intercalated green qtz-fsp-chl laminae.

Schist; light green, weakly crenulated; qtz-fsp-chl-musc.

Schist; gray and white, very finely laminated; specular hematite.

Amphibolite; mottled dark green and white; qtz-fsp-carb-bio-amph.

Arkose; pink-gray, fine-grained; qtz-fsp rich.

Schist; soft, green; fsp-qtz-epi-amph-carb-bio-chl; meta-volcanic; carbonate veinlets.

Schist; gray and white, finely laminated, fine-grained; specular hse-qtz, laminae defined by varying hse-qtz content.
Hole 32-104, p.2

Schist; gray and white, finely laminated; specular hematite.

Schist; light gray, fine-grained; quartz-feldspar-chlorite.

Schist; green, foliated and lineated; feldspar-quartz-biotite.

Schist; gray and white, finely laminated; specular hematite. Small folded and broken quartz vein.

Feldspathic quartzite; vitreous gray, fine-grained, massive; mica-quartz-feldspar.
Hole 32-105, p.1

Glacial overburden.

- Schist; gray & white, well foliated, thinly laminated; specular hematite; slightly oxidized.

- Shear zone; soft and powdery, yellow iron-oxides.

- Schist; white-gray, crenulated; muscovite.

- Schist; thinly laminated; hematitic; alternating with feldspathic clastic-dominated laminae.

- Schist; greenish-gray, fine-grained; chlorite-hematite.

- Schist; green-gray, crenulated; muscovite

- Schist; green-gray; tra-mag-fsp-musc-qtz-chl.

- Schist; greenish-gray; quartz-feldspar-chlorite.

- Schist; gray, thinly laminated; hematite.

- Shear zone; chloritized - brecciated.

- Schist; feldspar-muscovite.
Hole 32-105, p.2

Schist; green & white, fragmental clastic texture (Metagraywacke); sericite-carbonate-amphibole-chlorite.
Soft sediment Z-folds

Iron-formation; gray, fsp-qtz-mag laminae alternating with green, clastic-dominated chl-qtz-fsp laminae.

Arkosic metasediments; grayish-white, massive to weakly bedded; quartzo-feldspathic, tourmaline-bearing.

Arko; white; qtz-fsp-sc, with black tourmalinite laminae.

Vein; thin, carbonate.

Missing Interval

Schistose meta-graywacke; grayish-green; quartz-feldspar-chlorite.

Metagraywacke/arkose; light gray-green; epidote-quartz-feldspar.
Hole 32-106

Glacial overburden.

Arkose; grayish-green, fine-grained; mica-qtz-fsp.
Arenite; grayish-white; trm-epi-mica-qtz-fsp.

Schist; green-gray, thinly laminated; fsp-qtz-mag-chl; intercalated with gray, qtz-mag iron-formation.

Arkosic sandstone and/or siltstone; grayish-green, fine-grained; mica-epi-trm-qtz-fsp.

Arkose; darker gray than above.

Arkose or feldspathic arenite; pink-gray, fine-grained; trm-qtz-fsp.
Hole J2-107, p. 1

Glacial overburden.

Iron-formation; greenish-gray, laminated; qtz-mag laminae intercalated with greenish qtz-fsp-chl laminae.

Iron formation; grayish than above; less of the qtz-fsp-chl laminae.

Schist; green, finely laminated; qtz-fsp-chl-bio.

Metasiltstone; gray, massive; qtz-fsp rock; minor pyrite.

Iron-formation; gray, laminated; qtz-mag.

Schist; dark green, weakly laminated, sheared; py-fsp-qtz-chl.

Schist, green, fine-grained; mica-fsp-qtz; metasiltstone (?)?

Argillite, gray, fine-grained; magnetite-rich; fsp-mag-mica-qtz.

Meta-siltstone or sandstone; gray, weakly schistose; tra-py-mica-fsp-qtz.

Arkose; pinkish-gray, medium-coarse grained; qtz-fsp, probable tra.

Brecciated; quartz healings and veins.

Arkose; pink, fine-wed grained; tourmalined-bearing; qtz-fsp.
Hole 32-108.p.1

Glacial overburden.

Schist; brownish-green, laminated, slightly oxidized; tourmaline-magnetite-quartz-albite-chlorite-biotite.

Schist; as above grading into iron-formation below.
Iron formation; greenish-gray, weakly laminated; fsp-chl-qtz-mag.

Iron-formation; gray and white; less fsp and chl than above; more qtz-mag.

Schist; greenish-white; chl-qtz-fsp.
Iron-formation; green-gray; chl-qtz-mag.
Schist; green; tourmaline-bearing, similar to top unit.
Iron-formation; gray and white, laminated; fsp-qtz-chl-mag.
Schist; gray-green, laminated; tra-mag-musc-fsp-qtz-chl, muscovite porphyroblasts.
Iron-formation; gray, weakly laminated; fsp-chl-qtz-mag.

Schist; green-gray, laminated; tra-mag-musc-fspar-qtz-chl, muscovite porphyroblasts; gradational from the iron-formation above.

Schist; grn, f.gr., well foliated; epi-carb-fsp-qtz-chl; maf. mvolc.
Schist; grn-gry, f.gr., well foliated; tra-mag-bgr-fsp-qtz-chl, musc porphyroblasts; also flattened quartz-sericite volcanic rock fragments.
Arkose or feldspathic arenite; pinkish-gray, fine-grained; quartz-Na plagioclase (Rud. Tuff?).

Iron-formation; dark green, fine-grained; fsp-chl-qtz-mag.

Arkose; similar to above, but darker gray with chloritized shear zones and chloritic laminae.
Schist; soft, green, sheared, foliated and lineated; mag-chl-carb.

Arkose or feldspathic arenite; pinkish-gray, fine-grained, massive to weakly bedded; possibly a recrystallized tuffaceous rock; quartz-feldspar-sericite.

Missing interval.
Schist; dark green, soft, well foliated, sheared; mag-qtz-chl-carb; possible feldspar or volcanic rock fragments.
Interlaminated amphibolite, iron-formation, and arkose.
Arkose; pinkish-gray, micaceous, fine-grained; qtz-chl-fsp.

Schist; light green, fine-grained; fsp-epi-qtz-carb-chl; metavolcanic or mafic tuff; minor pyrite.

Arkosic; pinkish-gray to gray, fine-grained; chl-fsp-qtz; finer grained than the above arkosic units.
Schist; green, thin chloritic laminae within arkose.

Arkose; gray-pink, weakly magnetic; magnetite-quartz-feldspar

Slightly brecciated.

Schistose arkosic rock; grayer and more micaceous than above.

Metasiltstone; gray, fine-grained; mica-fsp-qtz.

Iron-formation; gray; fsp-qtz-mag, smeared into green epidote-chlorite schist.

Schist; green-gray, fine-grained, laminated; tra-mag-fsp-qtz-mica; similar to that at top of hole.

Quartz-carbonate veinlets.

Iron formation; gray and green, thickly laminated, fine-grained; gray fsp-qtz-chl-mag beds and green mag-fsp-qtz-chl-epi beds; broken and sheared.

Missing interval.

Schist; light gray, laminated, fine-grained; trm(?)-mag-qtz-fsp-mica.

Iron-formation; gray; laminated; laminations defined by varying shades of gray (qtz-mag variations); fsp-qtz-mag.

Carbonate veinlets.

Quartz veins.
Meta-siltstone (arkose); gray, fine-grained, mica-quartz-feldspar.
Iron-formation; gray, laminated; qtz-mag as above.

Schist; gray, crenulated; muscovite w/magnetite porphyroblasts; minor pyrite.
Iron-formation; black and white, laminated, weakly brecciated; qtz-mag as above.

Schist; green, well foliated; chlorite-carbonate, metavolcanic.
Schist, light gray, crenulated; muscovite; as above.
Iron-formation, black and white, laminated, fine-grained; qtz-mag, similar to above.

Quartz-epidote veins.
Iron-formation; gray and white, laminated, fine-grained; carbonate-rich, bedded carb-mag and carb-healed fractures.

Schist; green, soft, well foliated; chl-carb; probably mafic metagraywacke; asymmetric 'Z' fold.
Iron-formation; black and white, thinly laminated; carbonate-rich; carbonate occurs as thin white laminae between dark gray-black qtz-mag rich laminae.

Iron-formation; dark gray, laminated; as above, but less carb.

Arkose; pinkish-gray, fine-grained; qtz-feldspar, with thin mag-rich laminae.

Iron-formation; gray, laminated, qtz-mag.
Arkosic rock; light greenish-pink, fine-grained; chl-qtz-fsp.
Schist; green, fine-grained, massive; chlorite-carbonate-epidote.
Iron-formation; grayish, massive; fsp-qtz-chl-mag.

Iron-formation; dk. gray, laminated, fine-grained; fsp-chl-qtz-mag.

Missing interval.
Schistose laminae within iron-formation; possible blocky feldspar crystals; mag-chl-fsp-qtz.
Iron-formation; black and white, thinly laminated; qtz-mag laminae interbedded with greenish fsp-qtz-chl laminae.

Iron-formation; black and white, thinly laminated; quartz-magnetite laminae intercalated with greenish fsp-chl-qtz.
Arkose; gray, sheared and veined; py-chl-qtz-fsp.

Schist; green, well-foliated, fine-gr.; mag-carb-chl-qtz-fsp-epi; mafic metavolcanic, relict igneous texture of an altered basaltic or andesitic rock is visible; quartz-carbonate veins.

Iron-formation; black and white, thinly laminated; fsp-qtz-mag.
Schist (metasiltstone); light green, trm-mag-epi-fsp-qtz-chl;
similar to 135-180', but less chlorite and magnetite and more quartz.
### Hole 32-A-2

<table>
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<tr>
<th>Layer</th>
<th>Sample</th>
<th>Depth (m)</th>
<th>Description</th>
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<tbody>
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<td>Glacial overburden.</td>
<td>Schist; grayish-green, weakly laminated, slightly weathered; muscovite-quartz-feldspar-chlorite.</td>
<td></td>
<td></td>
</tr>
<tr>
<td>NM</td>
<td>B</td>
<td>5</td>
<td>150</td>
</tr>
<tr>
<td>F</td>
<td>vt</td>
<td>100</td>
<td>Schist; greenish-gray, fine-grained; magnetite-chlorite-biotite. Iron ftn.; black &amp; white, laminated; mag-qtz; carb-lined fractures.</td>
</tr>
<tr>
<td>M</td>
<td>Sk</td>
<td>vt</td>
<td>Schist; slightly hematitic chlorite; chlorite-lined slickensides. Iron ftn.; black and white, laminated; weakly oxidized; qtz-chl-mag.</td>
</tr>
<tr>
<td>B</td>
<td>vt</td>
<td>200</td>
<td>Schist; grayish-green, laminated; fsp-bio-qtz-chl-mag.</td>
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<tr>
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<td>B</td>
<td>vt</td>
<td>250</td>
</tr>
<tr>
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<td>S</td>
<td>B</td>
<td>5</td>
</tr>
<tr>
<td>M</td>
<td>S</td>
<td>10</td>
<td>Schist; greenish-gray, qtz-fsp-chl-bio; contains laminae of dark gray magnetite-rich schist.</td>
</tr>
<tr>
<td>NM</td>
<td>M</td>
<td>B</td>
<td>7</td>
</tr>
<tr>
<td>S</td>
<td>B</td>
<td>3</td>
<td>Schist; grayish-green, laminated; qtz-fsp-chl interbedded with black &amp; white quartz-feldspar-chlorite-magnetite iron ftn. Iron formation; black &amp; white, laminated; quartz-magnetite.</td>
</tr>
<tr>
<td>M</td>
<td>B</td>
<td>7</td>
<td>Schist; dark green, slightly crenulated; qtz-fsp-chl-bio.</td>
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<tr>
<td>S</td>
<td>NM</td>
<td>300</td>
<td>Schist (Arkose); light green; epidote-chlorite-quartz-feldspar; darker and more magnetite-rich near bottom. Iron ftn; chloritic, possibly grades from above.</td>
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<tr>
<td>S</td>
<td>W</td>
<td>B</td>
<td>7</td>
</tr>
<tr>
<td>NM</td>
<td>B</td>
<td>7</td>
<td>ECH 339</td>
</tr>
</tbody>
</table>
Glacial overburden.

Schist; dark green, fine-grained, well foliated & slightly crenulated; quartz-feldspar-biotite-muscovite-chlorite.

Schist; darker green than above, crenulated; py-amph-bio-chl.

Schist; green; weakly crenulated; biotite-muscovite.

Schist; fine-grained, chlorite-muscovite; similar to that at top.

Schist; dark green, crenulated; biotite-muscovite.

Schist; green, crenulated chlorite-muscovite.

Schist; green, soft, fine-grained chloride w/euhedral pyrite.

Schist; dark green and crenulated; biotite.

Schist; gray-green, fine-grained; quartz-feldspar-biotite-chlorite. Chlorite porphyroblasts.

Schist; dark green; quartz-feldspar-chlorite-biotite. Vein; quartz-carbonate; pyrite in vein and matrix.

Schist; green, soft; chlorite-carbonate; w/very thin black schistose, magnetite-rich laminae; pyrite throughout.

Schist; green-black; chl-mag; more magnetite than above; carbonate-pyrite veins and stringers.

Schist; fine-gr., well foliated, muscovite-quartz-feldspar-chlorite.

Schist; bright green, slightly crenulated; pyrite-rich; chloride; intercalated thin magnetite-rich chloritic laminae.

Iron formation; dark gray, slightly crenulated, laminated; magnetite-chlorite, euhedral pyrite.

Schist; green, chlorite, with muscovite porphyroblasts and pyrite.

Schist; bright green; chlorite, with thin gray carbonate-magnetite rich chloritic schist laminae; pyrite throughout. Iron formation; gray and white, laminated; carbonate-magnetite.

Schist; gray-green, chlorite-biotite-quartz-feldspar; some dark green, more chloritic laminae.
Hole 32- A-3, p. 2

Schist; gray-green, crenulated; chlorite-biotite-muscovite

Iron formation; gray and green; thinly laminated; fsp-chl-qtz-ser; with gray-green, elastic, quartz-feldspar-mica laminae.

Vein; dark gray; quartz.
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<th>B vt</th>
<th>ED</th>
<th>Depth</th>
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<tr>
<td></td>
<td>B vt</td>
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<td>223</td>
<td>223</td>
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</tbody>
</table>

Hole 32-A-5

Glacial overburden.

- **Schist; greenish-brown, oxidized; hematite.**

- **Schist; yellow-brown, slightly oxidized; hematite.**

- **Schist; brown, slightly oxidized; hematite.**

- **Schist; greenish-gray, fine-grained; feldspar-magnetite-chlorite.**

- **Schist; soft, light green, fine grained; chlorite.**

Meta-siltstone; gray magnetite-feldspar-chlorite laminae alternating with greener feldspar-chlorite laminae.
Hole 5-2

Glacial overburden.

Schist; greenish-gray, thinly laminated, oxidized; magnetite-rich qtz-fsp-chl-musc laminae and tra-qtz-fsp-mica laminae.

Missing interval.

Iron ftn.; brown, oxidized; qtz-chl-mag laminae intercalated with mag-qtz-chl-fsp laminae.

Schist; gray-green, oxidized; qtz-fsp-mica strata alternating with qtz-fsp-chl-mag strata.

Schist; mottled dark green and white; qtz-fsp-bio-chl-hbl-carb; possibly a metagraywacke.

Missing Interval.

Varied lithology; feldspathic metagraywacke; crenulated ep-mag-qtz-bio-chl sch; black & white carb-mag iron-ftn with traces of pyrite, chalcopyrite, bornite.

Hole 5-1

Missing Interval

Arkose; greenish, med-grained; quartz-Wa plagioclase; possibly quartz-phyric meta-dacite (?)
Hole 5-7

Glacial overburden.

Schist; red-gray, thinly laminated; specular hematite.

Iron-formation; gray; fsp-qtz-chl-mag with green qtz-fsp-chl-epi laminae.
Iron-formation; green and black; chlorite-magnetite laminae interbedded with clean quartz-magnetite laminae.
Iron-formation; black and dusty; possibly manganiferous.
Iron-formation; light green-gray; fsp-qtz-chl-mag w/carbonate veins.

Hole 5-8

Glacial overburden.

Schist; light green, slightly crenulated; tmt-qtz-fsp-chl-musc & euhedral pyrite; similar to that at the top of hole 8 108, but lacks the abundant muscovite porphyroblasts and contains more pyrite.
Increasing volume of pyrite.
Metagraywacke; gray-green, fine-grained; qtz-chl-fsp-ep.

Hole 5-11

Glacial overburden.

Meta-arkose or feldspathic arenite; pinkish-gray, massive, fine-gr; qtz-fsp; possibly a quartz-phryic dacitic rock.
Hole 5-10

Glacial overburden.

Schist; gray; specular hematite w/red oxidation dust.

Schist; gray; hematitic; more mag-rich than above; minor quartz-pyrite veins.

Schist; reddish-gray, slightly oxidized; specular hematite.

Missing interval.
Iron-formation; less specular, vuggy, more mag-rich than above; streaky green and white laminae of fsp-qtz-epi-chl schist.

Iron-formation; gray; quartz-magnetite; fractured and brecciated, healed with carbonate.

Missing interval.
Schist (mafic metavolcanic); streaky dark green and white; chlorite-carbonate-quartz-feldspar-amphibole.
Hole 5-12

Glacial overburden.

Schist; red-gray, well foliated, laminated, fresh; chl-qtz-spec hematite laminae due to varying quartz/hectite/chlorite content.

As above.

Iron-formation; greener, stronger black and white than above, laminated; hematite-chlorite-chert.

Carbonate-filled veinlets and fracture fillings.

Schist; dark gray and green, laminated; qtz-hectite laminae alternating with fsp-chl-qtz-hectite laminae.

Schist; dark green, lineated; carb-hbl-bio-chl-fsp; metavolcanic.

Schist; gray, laminated; qtz-spec hematite laminae w/minor intercalated green fsp-chl-qtz laminae.

Missing interval.
5·12. p.2

Pyrite along quartz-carbonate veins.

Hole 5-9

Glacial Overburden.
Iron-fatty; gray-gray, laminated, sl. brecc., healed w/ carb; qtz-mag.
Hole 5-15

Glacial overburden.
Arkose (Red Tuff?); green-pink-gray, fine-grained, massive; qtz-fsp.

Missing interval.

Schist; streaky grn-gray and white, laminated; gray qtz-mag laminae
alternating with green qtz-fsp-chl laminae.
Missing interval.

Schist; gray-green, laminated, magnetite-rich; similar to the
schist above the missing interval.

Iron-formation; gray or mottled green-white, laminated, probable
tourmaline-bearing; (trm)-fsp-chl-qtz-mag; poorly represented.

Schist; green, possibly laminated (sheared); qtz-fsp-ep-chl.

Missing interval.

Iron-formation; gray, laminated; quartz-magnetite.

Missing interval.

Schist; mottled green & white, well foliated, lineated, probable
volcanic metagraywacke; qtz-carb-fsp-ep-chl; minor pyrite.
Appendix B

CROSS-SECTIONS AND CORE LOGS OF DRILL HOLES INTO PHILBROOK INTRUSION

Figure B-1. Location map of drill holes in Sec. 3, T. 133 N., R. 32 W. which penetrate igneous rocks.
Figure B-2. Cross-section of drill holes penetrative into igneous intrusive body. See Fig. B-1 and Fig. 1 for drill hole locations.
Figure B-3. Cross-sections of drill holes penetrative into igneous intrusive body. See Fig. B-1 and Fig. 1 for drill hole locations.
Lithological symbols for igneous rock core logs:
- Magnetite and ilmenite (Fe-Ti oxides)
- Hornblende and pyroxene
- Apatite
- Plagioclase

- Missing interval

41-47 Equigranular oxide gabbro; 1-3 mm.
47-67 Equigranular oxide gabbro; 1-3 mm; Fe-carbonate-quartz-pyrite fracture linings.

67-70 Equigranular oxide pyroxenite
74-80 Oxides occur in clots.
80-83 White carb. fracture lining.
83-86 Disseminated and fracture controlled Chalcopryite.
86-91 Weak igneous lamination def. by tabular hornblende.

55-60 Granular oxide-gabbro.; 1-3 mm.
72-77 Amphibole up to 7 mm.
85-90 Fe-Mg silicate is chlorite; apatite is clustered into zones.

130 Zones of oxides; thin white carb fractures; amphibole up to 8 mm.
140-145 White carb and cpy in fracs; apatite up to 1.5 mm in clusters.
150-155 Grain size and px/amph to oxide ratio varies greatly; 1mm - 1cm.
56-60  Shrd hbl-pxenite; carb and epid veins; trace sauss plag; < 2 mm.

78-85  Sheared; as above.

100-106 Pink, en echelon carb veins; sheared, soft.

119-122 Fine-grained, chlorite schist; sheared, soft.

138-140 Tabular, sauss plag; 2-6 mm.

179-184 Locally shrd pxenite; < 2.5 mm.

187-198 Shrd chlr sch; carb veins; 1-3 mm.

203-208 Oxide rock w/apat conc; 1-5 mm.

208-218 Oxide-apatite rock; < 4 mm.

218-220 Ox-apat-pxenite; 1-7 mm; tr sul.

230-245 Shrd pxenite, soft. Tr cpy & bn on fracs; < 3 mm.

45-50  Oxide-apatite pyroxenite; 1-3 mm.

50-55  Laminated tabular amph; 2-10 mm.

63    C. gr. hornblendite; 2-15 mm.

37-38  Grn & blk mag-hb-pxenite; 2-7 mm.

42-47  F.gr. black oxide rock; 1-3 mm.

57-62  Ox & apat conc into 1 cm clots in pxenite.
45-50 Tabular sauss plag; > 2mm.
55-60 Apat & sul-bearing pxenite.

37-77 Slightly sheared pyroxenite.
77-85 Strongly sheared chlorite schist.
85-95 Weakly sheared pyroxenite.

55-60 1-3 mm.
60-80 Tabular plag stained pink; <10 mm.
80-92 Cpy disseminated & in veins.

50-58 Feldspar stained pink; < 1 cm.

50-60 Feldspars < 1 cm; altered white.
60-70 Fresh oxide-pyroxenite; 1-6 mm.
70-74 Feldspar stained pink; 1-6 mm.

122-130 Slightly sheared hornblende-pyroxenite; 1-8 mm.
APPENDIX C

MICROPROBE ANALYSES, WHOLE ROCK GEOCHEMISTRY, AND ASSAY RESULTS
### Table C-1: Microprobe Analyses of Tourmaline

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<thead>
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<th>SAMPLE#</th>
<th>SiO₂</th>
<th>Al₂O₃</th>
<th>Na₂O</th>
<th>CaO</th>
<th>K₂O</th>
<th>FeO</th>
<th>MgO</th>
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<tr>
<td>T1-M</td>
<td>34.23</td>
<td>33.62</td>
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<td>30.78</td>
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<td>1.91</td>
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<td>30.3</td>
<td>31.38</td>
<td>30.65</td>
<td>32.15</td>
<td>29.98</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>24.88</td>
<td>27.63</td>
<td>24.90</td>
<td>28.24</td>
<td>23.37</td>
</tr>
<tr>
<td>Na₂O</td>
<td>1.93</td>
<td>2.17</td>
<td>1.96</td>
<td>2.33</td>
<td>2.60</td>
</tr>
<tr>
<td>CaO</td>
<td>2.06</td>
<td>1.62</td>
<td>2.05</td>
<td>1.28</td>
<td>1.95</td>
</tr>
<tr>
<td>K₂O</td>
<td>0.16</td>
<td>0.13</td>
<td>0.15</td>
<td>0.10</td>
<td>0.15</td>
</tr>
<tr>
<td>FeO</td>
<td>10.55</td>
<td>9.69</td>
<td>10.49</td>
<td>8.16</td>
<td>10.33</td>
</tr>
<tr>
<td>MgO</td>
<td>7.48</td>
<td>6.83</td>
<td>7.44</td>
<td>7.10</td>
<td>7.22</td>
</tr>
</tbody>
</table>

| Si     | 4.714  | 4.674   | 4.738   | 4.752   | 4.805   |
| Al     | 4.549  | 4.881   | 4.545   | 4.919   | 4.413   |
| Na     | 0.580  | 0.531   | 0.568   | 0.667   | 0.620   |
| Ca     | 0.343  | 0.260   | 0.340   | 0.202   | 0.334   |
| K      | 0.031  | 0.026   | 0.030   | 0.020   | 0.031   |
| Fe     | 1.369  | 1.215   | 1.356   | 1.008   | 1.384   |
| Mg     | 1.731  | 1.527   | 1.715   | 1.564   | 1.726   |
DESCRIPTIONS OF TOURMALINE FROM MICROPROBE ANALYSES

<table>
<thead>
<tr>
<th>Sample</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>T1-M</td>
<td>Greenish-black to greenish-brown pleochroic tourmaline core, in arkosic rock. DDH #A-1, 175' depth.</td>
</tr>
<tr>
<td>T1-M'</td>
<td>Midway between core and rim, same crystal as T1-M.</td>
</tr>
<tr>
<td>T1-M''</td>
<td>Greenish-black to greenish-brown tourmaline, in arkosic rock. DDH #A-1, 175' depth.</td>
</tr>
<tr>
<td>T1-M'''</td>
<td>Greenish-black to greenish-brown tourmaline, in arkosic rock. DDH #A-1, 175' depth.</td>
</tr>
<tr>
<td>T2-M</td>
<td>Greenish-black to greenish-brown tourmaline, in arkosic rock. DDH #A-1, 175' depth.</td>
</tr>
<tr>
<td>T2-175</td>
<td>Midway between core and rim, same crystal as T1-M.</td>
</tr>
<tr>
<td>T3-175</td>
<td>Greenish-black to greenish-brown tourmaline, in arkosic rock. DDH #A-1, 175' depth.</td>
</tr>
<tr>
<td>T7-175</td>
<td>Greenish-black to greenish-brown pleochroic tourmaline core, in arkosic rock. DDH #A-1, 175' depth.</td>
</tr>
<tr>
<td>T9-175</td>
<td>Rim of T7-175.</td>
</tr>
<tr>
<td>T15-175</td>
<td>Rim of T15-175.</td>
</tr>
<tr>
<td>T16-175</td>
<td>Greenish-black to greenish-brown tourmaline, in arkosic rock. DDH #A-1, 175' depth.</td>
</tr>
<tr>
<td>T19-175</td>
<td>Greenish-black to greenish-brown tourmaline, in arkosic rock. DDH #A-1, 175' depth.</td>
</tr>
</tbody>
</table>

1 All of microprobed tourmaline is intermediate schorl-dravite in composition.

2 See Appendix A for drill hole locations.
### TABLE C-2. Microprobe Analyses of Feldspar

<table>
<thead>
<tr>
<th>SAMPLE</th>
<th>P1-10</th>
<th>P1-10</th>
<th>P1-10</th>
<th>P1-06</th>
<th>P1-06</th>
<th>P2-06</th>
<th>P2-06</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>54.81</td>
<td>56.23</td>
<td>55.25</td>
<td>55.41</td>
<td>54.66</td>
<td>55.06</td>
<td>55.06</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>28.47</td>
<td>27.55</td>
<td>27.71</td>
<td>27.74</td>
<td>27.50</td>
<td>27.57</td>
<td>27.56</td>
</tr>
<tr>
<td>Na₂O</td>
<td>5.84</td>
<td>6.34</td>
<td>6.31</td>
<td>6.35</td>
<td>6.15</td>
<td>6.45</td>
<td>6.30</td>
</tr>
<tr>
<td>CaO</td>
<td>10.27</td>
<td>9.28</td>
<td>9.35</td>
<td>9.75</td>
<td>9.51</td>
<td>9.64</td>
<td></td>
</tr>
<tr>
<td>K₂O</td>
<td>0.16</td>
<td>0.13</td>
<td>0.14</td>
<td>0.21</td>
<td>0.15</td>
<td>0.13</td>
<td>0.14</td>
</tr>
<tr>
<td>FeO</td>
<td>0.16</td>
<td>0.10</td>
<td>0.15</td>
<td>0.24</td>
<td>0.03</td>
<td>0.00</td>
<td>0.13</td>
</tr>
<tr>
<td>TOTAL</td>
<td>99.82</td>
<td>99.74</td>
<td>99.34</td>
<td>98.32</td>
<td>99.54</td>
<td>99.14</td>
<td></td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Number of Ions based on 8 (D)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Si</td>
</tr>
<tr>
<td>Al</td>
</tr>
<tr>
<td>Na</td>
</tr>
<tr>
<td>K</td>
</tr>
<tr>
<td>Fe</td>
</tr>
</tbody>
</table>

| Si     | 2.527 | 2.543 | 2.490 | 2.544 | 2.544 | 2.503 | 2.511 | 2.025 |
| Al     | 1.453 | 1.452 | 1.494 | 1.470 | 1.450 | 1.485 | 1.484 | 1.165 |
| Na     | 0.530 | 0.556 | 0.628 | 0.570 | 0.616 | 0.556 | 0.556 | 0.500 |
| Ca     | 4.441 | 4.445 | 4.487 | 4.454 | 4.422 | 4.474 | 4.423 | 0.122 |
| K      | 0.010 | 0.007 | 0.009 | 0.008 | 0.006 | 0.009 | 0.008 | 0.004 |
| Fe     | 0.012 | 0.009 | 0.120 | 0.000 | 0.000 | 0.009 | 0.002 | 0.000 |

| TOTAL  | 100.01 | 100.75 | 98.00 | 97.92 | 98.79 | 98.62 | 98.77 | 98.06 |

### (Quartz)

<table>
<thead>
<tr>
<th>Sample</th>
<th>PTH2-06</th>
<th>PTH3-06</th>
<th>P16-PA</th>
<th>P17-PA</th>
<th>P18-PA</th>
<th>P19-PA</th>
<th>P19-T175</th>
<th>QDB-2</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>65.95</td>
<td>67.38</td>
<td>50.73</td>
<td>52.26</td>
<td>52.70</td>
<td>50.24</td>
<td>65.79</td>
<td>93.14</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>20.46</td>
<td>19.56</td>
<td>27.54</td>
<td>28.42</td>
<td>28.05</td>
<td>28.06</td>
<td>28.66</td>
<td>89.11</td>
</tr>
<tr>
<td>Na₂O</td>
<td>11.37</td>
<td>11.87</td>
<td>5.33</td>
<td>5.30</td>
<td>5.74</td>
<td>3.91</td>
<td>10.67</td>
<td>0.10</td>
</tr>
<tr>
<td>CaO</td>
<td>1.42</td>
<td>0.63</td>
<td>10.66</td>
<td>10.60</td>
<td>9.33</td>
<td>10.56</td>
<td>1.22</td>
<td>0.14</td>
</tr>
<tr>
<td>K₂O</td>
<td>0.11</td>
<td>0.18</td>
<td>0.15</td>
<td>0.16</td>
<td>0.15</td>
<td>0.16</td>
<td>0.18</td>
<td>0.10</td>
</tr>
<tr>
<td>FeO</td>
<td>0.00</td>
<td>0.00</td>
<td>0.06</td>
<td>0.07</td>
<td>0.14</td>
<td>0.00</td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td>TOTAL</td>
<td>99.31</td>
<td>99.72</td>
<td>93.81</td>
<td>96.49</td>
<td>94.04</td>
<td>94.17</td>
<td>95.78</td>
<td>93.48</td>
</tr>
</tbody>
</table>

| Si     | 2.920   | 2.965   | 2.442  | 2.479  | 2.522  | 2.414  | 3.005    | 3.993 |
| Al     | 1.068   | 1.020   | 1.565  | 1.565  | 1.469  | 1.589  | 0.974    | 0.000 |
| Na     | 0.976   | 1.013   | 0.978  | 0.978  | 0.933  | 0.967  | 0.944    | 0.009 |
| Ca     | 0.067   | 0.050   | 0.519  | 0.550  | 0.479  | 0.543  | 0.046    | 0.006 |
| K      | 0.004   | 0.010   | 0.009  | 0.010  | 0.010  | 0.010  | 0.010    | 0.005 |
| Fe     | 0.000   | 0.000   | 0.000  | 0.003  | 0.003  | 0.006  | 0.000    | 0.000 |
### DESCRIPTIONS OF FELDSPAR FROM MICROPROBE ANALYSES

<table>
<thead>
<tr>
<th>Sample</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>P11-10</td>
<td>Andesine, in mesocratic diorite, from outcrop in NE1/4 SE1/4 NE1/4 SE1/4 Sec. 34, T.133N., R.32W.</td>
</tr>
<tr>
<td>P13-10</td>
<td>Andesine, in mesocratic diorite, from outcrop in NE1/4 SE1/4 NE1/4 SE1/4 Sec. 34, T.133N., R.32W.</td>
</tr>
<tr>
<td>P14-1</td>
<td>Andesine, in Melanocratic diorite, from outcrop in NW1/4 SW1/4 NE1/4 Sec. 34, T.133N., R.32W.</td>
</tr>
<tr>
<td>P1-0G</td>
<td>Core of andesine crystal from melanocratic diorite outcrop in SE1/4 SW1/4 NE1/4 Sec. 34, T.133N., R.32W.</td>
</tr>
<tr>
<td>P1-0G'</td>
<td>Rim of P1-0G</td>
</tr>
<tr>
<td>P2-0G</td>
<td>Andesine crystal included within the andesine crystal of P1-0G.</td>
</tr>
<tr>
<td>P2-0G'</td>
<td>Core of Andesine crystal, from same outcrop as P1-0G.</td>
</tr>
<tr>
<td>P3-0G</td>
<td>Midway between core and rim of P7-0G.</td>
</tr>
<tr>
<td>P5-0G</td>
<td>rim of P1-0G</td>
</tr>
<tr>
<td>P5-0G'</td>
<td>Interstitial oligoclase feldspar, same sample as P1-0G.</td>
</tr>
<tr>
<td>P6-0G</td>
<td>Interstitial albite, same sample as P1-0G.</td>
</tr>
<tr>
<td>P16-PA</td>
<td>Andesine to labradorite, in anorthositic zone from outcrop in SW1/4 SE1/4 SE1/4 NW1/4 Sec. 34, T.133N., R.32W.</td>
</tr>
<tr>
<td>P17-PA</td>
<td>Albite, from tourmaline-bearing feldspathic arenite; DDH #A-1, 175' depth; Sec. 32, T.133N., R.32W.</td>
</tr>
<tr>
<td>P18-PA</td>
<td>Interstitial quartz, same sample as Pth1-0G.</td>
</tr>
<tr>
<td>P19-PA</td>
<td>QOG-2</td>
</tr>
</tbody>
</table>
TABLE C-3. Microprobe Analyses of Sphene, Calcite, and Biotite

<table>
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<tr>
<th>SAMPLE</th>
<th>SP04-14</th>
<th>SP04-11</th>
<th>SP02-136</th>
<th>CB00-25</th>
<th>BI-DG6</th>
<th>BI1-27</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>28.33</td>
<td>30.32</td>
<td>32.58</td>
<td>0.00</td>
<td>34.53</td>
<td>35.01</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>0.31</td>
<td>0.24</td>
<td>0.57</td>
<td>0.00</td>
<td>17.20</td>
<td>17.04</td>
</tr>
<tr>
<td>Na₂O</td>
<td>0.10</td>
<td>0.09</td>
<td>0.00</td>
<td>0.11</td>
<td>0.17</td>
<td>0.12</td>
</tr>
<tr>
<td>CaO</td>
<td>28.81</td>
<td>29.16</td>
<td>28.96</td>
<td>54.99</td>
<td>0.30</td>
<td>0.16</td>
</tr>
<tr>
<td>K₂O</td>
<td>0.09</td>
<td>0.10</td>
<td>0.00</td>
<td>0.08</td>
<td>9.69</td>
<td>10.14</td>
</tr>
<tr>
<td>FeO*</td>
<td>1.41</td>
<td>1.49</td>
<td>1.06</td>
<td>0.95</td>
<td>23.17</td>
<td>22.04</td>
</tr>
<tr>
<td>MgO</td>
<td>0.13</td>
<td>0.12</td>
<td>0.00</td>
<td>0.23</td>
<td>8.32</td>
<td>8.36</td>
</tr>
<tr>
<td>TiO₂</td>
<td>37.20</td>
<td>36.97</td>
<td>38.46</td>
<td>0.00</td>
<td>1.79</td>
<td>2.33</td>
</tr>
<tr>
<td>MnO</td>
<td>0.02</td>
<td>0.04</td>
<td>0.00</td>
<td>0.36</td>
<td>0.12</td>
<td>0.11</td>
</tr>
<tr>
<td>TOTAL</td>
<td>96.41</td>
<td>98.52</td>
<td>101.63</td>
<td>56.72</td>
<td>95.30</td>
<td>95.34</td>
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</table>

Number of Cations based on $x$ (Q)

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<th></th>
<th>Si</th>
<th>Al</th>
<th>Na</th>
<th>Ca</th>
<th>K</th>
<th>Fe</th>
<th>Mg</th>
<th>Ti</th>
<th>Mn</th>
</tr>
</thead>
<tbody>
<tr>
<td>Si</td>
<td>3.88</td>
<td>4.05</td>
<td>4.18</td>
<td>0.00</td>
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<td>5.94</td>
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<td></td>
<td></td>
</tr>
<tr>
<td>Al</td>
<td>0.05</td>
<td>0.04</td>
<td>0.09</td>
<td>0.00</td>
<td>3.46</td>
<td>3.47</td>
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<td></td>
<td></td>
</tr>
<tr>
<td>Na</td>
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<td>0.02</td>
<td>0.00</td>
<td>0.02</td>
<td>0.55</td>
<td>0.04</td>
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<td></td>
</tr>
<tr>
<td>Ca</td>
<td>4.23</td>
<td>4.18</td>
<td>3.98</td>
<td>5.80</td>
<td>0.54</td>
<td>0.03</td>
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</tr>
<tr>
<td>K</td>
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<td>0.02</td>
<td>0.00</td>
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<td>2.15</td>
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</tr>
<tr>
<td>Fe</td>
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<td>0.17</td>
<td>0.12</td>
<td>0.08</td>
<td>3.31</td>
<td>3.13</td>
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</tr>
<tr>
<td>Mg</td>
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<td>0.02</td>
<td>0.00</td>
<td>0.03</td>
<td>2.12</td>
<td>2.11</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Ti</td>
<td>3.83</td>
<td>3.72</td>
<td>3.71</td>
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<td>0.23</td>
<td>0.30</td>
<td></td>
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</tr>
<tr>
<td>Mn</td>
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<td>0.00</td>
<td>0.03</td>
<td>0.02</td>
<td>0.01</td>
<td></td>
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<td></td>
</tr>
</tbody>
</table>

'x' = 20 for SP (sphene); 6 for CB (calcite); and 24 BI (biotite).

DESCRIPTIONS OF SPHENE, CALCITE, AND BIOTITE:

Sample  | Description                                                                 |
---------|-----------------------------------------------------------------------------|
SP14-74  | Interstitial sphene, in magnetite pyroxenite. DDH #1, 74' depth, Sec. 3, T.132N., R.32W. |
SP11-74  | Same as SP14-74.                                                             |
SP02-136 | Interstitial sphene, in oxide rock. DDH #3, 136' depth; Sec. 3, T.132N., R.32W. |
CB-50-25 | Calcite, in magnetite hornblende. DDH #6, 50' depth. Sec. 3, T.132N., R.32W.   |
BI-DG6   | Biotite, from outcrop of melanocratic diorite in SE1/4 SW1/4 NE1/4 NW1/4 Sec. 34, T.133N., R.32W. |
BI1-27   | Deep brown biotite, from melanocratic diorite outcrop in NW1/4 NW1/4 NE1/4 SE1/4 Sec. 34, T.133N., R.32W. |
### TABLE C-4  Microprobe Analyses of Oxides

<table>
<thead>
<tr>
<th>Sample</th>
<th>3-136-LI1</th>
<th>3-136-MG1</th>
<th>3-136-LI2</th>
<th>3-136-MG2</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>0.33</td>
<td>0.49</td>
<td>0.56</td>
<td>0.50</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>0.00</td>
<td>0.08</td>
<td>0.03</td>
<td>0.06</td>
</tr>
<tr>
<td>FeO</td>
<td>45.77</td>
<td>92.09</td>
<td>46.15</td>
<td>92.57</td>
</tr>
<tr>
<td>TiO₂</td>
<td>50.45</td>
<td>0.47</td>
<td>50.95</td>
<td>0.96</td>
</tr>
<tr>
<td>CaO</td>
<td>2.22</td>
<td>6.62</td>
<td>2.11</td>
<td>0.05</td>
</tr>
<tr>
<td>MnO</td>
<td>1.03</td>
<td>0.00</td>
<td>5.63</td>
<td></td>
</tr>
<tr>
<td>O*</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Total</td>
<td>100.01</td>
<td>100.02</td>
<td>100.00</td>
<td>99.99</td>
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</table>

**Number of Ions based on 12 (O)**

<table>
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<tr>
<th></th>
<th>Sample 9-52-LI1</th>
<th>6-52-LI2</th>
<th>5-52-MG3</th>
<th>8-52-LI4</th>
<th>7-52-MG4</th>
</tr>
</thead>
<tbody>
<tr>
<td>Si</td>
<td>0.034</td>
<td>0.074</td>
<td>0.057</td>
<td>0.075</td>
<td></td>
</tr>
<tr>
<td>Al</td>
<td>0.000</td>
<td>0.013</td>
<td>0.004</td>
<td>0.011</td>
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<td>Fe</td>
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<td>0.037</td>
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<tr>
<td>Mn</td>
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<td>0.000</td>
<td>0.181</td>
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<tr>
<td>Solution</td>
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<td>1.4% U</td>
<td>2.25% H</td>
<td>2.8% U</td>
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**Solid Solution**

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<th>6-52-LI2</th>
<th>5-52-MG3</th>
<th>8-52-LI4</th>
<th>7-52-MG4</th>
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<td>0.070</td>
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<td>1.87% H</td>
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<td>0.62% H</td>
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D* IS THE DIFFERENCE IN OXYGEN REQUIRED TO MAKE THE ANALYSES EQUAL 100%, DUE TO Fe²⁺/Fe³⁺ VARIATIONS.

H = hematite
U = ulvospinel
DESCRIPTIONS OF OXIDES FROM MICROPROBE ANALYSES
(All samples from oxide-apatite rock)
Sample
(DDH #3, 136' depth; Sec. #, T.132N., R.32W.)
3-136-L11 Exsolved lamellar ilmenite.
3-136-MG1 Magnetite containing lamellar ilmenite.
3-136-L12 Exsolved lamellar ilmenite.
3-136-MG2 Magnetite containing lamellar ilmenite.

(DDH #1, 52' depth; Sec., T.132N., R.32W.)
5-52-CI Interstitial composite ilmenite.
6-52-L13 Exsolved lamellar ilmenite.
5-52-MG3 Magnetite containing lamellar ilmenite.
6-52-L14 Exsolved lamellar ilmenite.
7-52-MG4 Magnetite containing lamellar ilmenite.
### TABLE C-5. Microprobe Analyses of Pyroxene

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<th>SAMPLE NUMBER</th>
<th>PX2-0G</th>
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<th>PX5-74</th>
<th>PX6-74</th>
<th>PX7-74</th>
<th>PX8-74</th>
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<tr>
<td><strong>Si O₂</strong></td>
<td>51.35</td>
<td>50.08</td>
<td>50.12</td>
<td>51.18</td>
<td>50.82</td>
<td>51.22</td>
<td>50.25</td>
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<tr>
<td><strong>Al₂O₃</strong></td>
<td>2.85</td>
<td>2.41</td>
<td>3.13</td>
<td>0.14</td>
<td>0.03</td>
<td>0.04</td>
<td>0.05</td>
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<tr>
<td><strong>Na₂O</strong></td>
<td>0.49</td>
<td>0.51</td>
<td>0.47</td>
<td>0.18</td>
<td>0.18</td>
<td>0.18</td>
<td>0.17</td>
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<tr>
<td><strong>CaO</strong></td>
<td>19.73</td>
<td>20.50</td>
<td>17.51</td>
<td>23.03</td>
<td>23.35</td>
<td>23.84</td>
<td>23.94</td>
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<td><strong>K₂O</strong></td>
<td>0.12</td>
<td>0.11</td>
<td>0.15</td>
<td>0.10</td>
<td>0.10</td>
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<td>0.10</td>
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<td><strong>FeO</strong></td>
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<td>11.57</td>
<td>13.65</td>
<td>15.35</td>
<td>14.10</td>
<td>14.39</td>
<td>12.21</td>
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<tr>
<td><strong>MgO</strong></td>
<td>12.52</td>
<td>11.96</td>
<td>12.24</td>
<td>9.30</td>
<td>9.84</td>
<td>9.17</td>
<td>10.67</td>
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<tr>
<td><strong>TiO₂</strong></td>
<td>0.65</td>
<td>0.93</td>
<td>0.47</td>
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<td>0.00</td>
<td>0.04</td>
<td>0.00</td>
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<tr>
<td><strong>MnO</strong></td>
<td>0.30</td>
<td>0.27</td>
<td>0.23</td>
<td>0.39</td>
<td>0.36</td>
<td>0.47</td>
<td>0.42</td>
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<tr>
<td><strong>TOTAL</strong></td>
<td>99.66</td>
<td>98.35</td>
<td>97.98</td>
<td>99.71</td>
<td>98.76</td>
<td>99.44</td>
<td>97.83</td>
</tr>
</tbody>
</table>

### Number of Ions Based on 6(O)

| **Si**        | 1.940  | 1.930  | 1.930  | 1.990  | 1.980  | 1.990  | 1.970  |
| **Al**        | 0.127  | 0.109  | 0.142  | 0.005  | 0.001  | 0.018  | 0.023  |
| **Na**        | 0.036  | 0.038  | 0.034  | 0.014  | 0.014  | 0.014  | 0.012  |
| **Ca**        | 0.796  | 0.844  | 0.724  | 0.959  | 0.976  | 0.992  | 1.010  |
| **K**         | 0.006  | 0.006  | 0.008  | 0.004  | 0.005  | 0.004  | 0.005  |
| **Fe**        | 0.367  | 0.372  | 0.440  | 0.499  | 0.460  | 0.460  | 0.401  |
| **Mg**        | 0.702  | 0.685  | 0.704  | 0.539  | 0.574  | 0.531  | 0.624  |
| **Ti**        | 0.020  | 0.030  | 0.010  | 0.000  | 0.000  | 0.001  | 0.000  |
| **Mn**        | 0.008  | 0.007  | 0.006  | 0.012  | 0.012  | 0.015  | 0.014  |

### 100 Mg

<table>
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<th>65.1</th>
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<tr>
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<td>44</td>
<td>39</td>
<td>48</td>
<td>48</td>
<td>50</td>
<td>49</td>
</tr>
<tr>
<td><strong>Mg</strong></td>
<td>37</td>
<td>36</td>
<td>38</td>
<td>27</td>
<td>29</td>
<td>27</td>
<td>31</td>
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<tr>
<td><strong>Fe</strong></td>
<td>20</td>
<td>20</td>
<td>23</td>
<td>25</td>
<td>23</td>
<td>23</td>
<td>20</td>
</tr>
</tbody>
</table>

---

**Px-2-0G**  Brownish augite, heavily mantled and replaced by hornblende, contains abundant exsolved rods; diorite outcrop

**Px-3-0G**  Colorless, clean, calcium-rich augite; weakly uralitized; Pyroxenite; DDH 1, 74-80

---

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### TABLE C-6. Microprobe Analyses of Amphibole

<table>
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<tr>
<td><strong>SiO₂</strong></td>
<td>37.88 38.22 47.45 43.39 47.57 48.74 50.63 37.52 46.98 36.91 40.58 37.34</td>
</tr>
<tr>
<td><strong>Al₂O₃</strong></td>
<td>13.54 13.71 4.00 7.90 3.32 1.78 0.42 13.16 2.54 12.39 12.38 12.69</td>
</tr>
<tr>
<td><strong>Na₂O</strong></td>
<td>1.93 2.30 0.93 1.42 0.69 0.37 0.16 2.52 0.53 2.45 2.45 2.21</td>
</tr>
<tr>
<td><strong>CaO</strong></td>
<td>11.46 11.49 11.51 11.57 11.86 12.23 12.24 11.38 11.80 11.21 11.47 11.38</td>
</tr>
<tr>
<td><strong>K₂O</strong></td>
<td>1.05 0.27 0.64 0.33 0.18 0.12 1.16 0.23 1.46 1.43 1.23</td>
</tr>
<tr>
<td><strong>MgO</strong></td>
<td>4.24 4.43 9.82 7.17 9.32 10.22 10.02 5.37 9.26 6.18 6.05 5.83</td>
</tr>
<tr>
<td><strong>TiO₂</strong></td>
<td>1.71 1.44 0.22 0.46 0.09 0.07 0.00 2.38 0.11 2.54 2.49 1.99</td>
</tr>
<tr>
<td><strong>H₂O</strong></td>
<td>0.35 0.26 0.32 0.39 0.38 0.27 0.27 0.26 0.35 0.25 0.25 0.23</td>
</tr>
<tr>
<td><strong>Total</strong></td>
<td>97.24 97.81 95.75 96.67 95.23 94.56 94.69 96.75 93.29 95.16 99.29 95.76</td>
</tr>
</tbody>
</table>

**Number of Ions Based on 23 (O)**

<table>
<thead>
<tr>
<th>Sample</th>
<th>HB16-74 ACT1-74 ACT2-74 HB1-08 HB2-08 HB3-08 HB4-08 HB5-08 HB5A-08 HB5-10 HB6-10 HB55-14 HE36-14</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>SiO₂</strong></td>
<td>46.82 50.95 51.25 39.36 40.25 39.02 56.77 39.79 50.21 50.08 41.74 39.98</td>
</tr>
<tr>
<td><strong>Al₂O₃</strong></td>
<td>1.63 1.22 1.16 14.49 17.56 15.51 17.93 2.39 15.63 3.44 3.40 14.27 16.14</td>
</tr>
<tr>
<td><strong>Na₂O</strong></td>
<td>0.43 0.38 0.34 1.64 1.64 1.63 1.69 0.45 1.76 0.65 0.63 1.70 1.86</td>
</tr>
<tr>
<td><strong>CaO</strong></td>
<td>11.61 11.97 11.78 11.60 11.57 11.22 11.53 10.67 11.52 12.28 12.22 11.61 11.45</td>
</tr>
<tr>
<td><strong>K₂O</strong></td>
<td>0.23 0.20 0.15 0.76 0.85 0.93 0.78 0.14 0.69 0.19 0.20 0.54 0.65</td>
</tr>
<tr>
<td><strong>MgO</strong></td>
<td>9.92 10.61 10.83 5.31 4.71 5.52 4.91 10.78 5.64 12.45 12.71 7.04 5.82</td>
</tr>
<tr>
<td><strong>TiO₂</strong></td>
<td>0.11 0.06 0.08 0.23 0.30 0.20 0.23 0.21 0.71 0.48 0.08 0.36 0.25</td>
</tr>
<tr>
<td><strong>H₂O</strong></td>
<td>0.40 0.32 0.44 0.20 0.22 0.20 0.23 0.21 0.18 0.24 0.14 0.23 0.23</td>
</tr>
<tr>
<td><strong>Total</strong></td>
<td>92.52 96.05 96.02 96.73 99.32 96.29 97.77 95.80 97.60 96.34 95.89 98.42 97.69</td>
</tr>
</tbody>
</table>

| Al     | 0.310 0.325 0.206 2.680 3.172 2.662 3.258 0.409 2.841 0.607 0.603 2.547 2.915 |
| Na     | 0.134 0.113 0.100 0.498 0.481 0.496 0.501 0.128 0.508 0.190 0.195 0.498 0.552 |
| Ca     | 2.005 1.947 1.923 1.950 1.872 1.882 1.892 1.661 1.904 1.970 1.997 1.884 1.880 |
| K      | 0.046 0.039 0.030 0.152 0.164 0.185 0.152 0.026 0.137 0.306 0.039 0.194 0.128 |
| Fe     | 2.881 2.570 2.560 3.047 2.826 2.876 2.737 1.717 2.856 2.062 2.062 2.652 2.732 |
| Mg     | 2.385 2.400 2.461 1.242 1.060 1.289 1.122 2.337 1.296 2.779 2.280 1.589 1.329 |
| Ti     | 0.013 0.006 0.000 0.018 0.017 0.024 0.046 0.040 0.025 0.021 0.023 0.029 0.029 |
| Mn     | 0.054 0.041 0.056 0.027 0.029 0.030 0.028 0.025 0.031 0.023 0.017 0.029 0.030 |
Description of amphiboles from microprobe analyses.

Hb-19-50 Red-brown ferropargasitic hornblende core in hornblendite; DDH 6, 50-53'
Hb-20-50

Hb-21-50 Light green ferroactinolitic hornblende rim on above; DDH 6, 50-53'
Hb-22-50 Dark green ferroedinitic hornblende rim on above; DDH 6, 50-53'
Hb-23-50 Light green ferroactinolitic hornblende rim on above; DDH 6, 50-53'
Hb-24-50 Dark green ferroactinolite rim on above; DDH 6, 50-53'
Hb-30-50 Red-brown ferropargasitic hornblende core in hornblendite; DDH 6, 50-53'
Hb-31-50 Colorless ferroactinolite rim on above; DDH 6, 50-53'
Hb-13-74 Red-brown ferropargasitic hornblende core, in pyroxenite; DDH 1, 74-80'
Hb-13'-74
Hb-14-74 Colorless ferroactinolite rim on above; DDH 1, 74-80'

Act-1-74 Colorless to pale green actinolite, with magnetite; as fibrous mats
Act-2-74 replacing pyroxene, in pyroxenite; DDH 1, 74-80'

Hb-5A-0G Fibrous pale green ferroactinolite replacing Pyroxene; diorite outcrop
Hb-1-0G
Hb-2-0G Blue-green ferropargasitic hornblende, forms prismatic rims over pyroxenes which often penetrate feldspar; Diorite outcrop sample
Hb-3-0G
Hb-3A-0G
Hb-4-0G

Hb-5-10 Dark army-green actinolite, apparently deuteric; diorite outcrop
Hb-6-10

Hb-35-1* Blue-green inclusion laden ferropargasitic hornblende; diorite outcrop
Hb-36-1* Secondary blue-green ferropargasitic hornblende, with biotite; diorite
**ASSAY RESULTS**

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**Table C-7.** All values reported in PPM, except Fe(%) and Au (PPB). ( ) denotes a concentration below the detection limit for that element. Samples TO-13175 to TO 13181 funded by the Minnesota Dept. of Natural Resources Minerals Division, assay (N/C plasma) by Bondar-Clegg; the rest of the samples were funded by the Minnesota Geological Survey, assay (inductively coupled plasma, atomic emission spectrometry; ICP/AES) by Geochemical Services, Inc. na.=not assayed for; wh=see whole rock analyses.
Table C-7 continued. Assay sample descriptions.
TO-13175 - Iron-formation, oxidized, brecciated, red-brown. DDH #4, sample depth 280', T. 132 N., R. 32 W., sec. 5, Todd County.

TO-13176 - Schist, green, cpy-py-mag-chl, quartz veining; DDH #1, sample depth 225'; T.132N., R.32W., sec 5, Todd County.

TO-13177 - Schist, green, qtz-plag-carb-chl-py; DDH #8, sample depth 150', T.132N., R.32W., sec. 5, Todd County.

TO-13178 - Iron-formation; black with semi-massive pyrite; DDH #12, T.132N., R.32W., sec. 5, Todd County.

TO-13179 - Schist, gray-green, tourmaline-bearing, DDH #15, T.132N., R.32W., sec. 5, Todd County.

TO-13180 - Schist, dark green, micaceous; DDH #108, T.132N., R.32W., sec. 32, Todd County.

TO-13181 - Metabasalt, gray-green, amygdular, carbonate-rich; DDH #108, 1145' sample depth, T.132N., R.32W., Todd County.

Table C-7. Continued.
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**Table C-8.** Visual estimates of modal mineralogy, estimated from thin section. (DDH #108 Sec. 32, T.133 N., R.32 W.)
APPENDIX D

ROCK GEOCHEMISTRY OF METASEDIMENTARY AND IGNEOUS ROCKS NEAR PHILBROOK
### Table D-1. CHEMICAL ANALYSES OF SEDIMENTARY AND VOLCANIC ROCKS

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**Total** 39.2 38.9 26.3 31.22 36.1

**Fe₂O₃** 2.26 2.17 2.47 14.1 10.6

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DESCRIPTIONS OF IGNEOUS ROCKS FROM WHOLE ROCK ANALYSES:

**Pb-1**
Medium-coarse grained, dark green apatite-rich melanocratic diorite from outcrop in SW 1/4 NE 1/4 SW 1/4 NW 1/4 Sec. 34, T.133N., R.32W.

**PB-3-1-74**
Medium-coarse-grained, dark green hornblende pyroxenite, DDH #1, 74', depth, Sec. 3, T.132N., R.32W.

**PB-3-5-179**
Oxide-apatite rock, DDH #8, 179', Sec. 3, T.132N., R.32W.

Table D-4. Chemical analyses of mafic and ultramafic igneous rocks.
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