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Robert L. Kastelic

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PRECAMBRIAN GEOLOGY AND
MAGNETITE DEPOSITS
OF THE
NEW JERSEY HIGHLANDS IN
WARREN COUNTY, NEW JERSEY

by

Robert L. Kastelic Jr.

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Master of Science
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CERTIFICATE OF APPROVAL

This thesis is accepted and approved in partial fulfillment of the requirements for the degree of Master of Science.

Aug. 17, 1979
(date)

Professor in Charge

Chairman of Department

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TABLE OF CONTENTS

TABLE OF CONTENTS

LIST OF TABLES

LIST OF FIGURES

LIST OF ILLUSTRATIONS

ABSTRACT

The Precambrian rocks of the New Jersey Highlands in the Belvidere and Washington quadrangles consist of marble, amphibolite, quartzo-feldspathic gneiss, and syntectonic granite which developed during regional metamorphism under granulite-facies conditions.

Field relationships, petrographic data, and bulk chemical analyses support the view that the quartz-potassium feldspar gneiss, quartz-epidote gneiss, marble, and garnet-biotite gneiss are of metasedimentary origin. The quartz-potassium feldspar gneiss and quartz-epidote gneiss were probably derived from arkosic sediments which contained intercalated limestones. The compositional layering in the metamorphic rocks appears to reflect original sedimentary bedding.

Chemical analyses indicate that the amphibolite was derived from sedimentary rocks. Chemical analyses support the conclusion that the premetamorphic parent of the oligoclase-quartz gneiss was probably an ash flow or keratophyric tuff.

Perthitic granite bodies do not show either intrusive contacts with the country rock or chilled margins. Small bodies of granite are commonly enclosed by gneiss and may have developed by partial fusion

(in situ anatexis) of metasedimentary quartz-potassium feldspar gneiss. Syenite bodies are interpreted as feldspar-rich residues resulting from filter pressing in areas of tight isoclinal folding during anatexis. Sodium-rich granitic rocks apparently developed by anatexis of oligoclase-quartz gneiss.

The magnetite deposits in the study area are massive granular mixtures of magnetite and quartz with minor pyrite. The classical hypothesis for the origin of these deposits is that the unique and distinctive pyroxenite associated with the gneiss-enclosed deposits is a contact-metasomatic skarn developed by fluids emanating from alaskite intrusions and that the ores are, therefore, of pneumatolytic origin since they are spatially and genetically associated with these pyroxene-rich skarns. However, the occurrence of magnetite concentrations in sharply bounded tabular-shaped bodies, the absence of large igneous intrusions which could have been the source of residual iron-bearing fluids, the absence of minerals which are characteristic of contact-metasomatic skarns, and the high Mg/Fe and relatively high Na content of the clinopyroxene of the pyroxenite are not consistent with the pneumatolytic hypothesis. These observations, however, are compatible with the hypothesis that the iron deposits are of metasedimentary origin. If the latter is correct, then the original mineralogical expression of the iron could have been siliceous siderite layers in a sedimentary sequence of impure carbonate rocks. During granulite-facies metamorphism, magnetite would have developed by decarbonation of siderite, and the pyroxenite could have developed by dedolomitization of a siliceous dolomitic marl.

Granite pegmatite occurs along the footwall and the hanging wall of the magnetite orebody at the Washington mine and separates the magnetite ore from the adjoining quartz-potassium feldspar gneiss. This pegmatite was probably not, as earlier supposed, the ore bringer. Rather, it was probably developed by interface melting of the quartz-potassium feldspar gneiss during granulite-facies metamorphism where the quartz-potassium feldspar gneiss was in contact with the ore, because the effect of iron is to lower markedly the temperature of fusion of the low-melting compositions in the system quartz-orthoclase-albite.

INTRODUCTION

The New Jersey Highlands in Warren County consist of a series of northeast trending ridges underlain by Precambrian rocks and of parallel valleys underlain by Cambrian and Ordovician sedimentary rocks. Collectively, this area is part of the Reading Prong which extends from western Connecticut southwestward to the vicinity of Reading, Pennsylvania. Locally it is known as the Hudson Highlands in New York, the Ramapo Mountains and the Highlands in New Jersey, and the Reading and Durham Hills in Pennsylvania.

The origin and metamorphic history of the various gneisses and the origin of the associated magnetite deposits has long been a subject of study and debate. Recent investigators have proposed that the gneisses of the Highlands are the products of granulite-rank regional metamorphism of sedimentary and volcanic rocks which have been intruded by a series of syntectonic granites (Sims, 1958; Drake, 1969, 1970; and Baker and Buddington, 1970). Recent hypotheses regarding the genesis of the magnetite deposits are diverse and include 1) introduction of iron by residual fluids from hornblende granite and alaskite to form contact-metasomatic skarn deposits (Sims, 1953, 1958); 2) release, transport, and concentration of iron through recrystallization and deformation of amphibolite during regional metamorphism (Hagner and Collins, 1955; Collins, 1969a, 1969b); and syngenetic concentration of iron by sedimentary processes prior to metamorphism (Baker and Buddington, 1970).

OBJECTIVES OF THIS STUDY

Previous work in the New Jersey Highlands in Warren County has been only of a reconaissance nature (Bayley, 1910, 1941; Bayley et al., 1914), and very little detailed mapping has been done in the area of this study. The magnetite deposits in Warren County were described by Bayley (1910, 1941) but they have not been studied in detail with an emphasis on petrology and ore genesis.

The overall objectives of this study were to develop new insight into the nature of the premetamorphic rocks which are now gneisses of granulite-rank, and to generate an hypothesis regarding the genesis of the magnetite deposits which is compatible with both field and petrographic observations.

The specific objectives of this study were as follows:

1) To carry out detailed field mapping of the

Precambrian rocks in the Belvidere and Washington 7½ minute quadrangles to provide the areal distribution of mappable petrologic units as well as structural data.

2) To determine the nature and identity of the progenitors of the various gneisses and amphibolites using whole-rock major- and minor-element analyses in conjunction with field observations and petrography.
3) To develop a coherent hypothesis regarding the genesis of the magnetite deposits using field relationships, new whole-rock major- and minor-element analyses of rocks associated with the magnetite ores, and modern petrogenetic theory.

PREVIOUS WORK

The New Jersey Highlands have been of economic interest because of the occurrence of iron and zinc deposits since the early 18th century. The first description of Precambrian rocks in New Jersey was by Pierce (1822), who mentioned the presence of granite and schist. Rodgers (1836, 1840) gave more detailed descriptions of Highlands geology and of mining in general. He found that the Precambrian rocks were overlain by lower Paleozoic rocks which were presumably downfolded between Precambrian ridges. He considered the magnetite deposits to be of igneous origin. Since that time, most of the detailed geological work in the Precambrian terrain of New Jersey has been concentrated in the northeastern part. Kitchell (1856) was the first to conclude that the magnetite deposits of the Highlands were sedimentary in origin based on their association with metasedimentary rocks. Nason (1889), against the widespread view at the time, proposed that the Highland rocks were not metamorphosed sedimentary rocks but were of igneous origin.

The U.S. Geological Survey began reconaissance studies in the New Jersey Highlands in the mid-1890's. Spencer (1908) proposed three new formation names in addition to the Franklin limestone, namely, the Byram gneiss in which potassium feldspar was dominant, the Losee gneiss containing dominant plagioclase, and the Pochuck gneiss which includes the mafic rocks. The Byram and Losee gneisses were considered to be metamorphosed igneous intrusions, whereas the Pochuck gneiss and Franklin limestone were considered to be metasedimentary. These names were subsequently used by investigators such as Bayley et al. (1914), who mapped in the area of this study, and by Lewis and Kümmel (1912) on the geological map of New Jersey.

More recent workers such as Sims and Leonard (1952), Hotz (1953, 1954), Sims (1953, 1958), and Buddington (1966) abandoned the old formation names in favor of more precise mineralogically based lithologic units. They considered the magnetite deposits to be of igneous origin because they did not resemble banded iron formations of the Lake Superior type (Buddington, 1966).

In the 1960's, geologists of the U.S. Geological Survey carried out detailed mapping across the Reading Prong along the Delaware River (Davis et al., 1967; Drake, 1967a, 1967b, 1969, 1970, 1978; Drake and Epstein, 1967; Drake et al., 1961, 1967, 1969).

Dodd (1965), Dallmeyer (1974a, 1974b), and Dallmeyer and Dodd (1971) have studied the tectonic and metamorphic history of the rocks of the northeastern part of the Reading Prong in New Jersey and New York.

GENERAL GEOLOGY OF THE AREA

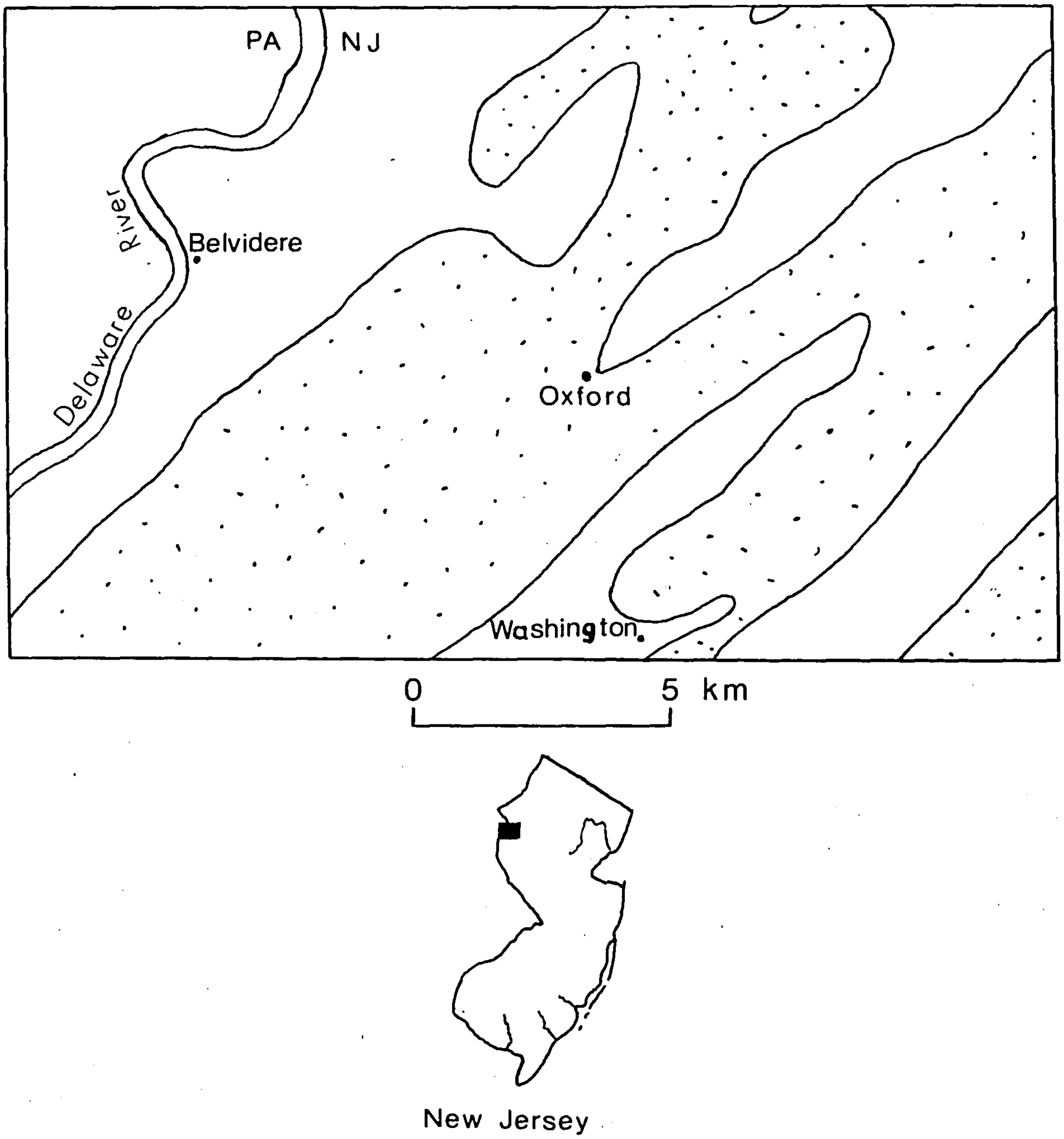
The Belvidere and Washington quadrangles contain northeast trending ridges of Precambrian rocks separated by valleys underlain by Cambrian and Ordovician limestones and shales (Fig. 1).

The rocks of the map area are covered in part by the Wisconsin terminal moraine. All the bedrock, however, is deeply weathered and outcrops are sparce and restricted to areas of high relief and artificial exposures such as roadcuts or construction work. Consequently, it was necessary to base much of the field mapping on the statistical occurrence of float of the mappable units.

The Precambrian rocks contain gneisses of metasedimentary and volcanic origin intruded by syntectonic granites. The occurrence of migmatites suggests that both partial melting of metasedimentary rocks and intrusion of microperthite alaskite into amphibolite took place.

Radiometric ages based on zircon (U-Pb) throughout the N.J. and N.Y. Highlands indicate that the last peak of regional metamorphism occurred 1060 million years ago (Dallmeyer et al., 1975). $^{40}Ar/^{39}Ar$ ages of 900 my and 790 my on hornblende and biotite respectively (Dallmeyer et al., 1975), represent cooling retention ages after the peak of Grenville metamorphism. It has been suggested by Dallmeyer and Sutter (1976) that low $^{40}Ar/^{39}Ar$ ages of 450 my for hornblende and 390 my for

Figure 1. Location of the study area. Precambrian rocks, stippled; Paleozoic rocks, unpatterned. Modified from Drake (1969).


PA
NJ
Delaware River
Belvidere
Oxford
Washington
0
5 km
New Jersey

biotite are due to argon loss during Taconic orogenic activity.

Mose and Hayes (1974) reported Rb/Sr ages of about 1500 my for granites in the Reading Prong, which suggest that granites are not synchronous with the main Grenville event of about 1.0 by but were emplaced earlier. Murray et al. (1977) reported whole-rock Rb/Sr ages of 1125 my for paragneiss in the Reading Prong of New York. They determined, based on pyroxene geothermometry and garnet-cordierite geobarometry, that temperatures of 750-800 degrees C and pressures of .3-.5 GPa were reached.

Recently, Carvalho (1978) and Carvalho and Sclar (1979) determined that the minimum temperature of metamorphism at Sterling Hill, N.J. based on gahnite-franklinite exsolution intergrowths in the stratiform zinc ores enclosed in the Franklin marble is 760 degrees C.

The magnetite ores were probably deposited before Grenvillian metamorphism inasmuch as they conform to the local structure of the surrounding rocks. They are commonly tabular bodies with a large strike dimension (Sims, 1958) and their long axes are usually parallel to the lineation of the surrounding rocks, thus suggesting that they were deformed during regional metamorphism.

Recent structural studies along the Delaware River by Drake (1969, 1970, 1978) have shown that the Precambrian rocks of the Reading Prong are part of a large nappe de recouvrement which has been thrust over lower Paleozoic rocks and refolded and faulted during Paleozoic orogenic events.

THE PRECAMBRIAN ROCKS

The rock names used in this study are based on the mineralogical and textural properties of the rock units and are those used by recent workers in the Reading Prong. Even with this nomenclature, not all the rocks which occur in the Highlands fit neatly into these specific categories. Changes in lithology occur on a scale of meters or a few centimeters and must be generalized in order to be accomodated to the map scale used in this study.

The Older Metasedimentary-Metavolcanic Series

Although the quartzofeldspathic gneisses have traditionally been considered intrusive, it is now realized that paragneisses occur in greater abundance than intrusive rocks. These are described below.

Amphibolite

The term amphibolite is used for rocks which consist predominantly of plagioclase (An_{12-40}) and hornblende. Biotite, chlorite and pyroxene may be present in substantial amounts.

Amphibolite occurs within most of the mappable rock units in the area including both the metasediments and the plutonic rocks in masses too small to be mapped and as lenses large enough to be mappable. A belt of amphibolite has been mapped along the northwestern edge of Scotts Mountain in the Belvidere quadrangle, and smaller lenses are scattered throughout the study area.

Amphibolite interlayered with oligoclase-quartz gneiss occurs at the Oxford Stone Quarry, southeast of Bridgeville in the Belvidere quadrangle. It also is found as small isolated ellipsoidal masses within the marble belt.

Amphibolite is commonly intruded by microperthite alaskite and pegmatitic alaskite at many places and may be migmatitic. These migmatitic amphibolites, however, are usually not developed on a scale sufficient to permit mapping.

Amphibolite is typically grayish-green to jet black. The grain size ranges from less than .1 mm to greater than 2 mm. The hornblende is commonly pleochroic green to yellow-green. Where biotite is present, it is brown to reddish-brown. Clinopyroxene is uncommon in most samples and orthopyroxene has not been found. Accessory minerals include sphene, magnetite, quartz, apatite, epidote, and calcite. The texture is granoblastic, but locally the rock has been sheared and recrystallized so that the texture is cataclastic. In the Oxford Stone Quarry, sheared-out lenses of optically oriented quartz grains up to 10 mm long and 1 mm thick occur in both the amphibolite and the adjoining oligoclase-quartz gneiss. Where amphibolite occurs near fold hinges, the grain size is typically increased with no change in mineralogy or texture which gives the rock the appearance of diorite.

The dominant type of foliation is crystallization foliation due to the alignment of hornblende and biotite. The sheared rocks are also foliated due to the orientation of crushed hornblende, biotite, and locally, quartz lenses. Retrograde effects are revealed by the partial replacement of clinopyroxene by hornblende, and by the partial replacement of hornblende by biotite and chlorite.

The ratio of plagioclase to the sum of the mafic minerals is quite variable, as shown in Fig. 2. The origin of an amphibolite cannot be determined by petrography, but the field relations suggest that there are both sedimentary and volcanic progenitors in this area. This will be discussed later in this report.

Marble

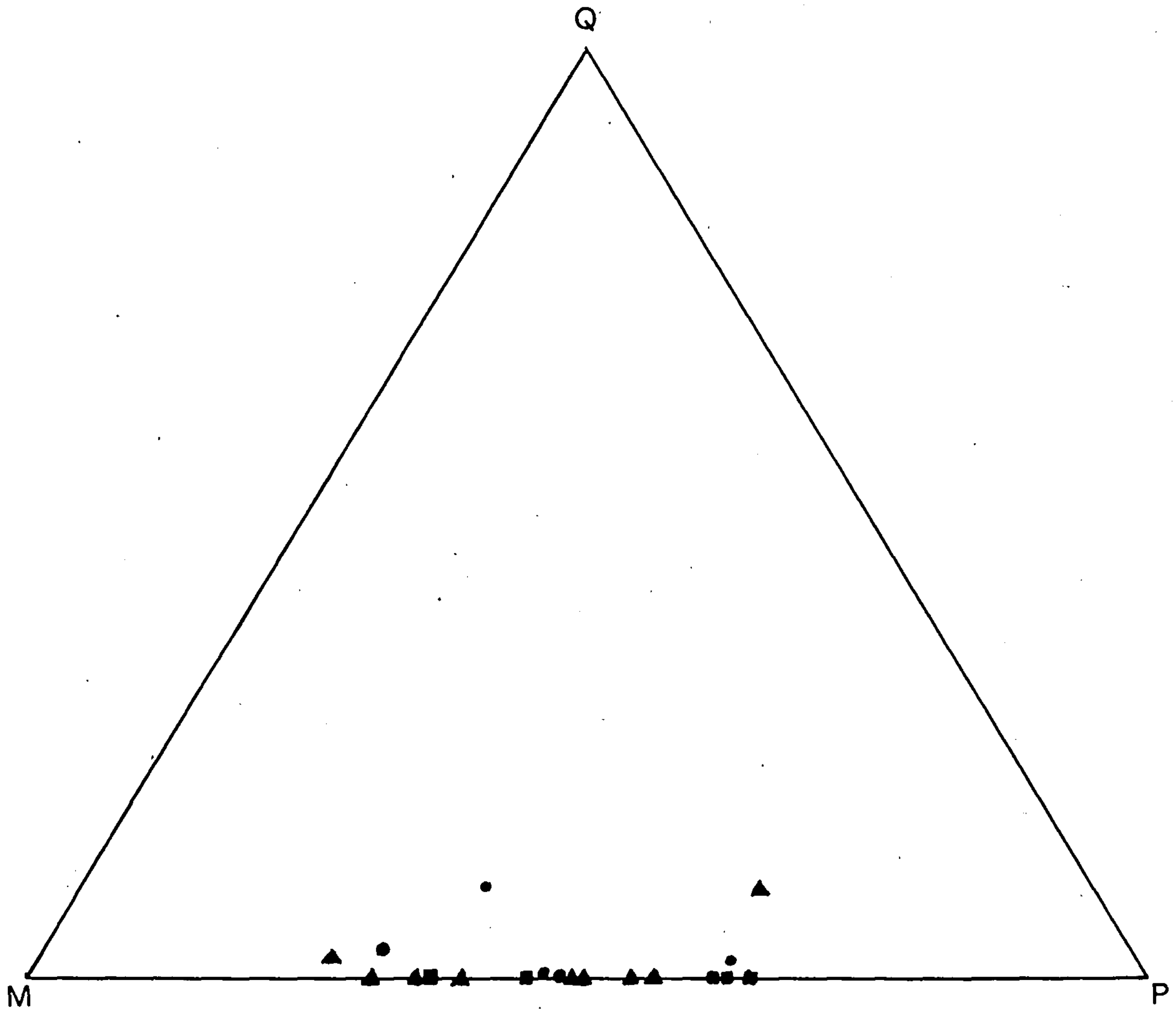
Marble occurs in a triangular area about 4 km in length and about 1.2 km in width, in the eastern part of the Belvidere quadrangle. This is one of the largest areas of marble in the western Reading Prong. Many bodies of marble, too small to map, occur interleaved with amphibolite and quartz-potassium feldspar gneiss.

Several marble quarries and abandoned magnetite mines occur within the marble belt. Chemical analyses obtained from the Oxford Stone Company coupled with X-ray powder diffraction analysis show that the marble consists of almost pure calcite.

The marble belt is bounded on the north by amphibolite. Isolated blocks of amphibolite are found within the marble, and appear to be boudins developed by differential plastic flow during tectonism.

The marble consists of granoblastic grains of

Figure 2. Plot of amphibolite modes. End members are quartz, Q; sum of mafic minerals, M; and plagioclase, P. Triangles are data from Baker and Buddington (1970); squares are data from Drake (1969); circles are new data from this study.


Q
M
P

twinned calcite, and contains minor amounts of graphite. The grain size ranges from .03 mm along grain boundaries with well developed mortar structure to greater than 8 mm in uncrushed grains. Foliation is poorly developed, and, where visible, is defined by alignment of tiny graphite grains. Locally, layers rich in magnetite define a planar structure in the marble. On weathered surfaces, the layering in the marble may be seen more clearly.

Quartz-Epidote Gneiss

A belt of gneiss composed dominantly of quartz, epidote and chlorite occurs on High Rock Mountain in the Washington quadrangle. This rock is characterized by a marked segregational layering of quartz and various dark minerals. Feldspars are locally sparce but occur in variable amounts throughout the area. Calcite, sphene, apatite and tourmaline are found in variable but minor amounts. A yellow isotropic mineral was present in one thin section, and upon X-ray powder diffraction proved to be sphalerite. Its occurrence in the quartz-epidote gneiss is minor.

On weathered surfaces, the rock shows a characteristic layering on a scale of a few millimeters which may be a relict sedimentary structure. It resembles migmatite found elsewhere in the study area, but the variability of feldspar in the quartz-epidote gneiss is a distinctive feature.

Poikiloblastic calcite grains often contain inclusions of chlorite and epidote. Calcite also occurs as fine-grained aggregates with chlorite and epidote. Epidote perhaps formed at the expense of potassium feldspar, although no direct evidence was found. Chlorite probably formed as an alteration product of a mafic constituent such as pyroxene.

Pyroxenite*

Lens-shaped bodies up to 10 cm in width and up to 20 cm in length of a massive green rock which consists almost entirely of pyroxene occur within the quartz-epidote gneiss in High Rock Mountain in the Washington quadrangle. Pyroxenite also occurs in a large 10 m thick tabular body at least 50 m in length at the Pequest mine as well as on the dump at the Washington mine.

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* The term pyroxenite is used in this report as a purely descriptive term for the coarse-grained massive aggregates of diopsidic clinopyroxene which are commonly associated with magnetite concentrations. No genetic implications (e.g. igneous) are indicated. Earlier workers referred to these rocks as "skarn" rocks with the full genetic significance of that term.

The rock consists almost exclusively of green diopsidic clinopyroxene. Minor amounts of magnetite in the form of small (.1 mm) disseminated grains or as lens-shaped aggregates 10 mm wide and 50 mm long are locally present.

The pyroxene grains are pleochroic from bluish-green (X, Y) to pale-yellow (Z). Commonly, the grains are color zoned, the rims being more strongly colored than the cores. $Z \wedge C$ on a section parallel to (010) is about 45 degrees. The optic sign is positive and the optic axial angle (2V) is about 60 degrees.

The texture of the rock is granoblastic. Some grains are prismatic and show polysynthetic multiple twinning. All the grains are subhedral to anhedral, and triple points are common.

The spatial association of pyroxenite bodies with magnetite concentrations may be of genetic significance, and is discussed more completely in a later section.

Quartz-Potassium Feldspar Gneiss

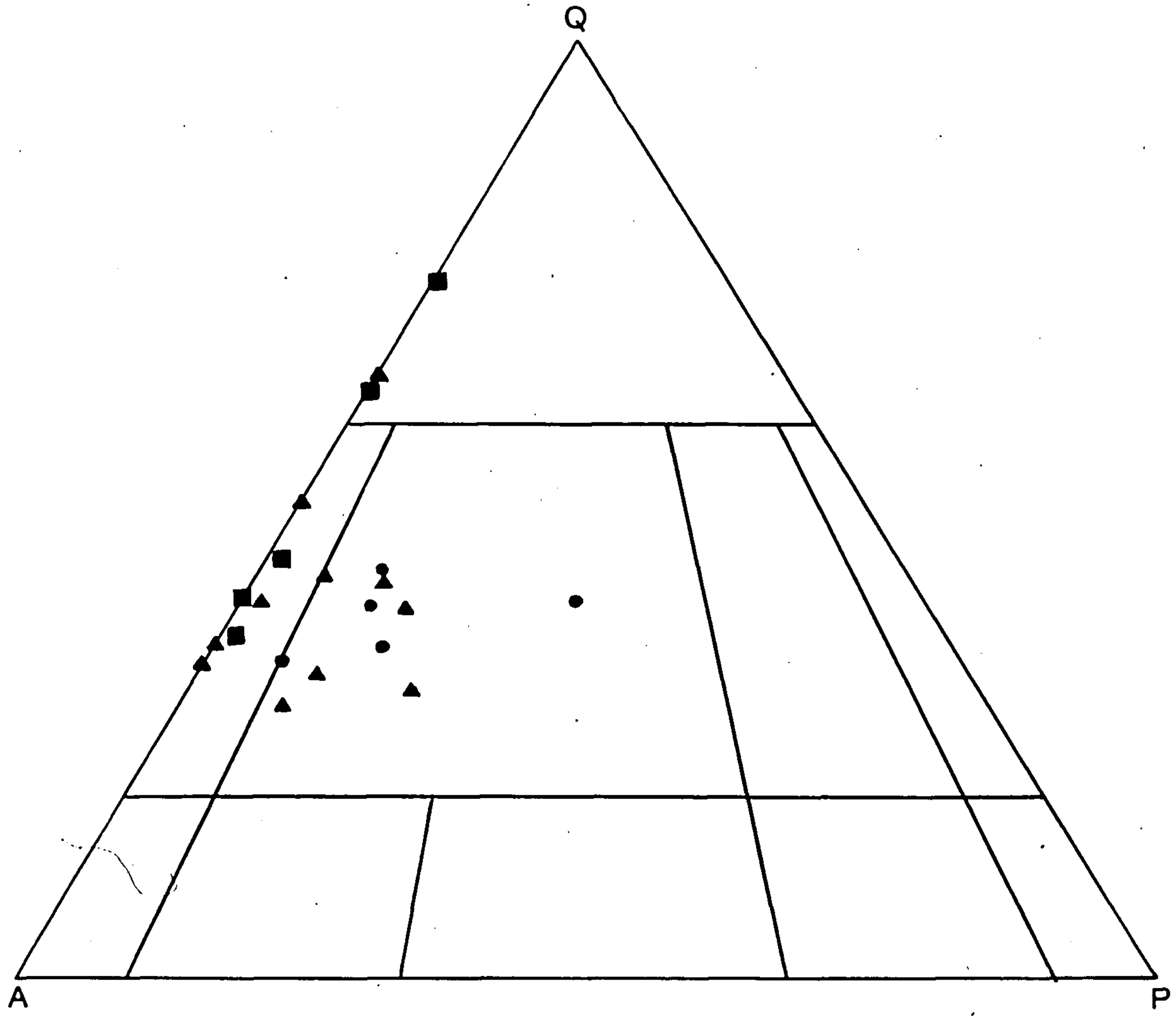
Bodies of microcline-bearing quartzo-feldspathic gneiss are common on Scotts Mountain in the Belvidere quadrangle. This unit contains layers and lenses of amphibolite, some of which are mappable.

The rock has a characteristic pink to pinkish-gray color. Layers of quartzite are found within the unit. Small bodies of microperthite alaskite are interlayered with the quartz-potassium feldspar gneiss both on a map and hand-specimen scale.

The quartz-potassium feldspar gneiss consists of variable amounts of quartz and microcline, and subordinate plagioclase (An_{8-16}), biotite, and magnetite. Accessory minerals include chlorite, apatite, zircon, sphene and hornblende. The texture is granoblastic, and foliation is poorly developed except where biotite, chlorite, or hornblende are present. The grains size ranges from .7 mm to 3.3 mm. Porphyroblasts of quartz are the largest grains and are commonly elongated parallel to the foliation. Anhedral grains of microcline are enclosed poikiloblastically in quartz.

Modes are given in Appendix A, Table 2. Fig. 3 shows modal data plotted on a QAP diagram. The broad range in composition is revealed both by the Q/A ratio and the microcline to plagioclase ratio. Although many of the data points fall in the field of alkali granite and granite, chemical analyses (Table 3) show marked differences from granites. It appears, therefore, that the quartz-potassium feldspar gneiss is probably a metamorphosed arkosic sediment and is not a plutonic

Figure 3. Plot of modes of quartz-potassium feldspar gneiss. Triangles are data from Baker and Buddington (1970); squares are data from Drake (1969); closed circles are data from this study. Q, quartz; A, alkali feldspar including albite up to An_5; and P, plagioclase.


Q
A
P

granite as has been suggested by some earlier investigators.

This rock may be distinguished from rocks of similar composition and appearance by its finer grain size, the presence of quartzite layers, and the virtual absence of perthitic feldspar.

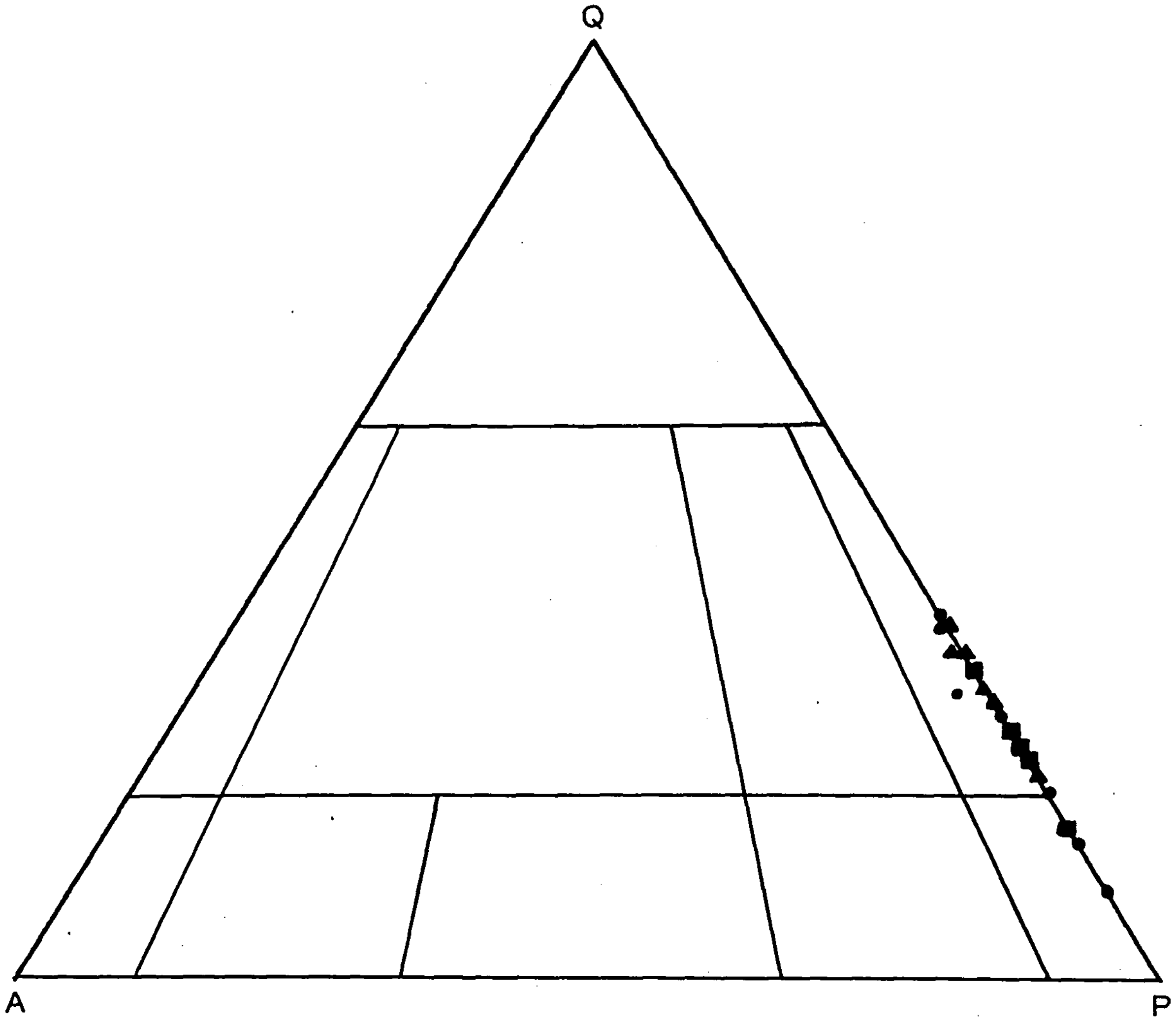
Oligoclase-Quartz Gneiss

This unit was termed the Losee Diorite gneiss, but the name currently used for rocks consisting principally of quartz and plagioclase (An_{10-14}) is oligoclase-quartz gneiss. The rock is light colored and usually greenish-gray to grayish-green. It is abundant and underlies much of the map area. The rock is massive, except where chlorite and hornblende are present in sufficient amounts to impart foliation and lineation, respectively. Thin layers of amphibolite of limited strike length which are generally less than 1 meter thick, are found in this unit. The grain size ranges from .07 mm in sheared samples to 3 mm in crystalloblastic samples. Porphyroblasts of strained quartz occur sporadically.

Clinopyroxene is sparce but occurs locally, and, in places, is altered to magnetite and chlorite. Pink garnet was found at one locality, and it appears to have formed by interface reaction of magnetite with plagioclase. Garnet-bearing oligoclase-quartz gneiss was reported in the Franklin and Hamburg quadrangles by Baker and Buddington (1970) who described a garnetiferous layer 300 m thick. Modes of the oligoclase-quartz gneiss are given in Appendix A, Table 3, which show that the proportion of plagioclase is far in excess of that of quartz. Fig. 4 shows the modal data graphically, and illustrates the variability of quartz and the very low content of microcline.

Locally the grain size of the oligoclase-quartz gneiss increases and the rock gradually grades into albite-oligoclase granite and microantiperthite alaskite. The latter apparently developed as a result of metasomatic activity which resulted in antiperthitic plagioclase with rims of albite. Apparently, both sodic and potassic solutions were active (Drake, 1969). Little evidence of anatexis was found in the area, but previous workers such as Drake (1969) have found relict folded amphibolite layers in albite-oligoclase granite which strongly suggests that the granite was not intruded but formed in place.

Figure 4. Plot of modes of oligoclase-quartz gneiss. Triangles are data from Baker and Buddington (1970); squares are data from Drake (1969); closed circles are data from this study. Q, quartz; A, alkali feldspar including albite up to An_5; and P, plagioclase.


Q
A
P

Garnet-Biotite Schist

A tabular body of garnet-bearing biotite schist 10 m thick occurs in the Belvidere quadrangle, about .2 km south of Hazen, N.J. The rock contains coarse-grained biotite and variable amounts of garnet, microcline, and magnetite. The biotite is pleochroic (pale-yellow to greenish-brown) and occurs as flakes up to 15 mm in diameter. The cleavage is often bent. The rock has an excellent foliation due to the alignment of biotite. Garnet is not confined to layers, and occurs as poikiloblasts enclosing microcline and biotite and as grains within poikiloblastic microcline. It appears, based on petrographic observations, that garnet is replacing biotite, as it cuts the foliation marked by the biotite. The garnet is pink, and it is probably rich in iron.

Magnetite occurs in grains elongated parallel to the foliation and in discrete grains. Layers of magnetite ranging in thickness from 10 mm to 100 mm are present. During 1882-1890, the Kaiser mine produced magnetite from an orebody 4 m thick in this rock (Bayley, 1910, 1941).

The garnet-biotite schist is interlayered with amphibolite, which, in turn, is interlayered with quartz-potassium feldspar gneiss and marble.

Plutonic Igneous Rocks

The igneous rocks of the study area include microperthite alaskite, pegmatitic alaskite, hornblende granite, biotite granite, syenite, albite-oligoclase granite, and microantiperthite alaskite. The feldspars of these rocks are dominantly perthitic, although smaller amounts of non-perthitic feldspar may be present. The feldspars of albite-oligoclase granite and microantiperthite alaskite, however, are dominantly non-perthitic although lesser but variable amounts of perthite are present.

Microperthite-Bearing Rocks

Granitic rocks having microperthite as the dominant feldspar are the most abundant plutonic rocks. These were termed Byram gneiss by early investigators. Microperthite alaskite and hornblende granite are the most common of the microperthite-bearing rocks.

Microperthite Alaskite

Belts of microperthite alaskite are found on County House Mountain, Mount Mohepinoke, Pohatcong

Mountain, and Musconetcong Mountain in the Washington quadrangle. Smaller isolated lenses crop out in the Belvidere quadrangle. Alaskite grades into hornblende granite and hornblende syenite on County House Mountain in the Washington quadrangle.

The unit is massive except where porphyroblasts of quartz are in planar alignment or oriented grains of disseminated hornblende impart a crude foliation. Locally, thin oriented layers of quartz-potassium feldspar gneiss, hornblende granite, and amphibolite give the rock a planar structure.

In the field, microperthite alaskite is gray to pinkish-gray. It can be distinguished from quartz-potassium feldspar gneiss by its coarser grain size and poorly developed foliation. The feldspars weather pale brown due to iron-oxide stains resulting from the oxidation of disseminated magnetite. Round quartz grains portrude from the weathered surface.

Some samples contain porphyroblasts of quartz and perthitic feldspar in a matrix of microcline, plagioclase and quartz. String mesoperthites composed of thin layers of potassium feldspar and plagioclase in approximately equal amounts are common in alaskite. Rocks containing string mesoperthites are usually equigranular and have a granoblastic texture. The grain size ranges from .2 mm to 4 mm with an average of approximately 2 mm. In cataclastically deformed samples, the feldspar is less perthitic in the groundmass inasmuch as it has exsolved more completely during cataclasis. Primary igneous textures are absent. Myrmekitic intergrowths of quartz and plagioclase are found in minor amounts in most samples.

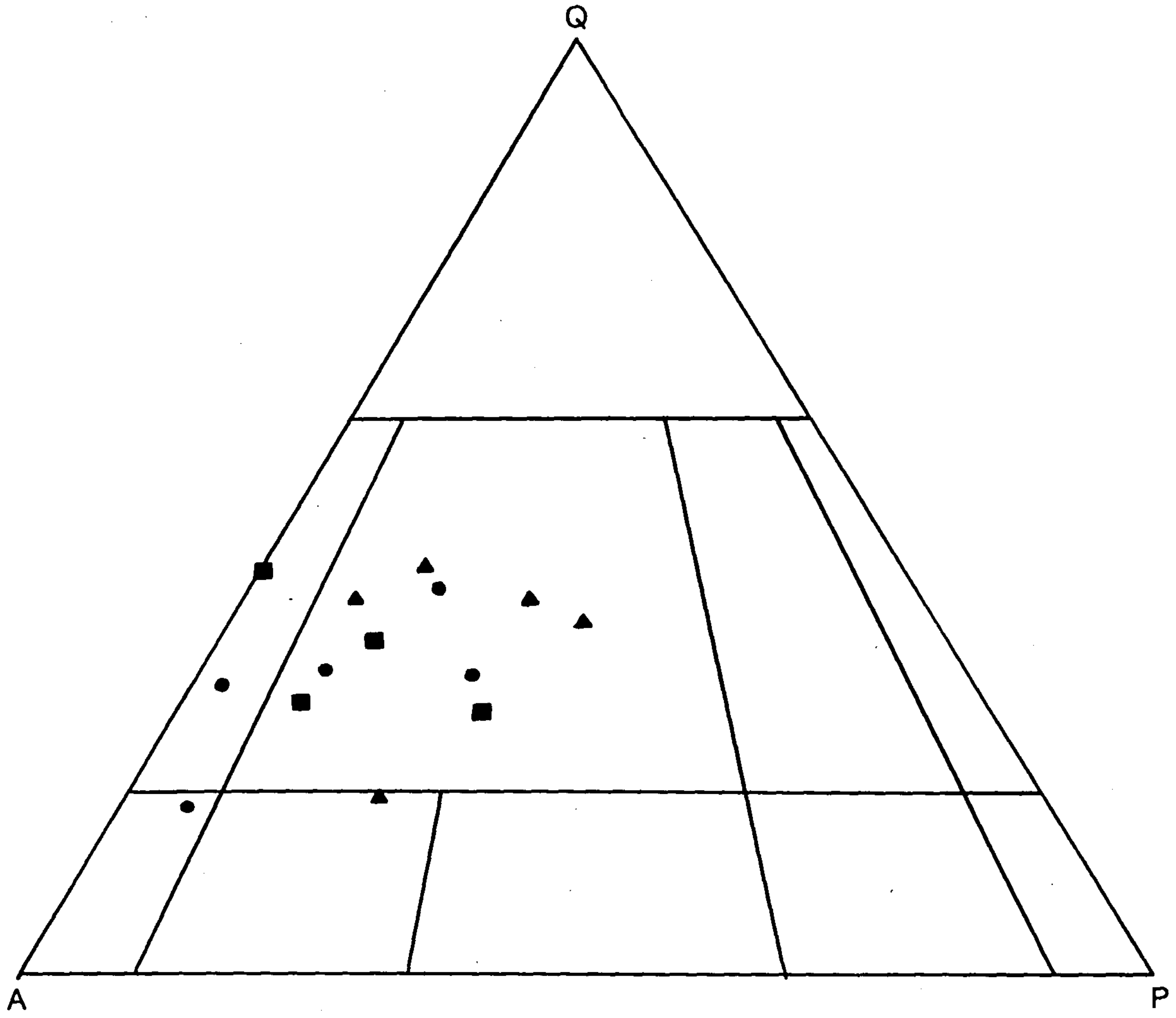
Modes of alaskite are given in Appendix A, Table 4. Fig. 5 shows the modal data plotted on a QAP diagram, with A representing alkali feldspar including perthite. The variable amounts of plagioclase may reflect different degrees of unmixing from mesoperthite.

Hornblende Syenite

A small body of hornblende-bearing syenite crops out on the southwestern edge of Oxford Mountain about 1.5 km south of the town of Oxford, N.J. The rock is characterized by its coarse grain size (up to 4 mm) and the abundance of dark hornblende. Locally, the hornblende is strongly lineated. The color of the fresh rock is bluish-gray to dark-greenish-gray; on weathered surfaces it is light brown,

The essential minerals are mesoperthite (4 mm in grain size) and hornblende. The feldspars are dominantly

Figure 5. Modes of microperthite alaskite. Triangles are data from Baker and Buddington (1970); squares are from Drake (1969); circles are new data from this study. Q, quartz; A, alkali feldspar including albite up to An_5; and P, plagioclase.


Q
A
P

string perthites. Microcline patch perthites having blebs of plagioclase enclosed in microcline occur along grain boundaries and between large mesoperthites. These grains are smaller than the mesoperthite and apparently represent crushed and recrystallized material. Quartz (.8 mm in grain size) is confined to areas between feldspar grains but also is enclosed poikiloblastically in mesoperthite. Quartz porphyroblasts are rare. Hornblende is commonly altered to biotite and chlorite. Apatite is a common accessory and occurs in grains up to .2 mm in diameter.

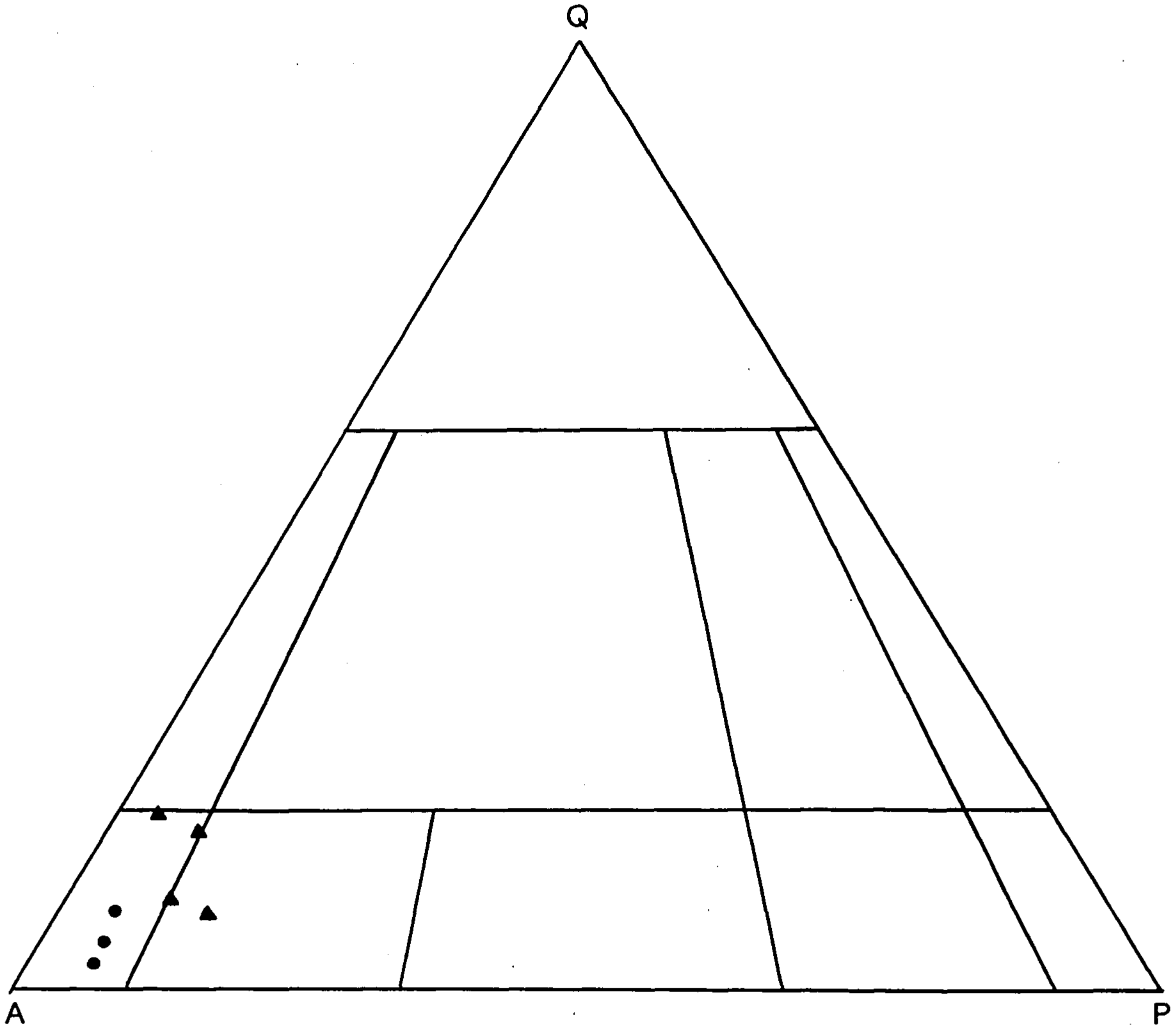
Modal analyses of the hornblende syenite are given in Appendix A, Table 5. The QAP plot in Fig. 6 shows the syenitic character of these rocks compared to similar rocks found in the Franklin and Hamburg quadrangles by Baker and Buddington (1970).

Hornblende Granite

Hornblende granite crops out along the northwestern edge of County House Mountain in the Washington quadrangle. It is associated with hornblende syenite and microperthite alaskite.

In the field, hornblende granite closely resembles hornblende syenite, the only difference being the quartz

Figure 6. Modes of syenite. Triangles are data from Baker and Buddington (1970); closed circles are data from this study. Q, quartz; A, alkali feldspar including albite up to An$_5$; and P, plagioclase.


Q
A
P

content. It differs from alaskite only in the amount of mafic minerals present. This is a gray to bluish-gray rock, is coarse grained (about 5 mm grain size), and it has a good foliation as defined by the planar orientation of hornblende.

The feldspars are typically perthitic and are string mesoperthites. Some samples have little perthite and contain instead single-phase microcline and albite-oligoclase. Hornblende is commonly altered to biotite and chlorite. Modes are given in Appendix A, Table 6.

An outcrop of hornblende granite near the town of Oxford on the Washington quadrangle is characterized by its non-uniform grain size, structure, and composition. The dip and strike of the foliation changes within distances of a meter, and the amount and size of hornblende grains varies markedly over the outcrop area of about 100 square meters.

Biotite Granite

A small body of biotite granite crops out on Mount Mohepinoke and County House mountain in the Washington quadrangle along the contact with microperthite alaskite and hornblende granite.

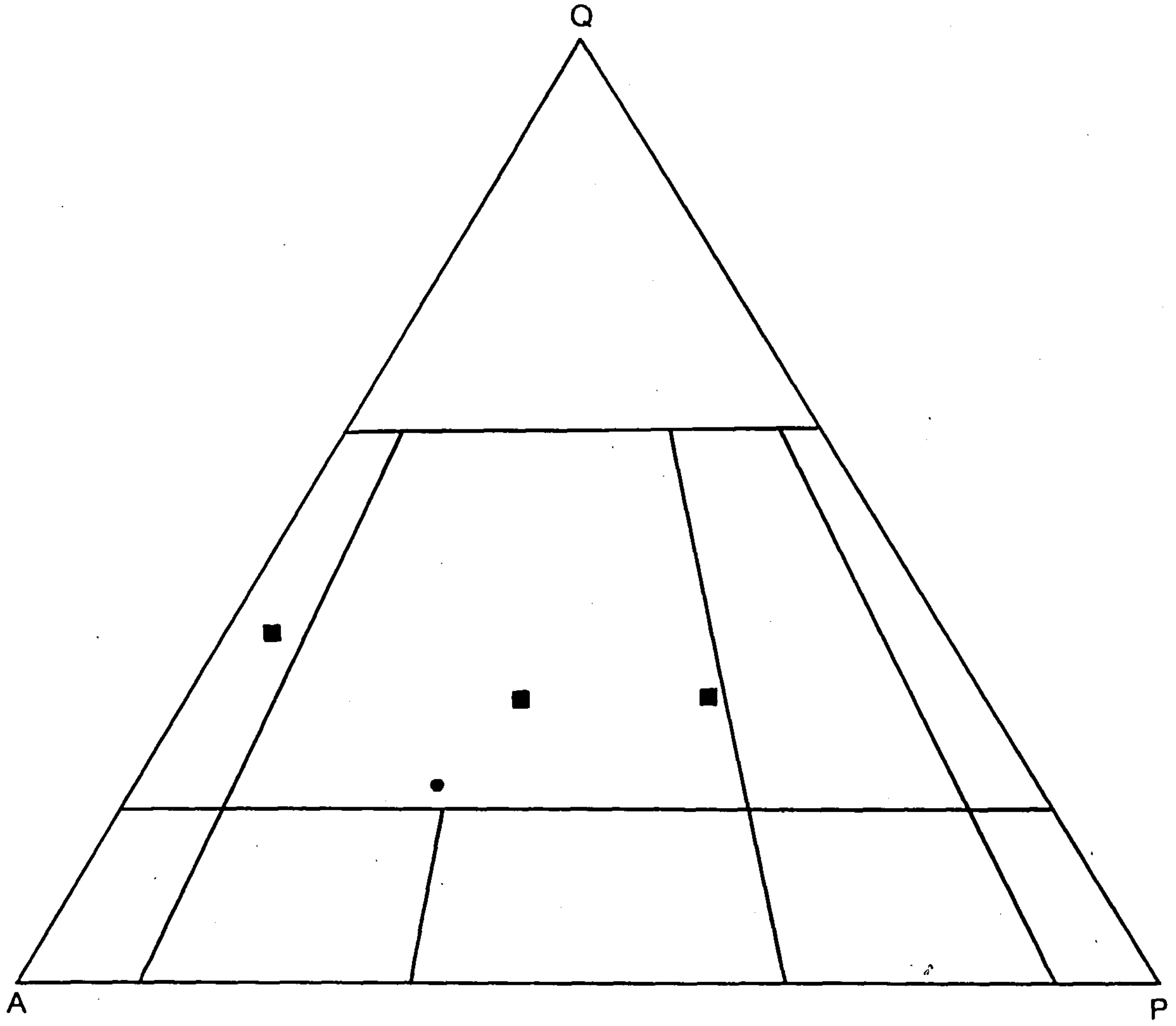
This rock is composed of quartz, perthitic microcline, and plagioclase plus about 6 percent of biotite. The feldspars occur as microcline patch perthite, mesoperthite, and antiperthite. Quartz is found as round grains between and enclosed by the feldspars. Hornblende is present in minor amounts, and is commonly replaced by chlorite. Accessory minerals include apatite, sphene and zircon.

Fig. 7 consists of modal data plotted on a QAP diagram, which shows the variation in composition of the biotite granite. Drake (1969) concluded, based on field relationships, petrographic data and chemical analyses, that the biotite granite formed by the assimilation of biotite-quartz-plagioclase gneiss by alaskite, or by the addition of potassium-rich fluids to the biotite-quartz-plagioclase gneiss.

Alaskite Pegmatite

Locally, in masses too small to map, pegmatites with the composition of microperthite alaskite are found. The rock is characterized by its pink color, in places it may be green due to the presence of plagioclase, epidote and chlorite. Where pegmatite is in contact with amphibolite, veinlets of granitic material were intruded along the foliation. The amphibolite is clearly

Figure 7. Modes of biotite granite. Squares are data from Drake (1969); circle, from this study. Q, quartz; A, alkali feldspar including albite up to An_5; and P, plagioclase.


Q
A
P

altered by emanations from a pegmatite intrusion at Oxford Church near Hazen, N.J. in the Belvidere quadrangle. Hornblende was transformed to biotite and quartz was introduced. Along joints, large (20 mm) grains of hornblende are found, presumably formed as a result of the introduction of warm fluids from the pegmatite.

Pegmatites are composed of variable amounts of microcline and albite-oligoclase plus quartz. At most places, feldspars are non-perthitic. Minor minerals include pale green clinopyroxene, chlorite, and epidote. The pyroxene is slightly pleochroic, and is probably a diopsidic augite. Apatite, sphene and magnetite constitute the accessory minerals. Sphene occurs in euhedral grains up to 1.1 mm in length. Apatite is also unusually coarse.

Although no primary igneous textures were observed, it is clear, from field observation, that these rocks are intrusive into the metasedimentary-metavolcanic sequence. The textures of the pegmatites are both granoblastic and cataclastic. In the latter, granulated quartz and feldspar grains form a relatively fine groundmass containing porphyroclastic quartz, plagioclase and microcline. Mortar structures further attest to the presence of shearing stresses subsequent to intrusion.

Because of the large grain size of the pegmatite, no modal analyses were made. It is clear, however, that their composition is not uniform throughout the samples observed, and does not necessarily fall on the ternary minimum in the system quartz-orthoclase-albite. This fact is important when discussing petrogenesis of the igneous rocks in a later section.

Sodic Plutonic Rocks

Albite-Oligoclase Granite and Related Pegmatite

A few small lens-shaped bodies of albite-oligoclase granite crop out on Scotts Mountain in the Belvidere quadrangle and Musconetcong Mountain in the Washington quadrangle. These are surrounded by oligoclase-quartz gneiss, and the contacts are gradational. It is not clear from field evidence whether this granite is intrusive or anatectic. Drake (1969) found evidence to support anatexis.

The rock is recognized by its coarser grain size as compared with that of the oligoclase-quartz gneiss. It has, however, the same mineralogy.

A small lens of albite pegmatite was found on

Mount No More in the Belvidere quadrangle. This is a very coarse equivalent of albite-oligoclase granite, and contains albite crystals as long as 15 mm.

Microantiperthite Alaskite

Small bodies of antiperthite-bearing alaskite are found on Scotts Mountain in the Belvidere quadrangle. These lens-shaped intrusions are surrounded by oligoclase-quartz gneiss.

Antiperthitic alaskite is recognized in the field by lenticular quartz grains which often have a brownish tint. The feldspars weather chalk white. Foliation and especially lineation are well developed as shown by the orientation of elongated rod-shaped aggregates of quartz.

The essential minerals are quartz, microcline and plagioclase (An_{10-12}). Some of the feldspars are perthitic. Quartz occurs as poikiloblasts which enclose round grains of feldspar as well as quartz that has a different optical orientation. Microcline, antiperthite, mesoperthite, and microcline microperthite occur as smaller grains which apparently represent crushed and recrystallized material. Varietal minerals include chlorite, biotite, and yellow-green slightly pleochroic clinopyroxene. Pyroxene commonly is altered to chlorite and magnetite. Modes are given in Appendix A, Table 7. These are compared with microantiperthite alaskites from Drake (1969) in Fig. 8. "A" represents alkali feldspar and includes microantiperthite.

PETROGENESIS

The rocks found in the study area have been derived from granulite-rank metamorphism of a series of sedimentary and volcanic rocks. Plutonic rocks, derived from sedimentary and volcanic rocks by anatexis occur in the area of this study.

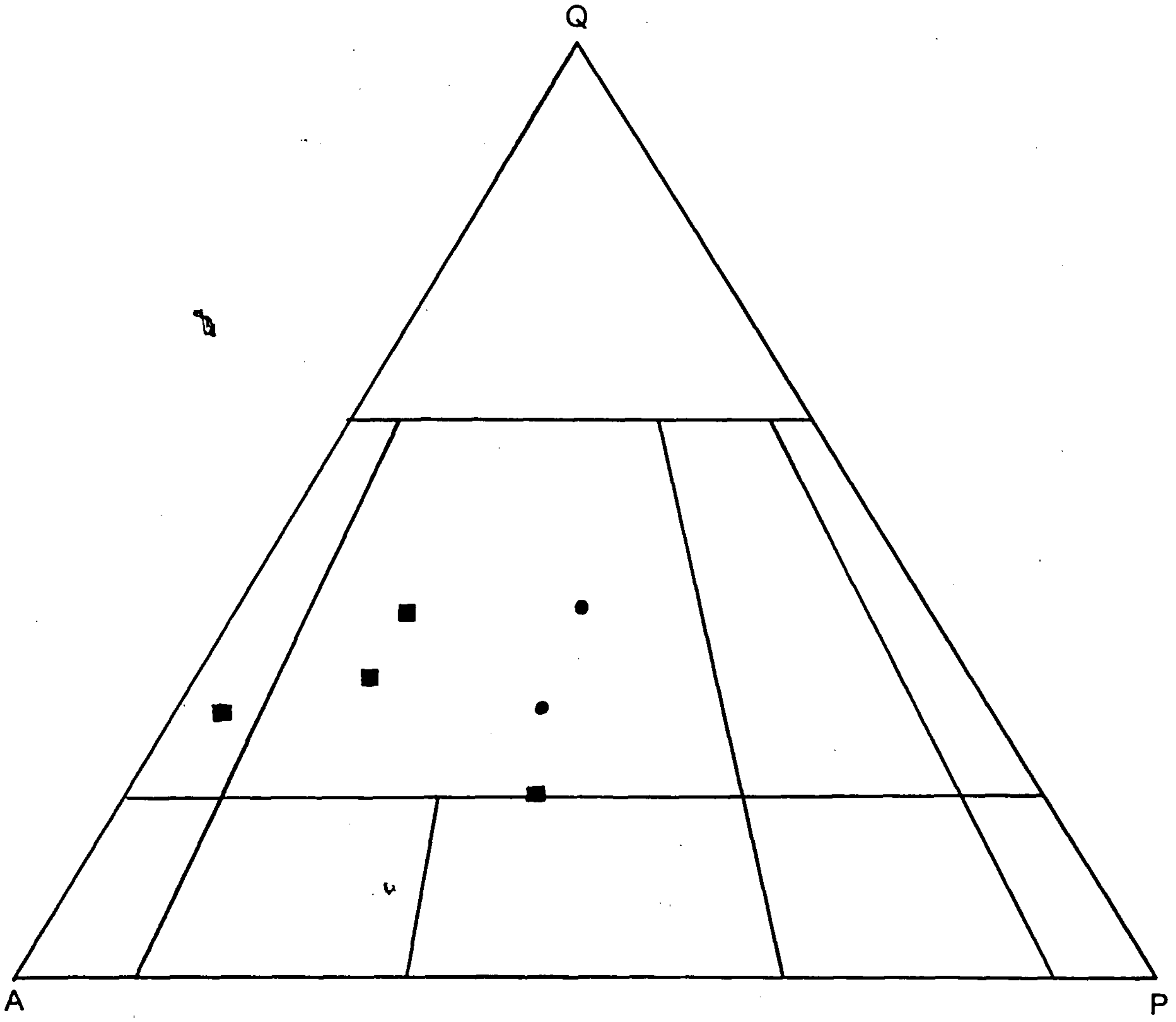
The Metasedimentary-Metavolcanic Sequence

Quartz-potassium feldspar gneiss, quartz-epidote gneiss, marble, and amphibolite in the study area are believed to be of sedimentary origin. Oligoclase-quartz gneiss is believed to be volcanic in origin. These conclusions are consistent with those of recent investigators such as Drake (1969), and Baker and Buddington (1970).

Amphibolite

As shown earlier, amphibolites of the Reading Prong in general and those in the area of this study in

Figure 8. Modes of microantiperthite alaskite. Squares are data from Drake (1969); circles are data from this study. Q, quartz; A, alkali feldspar including albite up to An_5; and P, plagioclase.


Q
A
P

particular are variable in composition. Table 1 shows a considerable variation in silica, lime, and magnesia. These variations may be due to either 1) differences in the unmetamorphosed parent rocks or 2) chemical changes during subsequent metamorphism. In order for geochemical studies to be meaningful, it is necessary to assume that little if any material has been added or removed prior to their metamorphism. Samples 3, 4 and 5 in Table 1 may have had material added inasmuch as they are located close (within 10 m) to an intrusion of pegmatitic alaskite. Sample 5 contains much more silica and sample 4 contains much more normative orthoclase than the average amphibolite which must result from the addition of granitic material to the amphibolite.

Baker and Buddington (1970) observed anorthositic and gabbroic layers in amphibolite in the Dover quadrangle, N.J. They also interpreted certain textures as primary igneous textures. Hague et al. (1956) report pillow structures in amphibolite in the Franklin-Sterling area which suggest that basaltic flows were the precursor of these rocks.

Sims (1958), on the basis of their association with obvious metasedimentary rocks, concluded that most

Table 1. Chemical analyses and CIPW norms of Amphibolite from the Reading Prong

	1*	2*	3*	4*	5*
SiO_2	53.2	48.4	51.3	56.8	68.3
Al_2O_3	14.4	16.7	12.3	15.3	11.6
Fe_2O_3	2.9	7.3	3.8	3.2	2.4
FeO	8.4	3.6	6.0	3.7	2.6
MgO	5.6	6.6	7.6	4.8	2.9
CaO	7.7	8.8	13.0	4.8	4.5
Na_2O	3.4	3.3	1.7	3.3	2.3
K_2O	1.6	1.8	2.6	5.0	2.8
P_2O_5	.27	.28	.25	.18	.09
TiO_2	1.5	1.0	.46	.86	.35
MnO	.24	.14	.25	.16	.10
H_2O^+	1.5	1.9	1.4	1.3	1.3
H_2O^-	.13	.13	.17	.19	.20
CO_2	.06	.07	.44	.28	.99
Sum	100.9	100.0	100.3	99.9	100.4

CIPW norms

	1*	2*	3*	4*	5*
Q	1.94	---	---	2.78	32.62
Ne	---	5.11	.94	---	---
Or	9.46	10.63	15.36	29.56	16.53
Ab	28.79	18.41	12.64	27.90	19.46
An	19.28	25.46	18.25	12.19	13.08
Wo	7.07	6.71	18.56	3.66	1.03
En	13.94	16.43	18.92	11.96	7.22
Fs	10.99	5.22	7.57	3.03	2.40
Il	2.85	1.90	.88	1.64	.67
Mt	4.21	---	5.51	4.63	3.47
Hm	---	7.30	---	---	---
Ap	.59	.62	.56	.40	.19
Cc	.14	.16	1.00	.64	2.25
Water	1.63	2.03	1.57	1.49	1.50
Sum	100.89	99.98	101.76	99.88	100.42

Table 1. (continued)

	6**	7***	8****
SiO_2	43.98	45.40	42.80
Al_2O_3	12.01	15.10	12.03
Fe_2O_3	6.60	3.35	6.56
FeO	12.20	9.75	13.25
MgO	5.46	7.90	4.67
CaO	11.99	9.30	11.76
Na_2O	2.93	2.50	2.91
K_2O	1.10	2.00	1.13
P_2O_5	.28	---	.17
TiO_2	2.25	1.62	2.24
MnO	.05	.15	.05
H_2O	1.33	2.37	1.33
CO_2	.18	.33	.20
Sum	100.36	99.77	100.10
		CIPW norms	
Or	6.5	11.8	6.67
Ab	14.3	14.5	15.20
An	16.3	24.0	16.40
Ne	5.7	3.6	5.11
Di	34.1	18.0	33.05
Ol	7.4	17.3	7.72
Mt	9.6	4.8	9.51
Il	4.3	3.1	4.26
Ap	.7	---	.40
Sum	98.9	97.1	98.32

---

* New analyses for this report. ** From Bayley et al. (1914). *** From Buckwalter (1962). **** From Darton et al. (1908). See Appendix B, Table 1 for sample locations.

amphibolites in the Dover district are metamorphosed impure calcareous or magnesian sedimentary rocks; however, he admitted that some may have had an igneous origin.

Amphibolite of almost certain sedimentary origin was found by van de Kamp (1971) in the Grenville Province of Ontario. This rock is a conglomerate having an amphibolite matrix which encloses pebbles of granite and gneiss. This amphibolite provides an unusual opportunity for comparing the chemistry of amphibolites of sedimentary and igneous parentages.

Rivalenti and Sighinolfi (1969) proposed that graywackes may become amphibolites upon metamorphism. It has since been shown by Van Alstine (1971), however, that graywackes are not likely progenitors of amphibolites.

In this study, three types of amphibolites were found: 1) those interlayered with quartzo-feldspathic gneisses, 2) belts of massive amphibolite, and 3) amphibolite interlayered with marble. No primary features such as igneous textures or structures were found in any of the amphibolites. Amphibolites may be derived through regional metamorphism of at least four different rock types: 1) basic volcanic flows, 2) basic tuffs, 3) differentiated layered intrusions, and 4) impure carbonate sediments.

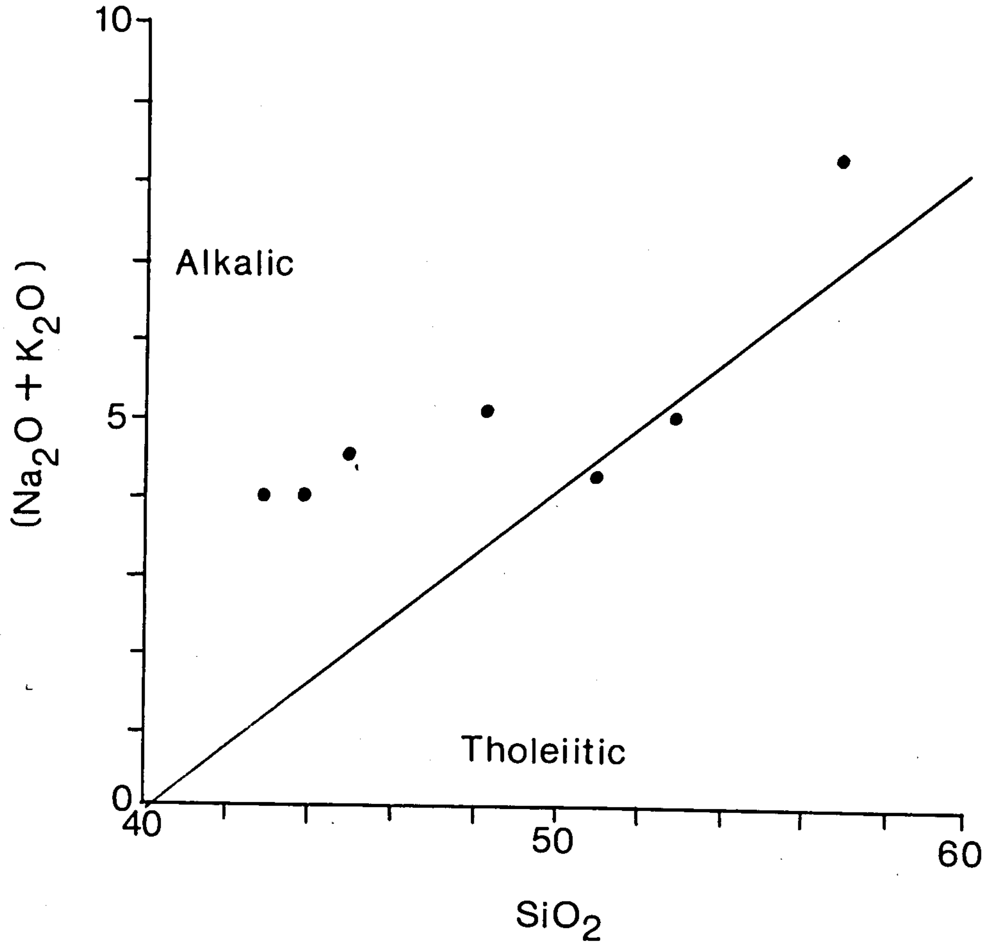
Inasmuch as no evidence of igneous layering was found in the study area, the problem, then, is to differentiate between amphibolites which are the products of the metamorphism of basic flows, tuffs, or sediments containing mixtures of carbonate and pelitic material.

As seen in Fig. 9, most of the amphibolites in the Reading Prong are slightly alkalic including those containing high silica. None of the samples analyzed fall in the field of tholeiitic basalt.

Fig. 10 shows that Reading Prong amphibolites generally follow the igneous trend. This in itself is not conclusive, because mixtures of dolomitic and pelitic sediments could also fall on the igneous trend.

To determine more closely the origin of the amphibolites from the study area, trace elements such as Ba, Cr, and Ni were considered. Figs. 11 and 12 show Cr and Ni trends respectively. For igneous rocks, there is a positive correlation between Niggli mg and Cr and Ni. A negative correlation would suggest a sedimentary trend (Van Alstine, 1971). As shown in Figs. 11 and 12, amphibolites from the study area show a negative correlation of Cr and Ni with respect to Niggli mg, and, in fact, transect the igneous trend.

Figure 9. Weight percent SiO_2 against total alkali, showing the slightly alkalic nature of amphibolite from the study area.



10
5
0
40
50
60
Alkalic
Tholeiitic
$(Na_2O + K_2O)$
SiO_2

Figure 10. Niggli c against Niggli mg for Reading Prong amphibolites (dots). Arrow, which indicates the igneous trend of the Karoo dolerites, and areas of pelites and dolomites are shown, from Leake (1964). Niggli values for Reading Prong amphibolites and the procedure used in calculating them are shown in Appendix C.

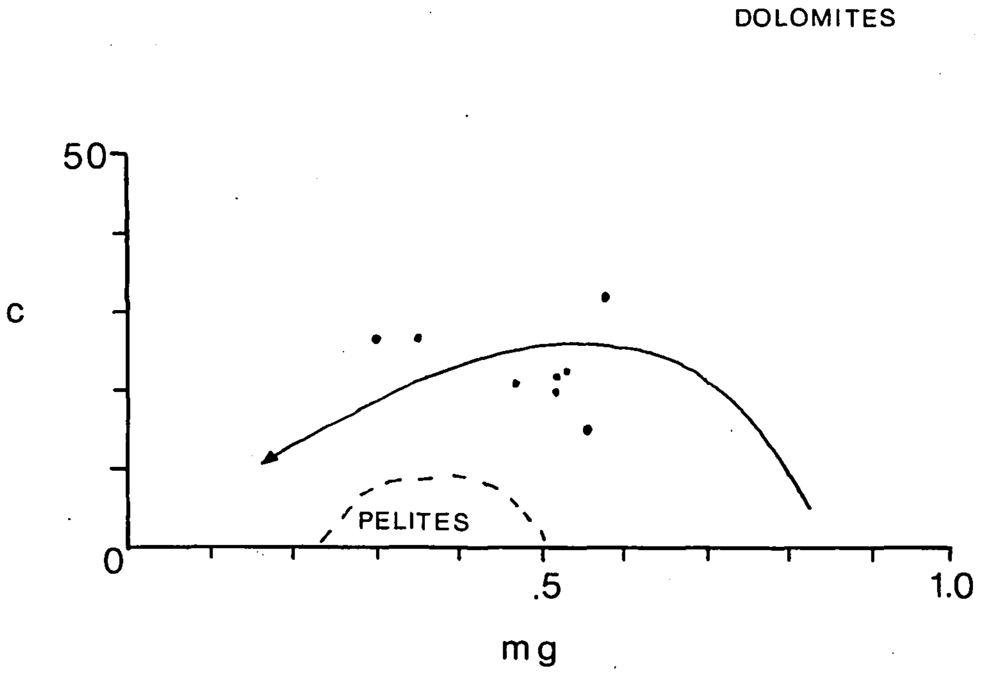

DOLOMITES
50
c
0
.5
1.0
mg
PELITES

Figure 11. Niggli mg against Cr. Arrow indicates igneous trend of amphibolites from Van Alstine (1971).

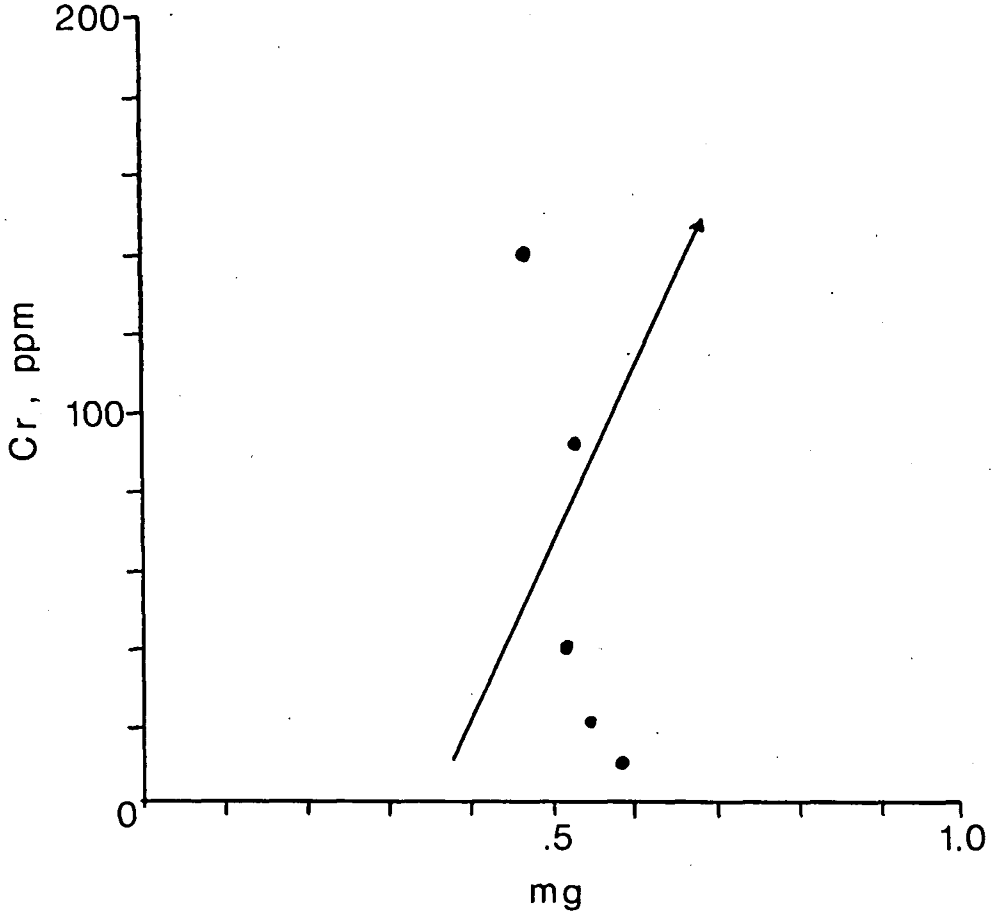
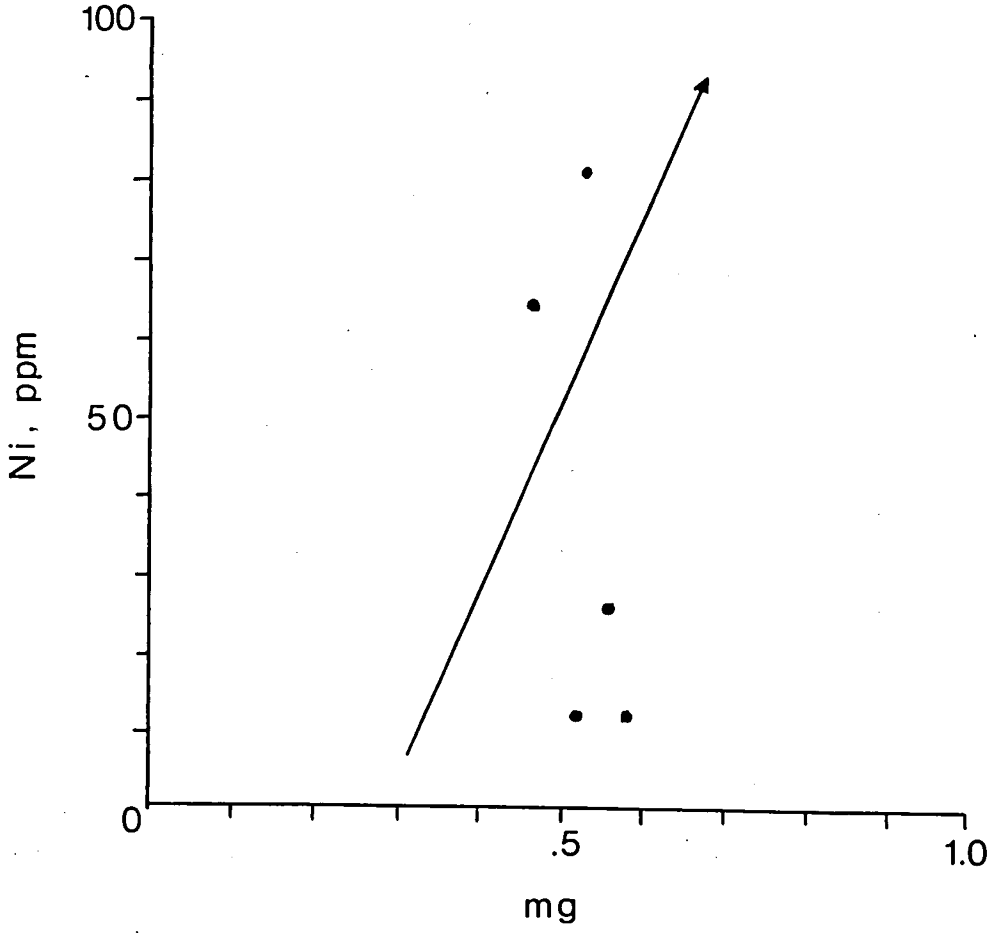

200
100
0
Cr , ppm
.5
1.0
mg

Figure 12. Niggli mg against Ni. Arrow indicates igneous trend of amphibolites from Van Alstine (1971).


100
50
0
Ni , ppm
.5
1.0
mg

Ba is also an indicator of the sedimentary vs. igneous origin of amphibolite. Van de Kamp (1968, p. 1348, 1352-1353) shows that orthoamphibolites from Ontario contain between 100 and 400 ppm Ba with an average of 173 ppm, whereas paraamphibolites contain between 100 and 1000 ppm Ba with an average of 355 ppm. Paraamphibolites clearly have a broader range and a greater average concentration of Ba, as would be expected from sedimentary rocks. Amphibolites from the study area contain between 230 and 1000 ppm Ba with an average of 562 ppm of Ba (see Table 2).

La/Ce ratios are also useful in differentiating ortho- from paraamphibolites. Van de Kamp (1968) found that La/Ce is usually greater than 0.4 for sedimentary rocks and less than 0.4 for basic igneous rocks. The ratio of La to Ce in two amphibolites from the study area are both greater than 0.4 as shown in Table 2. The trace element data thus strongly indicate that the amphibolites in the study area are of sedimentary origin.

Field evidence, especially at the Oxford Stone Quarry, show that the amphibolite at this locality is interlayered with oligoclase-quartz gneiss which is probably a metamorphosed volcanogenic rock. The parent rocks for the amphibolites were probably mixtures of

Table 2. Minor-element analyses of Amphibolite from the study area, ppm*

Ba	400	230	690	990	500
V	160	180	260	150	68
Cr	140	93	9.5	21	40
Ni	64	81	12	26	12
Cu	20	4.4	11	12	3.5
Zn	130	140	550	100	90
Sr	300	710	360	250	280
Y	53	23	22	49	37
Ce	n.d.	n.d.	n.d.	150	98
La	29	25	33	83	53
Zr	230	73	57	140	52
Nb	7.5	4.4	5.1	11	6.2
Pb	11	11	9.5	9.4	7.8
Ga	23	19	16	25	19
Nd	n.d.	n.d.	n.d.	110	n.d.
Co	18	34	17	16	8.6
La/Ce	---	---	---	.53	.54

---

* Emission spectrographic analyses by L. Mei, U.S. Geological Survey. The error limits on each number are plus 50% and minus 33%.

n.d.- not detected.

Sample locations given in Appendix B, Table 1.

pelitic and carbonate material.

The Sodium-Rich Rocks

The sodium-rich rocks include oligoclase-quartz gneiss, albite-oligoclase granite and albite pegmatite. Oligoclase-quartz gneiss is by far the most areally abundant of these rocks and underlies a large part of the map area. These rocks are unique in that they typically have high Na_2O/K_2O weight ratios, ranging from 2.9 to 5.6.

Field relations strongly suggest that albite-oligoclase granite and pegmatite are rheomorphic products of the oligoclase-quartz gneiss and were not intruded from an unknown magma source at depth (Drake, 1969). The problem, then, is to determine the parentage of the oligoclase-quartz gneiss.

Sims (1958) reported the presence of biotite-bearing oligoclase-quartz gneiss in the Dover district. This rock has essentially the same bulk chemical and mineralogical composition as the oligoclase-quartz gneiss found in the area of this study and along the Delaware River. Sims (1958) suggested that the progenitor of this rock was a graywacke. Drake (1969), however, pointed out that the Na_2O/K_2O ratios of most graywackes are too low to account for the high sodium content in the oligoclase-quartz gneiss.

Baker and Buddington (1970) showed, on the basis of field studies, that the weight of evidence for either a magmatic or a sedimentary origin for this unit is about the same. On a small scale, the oligoclase-quartz gneiss has discordant contact relationships with the surrounding rocks, but on a large scale it is concordant. Hague et al. (1956) considered this unit to be a phacolithic intrusion of dioritic magma.

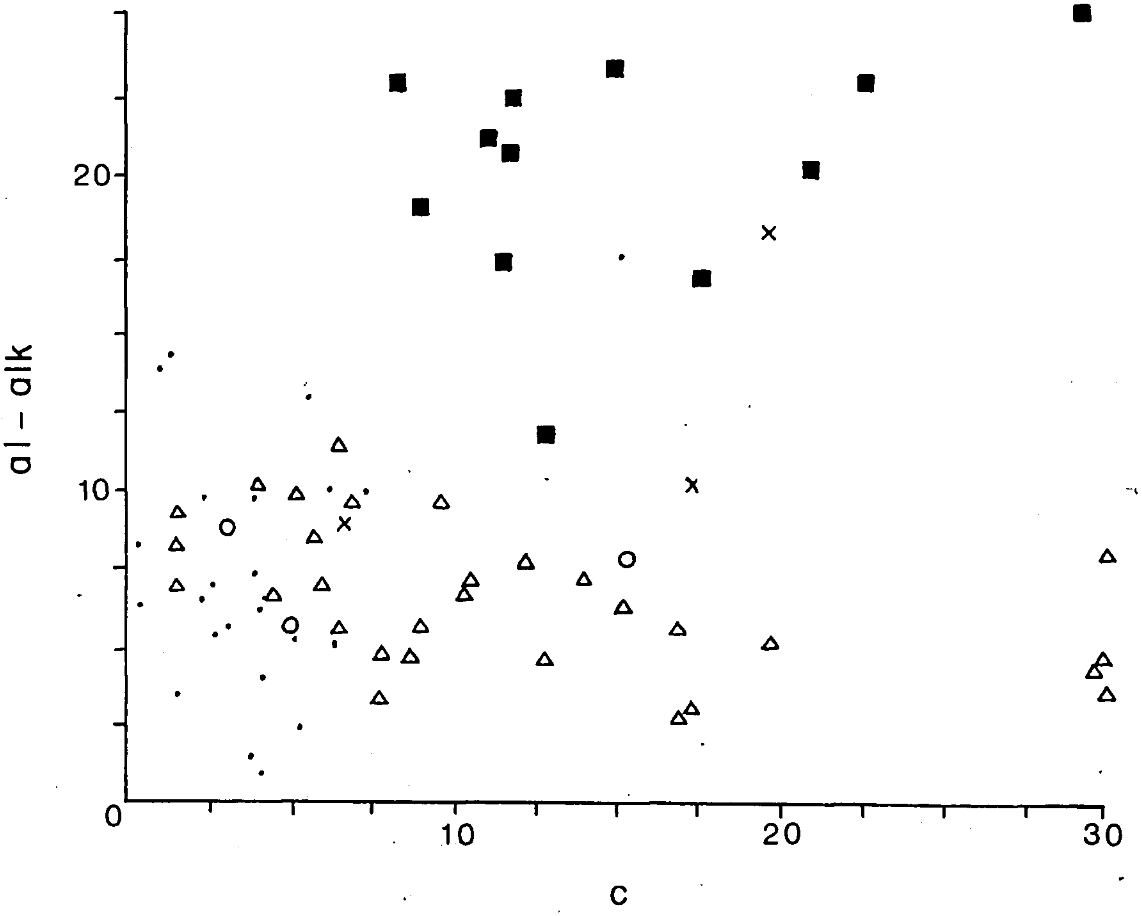
Hague et al. (1956) and Baker and Buddington (1970) pointed out the occurrence of garnet in a belt of the oligoclase-quartz gneiss in the Franklin area which would be consistent with a sedimentary origin. However, no compositional layering which might reflect sedimentary layering was found. Baker and Buddington (1970) concluded that the oligoclase-quartz gneiss was of uncertain origin.

The geochemical characteristics of the oligoclase-quartz gneiss may be indicative of its origin. Sodium-rich sedimentary rocks are relatively rare but are not unknown. Excellent examples are the Triassic Lockatong Argillite of New Jersey and eastern Pennsylvania and the Eocene Green River Formation of Wyoming, Utah, and Colorado. These formations are lacustrine deposits. The Lockatong Formation has a Na_2O/K_2O ratio similar to that of the oligoclase-quartz gneiss, but the SiO_2 content is much less in the Lockatong. Correspondingly, the Green River Formation contains more sodium than potassium (Milton and Eugster, 1959).

Fig. 13 shows the chemical trends of the Lockatong Argillites as compared to the trends of albite granites (Kinkel et al., 1956; Coleman and Peterman, 1975; and Brown et al., 1979), keratophyres (Gilluly, 1935; Bartrum, 1929, 1936; Battey, 1955; Albers, 1959; Dickinson, 1962; and Brown et al., 1979), and sodic gneisses of the Reading Prong (Drake, 1969; Bayley, 1941; Baker and Buddington, 1970; and Sims, 1958). There is a clear separation between the Lockatong rocks and the plutonic sodic igneous rocks. The oligoclase-quartz gneiss and related rocks from the Highlands of New Jersey clearly follow the igneous trend and overlap the keratophyre field and the albite granite field, and only part of the Lockatong field.

Keratophyres and albite granites are associated with ophiolites on Fidalgo Island, Washington (Brown et al., 1979). Here, albite granite dikes intrude gabbro and keratophyric tuffs and breccias. The breccias contain clasts of pelagic argillite and turbidite. They concluded that the sodic magma was not a

Figure 13. Plot of Niggli c against Niggli al-alk. Dots are keratophyres, squares are albite granites and plagiogranites, triangles are Lockatong Argillites, open circles are albite-oligoclase granites from the Reading Prong, crosses are oligoclase-quartz gneisses from the Reading Prong. Niggli c, al, and alk represent lime, alumina, and alkali contents of these rocks respectively. See Appendix C for Niggli values and the method used in calculating them.


al - alk
20
10
0
10
20
30
c

differentiation product of the gabbroic magma on the basis of the volume ratio of gabbro to albite granite dikes. Instead, this magma was intruded into the gabbro and subsequently reacted with sea water so that K was exchanged with Na. The sodic rocks, they concluded, were thus formed on the sea floor. Carmichael et al. (1974) also state that keratophyres and related sodium-rich rocks are commonly related to spilites, and, therefore, probably formed on the sea floor.

It is not clear if the oligoclase-quartz gneiss is in any way related to ophiolites inasmuch as contacts commonly are covered and the structure of the area is very complex due to the high rank of metamorphism and successive deformations. It may be related to ophiolites, but it is also possible that the high sodium content may be due to metasomatism of ordinary calc-alkalic rhyolites, andesites, and dacites (Gilluly, 1933; Carmichael et al., 1974). However, no evidence for sodium metasomatism in the oligoclase-quartz gneiss has been forthcoming from studies in the Reading Prong. The sodic oligoclase-quartz gneiss thus remains somewhat enigmatic in terms of its origin, but the chemical evidence strongly indicates that it is of volcanogenic parentage.

Quartz-Potassium Feldspar Gneiss and Associated Quartz-Epidote Gneiss

This unit is rich in quartz and, in places, contains garnet, sillimanite, epidote, and calcite. The mineralogy, therefore, indicates that it is a metasedimentary rock. Table 3 gives two chemical analyses of quartz-potassium feldspar gneiss which show both the high silica content and the variable Na_2O/K_2O ratio.

The likely progenitor for this rock is an arkosic sandstone and not a pelitic sediment. Pelites typically are poorer in silica and higher in alumina (Shaw, 1956; Pettijohn and Bastron, 1959; Nanz, 1953; and White, 1959). Arkoses, on the other hand, contain silica and alumina concentrations similar to that of the quartz-potassium feldspar gneiss (Pettijohn, 1963).

The quartz-epidote gneiss is spatially related to the quartz-potassium feldspar gneiss, and although no chemical analyses are available, they both contain moderate to large amounts of quartz. Epidote is probably an alteration product of potassium feldspar, which may be present in moderate amounts. Diopsidic

Table 3. Chemical analyses and CIPW norms of Quartz Potassium-Feldspar Gneiss from New Jersey.

	1*	2**
SiO_2	68.0	71.0
Al_2O_3	14.5	13.0
Fe_2O_3	3.8	3.0
FeO	2.1	1.3
MgO	.91	.32
CaO	1.4	.07
Na_2O	4.7	.95
K_2O	3.5	9.3
H_2O+	.42	.74
H_2O-	.15	.04
TiO_2	.72	.33
P_2O_5	.22	.06
MnO	.04	.03
CO_2	.01	---
Sum	100.5	100.1
	CIPW norms	
Qz	23.52	29.3
An	5.54	---
Ab	39.75	8.4
Or	20.71	55.6
Cd	.95	1.0
En	2.27	---
Fs	---	.8
Mt	4.68	3.2
Ap	.50	---
Il	1.37	.6
Hm	.58	.8
Cc	---	---
Sum	99.87	99.7

* New analysis for this report. Sample taken .25km S45W from Oxford Church, Belvidere quadrangle. Rapid Rock analysis by K. Coates and H. Smith, U.S. Geological Survey.

** From Drake (1969).

pyroxene in the quartz-epidote gneiss probably represents limy layers in the original sediment.

The Plutonic Igneous Series

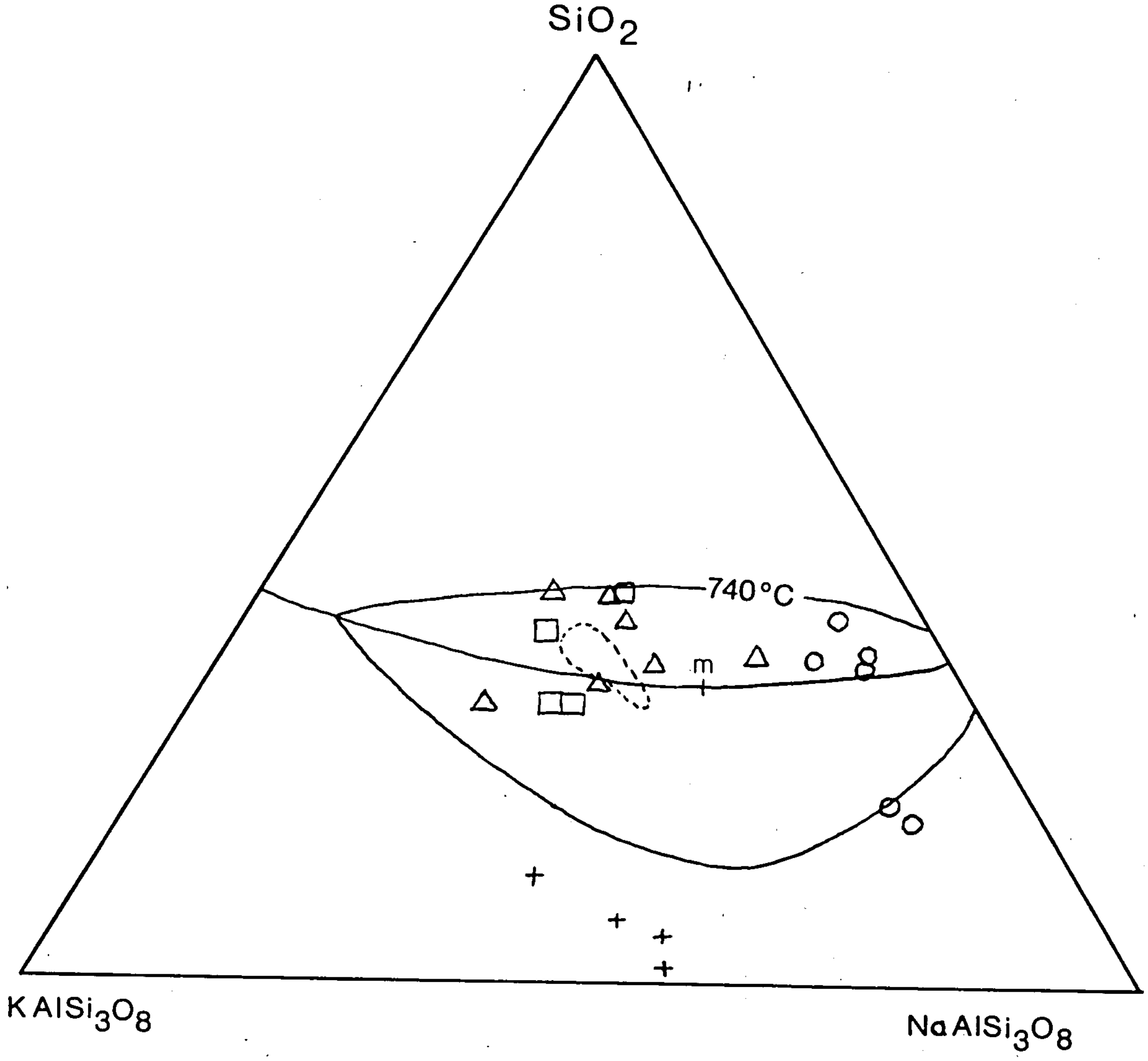
The Granitic and Syenitic Rocks

Hornblende granite and microperthite alaskite are considered to be magmatic for the following reasons: 1) variations in chemical and mineralogical composition are constrained within narrow limits; 2) the feldspars are dominaltly perthitic; 3) zircon and sphene are commonly elongated and subhedral to euhedral; and 4) general field relationships.

Baker and Buddington (1970) were able to find additional evidence for the intrusive nature of these rocks north of the Wisconsin terminal moraine where contacts are visible. Xenoliths of paragneiss enclosed in granite were found by Baker and Buddington (1970) and by Drake (1969), and cross-cutting (intrusive) relationships were found by Sims (1958) and Hotz (1952). Such evidence is lacking in the area of this study.

Fig. 14 is a ternary plot of the CIPW norms of the varieties of plutonic igneous rocks in the New Jersey Highlands. It shows that microperthite alaskite and hornblende granite do not fall on the ternary

Figure 14. Ternary plot of CIPW norms of plutonic rocks from New Jersey. Triangles are microperthite alaskites; squares are hornblende granites; circles are sodic rocks; crosses are syenitic rocks. Point m is the ternary eutectic at .3 GPa water vapor pressure at 670 degrees C, from Tuttle and Bowen (1958). Dotted outline is the field of 571 analyzed granites from Tuttle and Bowen (1958).


SiO2
740°C
m
KAlSi3O8
NaAlSi3O8

minimum for a .3 GPa water saturated melt, and do not necessarily fall in the field of analyzed granitic rocks. All of the analyzed alaskites and hornblende granites do, however, fall inside the 740 degrees C contour for water saturated melt. It has been shown by Carvalho (1978) that peak metamorphic temperatures in the Highlands were at least 760 degrees C, which would permit melting of all analyzed alaskites and hornblende granites in a water saturated environment. The presence of hydrous minerals in granitic igneous rocks from the Highlands indicates that some water was present, although it is not known if saturated melts existed.

The variability in composition of the granitic igneous rocks in the Highlands, it is concluded, reflects the variability in composition of the metasedimentary gneisses which were subsequently melted.

The presence of microperthite in alaskite, hornblende granite, and syenite, the chemical evidence cited above, plus the spatial association of these rocks with quartz-potassium feldspar gneiss, leads to the conclusion that the granitic rocks are partial melt products of the metasedimentary sequence and are not the products of alkali metasomatism as has been suggested by Baker and Buddington (1970).

Pegmatites, which are clearly intrusive, probably represent late fluids from alaskitic magma.

Young (1978) proposed that quartz syenite magmas in the Reading Prong were generated by partial melting of paragneiss at deep levels. He also concluded that the most syenitic rocks were generated by partial melting of anorthositic rocks at depth. There is no evidence for such a history in this area. There is a possibility, however, that the hornblende syenite is a refractory residue from the partial melting of paragneiss, the representing upon crystallization the alaskite and the hornblende granite.

As shown in the geologic map of the Washington quadrangle (Plate 2), the hornblende syenite is located in the hinge of an antiform. Although no textural evidence was found, granitic liquids could have been squeezed out of the rock during Grenville tectonism thus leaving the syenitic residue behind. Subsequent movement and recrystallization would have destroyed any textural evidence of anatexis.

Migmatitic hornblende granite occurs near the hornblende syenite body, and this may represent an intermediate stage in the partial melting which ultimately gave rise to a hornblende granite magma and a syenitic residue. More work is needed to fully understand this problem.

The Sodic Plutonic Rocks

Microantiperthite alaskite and albite-oligoclase granite are related spatially to oligoclase-quartz gneiss, and are probably mobilized portions of this gneiss (Drake, 1969).

Carl and Van Diver (1975) concluded that alaskites in the northwest Adirondacks are volcanic in origin based on Na_2O-K_2O-CaO and AFM ternary diagrams. Robinson and Leake (1975), however, have shown that igneous and sedimentary trends are identical on AFM diagrams along the A-F join. It is therefore impossible to distinguish chemical trends of sedimentary and igneous rocks unless data points fall along the F-M join as well.

With respect to this area, field relations show that alaskite and albite-oligoclase granite are spatially and perhaps genetically related to oligoclase-quartz gneiss which is probably volcanic on origin. Thus the microantiperthite alaskite and albite-oligoclase granite may have been derived from rocks of volcanic origin.

METAMORPHISM

Rocks of the study area are of high metamorphic rank as shown by the presence of migmatites and anatectic granites. Mineral assemblages in quartzo-feldspathic gneiss include quartz-microcline-sillimanite-almandine and that of amphibolites include plagioclase-hornblende-augite. These assemblages are characteristic of the sillimanite-almandine subfacies of the almandine amphibolite facies, and the hornblende granulite subfacies of the granulite facies (Fyfe et al., 1958; and Turner, 1968). In some amphibolite thin sections, hornblende clearly replaces clinopyroxene. The presence of clinopyroxene thus indicates that locally at least, metamorphic conditions were those of the pyroxene granulite facies. The presence of perthitic feldspar, even in some of the metasedimentary rocks, is compatible with hornblende granulite facies metamorphism.

Evidence for retrograde metamorphism to the greenschist facies is widespread, but is more apparent in some rocks than others. The quartz-epidote gneiss occurs over a large area (see Plate 2) and contains epidote-rich layers.

Biotite replaces hornblende in some amphibolites, which may indicate mesodiaphthoretic effects.

Epidote is also found locally along joints and shear zones, and is clearly not related to the main Grenville metamorphic event.

STRUCTURAL GEOLOGY

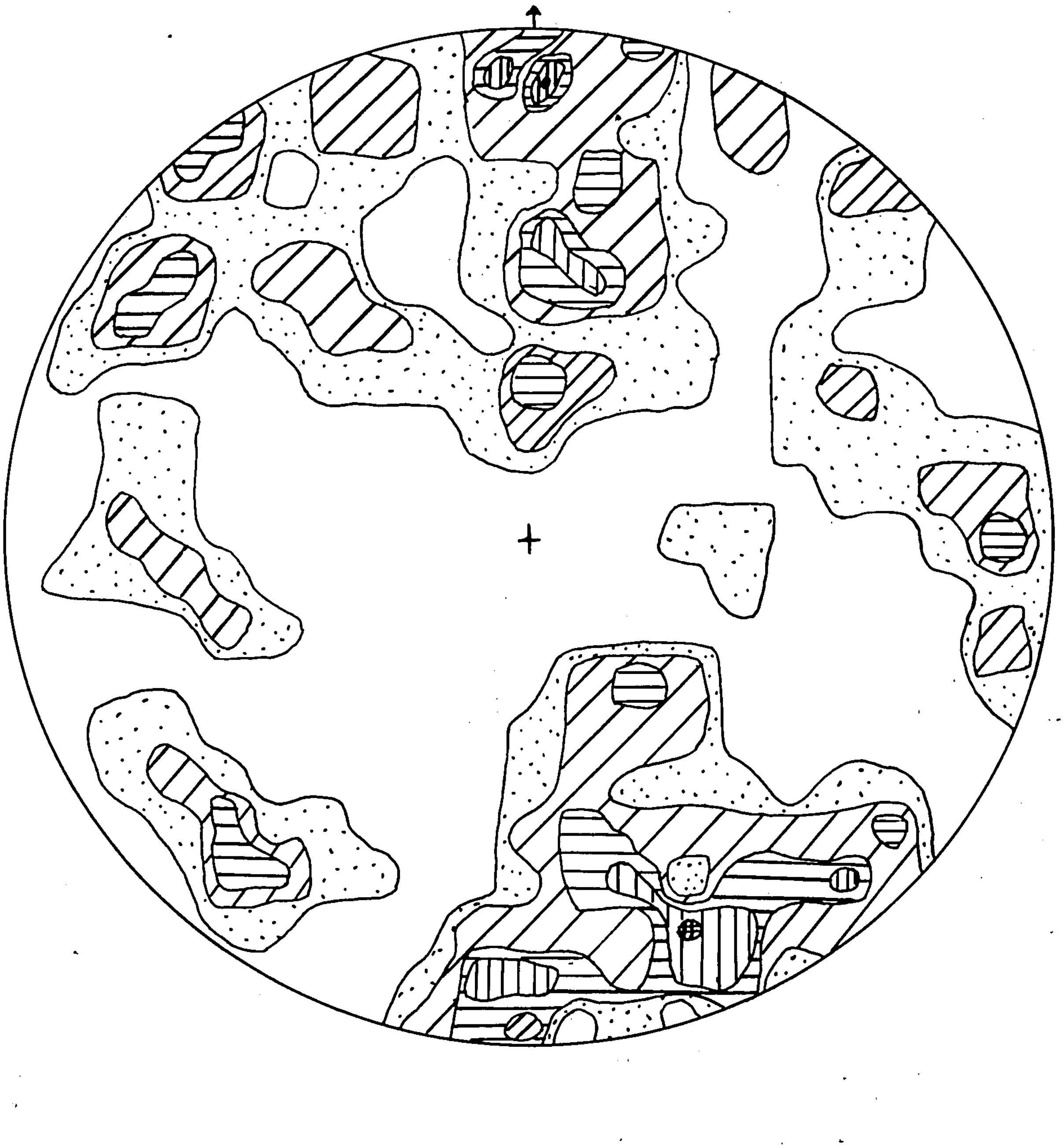
The Precambrian rocks in the study area constitute part of four main structural units, namely, Scotts Mountain, Upper Pohatcong Mountain, Oxford Mountain-County House Mountain, and Musconetcong Mountain. The general strike of crystallization foliation is to the northeast, and dips are steep to the northwest and southeast (Fig. 15). Locally there are variations in the strike due to subsequent periods of folding. The general northeast trend is due to tight isoclinal folding. Paleozoic structures have been superimposed on the Precambrian rocks, but were developed by brittle failure rather than plastic deformation.

Foliation

The Precambrian rocks have a foliation marked by the alignment of quartz and platy minerals such as hornblende and biotite, and by concentration of minerals into lenses and streaks. Most workers in the Reading Prong believe that foliation is mimetic after layering

Figure 15. Equal area contour diagram of 100 poles to foliation planes in the study area. The general trend is to the northeast, and dips are steep. Minor northwest trends are present which suggest multiple tectonic events.

░	▨	▤	▥	▦	■
1-2%	2-3%	3-4%	4-5%	5-6%	6-7%



in the premetamorphic rock, whether volcanogenic or sedimentary or both. Drake (1969) has found that foliation and compositional layering are not always sub-parallel, and that foliation is mimetic after cleavage. Foliation then transects the layering, and is seen in outcrop to be axial planar to tight isoclinal folds.

Foliation in the granitic rocks is parallel to that in the gneisses, and was probably formed during syntectonic emplacement. Commonly no structures are visible in the granites, and, where foliation is present, it is usually near a contact with gneiss.

Lineation

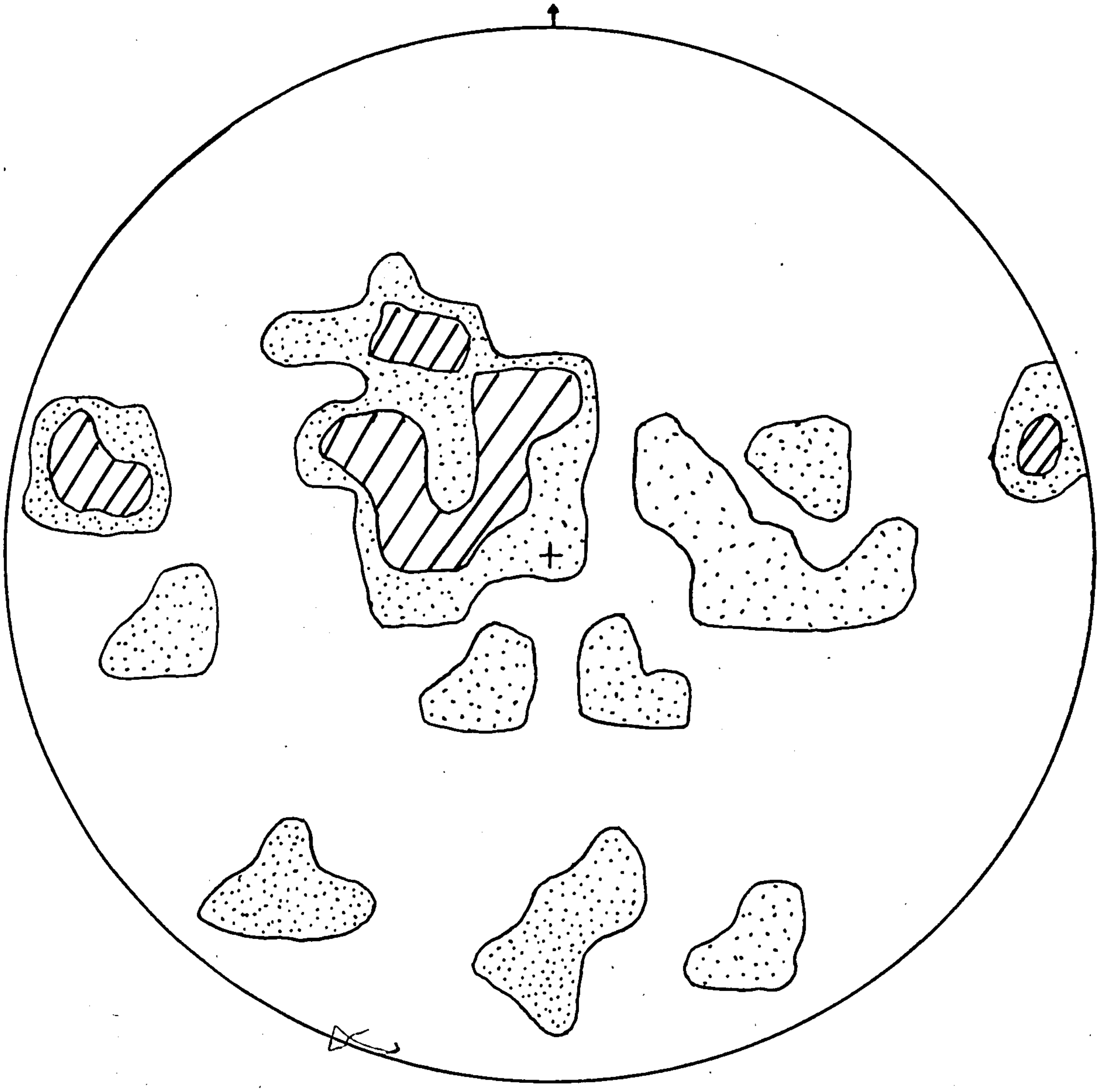
Lineation is well marked in rocks that contain abundant hornblende and biotite. In other rocks, quartz imparts a lineation by its occurrence in streaks and rod-shaped grains. Lineation is most commonly found in amphibolite, but is also weakly developed in some of the quartzo-feldspathic gneisses where prismatic minerals are locally abundant.

A stereographic plot of twenty six measurements of lineations (Fig. 16) in the rocks of this area shows a scatter of directions, but three strong maxima are

Figure 16. Contour diagram of equal area projection of 26 lineations from the study area. Three strong trends are present.

░ 4-8%

▨ 8-10%



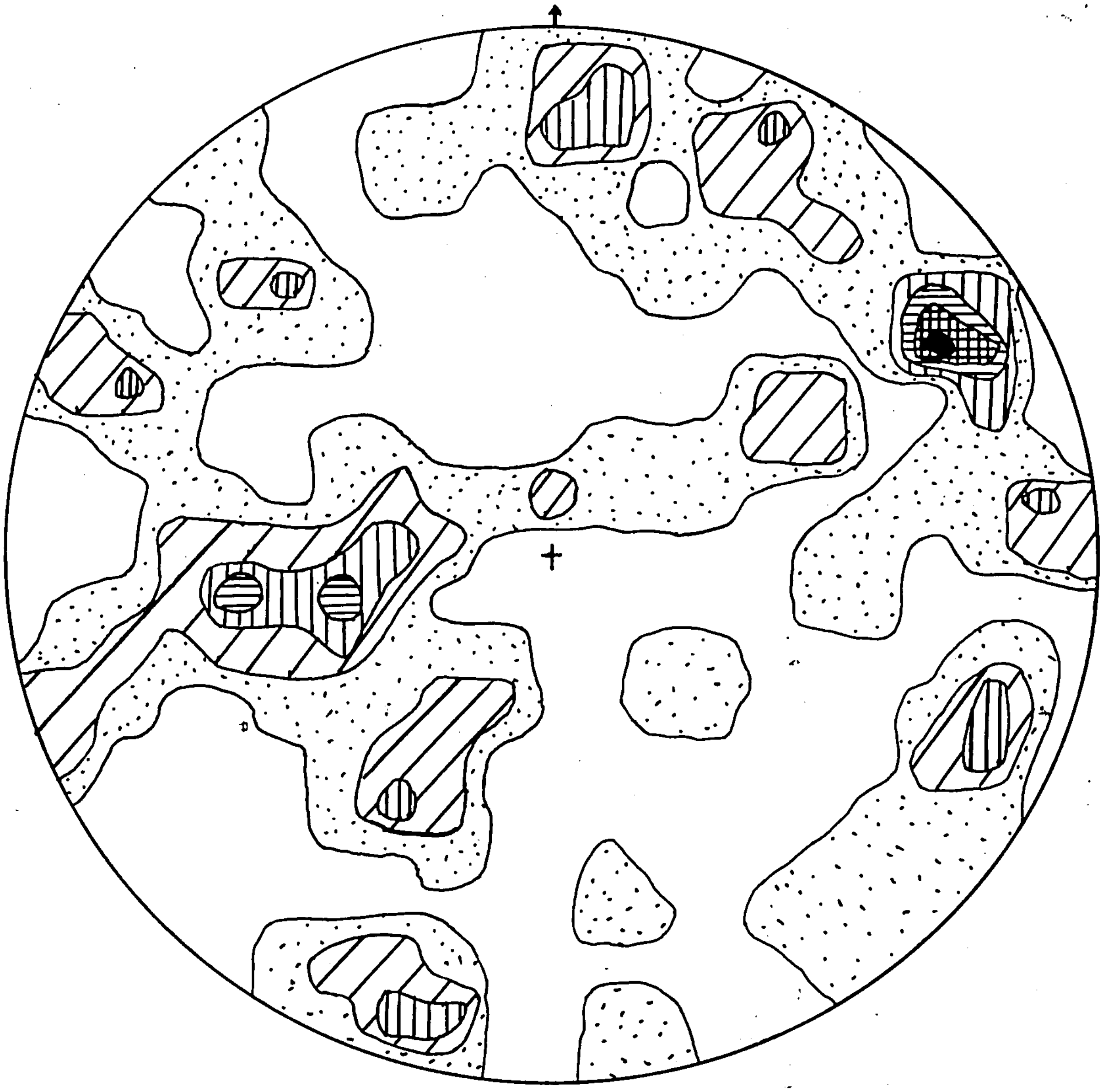
present. Drake (1969) found in the rocks of the Delaware Valley a statistical maximum for measured lineations which plunges gently to the northeast. The present study has also shown this, plus two additional maxima plunging northwest, one steeply and the other gently. Because lineation cannot be observed at most outcrops, it is difficult to recognize specific phases of folding, although it is obvious that the rocks have a complex tectonic history.

Joints

The Precambrian rocks contain well developed joint sets. Fig. 17 shows that most joints strike northwest and northeast and are steeply dipping. These are cross joints inasmuch as they are perpendicular to the general strike of the foliation. Longitudinal joints parallel to lineation are present but are far less common than cross joints. In places, joints are slickensided and filled with pegmatite and quartz veins. The geometrical relationships of joints to foliation were formed during the tectonic events which produced the foliation (Drake, 1969). Cross joints may be tensional fractures produced during uplift and nappe emplacement during lower Paleozoic time.

Figure 17. Contour diagram of equal area projection of 76 poles to joints. Cross joints, which transect foliation, are far more common than longitudinal joints which are parallel to foliation.

⁙	▨	▥	▤	▦	■
1-3%	3-4%	4-5%	5-8%	8-11%	11-12%



Folding

Folding is very difficult to recognize in the field. The lack of a known stratigraphic sequence of lithologic units in the study area along with the paucity of exposures has made detailed structural analysis impossible.

Minor folds having wavelengths of a few meters and micro folds having wavelengths of tens of millimeters are visible in a few outcrops, especially in fine-grained amphibolite. At least three and probably four phases of folding may be observed in one especially good outcrop at Oxford Church near Hazen, N.J. in the Belvidere quadrangle. Most measured fold axes plunge toward the northeast parallel to foliation, although a minor fold axis was observed to plunge southeast.

On a map scale, folds such as that associated with the hornblende syenite body near Oxford N.J. are revealed. A strongly developed lineation with a poorly developed foliation is visible at one outcrop which suggest a fold hinge. On a map view, this structure appears to be an antiform plunging east.

Another fold in amphibolite occurs southwest of Oxford. A core of very coarse-grained amphibolite is flanked by fine-grained amphibolite. The structure appears to be a southwest plunging synform which is overturned to the northwest.

Faults

In these high-grade crystalline rocks, faults are extremely difficult to locate because of poor exposures and the absence of stratigraphic markers. Faults are recognized in Precambrian rocks principally where they are in contact with Paleozoic rocks.

Locally, slickensides and ductile shear zones are recognized in the Precambrian rocks, but they are nearly impossible to trace for any great distance. These are probably related to Paleozoic deformation.

Drake (1969) found a fault along I-78 in the Bloomsbury N.J. quadrangle which cuts only Precambrian rocks and is filled with a lamprophyre dike of Precambrian age. He also found a flat thrust fault cutting Precambrian rocks at the same location.

A zone of blastomylonite about 10 m thick was found in the area of this study in Hazen, N.J., in the Belvidere quadrangle. It is characterized in the field by its aphanitic appearance, and its pale-pink to white color. It is spatially associated with a small magnetite deposit and its genetic significance with respect to iron mineralization is discussed in a subsequent section.

The entire sequence of Precambrian rocks has been involved in three major Paleozoic tectonic events. Drake (1969, 1970, 1978) concluded that the Precambrian rocks constitute a series of nappes de recouvrement which were thrust to the northwest along the Musconetcong fault. Evidence for this is found in the configuration of the lower Paleozoic rocks, from outcrops which show Precambrian rocks thrust over Paleozoic rocks, and from geophysical evidence.

Brock et al. (1976), however, disputed the refolded nappe idea of Drake (1969). They concluded that the Precambrian rocks in northwestern New Jersey were thrust over Paleozoic rocks and not emplaced as cores of nappes. Similar structural patterns from ridge to ridge and significant differences between Precambrian and Paleozoic structures were cited as evidence for this view.

It should be noted, however, that plastic deformation did not occur in the Precambrian rocks during Paleozoic tectonism. It is not surprising, then, that the structural patterns are different.

Similar structural features between Precambrian ridges do not by themselves refute the refolded nappe idea. The Precambrian rocks may have been emplaced as a block in the core of the nappe and later eroded to their present configuration (Drake, 1969, 1970, 1978).

MAGNETITE DEPOSITS

History and Production

The Oxford Iron district, which is located in the center of the study area, has been an important contributor to the total iron produced from the Precambrian Highlands of New Jersey.

The first mines supplied ore to the Oxford Furnace which began operations in 1743 and closed in 1925. Iron produced there was used for military equipment during the American Revolution. The early mines were small, and produced ore intermittently.

The Washington mine, located 1.7 km south of Oxford, N.J., was the largest single mine in the Oxford district and the production of ore began in 1854. A magnetic dip needle survey by Scranton (1879) showed that the orebody extended for about 1300 m, and had an average thickness of 5 m. There was no surface expression of the orebody.

The ore contained about 2.75 percent sulfur because of pyrite and had to be roasted in air prior to shipment to the smelter (Bayley, 1910). The Washington mine was worked long after all the others in the district had closed. It operated until 1965 when it closed, not for lack of ore, but because the costs of underground mining were too great (McInerney, oral commun., 1978).

Geology of the Magnetite Deposits

Magnetite concentrations are found in at least two rock types in the study area, namenly, 1) marble and 2) pyroxene-bearing quartz-epidote gneiss. Bayley (1941) reported that the ore in gneiss often pinches and swells, and that the thickness of such orebodies is quite variable. In marble, however, the orebodies are typically relatively uniform in thickness and continuous along strike.

Although no mines in the study area are currently operating, an abandoned pit which was the Pequest mine, in the Washington quadrangle .6 km east of Pequest, N.J. was examined. The ore in the Pequest mine was enclosed in a pyroxene-rich rock (here termed pyroxenite) composed almost entirely of green diopsidic pyroxene which, in turn, is associated with quartz-epidote gneiss. Green pyroxenite was also found on the dump pile of the Washington mine. Its bulk chemical analysis (Table 4) showed no similarities to pyroxenites of igneous derivation.

Mines in the Marble Belt

The Ahles Mine

The ruins of the Ahles mine were located in this study along with the dump pile from the mine. The Ahles mine is located in the east central part of the Belvidere quadrangle (see Plate 1). It is now impossible to observe the ore and host rocks in place. Bayley (1910) described the ore as:

> a mixture of soft brown limonite containing nodules of pyrolusite and crystals of magnetite, and often enclosing boulders of limestone, in which the limestone is partially replaced by limonite. There are also present in the ore great masses of chert or cherty limonite. Many of the large lumps of this material on the dump are gashed and jointed, and the surfaces are coated with cherty quartz.

The ore contained up to 11 percent MnO_2 (Bayley, 1910).

The mine was located at the contact of marble and quartz-potassium feldspar gneiss. The calcitic marble of the footwall contained colorless clinopyroxene and biotite, and, in places, oligoclase, microcline, and hornblende.

Other mines that were worked in the marble belt are no longer visible. However, it should be noted that they all produced iron ore containing very low concentrations of titanium, phosphorus, and sulfur, and relatively high concentrations of manganese (Bayley, 1910, 1941).

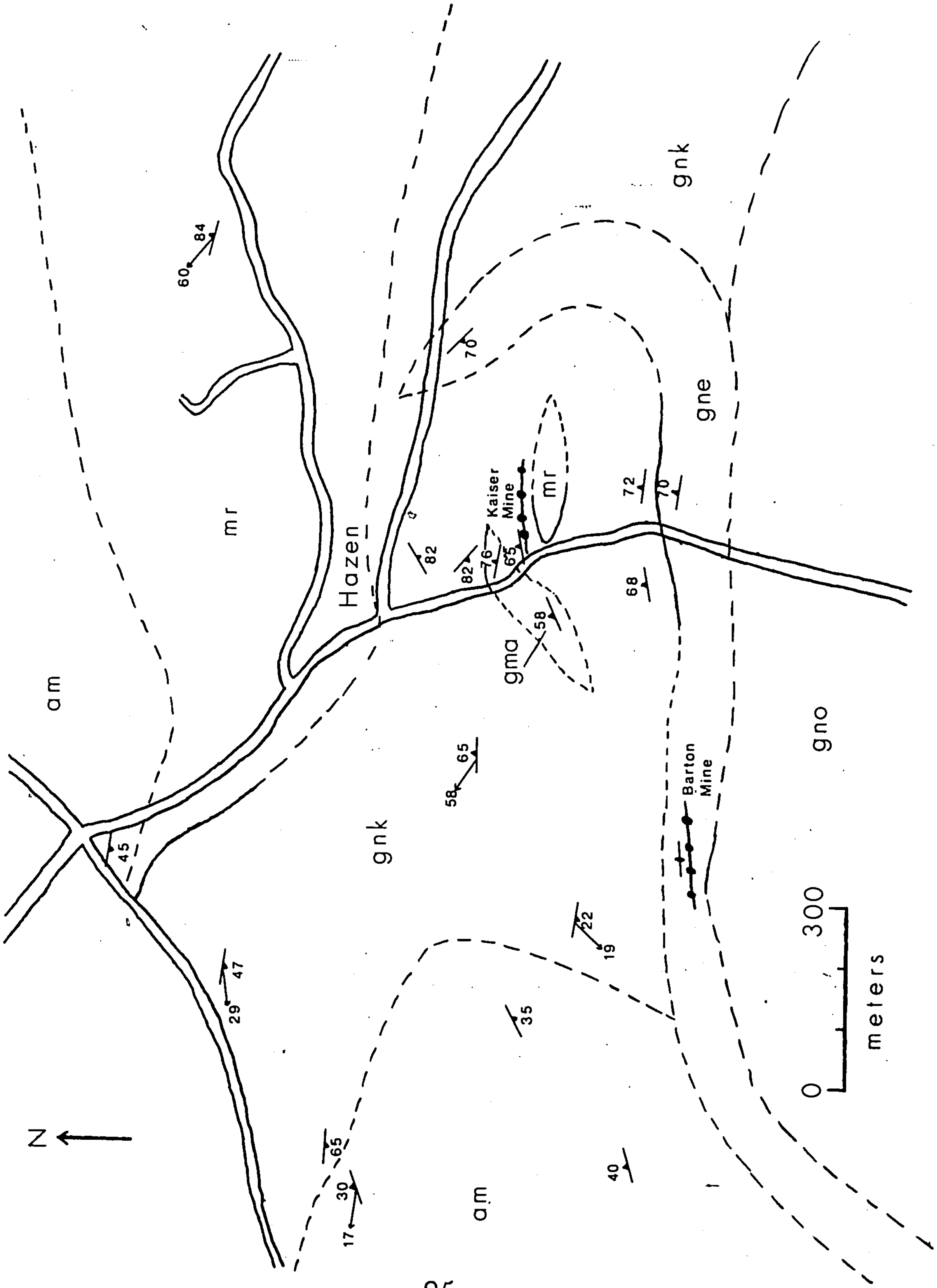
Mines in Gneiss

The Kaiser Mine

The shaft of the Kaiser mine, in the Belvidere quadrangle (see Fig. 18), is visible today although the mine was closed in 1891. The ore consisted of concentrations of magnetite in the form of layers parallel to the foliation of the host rock which is a garnet-biotite schist of variable composition (see description above). Pale-green clinopyroxene is present locally in the schist and, in places, may be associated with magnetite. The schist is bounded on the north by medium-grained amphibolite; the southern contact with marble is not exposed.

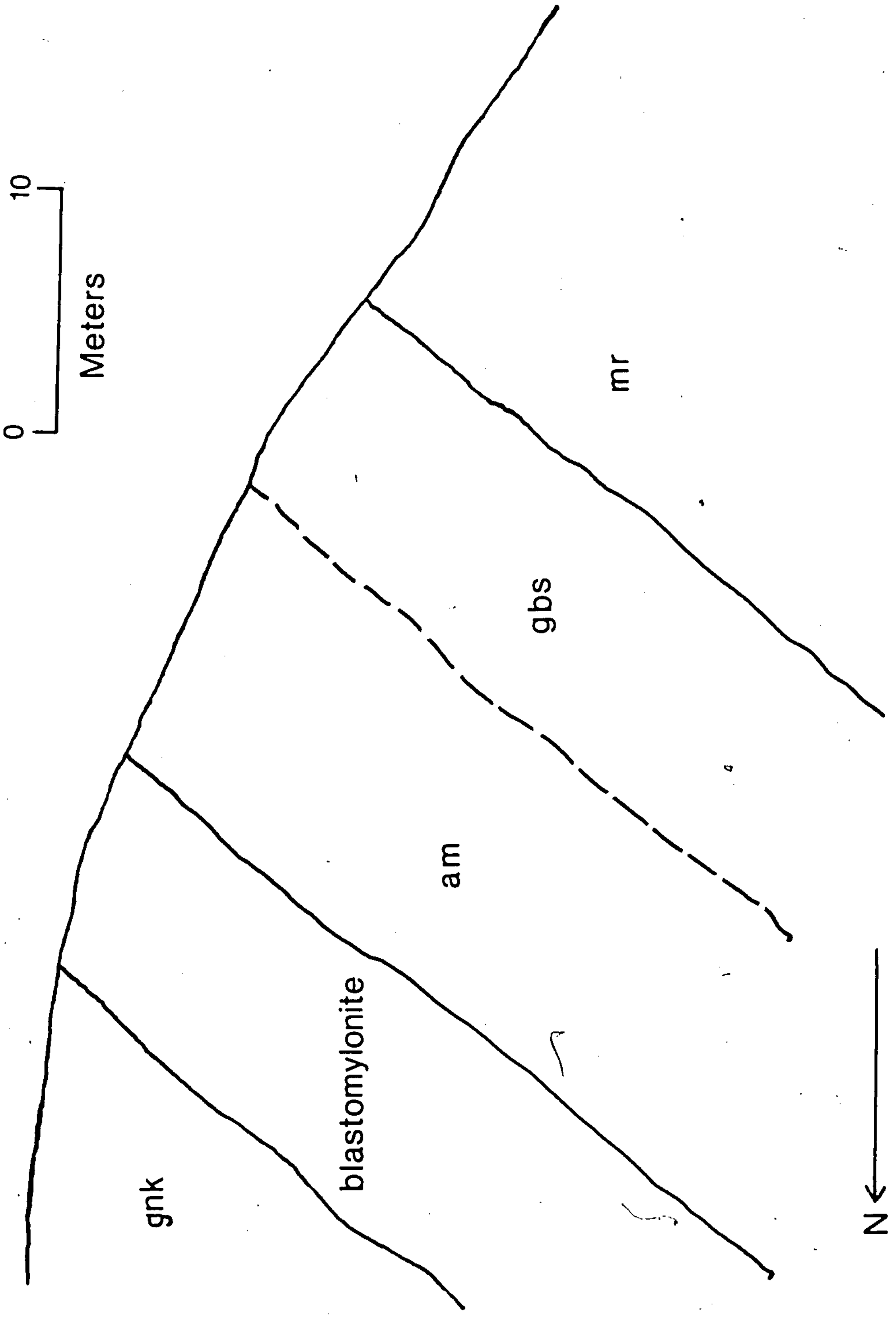
There is no surface expression of the schist, but by following the position of the shaft along strike it was found that the garnet-biotite schist is near the contact of amphibolite with a 10 m thick zone of blastomylonite. The foliation in the blastomylonite is parallel to the crystallization foliation in the surrounding less-sheared rocks (Fig. 19). The blastomylonite is composed of quartz, oligoclase, and

Figure 18. Detailed geologic map of the Hazen NJ area, showing the locations of the Kaiser and Barton mines. am - amphibolite, mr - marble, gne - quartz-epidote gneiss, gnk - quartz-potassium feldspar gneiss, gma - microantiperthite alaskite, gno - oligoclase-quartz gneiss. Contacts dashed where inferred.



am
mr
Hazen
gnk
gne
gno
gma
mr
Kaiser Mine
Barton Mine
60
84
70
82
82
76
65
72
70
68
58
65
58
45
22
19
47
29
35
65
30
17
40
N
0
300
meters

Figure 19. Generalized cross-section of the Kaiser mine area. gnk - quartz-potassium feldspar gneiss, am - amphibolite, gbs - garnet-biotite schist, mr - marble. Orebody is in gbs. Contact dashed where inferred, as the exact position of gbs is not known.

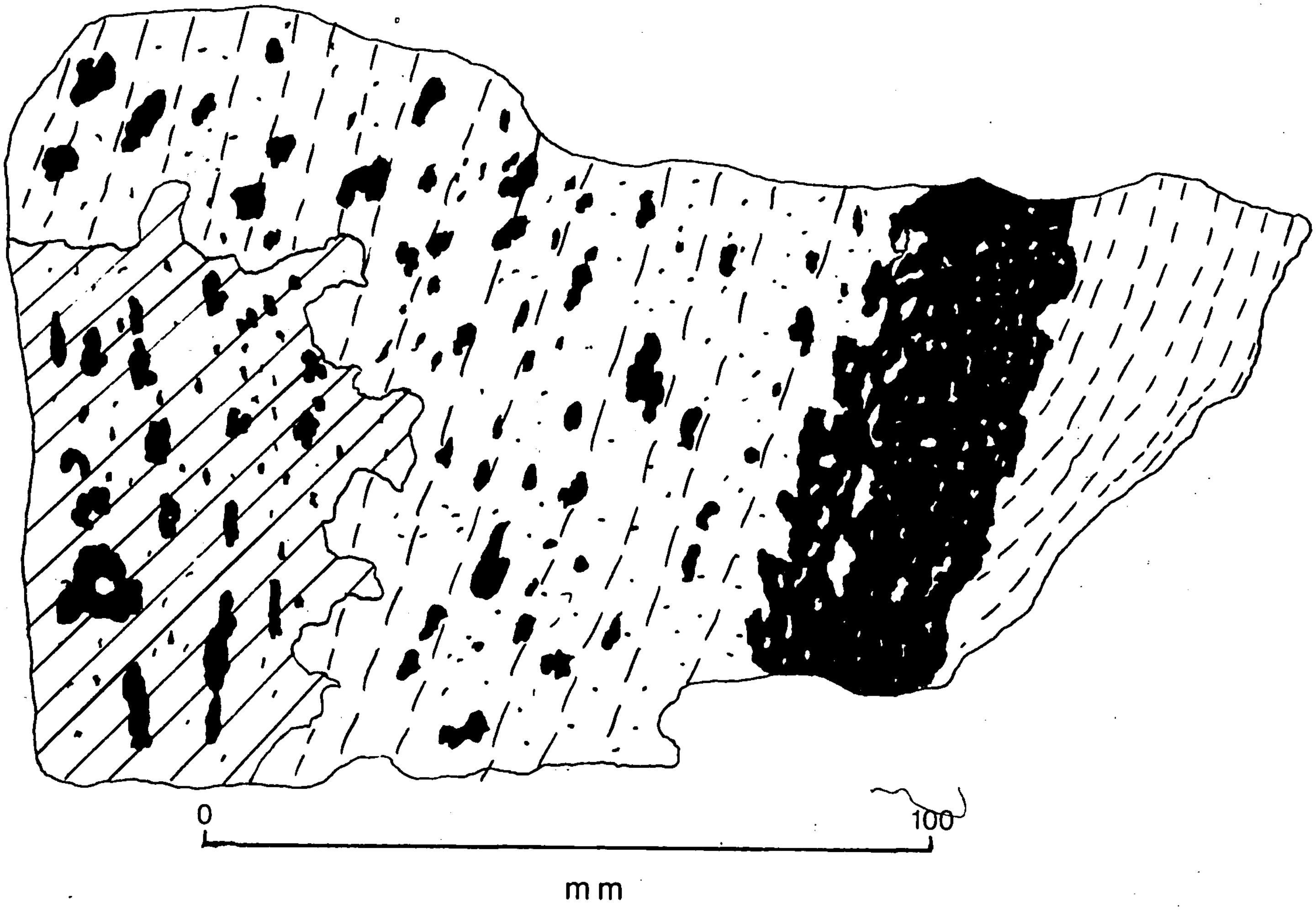

0
10
Meters
mr
gbs
am
blastomylonite
gnk
N

antiperthitic plagioclase and was originally microantoperthite alaskite. Less sheared alaskite occurs farther away from the mylonite zone. In hand-specimen, the mylonite is aphanitic. In thin section, quartz appears as round grains often completely enclosed by poikiloblastic plagioclase. Porphyroblasts of antiperthite occur as well as discrete grains of plagioclase and microcline which apparently exsolved completely as a result of cataclasis and recrystallization.

Fig. 20 shows that the garnet aggregates cut the foliation of the biotite-magnetite rock and therefore represent post-tectonic crystallization. The magnetite grains, however, are aligned with the foliation which suggests that they are pre-tectonic. Magnetite grains are embayed where they are in contact with garnet, suggesting that they were replaced by the garnet. Magnetite grains not associated with garnet appear relatively unattacked, and occur as equidimensional grains. Magnetite enclosed in biotite is typically euhedral. It appears that the garnet is a product of the post-kinematic prograde reaction of biotite and magnetite.

The magnetite occurs as discrete grains disseminated throughout a matrix of oriented biotite. In addition, it

Figure 20. Sketch of polished slab of magnetite-bearing garnet biotite schist. Black, magnetite; diagonal pattern, garnet; dashed pattern, biotite. Note that the garnet cuts the foliation.



0
100
mm

is also concentrated in layers about 30 mm thick. The magnetite-bearing zone was about 4 m thick, but a thickness of only 1.5 m of this zone was considered to be good ore (Bayley, 1910).

The magnetite itself is optically homogeneous. Oxidation lamellae of ilmenite are found in minor amounts, as are tiny oriented platlets of non-opaque spinels.

The Barton Mine

The Barton mine southwest of the Kaiser mine (Fig. 18), was operated between 1873 and 1877. Bayley (1910) described the ore as interlayered seams of hornblende, biotite, and magnetite.

Several pits are still visible and some magnetite-rich rock may be found on the dumps. The host rock is the quartz-epidote gneiss containing magnetite-rich layers parallel to the compositional layering of the gneiss. Dark-green pyroxene-rich layers are associated with the magnetite. Reddish-brown garnets were observed in the gneiss near the magnetite layers.

The Pequest Mine

The Pequest mine was located on Mount Mohepinoke in the Washington quadrangle, and was worked from 1869 to 1880. It was an open-pit operation which is still accessible today. Although the orebody was virtually mined out, the pit reflects its shape and size.

The north wall of the pit is composed of coarse grained (2.7 mm) pyroxenite that contains minor amounts of calcite and magnetite. The texture is granoblastic with many triple point grain boundaries. Along the contact of the orebody with the pyroxenite are large hornblende crystals, up to 100 mm in length. The magnetite is also very coarse-grained; grains up to 30 mm in diameter may be found.

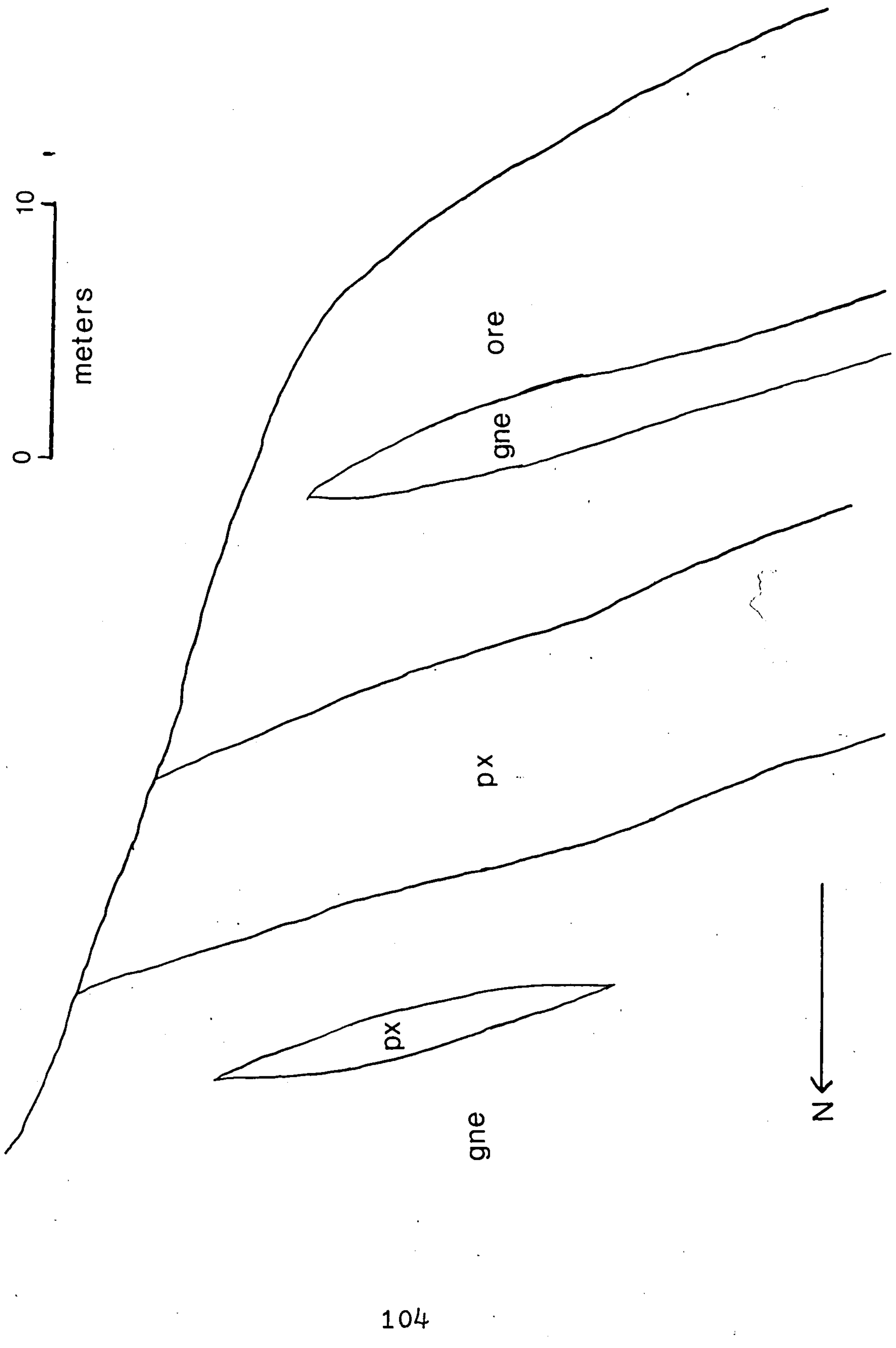
Fig. 21 shows a generalized cross-section of the mine as observed from the pit. The country rock surrounding the deposit is quartz-epidote gneiss which underlies much of Mount Mohepinoke (see Plate 2).

The Washington Mine

The Washington mine was an underground operation so that, at present, access to the orebody and host rock is limited to diamond-drill cores and the waste dumps.

Bayley (1910, 1941) described the wall rock as a

Figure 21. Cross-section of the east wall of the Pequest mine. gne - quartz-epidote gneiss, px - pyroxenite.


0
10
meters
ore
gne
px
px
gne
N

hornblende-bearing pegmatite. In the dump pile, many rock types are present, including alaskite pegmatite, green pyroxenite, amphibolite, quartz-potassium feldspar gneiss, and biotite schist having biotite crystals up to 40 mm in diameter. Green pleochroic diopsidic clinopyroxene occurs in the alaskite pegmatite and is commonly altered to chlorite. Magnetite grains are found in the pegmatite along grain boundaries between quartz and feldspar, and as discrete grains.

The pyroxenite is essentially monomineralic and is composed of diopsidic clinopyroxene containing minor but variable amounts of magnetite. The chemical analysis of a sample of pyroxenite, which contained very little visible magnetite in thin section, is given in Tables 4 and 5. The composition of the clinopyroxene, deduced from the whole-rock chemical analysis is $\mathrm{CATS}_{6.6}\mathrm{Di}_{75.2}\mathrm{Hd}_{13.4}\mathrm{Ac}_{4.8}$, and, when plotted on the pyroxene quadrilateral (Fig. 22), falls in the field of calcic salite. The green color and marked pleochroism of the pyroxene is probably due to the acmite content.

Amphibolite containing abundant disseminated magnetite is also present on the dump pile. Hornblende and colorless clinopyroxene are the mafic silicates in this rock. As shown by outcrops .7 km west of the mine,

Table 4. Chemical analysis and CIPW norm of pyroxenite from the Washington mine

	1*
	Chemical analysis, weight percent
SiO_2	52.9
Al_2O_3	2.5
Fe_2O_3	3.8
FeO	7.8
MgO	11.3
CaO	19.9
Na_2O	1.1
K_2O	.29
H_2O^+	.19
H_2O^-	.11
TiO_2	.05
P_2O_5	.02
MnO	.48
CO_2	.02
Sum	100.5
	CIPW norm
Qz	3.70
Ab	---
An	1.00
Or	1.73
Di	75.90
Hy	4.83
Jd	7.20
Mt	5.51
Ap	.03
Il	.09
Cc	.05
Water	.30
Sum	100.34

---

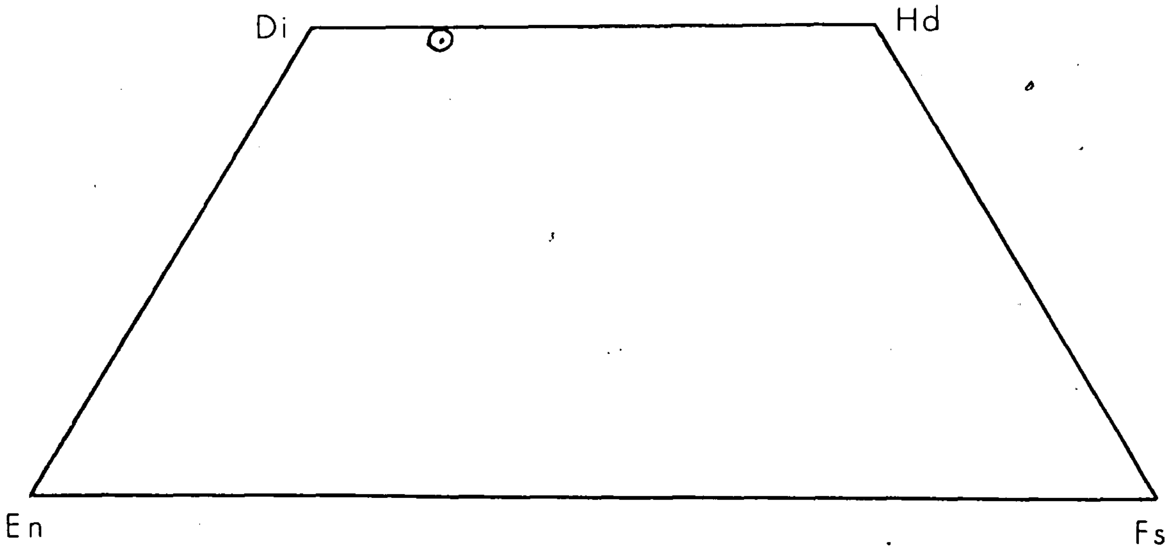
* Rapid rock analysis by K. Coates and H. Smith, U.S. Geological Survey.

Table 5. Minor element analysis of pyroxenite from the Washington mine, ppm*.

Ba	12
V	66
Cr	2.7
Ni	18
Zn	230
Sr	26
Y	21
La	32
Zr	42
Ga	6.1
Co	20

* Emission spectrographic analysis by L. Mei, U.S. Geological Survey. The error limits are plus 50% and minus 33%. No other elements were detected.

Figure 22. Plot of the composition of the clinopyroxene of the pyroxenite from the Washington mine on the pyroxene quadrilateral.

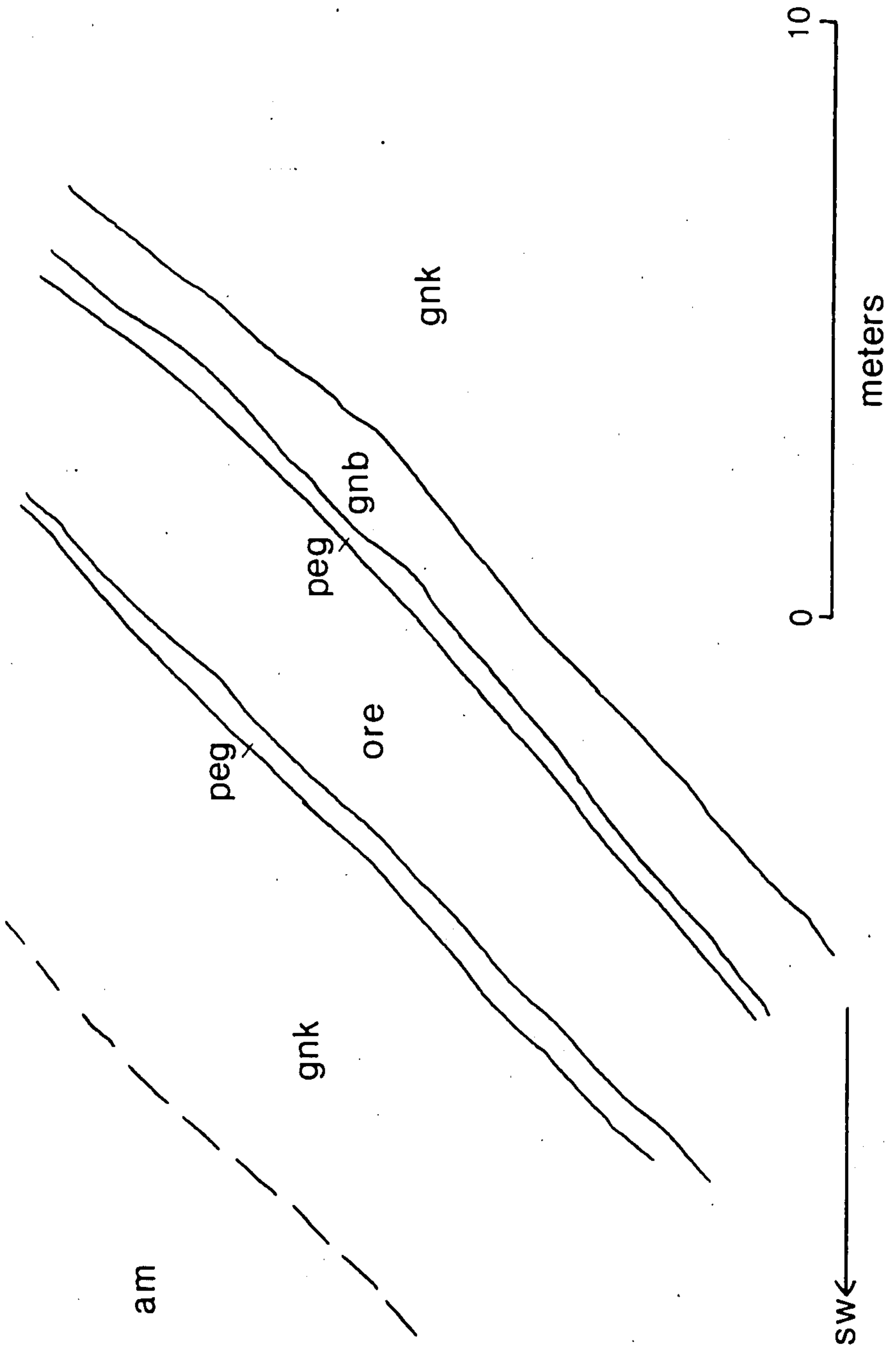

Di
Hd
En
Fs

amphibolite overlies quartz-potassium feldspar gneiss (Fig. 23). Amphibolite in the dump is texturally and mineralogically equivalent to the amphibolite which crops out west of the mine. The amphibolite has been interpreted as part of a northeast plunging overturned synform.

The tabular magnetite orebody was reportedly about 4 m thick; the thickness varying with depth. The ore was composed of magnetite, quartz, microcline, oligoclase, and minor pyrite. The average iron content was about 58 percent and the average sulfur content was about 2.75 percent.

A diamond-drill core through the orebody and the adjoining country rock obtained from the Alan Wood Steel Company was examined in this study. The ore intersection is about 4 m thick and the hanging wall and footwall are alaskite pegmatite. The alaskite pegmatite is composed of coarse-grained microcline and quartz which also occurs as stringers and layers in the magnetite orebody. The pegmatite ranges in thickness from 5 to 100 mm and grades into biotite-plagioclase-quartz gneiss on the footwall contact, which, in turn, grades into quartz-potassium feldspar gneiss. A cross-section of the mine generalized from this drill core is

Figure 23. Generalized cross-section of the Washington mine, reconstructed from a diamond-drill core. gnk - quartz-potassium feldspar gneiss, gnb - biotite-plagioclase-quartz gneiss, peg - alaskite pegmatite, am - amphibolite.


gnk
gnb
peg
peg
ore
gnk
am
10
0
meters
sw

shown in Fig. 23.

A rock sample which includes a contact between pyroxenite and tourmaline-bearing pegmatite was found on the dump and this suggests that pyroxenite may occur as discontinuous lenses contiguous with the pegmatite. No pyroxenite was observed in the drill core.

ORIGIN OF THE MAGNETITE DEPOSITS

Sims (1953, 1958) and Sims and Leonard (1952) found that many magnetite deposits in the Highlands of New Jersey are associated with dark-green pyroxenite and pyroxene-bearing gneiss. They interpreted these rocks to be skarns formed by contact metasomatism which was a consequence of the intrusion of granitic magma into carbonate rocks. Sims (1958) found that the diopsidic skarns did not contain mineable magnetite bodies but that the salite or ferrosalite skarns always contained relatively high concentrations of magnetite.

Skarn rocks, as described by Lindgren (1933), are contact-metasomatic bodies which develop as the result of the intrusion of siliceous magma into carbonate-rich host rocks. Skarns are characterized by andradite, hedenbergite, wollastonite, scapolite, axinite, adularia, albite, calcite, fluorite, specularite, magnetite, chalcocite, molybdenite, and various other heavy metal sulfides (Lindgren, 1933). It should be noted that hedenbergite is the characteristic pyroxene of skarns and not diopside, salite, or ferrosalite.

Both Sims (1958) and Baker and Buddington (1970) concluded that the magnetite deposits of the New Jersey Highlands are of contact metasomatic (pneumatolytic) origin. Although this interpretation is not unreasonable, neither is it based on unequivocal evidence. Sims (1958) argued for a pneumatolytic origin for the Dover magnetite ores because the ore was found in a variety of host rocks. He reasoned that sedimentary iron formations would more likely be confined to a common host rock of fairly constant composition. He concluded that salite- or ferrosalite-bearing skarns contain magnetite deposits because they were more susceptible to metasomatic replacement whereas diopsidic skarns were relatively refractory. The iron-rich fluids were presumed to be solutions which emanated from intrusive alaskitic magma upon crystallization. This, it was argued, would explain the proximity of alaskite bodies to ore deposits.

Skarns are characteristically enclosed in and derived from sedimentary carbonate rocks. Consequently, the pyroxene-rich rocks of the Highlands were considered to be derived through metasomatic reactions of iron- and silica-bearing solutions with marble. However, there is little detailed chemical or mineralogical similarity between these pyroxene-rich rocks and deposits of undisputed pyrometasomatic origin which are unmetamorphosed and associated with the relatively young intrusions in the Rocky Mountains of the western United States. Therefore, alternative views on the origin of the pyroxene-rich rocks in the Precambrian of New Jersey must be considered. In the area of this study, the ubiquitous association of pyroxenite and calcite-bearing pyroxenite with carbonate-bearing rocks both on a macro and micro scale suggests a genetic relationship. Siliceous dolomitic sediment (marl) would, upon regional metamorphism be transformed to diopsidic pyroxene and calcite through simple dedolomitization reactions. The concentration of sodium in the pyroxene suggests that the primary sediment was a lacustrine deposit in an intermontane basin and thus relatively sodium-rich like the Lockatong Formation of the Triassic of New Jersey and Pennsylvania or the Eocene Green River Formation of Wyoming and Colorado. The magnetite deposits may thus be viewed as iron-rich sedimentary deposits.

If the magnetite deposits were derived from iron-rich sediments intercalated with impure carbonate sediments, the next major question is the mineralogical occurrence of the sedimentary iron. The massive magnetite-quartz aggregates and minor amounts of pyrite which constitute the ore suggest the possibility that the primary iron-rich sediment was composed of fine-grained cherty siderite with disseminated pyrite which during regional metamorphism decarbonated to magnetite. It has been shown by French (1971) that decarbonation of siderite occurs at about 450 degrees C, far below the peak metamorphic temperature reached in the Highlands. Siderite and pyrite are stable under similar conditions of pH and Eh (Garrels and Christ, 1965; Berner, 1971), and, under near neutral pH and low Eh, may coexist. The sulfide activity levels must have been less than 10^{-10} M or precipitation of pyrite to the exclusion of siderite would have occurred (Garrels and Christ, 1965). Thus, the magnetite deposit at the Washington mine and other so-called "skarn" deposits of the Highlands might have had such a sedimentary origin.

The occurrence of granitic pegmatite in contact with and interlayered with the magnetite orebody at the Washington mine suggested to Bayley (1941) that the ore was of pneumatolytic-igneous origin. However, it should be noted that iron can reduce markedly the melting point of granitic magma. The minimum melting point (eutectic) in the dry system albite-quartz is about 1060 degrees C, but, with the addition of only about 2 weight percent FeO to a mixture of 34 mole percent quartz and 66 mole percent albite, the ternary minimum drops to about 980 degrees C (Bowen and Schairer, 1938). In the system orthoclase-quartz-fayalite, the addition of only about 3 weight percent fayalite to a mixture of 40 weight percent quartz and 60 weight percent potassium feldspar results in a lowering of the melting point (dry) from 990 to the ternary minimum which is a much lower but unspecified temperature (Roedder, 1951, 1978). Carmichael and MacKenzie (1963) have shown that similar lowering of the minimum melting point of granitic composition results from the addition of Fe_2O_3. Thus the effect of iron in lowering the minimum melting temperature of rocks might be operative in the development of granitic magma wherever the magnetite ore was in contact with the quartz-potassium feldspar gneiss whose bulk composition was near the ternary minimum. At the Washington mine, therefore, the granitic pegmatite may not have been the ore bringer but may have been the product of interface melting wherever the pre-existing magnetite deposit was in contact with quartz-potassium feldspar gneiss. Such a magma would, of course, be mobile and show intrusive relations against the magnetite body. Pegmatite is absent at the Pequest magnetite deposit which suggests that the surrounding metasedimentary quartz-epidote gneiss did not have the necessary bulk composition, where it was in contact with the iron-rich layer, to result in melting at the prevailing temperature even in the presence of iron. The occurrence of pegmatite at the Ahles mine appears to be due to local melting at the contact between the magnetite deposit and quartz-potassium feldspar gneiss.

In summary, the magnetite deposit at the Washington mine is interpreted as an iron-rich sediment probably composed originally of siderite, chert, and minor pyrite which was subsequently transformed during regional metamorphism into pyrite-bearing magnetite-quartz rock. The associated pyroxenite, which is now composed of Na-bearing diopsidic clinopyroxene, is interpreted as a metamorphosed siliceous dolomitic marl. The premetamorphic sediments may thus have been siliceous siderite mud intercalated with siliceous dolomite which upon regional metamorphism yielded beds of magnetite-quartz rock enclosed in silicated calcitic marble through decarbonation and dedolomitization reactions, respectively. The magnetite ores in the marbles are relatively high in manganese as compared with the magnetite ores in quartz-potassium feldspar gneiss and quartz-epidote gneiss and this lithologic control may also reflect differing conditions of chemical sedimentation during deposition of the sedimentary precursors of these rocks. Locally, melting occurred during regional metamorphism where metasedimentary quartz-potassium feldspar gneiss of appropriate bulk composition was in contact with the magnetite layer.

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Appendix A

Modal Analyses

Table 1. Modes of Amphibolites

	1	2	3	4	5
Quartz	1.5	---	---	3.3	8.6
Plagioclase	58.2	46.5	42.2	30.2	36.3
Hornblende	.9	36.6	41.2	59.9	45.1
Biotite	30.2	---	tr.	5.2	5.9
Magnetite	.7	.4	5.8	1.2	2.1
Apatite	tr.	.3	tr.	.2	tr.
Sphene	5.3	1.0	3.3	tr.	tr.
Clinopyroxene	2.9	15.2	7.5	---	---
Epidote	tr.	---	---	---	tr.
Chlorite	.3	---	---	---	2.0
Sum	100.0	100.0	100.0	100.0	100.0

1. Biotite amphibolite from railroad cut near Pequest, Washington quadrangle.

2-4. Scotts Mountain, Belvidere quadrangle.

5. Oxford Stone Quarry, Belvidere quadrangle.

Table 2. Modes of Quartz Potassium Feldspar Gneiss from Scotts Mountain, Belvidere quadrangle.

Quartz	43.8	39.3	35.2	40.0	36.2
Microcline	43.0	31.3	47.4	47.6	47.6
Plagioclase	10.6	25.3	5.0	10.9	13.7
Biotite	---	.3	8.0	---	---
Clinopyroxene	.2	---	---	---	---
Hornblende	---	2.5	---	---	---
Magnetite	2.3	.7	4.4	1.5	.6
Chlorite	tr.	.6	tr.	---	1.1
Apatite	tr.	tr.	tr.	tr.	.1
Sphene	tr.	tr.	---	tr.	.7
Zircon	.1	tr.	---	tr.	tr.
Sum	100.0	100.0	100.0	100.0	100.0

Table 3. Modes of Oligoclase-Quartz Gneiss from Scotts Mountain, Belvidere quadrangle.

	1	2	3	4	5	6	7
Quartz	31.6	30.1	38.1	39.0	15.0	8.3	15.4
Plagioclase	57.8	62.0	57.7	51.8	62.0	57.8	42.7
Microcline	3.0	tr.	.4	---	---	---	tr.
Chlorite	4.9	.8	.7	3.2	2.3	8.5	17.7
Clinopyroxene	1.2	---	---	---	---	---	---
Magnetite	.8	.1	.1	5.8	.2	6.8	3.1
Sphene	.7	tr.	---	---	---	---	---
Apatite	tr.	tr.	tr.	.2	.5	.5	.3
Hornblende	tr.	6.8	3.0	---	20.0	11.9	18.4
Zircon	tr.	.2	tr.	tr.	---	tr.	tr.
Garnet	---	---	---	---	---	6.2	2.4
Sum	100.0	100.0	100.0	100.0	100.0	100.0	100.0

Table 4. Modes of Microperthite Alaskite.

	1	2	3	4	5
Quartz	33.1	42.3	32.2	18.6	32.0
Microperthite	38.0	21.8	43.9	65.3	59.4
Microcline	.5	17.1	9.7	4.2	.9
Plagioclase	22.4	16.6	11.2	5.5	2.0
Hornblende	.6	.1	1.4	---	---
Chlorite	.5	1.8	1.6	5.2	3.6
Biotite	4.7	---	---	---	---
Magnetite	.1	.1	tr.	1.2	.5
Apatite	.1	.2	tr.	tr.	.5
Zircon	---	tr.	tr.	tr.	tr.
Sphene	---	---	---	---	.9
Calcite	---	tr.	---	---	---
Myrmekite	tr.	---	---	tr.	---
Clinopyroxene	---	---	---	---	.2
Sum	100.0	100.0	100.0	100.0	100.0

1. Scotts Mountain, Belvidere quadrangle.
2-4. Mount Mohepinoke, Washington quadrangle.
5. Pohatcong Mountain, Washington quadrangle.

Table 5. Modes of Hornblende Syenite from Oxford Mountain, Washington quadrangle.

	1	2	3
Quartz	5.6	8.8	3.3
Microperthite	64.9	76.9	73.9
Hornblende	13.7	1.9	7.8
Chlorite	8.2	7.1	8.9
Plagioclase	3.7	3.5	3.4
Microcline	1.4	.7	.3
Biotite	.9	.6	.1
Magnetite	1.6	.5	2.3
Apatite	tr.	tr.	tr.
Sphene	tr.	tr.	tr.
Zircon	tr.	tr.	tr.
Sum	100.0	100.0	100.0

Table 6. Modes of Hornblende Granite.

	1	2
Quartz	27.2	30.2
Mesoperthite	tr.	31.7
Plagioclase	32.2	19.9
Microcline	22.6	12.1
Biotite	6.4	---
Chlorite	1.7	2.4
Hornblende	9.4	3.0
Apatite	.3	.1
Magnetite	.2	.6
Myrmekite	tr.	---
Clinopyroxene	tr.	---
Sphene	tr.	tr.
Zircon	tr.	tr.
Sum	100.0	100.0

1. Mount No More, Belvidere quadrangle.
2. Mount Mohepinoke, Washington quadrangle.

Table 7. Modes of Microantiperthite Alaskite from Scotts Mountain, Belvidere quadrangle.

	1	2
Quartz	40.1	29.6
Antiperthite	24.4	21.5
Plagioclase	27.5	26.4
Microcline	3.7	9.5
Clinopyroxene	1.9	11.5
Magnetite	1.5	tr.
Chlorite	.5	.8
Biotite	.4	---
Apatite	tr.	.1
Sphene	tr.	.6
Sum	100.0	100.0

Appendix B

Sample Locations

for

Chemically analyzed Samples

Table 1. Location of Amphibolite Samples.

1. Black amphibolite from Oxford Stone Quarry, 1.4 km S20E from Bridgeville, N.J., Belvidere quadrangle. New analysis for this study by K. Coates and H. Smith, U.S. Geological Survey.

2. Grayish-green fine-grained amphibolite from Lommasons Glen, N.J., Belvidere quadrangle. New analysis for this study by K. Coates and H. Smith, U.S. Geological Survey.

3. Fine-grained gray amphibolite .7 km S 40E from Lommasons Glen, N.J., Belvidere quadrangle. Sample taken 4 meters from pegmatite contact. New analysis for this study by Z.A. Hamlin, U.S. Geological Survey.

4. Fine-grained gray amphibolite from same location as 3, but taken 1 meter from pegmatite contact. New analysis for this study by Z.A. Hamlin, U.S. Geological Survey.

5. Fine-grained amphibolite with granitic layers, from same location as 3, but sample was taken at pegmatite contact. New analysis for this study by Z.A. Hamlin, U.S. Geological Survey.

6. Amphibolite from northwest N.J., Raritan Folio, (Bayley et al., 1914).

7. Amphibolite from the Reading quadrangle, Pennsylvania, (Buckwalter, 1962).

8. Pochuck Gabbro Gneiss, Greenwood Lake quadrangle, N.J., (Darton et al., 1908).

Table 2. Chemical Analyses and CIPW norms of Plutonic rocks from the Study Area.

	1	2	3	4
SiO_2	72.8	74.0	74.2	71.3
Al_2O_3	15.0	12.7	13.6	13.3
Fe_2O_3	1.5	1.2	.32	1.5
FeO	.04	.20	.04	1.6
MgO	.33	.63	.06	.56
CaO	1.7	1.7	.19	1.0
Na_2O	4.9	2.6	2.8	.36
K_2O	2.5	4.7	7.2	5.5
H_2O^+	.96	.70	.40	.58
H_2O^-	.19	.10	.11	.16
TiO_2	.09	.12	.01	.31
P_2O_5	.05	.04	.02	.08
MnO	.03	.04	.02	.05
CO_2	.02	.02	.02	.01
Sum	100.1	98.8	99.0	99.6

CIPW norms

Q	30.60	36.46	29.92	26.26
An	8.43	8.04	.81	3.87
Ab	41.48	22.03	23.71	30.47
Or	14.75	27.78	42.53	32.51
Mt	---	.30	.12	2.18
Il	.09	.23	.02	.59
Ap	.12	.09	.03	.19
Cc	.01	.05	.02	.02
Cd	1.14	.39	.91	---
Hm	1.50	.99	.24	---
Wo	---	---	---	.22
En	.82	1.57	.15	1.40
Fs	---	---	---	1.19
H_2O	1.15	.80	.51	.74
Sum	100.09	98.73	98.97	99.64

Table 2 (continued).

	5	6
SiO_2	61.8	73.2
Al_2O_3	15.7	12.2
Fe_2O_3	1.4	1.7
FeO	3.5	1.3
MgO	.38	.18
CaO	4.9	1.9
Na_2O	5.1	4.1
K_2O	5.3	4.1
H_2O^+	.26	.28
H_2O^-	.20	.11
TiO_2	.68	.32
P_2O_5	.12	.03
MnO	.16	.03
CO_2	.08	.07
Sum	99.5	99.5

CIPW norms

	5	6
Q	3.35	30.64
An	4.28	2.78
Ab	43.16	34.72
Or	31.34	24.22
Mt	2.04	2.46
Il	1.29	.61
Ap	.25	.06
Cc	.21	.17
Cd	---	---
Hm	---	---
Wo	8.08	2.57
En	.94	.45
Fs	4.14	.46
H_2O	.46	.39
Sum	99.54	99.53

Table 2 (continued)

1. Alaskite pegmatite from Lommasons Glen, N.J., Belvidere quadrangle. New analysis for this study by Z.A. Hamlin, U.S. Geological Survey.
2. Alaskite pegmatite from same location as 1. New analysis for this study by Z.A. Hamlin, U.S. Geological Survey.
3. Alaskite pegmatite from Oxford Church, near Hazen, N.J., Belvidere quadrangle. New analysis for this study by Z.A. Hamlin, U.S. Geological Survey.
4. Microperthite alaskite from U.S. 46, .8 km southwest of Townsbury, N.J., Washington quadrangle. New analysis for this study by Z.A. Hamlin, U.S. Geological Survey.
5. Quartz syenite from south end of Van Nest Tunnel, 1.5 km southeast of Oxford, N.J., Washington quadrqngle. New analysis for this study by K. Coates and H. Smith, U.S. Geological Survey.
6. Microperthite alaskite from Stewarts Gap, Washington quadrangle. New analysis for this study by K. Coates and H. Smith, U.S. Geological Survey.

Table 3. Minor element analyses of plutonic rocks from the study area, ppm*.

	1	2	3	4	5	6
Ba	330	1300	120	1500	1000	650
V	9.5	32	1.9	17	2.1	2.5
Cr	2.8	8.5	1.8	3.5	3	1.5
Ni	31	6.5	1.5	2.4	2.6	n.d.
Cu	20	5.7	n.d.	23	30	24
Zn	n.d.	19	n.d.	14	74	16
Sr	450	580	25	120	330	110
Y	22	29	4.0	86	82	76
Ce	120	70	n.d.	250	200	300
La	67	52	n.d.	140	110	170
Zr	17	30	9.6	190	1200	660
Nb	7	2.9	n.d.	28	16	18
Pb	n.d.	8.5	33	n.d.	n.d.	n.d.
Ga	15	18	15	28	29	22
Nd	n.d.	n.d.	n.d.	150	86	190
Co	8.3	1.3	n.d.	4.1	3.3	2.3
La/Ce	.56	.74	---	.56	.55	.57

* Emission spectrographic analyses by L. Mei, U.S. Geological Survey. The error limits on each number are plus 50% and minus 33%.

n.d.- not detected.

sample locations given in table 3.

Appendix C

Niggli Values

Procedure for calculating Niggli values (from Johansen, 1931, p. 103).

Weight percentages of oxides are converted into molecular numbers. The molecular number of Fe_2O_3 is multiplied by 2 and added to that of FeO. The sum of Al_2O_3, (Fe, Mn)O + MgO, CaO, and K_2O + Na_2O is recalculated to 100, and is represented by al (for Al_2O_3), fm (for (Fe, Mn, Mg)O), c (for CaO) and alk ((for Na, $K)_2$ O). Values for k are calculated by the ratio of K_2O to the sum of the alkalis in alk. Values for mg are calculated by the ratio of MgO to the sum of FeO, MnO, and MgO in fm. Values for si, ti and p are obtained by reducing their molecular numbers in the same proportions as above.

Niggli values for Reading Prong amphibolites

	1	2	3	4	5
si	137.09	115.04	116.86	170.39	283.90
al	21.86	23.38	16.51	27.04	28.41
fm	45.76	43.87	44.23	38.37	34.86
c	21.26	22.41	31.73	15.43	20.04
alk	11.12	10.33	7.53	19.16	16.69
ti	2.91	1.79	.79	1.94	1.09
p	.29	.28	.24	.23	.16
k	.24	.26	.50	.50	.44
mg	.47	.53	.58	.56	.52
oxrt	21.85	62.15	33.90	41.19	42.78
al-alk	10.74	13.05	8.98	7.88	11.72

	6	7	8
si	93.95	100.59	92.54
al	15.11	19.71	15.32
fm	49.88	50.02	49.77
c	27.44	22.08	27.25
alk	7.56	8.19	7.66
ti	3.61	2.70	3.64
p	.25	.00	.16
k	.20	.34	.20
mg	.35	.52	.30
oxrt	30.46	21.77	28.62
al-alk	7.55	11.52	7.66

Niggli values for sodium-rich rocks from the Reading Prong

	1	2	3	4	5
si	416.83	478.14	316.70	288.19	292.98
al	41.47	49.41	49.29	40.30	40.77
fm	8.76	1.89	5.87	21.06	18.51
c	15.44	4.82	3.61	6.95	13.36
alk	34.33	43.87	41.23	31.69	27.36
ti	.83	.74	2.46	2.27	1.56
p	.16	.08	.14	.04	.29
k	.12	.11	.13	.15	.29
mg	.41	.00	.24	.60	.40
oxrt	44.75	53.79	65.66	12.50	17.64
al-alk	7.14	5.54	8.06	8.61	13.41

	6	7
si	309.78	349.23
al	42.38	39.21
fm	14.20	13.96
c	18.79	17.57
alk	24.63	29.26
ti	1.01	1.38
p	.36	.27
k	.19	.11
mg	.27	.50
oxrt	27.72	25.23
al-alk	17.75	9.95

---

1. Albite-oligoclase granite from Drake (1969).
2. Albite granite from Bayley (1941).
3. Albite-oligoclase granite from Sims (1958).
4. Biotite-quartz-oligoclase gneiss from Sims (1958).
5. Hypersthene-quartz-oligoclase gneiss from Baker and Buddington (1970).
6. Oligoclase-quartz gneiss from Drake (1969).
7. Oligoclase-quartz gneiss from Drake (1969).

VITA

Robert L. Kastelic Jr. was born on September 7, 1955 in Zanesville, Ohio. His parents are Robert and Carolyn Kastelic. In September, 1973 he entered the University of Dayton in Dayton, Ohio, where he was awarded the Bachelor of Science degree in Geology in July of 1977. Since September of 1977, he has undertaken graduate studies in the Department of Geological Sciences at Lehigh University where he has been a graduate research assistant and a graduate teaching assistant.

During 1977 and 1978, he was a graduate research assistant on a project sponsored by the National Aeronautics and Space Administration through the Lunar Sample Analysis Program.

As of August 20, 1979, Mr. Kastelic will assume the position of Exploration Geologist for Gold Fields Mining Corporation.

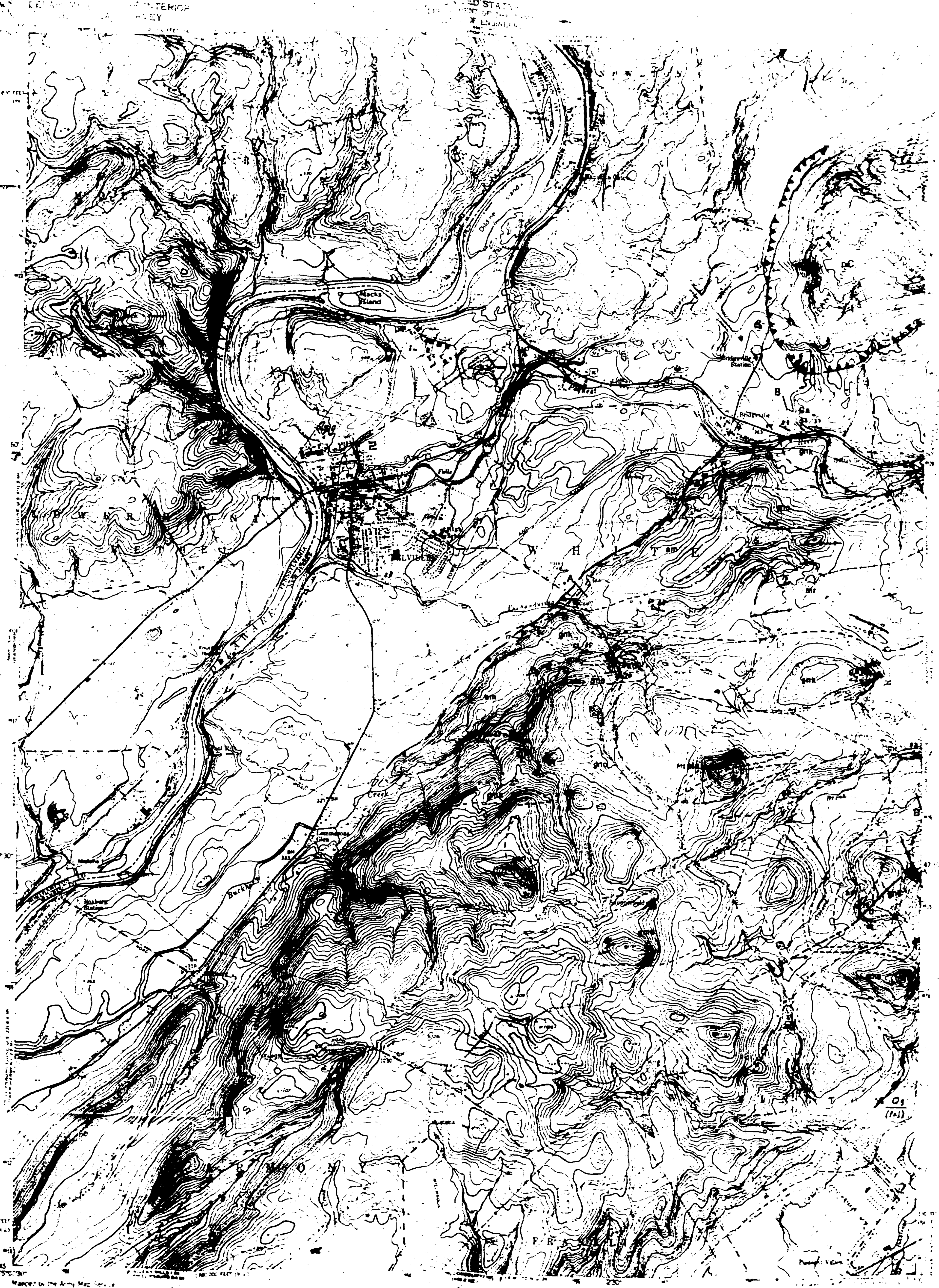
Mr. Kastelic is an associate member of Sigma Xi. His publications are as follows:

Sclar, C.B., Bauer, J.F., and Kastelic, R.L., 1978, Subsolidus shock-induced reduction of Fe^{+2} on the Lunar Surface: An experimental assessment; Lunar and Planetary Sci. IX, p. 1307-1309.

Sclar, C.B., Bauer, J.F., and Kastelic, R.L., 1979, Subsolidus shock-induced reduction of Fe^{+2} on the Moon: Evidence from experimentally shocked Ulvospinel; Lunar and Planetary Sci. X, p.1090-1092.

Sclar, C.B., and Kastelic, R.L., 1979, Iron in Anorthite: Synthesis and characterization of the $CaAl_2Si_2O_8$ - $CaFeSi_3O_8$ series; EOS, Trans. Amer. Geophys. Union, 60, 421.

Sclar, C.B., and Kastelic, R.L., 1979, Iron in Anorthite: An experimental study; Lunar and Planetary Sci. X, p. 1087-1089.



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SCALE 1:24000

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A FOLDER DESCRIBING TOPOGRAPHIC MAPS AND SYMBOLS IS AVAILABLE ON REQUEST

GEOLOGIC MAP OF PARTS OF THE BELVIDERE QUADRANGLE
NEW JERSEY – PENNSYLVANIA

By

Robert Louis Kas

1979

PLATE 1

[redacted]

[redacted]

PLATE 2

GEOLOGIC MAP OF PARTS OF THE WASHINGTON QUADRANGLE
NEW JERSEY

By
Robert Louis Kastelic, Jr.
1979

[redacted]

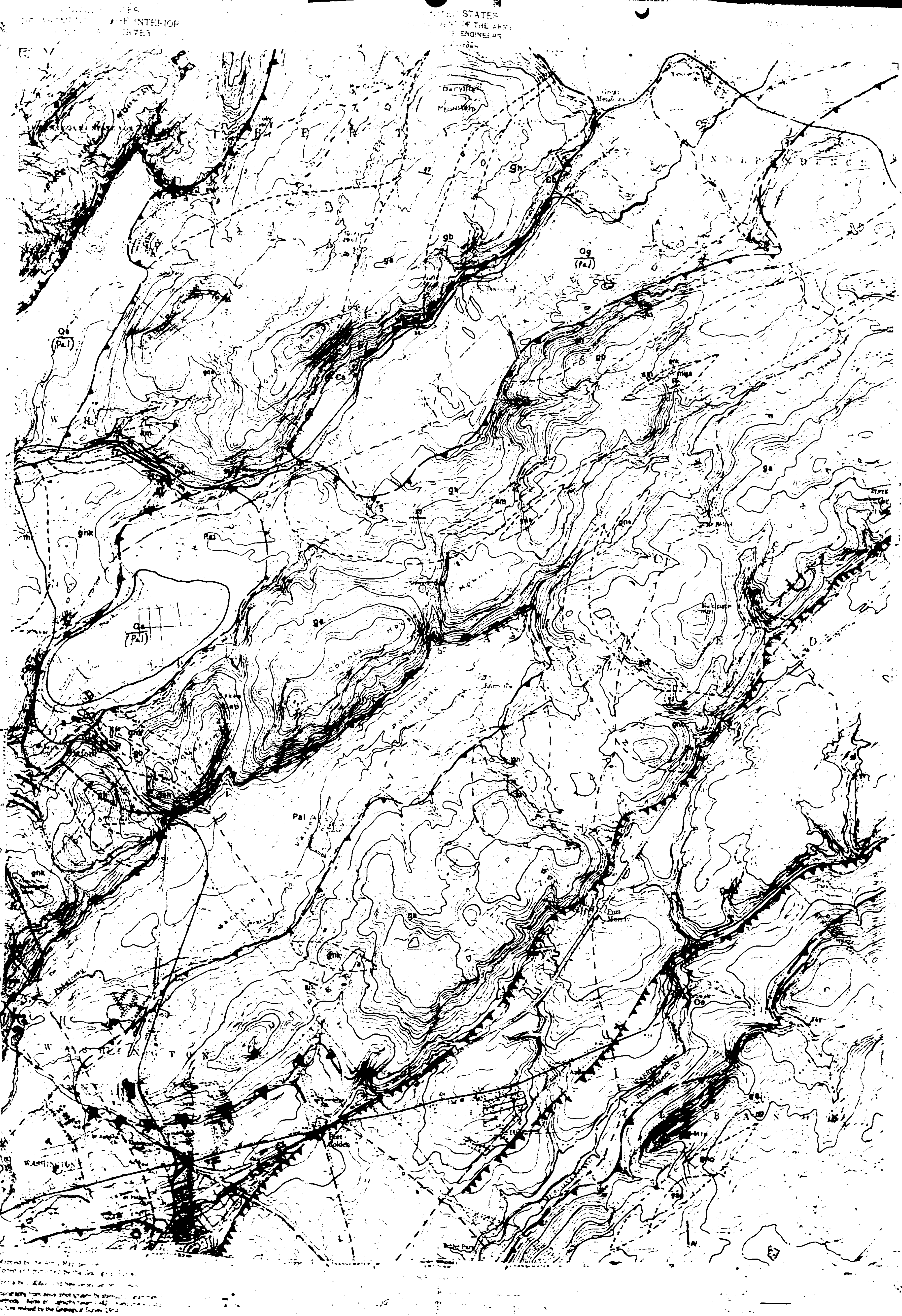
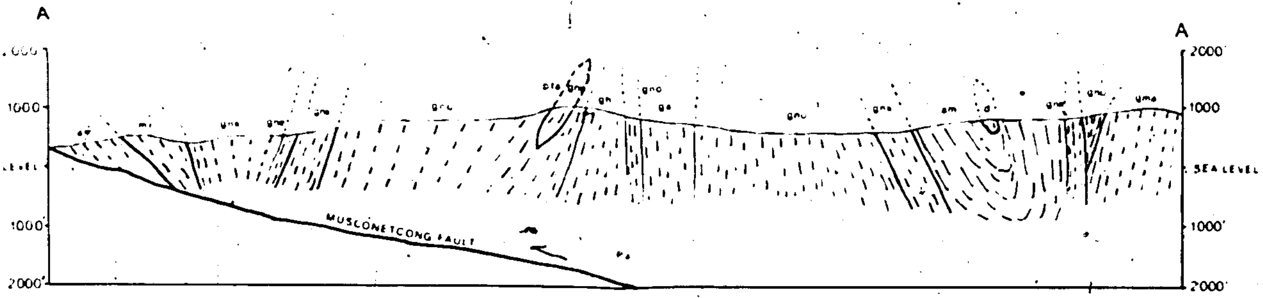
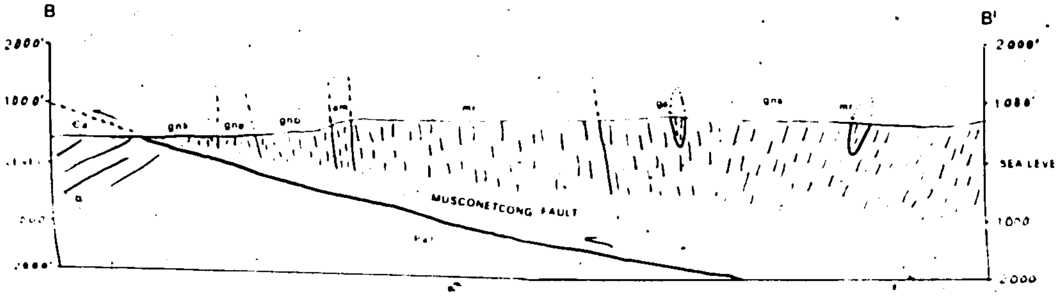


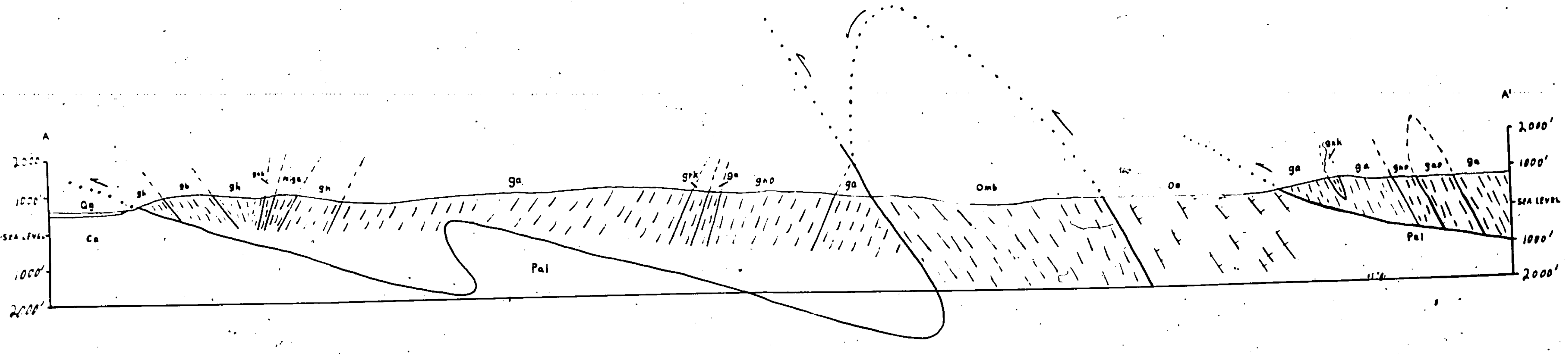
PLATE 3

Cross-Sections Through Parts of the Belvidere Quadrangle, New Jersey





CROSS-SECTIONS THROUGH PARTS OF THE WASHINGTON QUADRANGLE, NEW JERSEY



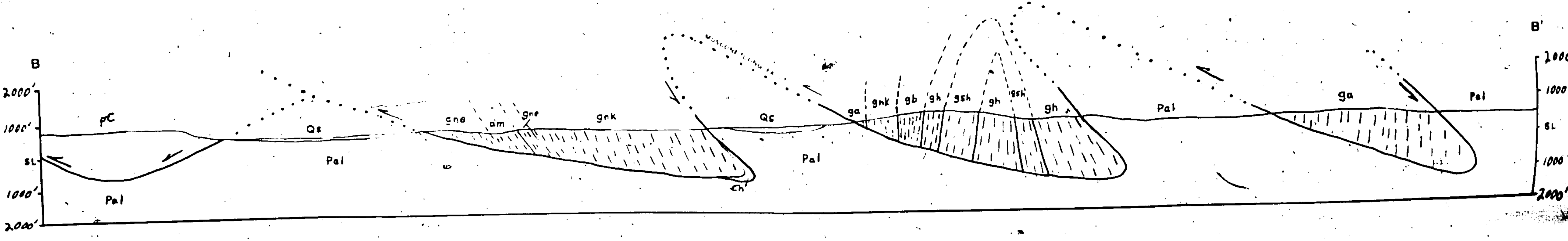


PLATE 5

GEOLOGIC MAP AND SECTIONS OF PARTS OF THE BELVIDERE AND WASHINGTON QUADRANGLES, NEW JERSEY

By

Robert Louis Kastelic, Jr.

1979

Explanation

[illegible]

[illegible]

Swamp de[illegible] overlyi[illegible]
Lower Paleozoic sedimentary rocks.
[illegible]

[illegible]

G[illegible]acial dep[illegible]
[illegible]

QUATERNARY

Pleistocene and Holocene

Swamp Deposits

clay, silt, fine gravel, and partly decomposed plant material

Pleistocene

og

Undifferentiated glacial deposits

sand, gravel, till, and ground moraine of Wisconsin age

ORDOVICIAN

Middle Ordovician

Omb

Martinsburg Formation

Omb, Bushkill Member. Thin bedded, dark-gray, medium-gray to yellowish-brown weathering claystone slate

Lower Ordovician

Oe

Epler Formation

~~Light~~ to medium-~~gray~~ limestone and light to dark-~~medium-gray~~ dolomite, containing lenses of orthoquartzite

CAMBRIAN

Upper Cambrian

Allentown Dolomite

[illegible] Very fine-to medium-grained, light gray [illegible] dolomite, containing argillaceous layers. Oolites are common.

Lower Cambrian

Ch

Hardyston Quartzite

Gray, brown-weathering quartzite, often arkosic.

Pal

Undifferentiated Lower Paleozoic rocks

unconformity

IGNEOUS ROCKS

ga / gma

Alaskite

ga, microperthite alaskite. Medium to coarse grained, light gray gneissoid granite and granite gneiss, composed mostly of microperthite, quartz, and oligoclase. Contains local unmappable bodies of amphibolite and potassium feldspar gneiss.

gma, microantiperthite alaskite. Differs from ga in that the feldspar is dominantly antiperthitic. Lenticular quartz is commonly present. The rock weathers chalk white.

[illegible]

Hornblende granite and associated biotite granite

gh, hornblende granite. Medium to coarse grained, pink to light gray gneissoid granite and granite gneiss, composed mainly of microperthite, quartz, oligoclase, and hornblende. Locally contains unmappable bodies of microperthite alaskite, amphibolite and potassium feldspar gneiss.

gb, biotite granite. Differs from gh in that biotite is the principal mafic constituent. Unit is variable in composition. The rock has a light greenish color, and weathers to a light brown color.

PRECAMBRIAN

gsh

Hornblende syenite

Dark gray to greenish gray, medium to coarse grained rock, consisting mainly of microperthite hornblende, with minor quartz.

METASEDIMENTARY AND (OR) METAVOLCANIC ROCKS

Age relations of the following rocks are unknown

am

Amphibolite

Fine-to medium-grained dark greenish-gray to black amphibolite with variable amounts of hornblende, biotite, and pyroxene.

gnk

Quartz-potassium feldspar gneiss

Fine-to medium-grained pink to pinkish-gray to gray gneiss with poor to fair foliation. The unit is composed largely of quartz and microcline with minor amounts of biotite, magnetite, and garnet. Local layers and lenses of feldspathic quartzite and oligoclase-quartz-gneiss are found.

gne

Quartz-epidote gneiss

Fine-to medium-grained gneiss with alternating green and gray layers, containing quartz, epidote, green pyroxene with local concentrations of potassium feldspar and magnetite. Lenses of green pyroxenite are found locally.

mr

Marble

Medium-to coarse-grained gray to bluish-gray rock containing isolated bodies of amphibolite, and lenses and layers of magnetite. Calcite is the dominant mineral, and accessory finely disseminated graphite is found. Layers of oligoclase-quartz gneiss are found locally.

miga

Migmatite

Mixed rock with lenses and veins of microperthite alaskite in amphibolite.

gno / pta / gao

Oligoclase-quartz gneiss and related albite oligoclase granite

gno, oligoclase-quartz gneiss. Fine-to medium-coarse-grained, light greenish-gray to grayish-green granoblastic gneiss composed mainly of oligoclase and quartz with minor biotite, magnetite, pyroxene, and, locally garnet. Unmappable layers of amphibolite, albite-oligoclase granite, and albite pegmatite are locally found.

gao, albite-oligoclase granite. Medium-to-coarse grained, light greenish gneissoid granite composed dominantly of albite-oligoclase and quartz.

pta, albite pegmatite. Coarse grained, light-greenish-gray rock composed largely of albite and quartz. Locally, large crystals of pyroxene, hornblende, and magnetite are found.

d

Pseudodiorite

Medium-to-coarse-grained dark-gray-to-dark-greenish-gray rock composed of hornblende, plagioclase, and augitic pyroxene. It is probably coarsened amphibolite.

pC

Undifferentiated Precambrian rocks

Amphibolites, marbles, gneisses, and granites of Grenville age.

Contact

[illegible] located

Thrust fault, showing dip

Dashed where approximately located. T, on upper plate.

Inclined

Overturned

[illegible]netcong fault

Contact of Precambrian core with limbs of the Musconetcong nappe. Sawteeth on core side of contact. Where overturned, sawteeth in direction of dip; bar on core side of contact. Dashed where approximately located.

[illegible]

[illegible]

Overturned antiform

Showing trace of axial surface and
[redacted]

Overturned synform

Showing trace of axial surface and direction of dip of limbs and plunge. Dashed where approximately located.

PLANAR FEATURES

Inclined

Strike and dip of beds

Inclined

Vertical

Strike and dip of compositional layering and crystallization folia

in Precambrian rocks.

Inclined

Strike and dip of cataclastic folia

in Precambrian rocks.

Inclined

Strike and dip of axial plane clea

Inclined

Strike and dip of essentially parallel bedding and cleavage.

Magnetite deposit

Showing direction of dip.

LINEAR FEATURES

(May be combined with any of the ab

features)

Bearing and plunge of mineral lineatio

in Precambrian rocks

Bearing and plunge of intersection of bedding and cleavage.