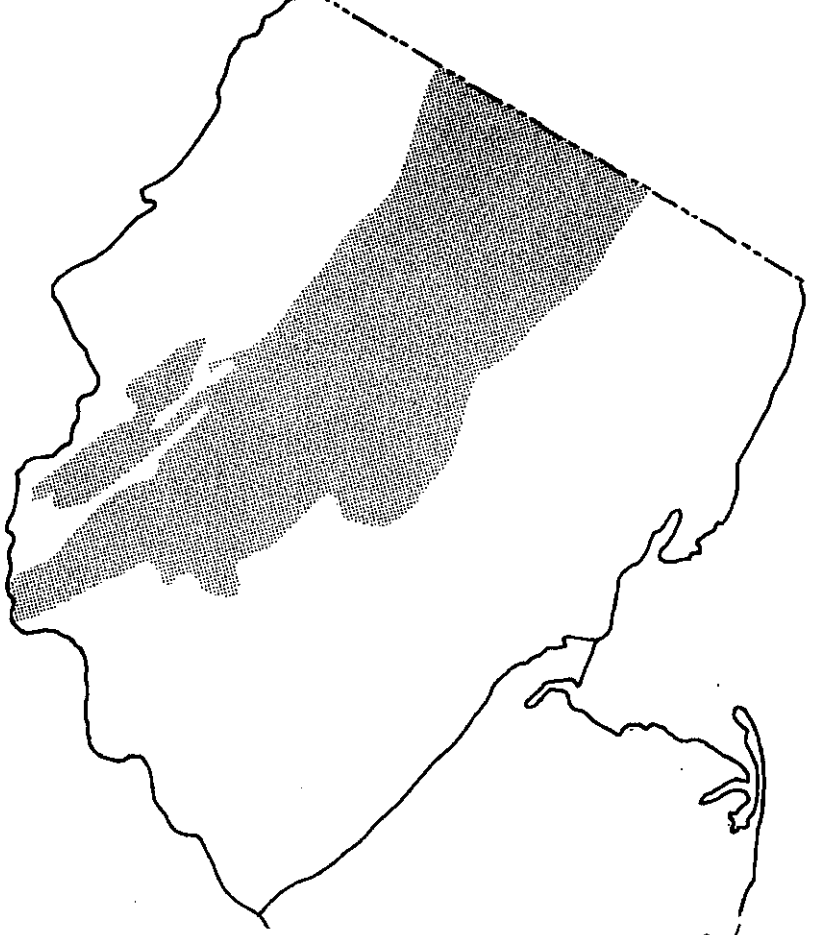

The Economic Geology of Northern New Jersey

Field Guide and Proceedings

Edited by

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Department of Applied Sciences
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Staten Island, NY 10314

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Department of Geology
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Fourteenth Annual Meeting of the Geological Association of New Jersey

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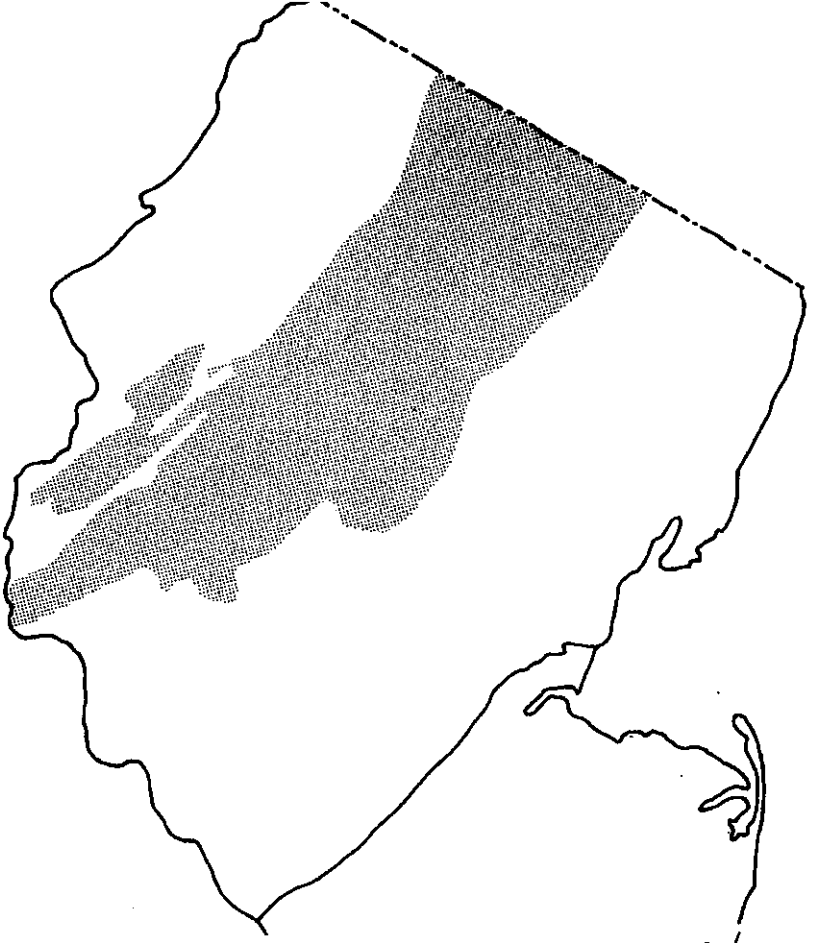

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FOREWARD

The 1997 and 1978 Geological Association of New Jersey (GANJ) meetings will focus on the Economic Geology of New Jersey. The theme of this year's GANJ is the "Economic Geology of Northern New Jersey" and the theme of next year's GANJ meeting will be "Economic Geology of Central and Southern New Jersey". We are very fortunate to have Robert Metsger, the chief geologist (retired) of the New Jersey Zinc Co. speaking on and leading the field trip to the Sterling Hill Zinc mine. We also thank the other speakers and field trip leaders, Richard A. Volkert, Craig A. Johnson, Donald H . Monteverde, Warren Cummings and J. Mark Zdepski. In addition, we thank Charles B. Sclar. Without their participation this meeting would be impossible. We are grateful to Medusa Minerals Inc. for granting permission for the November 1, 1997 site visit to the Lime Crest quarry. Finally, we hope that you will find this meeting both enjoyable and educational.

Alan I. Benimoff
John H. Puffer

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OCCURRENCE AND ORIGIN OF GRAPHITE DEPOSITS IN MIDDLE PROTEROZOIC ROCKS OF THE NEW JERSEY HIGHLANDS

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INTRODUCTION

Middle Proterozoic rocks of the New Jersey Highlands (fig. 1) consist of a heterogeneous assemblage of metavolcanic, metasedimentary, and granitoid rocks. The oldest of these are inferred to be plagioclase-rich gneiss and granite and metabasalt of the Losee Metamorphic Suite (Drake, 1984; Puffer and Volkert, 1991). They are associated spatially with quartz-rich and quartz-poor charnockitic (hypersthene-bearing) rocks with which they appear to be cogenetic (Volkert, 1993; Volkert and Drake, in press). Collectively, these rocks underlie approximately 20 percent of the Highlands.

The aforementioned rocks are unconformably overlain by a sequence of metasedimentary rocks that are comprised of quartzofeldspathic gneisses, calc-silicate rocks, metaquartzite, and marble. Quartzofeldspathic gneisses include potassic feldspar gneiss, microcline gneiss, biotite-quartz-feldspar gneiss, and hornblende-quartz-feldspar gneiss. Collectively, they underlie approximately 10 to 15 percent of the Highlands. Calc-silicate rocks include clinopyroxene-quartz-feldspar gneiss, pyroxene gneiss, clinopyroxene-epidote gneiss, quartz-scapolite gneiss, epidote gneiss, diopsidite, and hornblende-pyroxene skarn. Collectively they underlie approximately 7 to 10 percent of the Highlands. Metaquartzite is sparsely exposed, underlying less than 1 percent of the Highlands, but is a lithologically significant unit that provides an excellent stratigraphic and structural marker. Marble, although widespread, underlies only about 5 percent of the Highlands mainly west of the Green Pond Mountain region.

Amphibolite formed from different protoliths is ubiquitous and is associated with most Middle Proterozoic rocks.

Widespread granitoid rocks include the hornblende and biotite-bearing rocks of the Byram Intrusive Suite, the clinopyroxene-bearing rocks of the Lake Hopatcong Intrusive Suite, and the Mount Eve Granite. Collectively, granitoid rocks underlie approximately 55 percent of the Highlands, although the Mount Eve is confined to the extreme northern Highlands in Sussex County.

GRAPHITE OCCURRENCE

The occurrence of graphite in New Jersey is restricted principally to Middle Proterozoic rocks of the Highlands. It is absent elsewhere in the state except as rare detritus in younger rocks and sediments.

In Benimoff, A.I., and Puffer, J.H., (editors), The economic geology of northern New Jersey: Field Guide and Proceedings of the fourteenth annual meeting of the Geological Association of New Jersey, 1997, p 1-19.

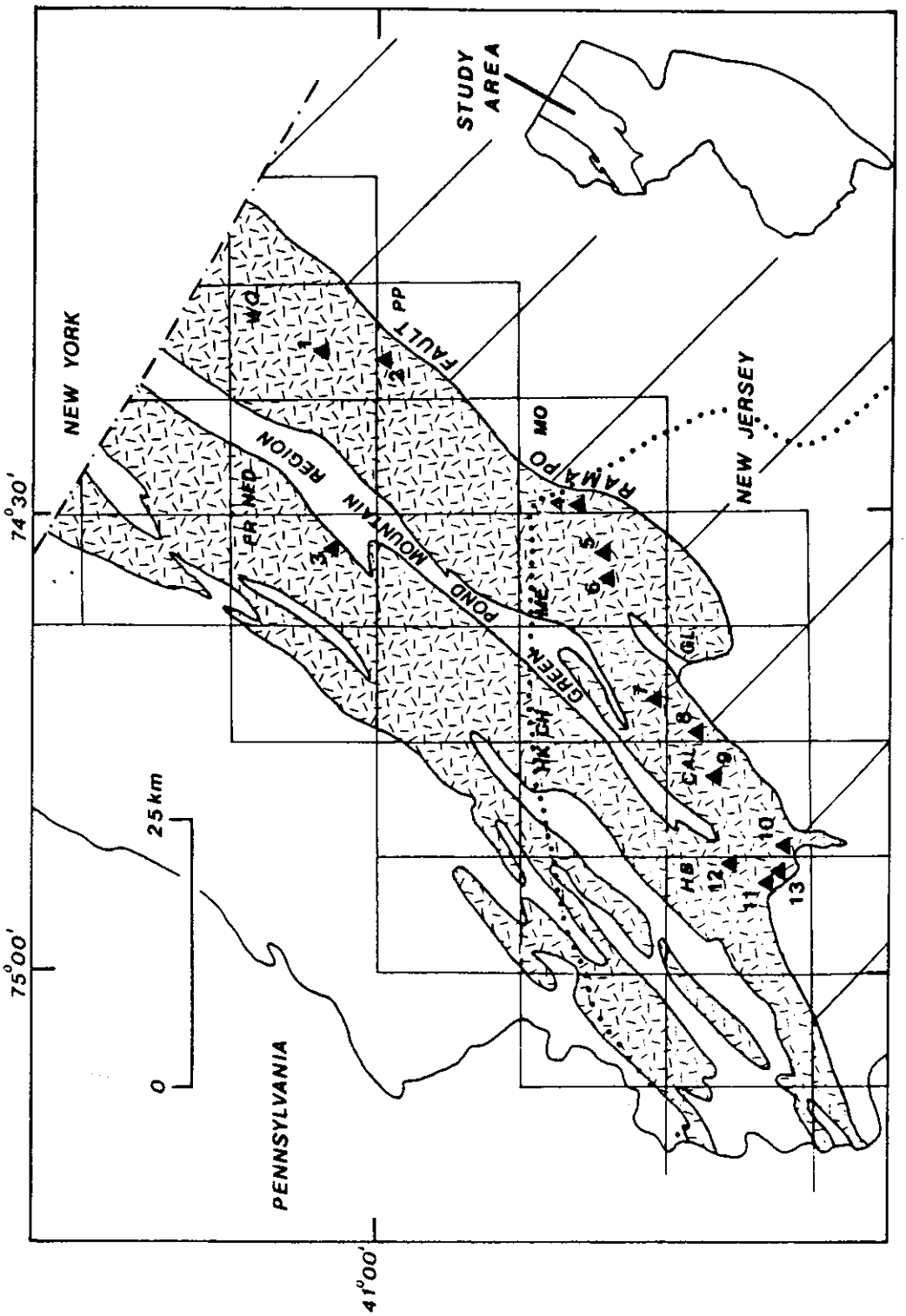

Figure 1. Generalized geologic map of northern New Jersey showing areal extent of Middle Proterozoic rocks of the Highlands (patterned). Dotted line marks the limit of Wisconsinan terminal moraine. Small inset map locates area shown. Triangles denote locations of graphite mines numbered as in table 1 and figure 2. Abbreviations of 7.5-minute quadrangles are: FR, Franklin; NFD, Newfoundland; WQ, Wanaque; PP, Pompton Plains; MO, Morristown; ME, Mendham; CH, Chester; HK, Hackettstown; HB, High Bridge; CAL, Califon; and GL, Gladstone.

Graphite is present in a number of Middle Proterozoic rocks, typically in trace amounts that seldom exceed a few tenths of a volume percent of the rock. Only biotite-quartz-feldspar gneiss and metaquartzite and some pegmatite contain graphite in sufficient quantity to constitute commercially exploitable ore deposits. Virtually all of the graphite deposits in the Highlands are confined to the southeast-central part (fig. 2) where they occur along similar trends in the High Bridge, Califon, Gladstone, Chester, Mendham, and Morristown quadrangles. Exceptions are deposits in the Pompton Plains, Wanaque, and Franklin quadrangles. The Franklin occurrence is the only known worked graphite deposit west of the Green Pond Mountain region. Middle Proterozoic rocks in the southwestern Highlands are sparsely graphitic and no deposits or concentrations of this mineral are known from there (A.A. Drake Jr., personal communication, 1997).

It is especially noteworthy that graphite deposits in southeastern Pennsylvania (Miller, 1912) and the Adirondack Mountains in New York (Alling, 1918) are predominantly hosted by the same sequence of Middle Proterozoic rocks as the deposits in New Jersey, namely biotite-quartz-feldspar gneiss, metaquartzite, and pegmatite. Miller (1912) may have been the first to recognize the similarity of the Pennsylvania and New York host rocks. His observation that metaquartzite is less abundant in Pennsylvania is consistent with the overall thinning of this lithology along strike from New York to Pennsylvania.

In most of the New Jersey Highlands deposits, graphite occurs as plates up to a few tenths of an inch in diameter, although very locally in pegmatite the plates may reach 1 inch or more in diameter. Except in pegmatite, graphite is concentrated mainly along foliation surfaces and, where present in abundance, occasionally imparts a schistose fabric to the rock. The following Middle Proterozoic rocks have been noted to contain graphite.

Metavolcanic Rocks

Hypersthene-quartz-plagioclase gneiss

Quartz-bearing charnockitic gneiss is composed of plagioclase (oligoclase to andesine), quartz, clinopyroxene, hornblende, biotite, hypersthene, minor potassic feldspar, and opaque minerals. Graphite occurs very locally in some exposures of this unit in the Franklin and Newfoundland quadrangles and rarely exceeds more than a few tenths of a volume percent of the rock.

Biotite-quartz-plagioclase gneiss

This unit contains quartz, plagioclase (oligoclase or andesine), biotite, and sparse hornblende. The quartz content is somewhat variable and in the Chester quadrangle this unit grades over short distances into biotite-plagioclase gneiss ± hornblende. Graphite in these rocks in the Newfoundland and Chester quadrangles occurs very locally and does not exceed a few tenths of a percent of the rock. In Newfoundland, graphitic biotite-quartz-plagioclase gneiss is conformably interlayered with graphitic hypersthene-quartz-plagioclase gneiss. In Chester, both quartz-bearing and quartz-deficient graphitic biotite-quartz-plagioclase gneiss are spatially associated with graphitic biotite-quartz-feldspar gneiss.

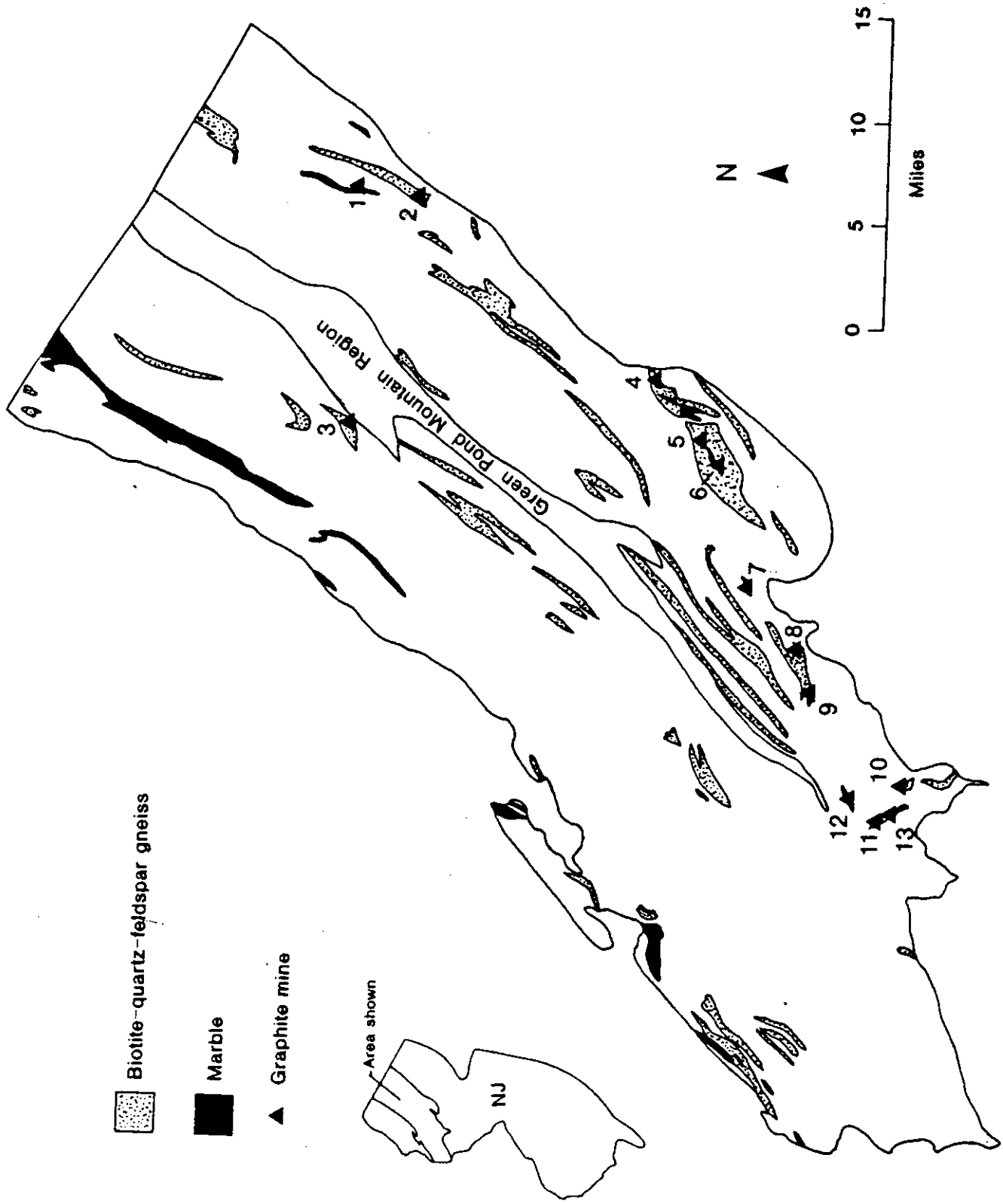

Figure 2. Spatial distribution of biotite-quartz-feldspar gneiss and marble in the New Jersaey Highlands (modified from Drake and others, 1996). Triangles locate known graphite mines. These are: 1) unnamed; 2) Bloomingdale; 3) unnamed; 4) Betts' exploration; 5) unnamed; 6) Dickson; 7) Englemann; 8) Fisher; 9) Sutton; 10) Annandale; 11) Beavers'; 12) "High Bridge"; and 13) Hackett.

Quartzofeldspathic Metasedimentary Rocks

Biotite-quartz-feldspar gneiss

Biotite-quartz-feldspar gneiss is composed principally of quartz, oligoclase and microcline in varying proportions, and biotite. Garnet and sillimanite are common accessory minerals, but graphite is exclusively confined to the variant that contains pyrite and weathers rusty. Graphite typically comprises less than 1 percent of the rock, but occasionally may exceed 30 percent or more where this unit hosts graphite deposits. The distribution of biotite-quartz-feldspar in the Highlands is shown in figure 2.

Locally graphitic, feldspathic metaquartzite layers generally less than 10 feet thick are common within the rusty-weathering type of biotite-quartz-feldspar gneiss. Thin, conformable, graphite-bearing calc-silicate layers are spatially associated with biotite-quartz-feldspar gneiss in the Gladstone, High Bridge, and Pompton Plains quadrangles.

The origin of biotite-quartz-feldspar gneiss, as well as the other graphitic metasedimentary rocks mentioned, will be discussed in more detail later in this paper.

Calc-Silicate Metasedimentary Rocks

Pyroxene gneiss

Pyroxene gneiss is composed of oligoclase, clinopyroxene, and variable amounts of quartz. Titanite is as a common accessory mineral. Graphite-bearing pyroxene gneiss is exposed in the Newfoundland, High Bridge, and Califon quadrangles. Hague and others (1956) describe an exposure of graphitic pyroxene gneiss in the Franklin quadrangle, and Offield (1967) mapped graphitic pyroxene gneiss in the Greenwood Lake quadrangle in southern New York that is gradational with graphitic quartzose gneiss. Graphite content in pyroxene gneiss rarely exceeds more than 1 percent of the rock.

Throughout the Highlands, pyroxene gneiss and rusty-weathering, sulfidic, biotite-quartz-feldspar gneiss commonly occur together in conformable contact.

Epidote gneiss

This unit is composed primarily of epidote, with minor amounts of bronze-colored mica (pholgopite?), scapolite, and trace amounts of titanite. The epidote is clinozoisite. Graphite contents range from a few tenths of a percent to 3 or 4 percent. However, Bayley and others (1914) describe a similar rock from the Englemann mine in the Chester quadrangle that is abundantly graphitic.

Graphitic epidote gneiss is best exposed in the Gladstone and High Bridge quadrangles where it is spatially associated with graphitic biotite-quartz-feldspar gneiss and metaquartzite. Epidote gneiss in the Gladstone quadrangle, although never mined, locally appears to contain up to 10 percent graphite. Bayley and others (1914) noted pieces of marble on the dump of the

prospect in the Chester quadrangle; therefore, graphitic epidote gneiss is spatially associated with both rock types.

Diopsidite

Diopsidite is sparsely exposed in the Highlands as thin lenses and layers that are typically associated with pyroxene gneiss, but in the Pompton Plains and High Bridge quadrangles it is associated with biotite-quartz-feldspar gneiss. Diopsidite is a medium-grained, light green, nearly monomineralic rock composed of clinopyroxene that has the composition of diopside. Geochemical analysis of diopside from pegmatite spatially associated with diopsidite at the Bloomingdale mine has the following composition in weight percent:

SiO_2	54.32
Al_2O_3	0.92
Feo^t	5.40
MgO	17.89
CaO	20.85
Na_2O	0.34
K_20	0.06
Total	99.78

Diopsidite occasionally contains up to 15 percent graphite where associated with rusty-weathering biotite-quartz-feldspar gneiss.

Other Metasedimentary Rocks

Metaquartzite

Metaquartzite occurs as thin lenses and layers scattered throughout the Highlands but is most abundant in the Wanaque and Mendham quadrangles. In Wanaque it is interlayered with pyroxene gneiss and marble, whereas in Mendham it occurs with rusty-weathering biotite-quartz-feldspar gneiss.

Metaquartzite associated with pyroxene gneiss and marble often contains plagioclase, scapolite and/or clinopyroxene and commonly contains graphite. Metaquartzite associated with biotite-quartz-feldspar gneiss contains feldspar, sometimes in sufficient quantity that the rock approaches a quartz-feldspar gneiss. Metaquartzite locally grades into biotite-quartz-feldspar gneiss with an increase in the biotite content. Its association with rusty-weathering biotite-quartz-feldspar gneiss is especially significant as these rocks appear to host many of the graphite deposits in the Highlands. Typical graphite content of metaquartzite is less than 1 percent, but locally ranges up to 40 percent where this unit hosts graphite deposits. Modal analyses of metaquartzite from some of these deposits are given in table 1.

Marble

Most marble is a calcitic to locally dolomitic rock that commonly contains graphite, phlogopite, chondrodite, and clinopyroxene. While graphite is nearly ubiquitous in marble as disseminated plates, it is not present in commercial quantities and seldom exceeds more than a few percent of the rock (Spencer and others, 1908). This is typical of the Franklin and Wildcat bands of marble mapped in the Franklin area and the northwest Highlands by New Jersey Zinc Company geologists (Hague and others, 1956).

Other pods, lenses, and layers of marble in the Highlands mainly east of the Green Pond Mountain region are sparsely graphitic and contain characteristic serpentine minerals. It is this type of marble that is associated with graphite deposits in the Highlands. Because these bodies of marble are spatially associated with the same rocks as marble in the Franklin area, Volkert and Drake (in press) chronocorrelate them with the Franklin, although it is recognized that they may not occur at the same stratigraphic level. The distribution of marble in the Highlands is shown in figure 2.

Intrusive Rocks

Byram Intrusive Suite

Granitoid rocks of the Byram Intrusive Suite consist of hornblende ± biotite, quartz in varying amounts, microcline microperthite and oligoclase. Graphite is present very locally in a few outcrops of hornblende granite near the Sutton mine in the Califon quadrangle, where it comprises less than 2 percent of the rock. Graphite has not been recognized in any rocks of the Lake Hopatcong Intrusive Suite or in the Mount Eve Granite.

Pegmatite

Two types of pegmatite that locally contain graphite occur in the Highlands. They are referred to here as metamorphic pegmatite, resulting from local melting of the host rock mainly *in situ*, and magmatic pegmatite, of which most are related petrogenetically to the Byram Intrusive Suite.

Metamorphic pegmatites are graphitic only where they are generated within, or intrude, graphitic metasedimentary rocks, most notably biotite-quartz-feldspar gneiss. These pegmatites form thin-to-thick, coarse-grained seams and layers that are mainly conformable to foliation and less often discordant. They are composed of quartz, microcline, oligoclase, and biotite. Graphite generally is disseminated throughout these pegmatites.

Magmatic pegmatites appear to contain graphite only where they intrude known graphite-bearing rocks. Magmatic pegmatites form thin-to-thick seams and veins that may be conformable to foliation but most often are discordant. They are primarily composed of quartz and microcline microperthite, with variable amounts of oligoclase and biotite ± hornblende. Graphite in magmatic pegmatites is seldom disseminated throughout the rock and is usually

confined to its contact with the intruded graphitic lithology. There it occurs in veins and masses, or as large plates covering, and slightly intergrown with, mineral grains in the pegmatite. Magmatic pegmatites are generally devoid of graphite more than a few feet away from this contact.

ORIGIN OF GRAPHITE IN THE HIGHLANDS

Geology of the host rocks

A fundamental issue in the genesis of graphite in the Highlands is the source of the requisite carbon. In order to address this we must first look at the rocks hosting the deposits. Graphite deposits in the Highlands are neither widespread, nor are they random occurrences. They appear to be restricted geographically as well as stratigraphically, and occur principally in marine sediments (metaquartzite and biotite-quartz-feldspar gneiss) and, to a lesser extent, pegmatite.

The marine origin of these sediments has been determined through detailed geochemistry and comparison with past and present sedimentary basinal analogs (Volkert, 1993; Volkert and Drake, in press). Further support for a marine origin comes from the recognition of relict pillow structures in amphibolite associated with the Franklin Marble (Hague and others, 1956) that has

Table 1. Modes of metaquartzite from Highlands graphite deposits [Based on 800 points. Analyst: R, Volkert; --, not present]

Mine Number[#]	Unnamed 1	Unnamed 5	Annandale 10	"High Bridge" 12
Quartz	54.6	54.1	48.7	61.1
K-feldspar[1]	25.8	14.1	19.2	10.9
Plagioclase	--	trace	4.9	trace
Graphite	17.2	30.9	23.1	23.2
Sillimanite	--	--	0.8	--
Biotite	--	0.1	trace	1.8
Hornblende	--	0.2	--	--
Garnet	--	--	1.2	--
Opaque[2]	0.3	trace	0.1	trace
Other[3]	2.1	0.6	1.9	3.0
Total	100.0	100.0	100.0	100.0

[#]Numbered as in figure 1;[1]Includes microperthite[2];Mainly iron sulfide
[3]Mainly alteration products

a mid-ocean ridge geochemical affinity (Volkert and others, 1986).

The majority of metaquartzite is metamorphosed impure quartzose sandstone. Abundant feldspar contents of some metaquartzite suggest that these rocks were locally fairly argillaceous. The pervasiveness of metaquartzite sequences hosting, or associated with, graphite deposits in the Highlands, as well as their conformable and gradational contact with biotite-quartz-feldspar gneiss argues against their origin as anatectic quartz segregations or as hydrothermal quartz. Biotite-quartz-feldspar gneiss has the composition of graywacke sandstone or siliceous shale that occasionally grades into calcareous variants of the same. Pyroxene gneiss has a composition that ranges from calcareous sandstone to shaly sandstone, occasionally grading into siliceous limestone. Diopsidite has been interpreted by Volkert and Drake (in press) to be metamorphosed cherty dolomite. Marble is metamorphosed, locally siliceous limestone and lesser dolomitic limestone.

This interlayered lithologic sequence represents sedimentation during the Middle Proterozoic in a marine shelf environment along the Laurentian (eastern North American) continental margin. A minimum age for this sequence is provided by a $^{207}Pb/^{206}Pb$ age of 1100 Ma from galena that cuts a gneiss fragment within the Franklin Marble at the Sterling Hill zinc mine (Metsger, 1977).

Metamorphic conditions and fluid composition

Middle Proterozoic rocks in the New Jersey Highlands were metamorphosed at granulite facies during Grenville orogenesis. Estimates of temperature (T) and pressure (P) during regional metamorphism of the metasedimentary rocks are reasonably constrained by their mineral assemblages. The equilibrium assemblage K-feldspar + sillimanite + garnet + biotite + plagioclase + quartz in biotite-quartz-feldspar gneiss hosting many of the graphite deposits is diagnostic of granulite facies. Local melting of this rock, and the generation of metamorphic pegmatite, define a lower limit for the P-T conditions within the sillimanite field to the granite minimum melt that occurs at ≈650°C at 5 kb (fig. 3).

Based on franklinite-gahnite intergrowths in zinc ore at the Sterling Hill mine in the Franklin quadrangle, Carvalho and Sclar (1988) obtained a temperature of 760°C during metamorphism of the ore hosted by the Franklin Marble. Using pyroxene, garnet-biotite, and sphalerite-pyrrhotite-pyrite geothermometry, Hewins and Yersak (1977) obtained temperatures of 680° to 750°C and a pressure of about 5 kb from the same study area. Based on Fe-Ti oxides from mineral assemblages at the Edison mine in the Franklin quadrangle, Puffer and others (1993) obtained a temperature of 732°C for the quartzofeldspathic gneiss host rock, 706° to 748°C for coexisting magnetite-ilmenite in the Fe ore, and water fugacity of 5044.4 atmospheres, or 5.1 kb. Other temperatures have been reported for Highlands granitoid rocks, but they likely record peak crystallization temperatures rather than metamorphic temperatures and have little relevance in the following discussion.

Using the above data for T and P, as well as experimental data, certain assumptions may be made regarding the oxygen fugacity (fo_2) and fluid composition that was in equilibrium with graphite during granulite facies metamorphism in the Highlands. Under a confining pressure of

about 5 kb and temperatures of 700°C to 760°C, the gas and fluid phases present would have coexisted as a supercritical fluid rather than separate phases. Given these conditions it is

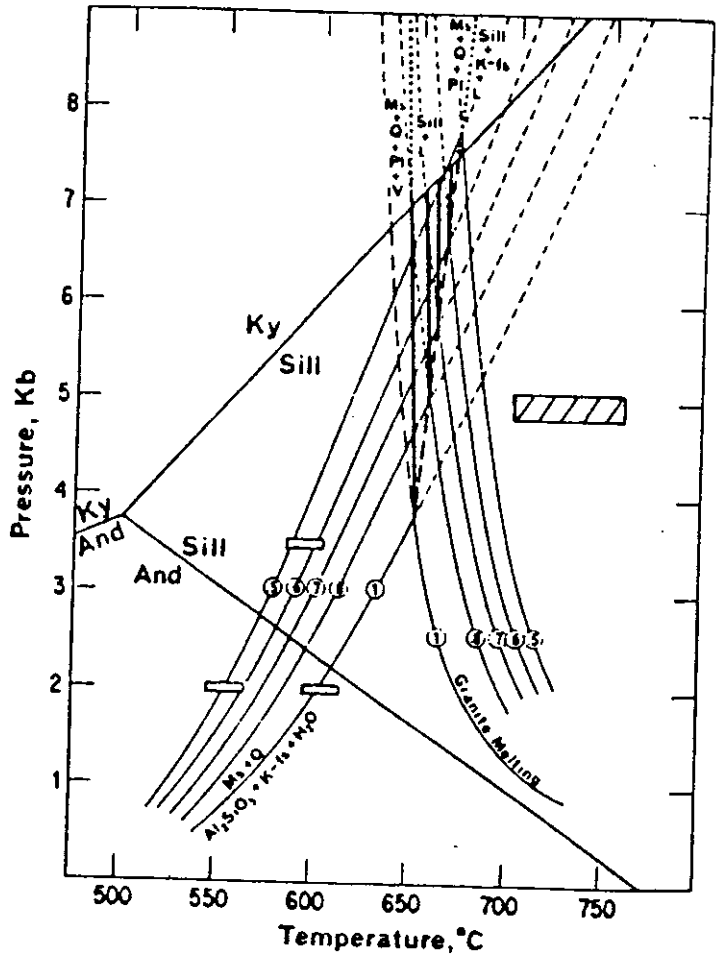

Figure 3. P-T diagram from Kerrick (1972) showing the phase boundaries for kyanite (Ky), andalusite (And), and sillimanite (Sill), and the granite minimum melting curve. Highlands metasedimentary rocks plot within the diagonal ruled field to the right of the sillimanite+K-feldspar+liquid curve at X_{H2O} <0.5.

reasonable to assume that the load pressure (P_{load}) _ fluid pressure (P_{fluid}) and that the P_{fluid} _ gas pressure (P_{gas}). That is, the total pressure was roughly equal to the sum of the partial pressures of the predominant gas phases in the fluid, namely $PH_2O + PCH_4 + PCO_2 + PCO + PO_2 + PH_2 + PH_2S$ and, therefore, about 5 kb.

Under granulite facies conditions the presence of graphite in a rock exerts a strong control on fo_2 and constrains it to the stability field of magnetite at a point on, or below, the quartz-fayalite-magnetite (QFM) buffer (fig. 4). Graphite is unstable under more oxidizing conditions above this curve and converts into a CO_2-rich phase. As seen on figure 4, the graphite stability

surface is both temperature and pressure dependent. With increasing T and decreasing P, the graphite-gas buffer curves intersect more reduced oxygen buffers. Using the following equation from French and Eugster (1965)

$$\log fo_2 = -(20{,}586/T) - 0.044 + \log P_{gas} - 0.028\,(P_{gas} - 1)/T$$

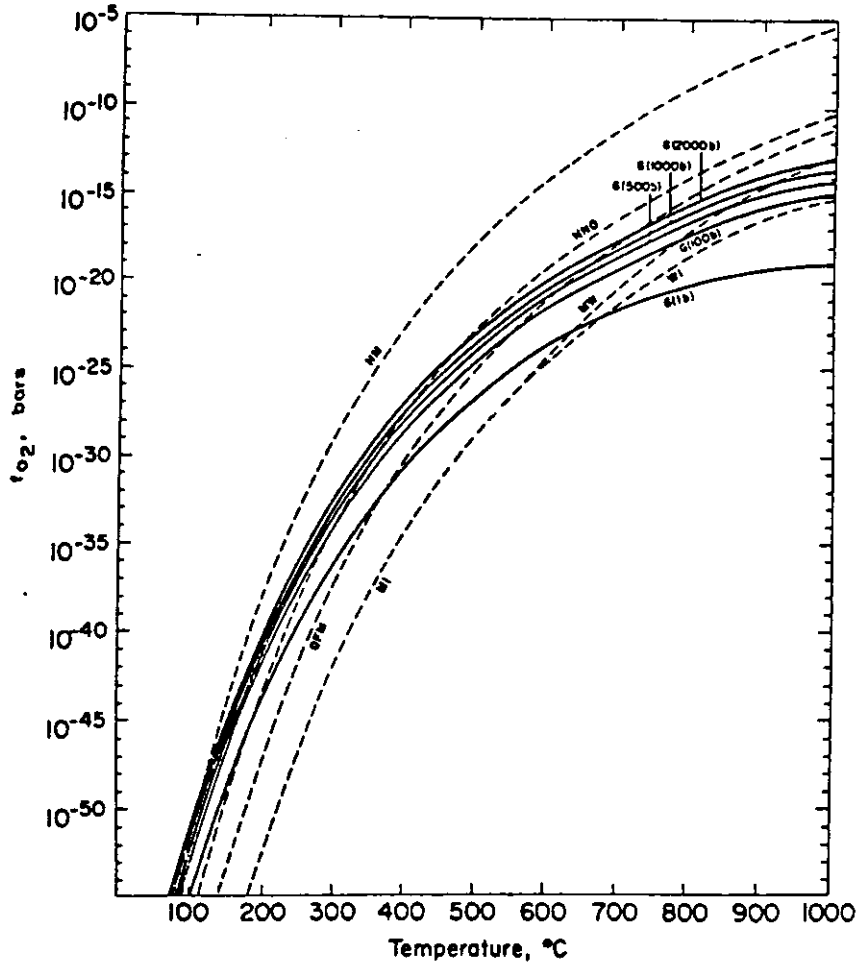

Figure 4. Log fo_2-temperature diagram from French and Eugster (1965) for the graphite-gas buffer at total gas pressures of 1 bar to 2 kb (solid lines). Oxygen buffer assemblages (dashed lines) are: HM, hematite+magnetite; NNO, nickel+bunsenite; QFM, quartz+fayalite+magnetite; MI, magnetite+iron; MW, magnetite+wustite; and WI, wustite+iron.

values of log fo_2 were calculated for a gas phase in equilibrium with graphite at T of 700° and 760°C and P_{gas} of 5 kb. These values are -17.27 and -16.05 respectively and, as shown in figure 4, they are conicidant with the QFM buffer curve. They represent the highest fo_2 values permissible for the stability of graphite under the T and P conditions inferred for the Highlands and probably were lower as the temperature increased and conditions were more reduced.

Common volatile species in the C-O-H-S system present in a fluid during regional metamorphism include H_2O, CO_2, CO, O_2, CH_4, H_2, and H_2S. Only a few of these occur in significant proportion and they are responsible for determining the P_{gas}. The principal species in equilibrium with graphite at temperatures <800°C and pressures >2kb are H_2O, CO_2, CH_4 and H_2S (French, 1966; Ohmoto and Kerrick, 1977; Frost, 1979; fig. 5). The presence of sulfide minerals coexisting with graphite in the Highlands (table 1) suggests that H_2S was present locally. It was likely generated internally through desulfidation reactions. H_2O was derived mainly from internal sources through devolitilization reactions during prograde metamorphism.

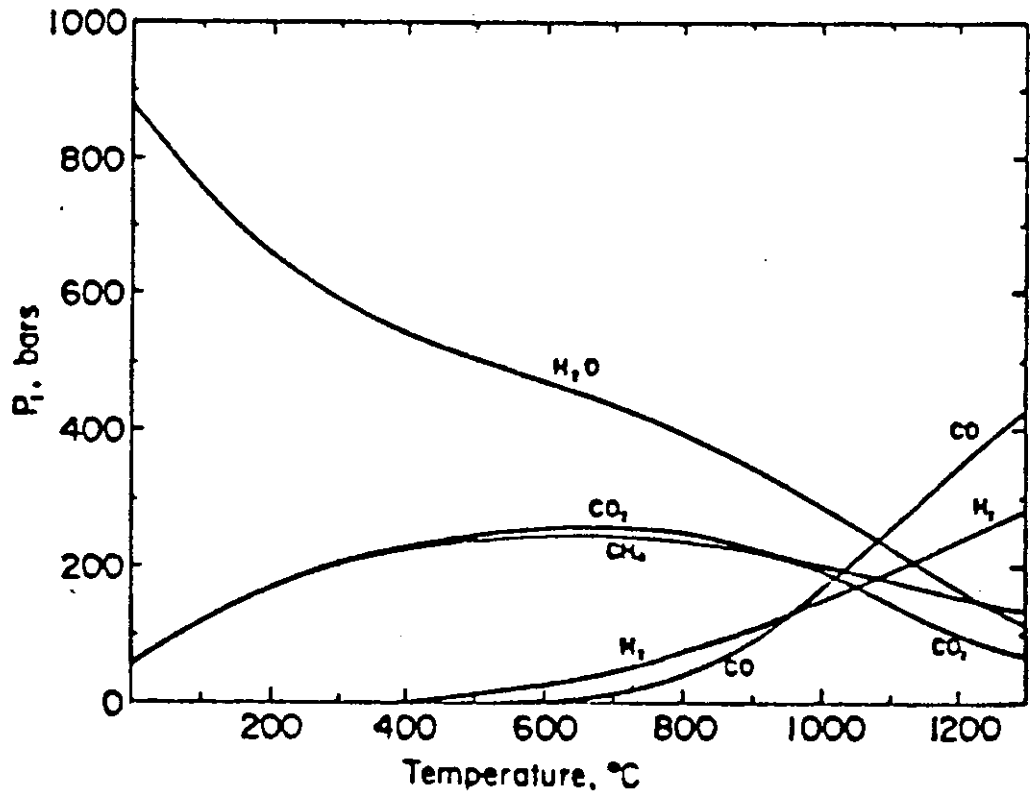

Figure 5. Relative proportion of species in a C-O-H fluid in equilibrium with graphite at total gas pressure of 1 kb. Diagram from French (1966).

CO_2 was derived externally through decarbonation reactions involving marble and/or calc-silicate rocks, and where they were absent, internally, where along with CH_4 it was produced through the interaction of H_2O with carbonaceous material in the rock. This is shown by the following reaction from Ohmoto and Kerrick (1977)

$2C+2H_2O \Rightarrow CO_2+CH_4$.

At relatively high values of fo_2 such as those calculated for the Highlands, the main fluid species present would have been a mixture of H_2O, CO_2, and CH_4 (fig. 5). Because higher temperature has the effect of increasing the amount of H_2S present (Poulson and Ohmoto, 1989), this was probably a locally important species also. The proportions of these species present in the fluid would have varied depending on the composition of the rock undergoing metamor- phism. For example, graphite-bearing marble and calc-silicate rock would have been in equilibrium with a mixture of predominantly CO_2 and H_2O, whereas sulfidic biotite-quartz-feldspar gneiss and metaquartzite would have been in equilibrium with H_2O and $CH_4 \pm CO_2$. Calcareous layers within, or adjacent to, the gneiss would have created a more open-system condition resulting in a fluid containing a mixture of these species. In summary, the volatile species in equilibrium with graphite in the Highlands are inferred to have been a mixture of mainly H_2O, CO_2, CH_4, and subordinate H_2S.

Sources of carbon and graphite genesis

Sedimentary rocks of Proterozoic age contain an abundant diversity of microorganisms mainly in the form of bacteria and algae (Salop, 1982; Cloud, 1983), and the average content of organic matter as reduced carbon in Proterozoic sediments does not differ significantly from that in Phanerozoic sediments (Schidlowski, 1982). This idea is not new and a biogenic origin for reduced carbon in some Proterozoic metasedimentary rocks has long been interpreted (e.g. Alling, 1918).

Graphite present in metasedimentary rocks in the Highlands is similarly inferred to be biogenic in origin. This is supported by the following: 1) the occurrence of graphite deposits primarily in a restricted sequence of marine sediments that represent organic-rich sands and muds; 2) the uniform occurrence of graphite in these rocks over strike lengths of several hundred feet to several miles; 3) the stratigraphically conformable relationship of graphite with the host lithologies; and 4) the apparent lack of field or petrographic evidence to suggest that graphite layers in the metasedimentary rocks formed by invading fluids.

Much of the organic matter in sediments is consumed by aerobic bacteria prior to compaction (Schidlowski, 1982). Therefore, to preserve organic matter an unhospitable, anoxic environment likely existed during the Proterozoic in this marine basin in the Highlands. The presence of abundant pyrite within graphitic metaquartzite and biotite-quartz-feldspar gneiss suggests that such an environment did locally exist, as these rocks grade into non-carbonaceous and non-sulfidic phases of the same units. Organic matter was able to accumulate in this setting under poorly oxygenated and strongly reducing conditions due to restricted circulation. Whether this restriction was due to tectonic or climatic factors, or to a combination of controls is unknown. Under reducing conditions in an anoxic environment the content of organic matter in sandstones may be equal to that of shale (Oehler and others, 1982) supporting the interpretation of this study that carbon in the metaquartzite and biotite-quartz-feldspar gneiss protoliths in the Highlands was primary and biogenic in origin. Likely products of anerobic bacterial decay in this euxinic environment were the generation of methane and hydrogen sulfide, as well as the

reduction of sulfate and iron in the sediments, which combined to form the pyrite commonly found in the graphitic metasedimentary rocks.

Biogenic carbon compounds have distinctive $\delta^{13}C$ isotopic values that range from -25 to -35‰ (Oehler and others, 1972). Although no isotopic analyses have been performed on graphite from New Jersey, graphite from the Middle Proterozoic Pickering gneiss (correlative with biotite-quartz-feldspar gneiss in the Highlands) and the Franklin Marble in southeastern Pennsylvania was analyzed by Crawford and Valley (1990). They obtained $\delta^{13}C$ values of -14.4 to -23.8‰ from graphite in the gneiss and concluded the source was organic matter in the sedimentary protolith.

During granulite facies metamorphism in the New Jersey Highlands, graphite in the gneisses and metaquartzite formed largely *in situ* from the organic-rich layers that now parallel compositional layering and foliation. Organic matter probably altered to graphite through the loss of hydrogen and subsequent recrystallization of carbon as graphite (e.g. Faure, 1986).

A different interpretation is required to explain the occasional presence of graphite in intrusive rocks of the Highlands, notably magmatic pegmatite, as well as in metavolcanic rocks (hypersthene-quartz-plagioclase gneiss and biotite-quartz-plagioclase gneiss). It is widely recognized that juvenile carbon may be a minor component of some igneous magma (Faure, 1986). However, it is unlikely that magmatic carbon was responsible for the graphite in the Highlands intrusive rocks. If this was the case, graphite should be a more widespread constituent of granitic rocks and magmatic pegmatite, whereas it is not and occurs mainly where these rocks intrude graphitic lithologies. Therefore, the evidence suggests that granitic rocks and magmatic pegmatite locally obtained carbon from carbonaceous metasedimentary rocks during intrusion and subsequently precipitated graphite along their contacts.

Several theories have been proposed to account for the genesis of graphite. Salotti and others (1971) propose a reaction whereby hydrogen diffuses into carbonate rocks at temperatures of 200° to 600°C, reacting with carbonate minerals to form methane. This is shown by the reaction

$$CaCO_3 + 4H_2 \Rightarrow CH_4 + H_2O + Ca(OH)_2.$$

The subsequent breakdown of methane produces carbon and hydrogen according to the reaction

$$CH_4 \Rightarrow C + 2H_2.$$

A problem with this mechanism in terms of the Highlands is the absence of graphite deposits where granite or pegmatite intrude exposures of marble. Also, if graphite was produced by this reaction then it should be present in marble adjacent to graphite deposits in the metasedimentary rocks, when in fact graphite generally is absent in the marble.

Glassley (1982) and Katz (1987) propose that graphite in high-grade metamorphic rocks forms through the interaction between CO_2-rich fluids and H_2O. The source of the CO_2 is postulated as being from the mantle or from metamorphic decarbonation reactions involving carbonate rocks. Graphite forms when the fluid composition is in equilibrium with the mineral assem-

blage of crustal rocks. This could account for the sparse amounts of graphite in the metavolcanic rocks, as they are not spatially associated with carbonate rocks or graphite deposits and they doubtfully contained primary organic carbon.

However, CO_2 flooding is an unlikely mechanism for other graphite genesis in the Highlands for the following reasons: 1) graphite should be more widespread and abundant in rocks of diverse composition; 2) the mineral assemblages of rocks hosting graphite deposits are relatively uniform and these rocks occur throughout the Highlands, yet they are often barren in areas proximal to graphite deposits; 3) there appears to be a lack of evidence in the Highlands of invasion by pervasive CO_2-rich fluids (Johnson and others, 1990); and 4) there is a documented absence in the Adirondack Mountains of pervasive CO_2-rich fluid migration during granulite facies metamorphism (McLelland and others, 1988 and references therein).

Weis and others (1981) suggest that graphite genesis occurs through a water gas reaction whereby superheated (700° to 900°C) water vapor reacts with organic matter in metasediments to produce carbon monoxide according to the following reaction

$$C+H_2O \Rightarrow CO+H_2.$$

Carbon monoxide then decomposes according to the Boudouard reaction

$$2CO \Rightarrow C+CO_2,$$

precipitating graphite at temperatures 50° to 100°C lower (Weis and others, 1981). Temperatures required for this reaction are in good agreement with those attained in the Highlands during granulite facies metamorphism.

Rumble and others (1986) propose a mechanism involving the mobilization and transport of carbon from sediments as carbon dioxide and methane during metamorphic devolitalization reactions. Graphite precipitation occurs when migrating fluids of differing CO_2 and CH_4 ratios mix in fractures or along permeable zones.

Although any of the above processes are capable of graphite formation, those proposed by Weis and others (1981) and Rumble and others (1986) appear more consistent with the known geologic relations in the Highlands. Reactions involving CO_2 and H_2 would have increased the amount of H_2O available to diffuse into adjacent wall rocks, accounting for the presence of hydrous minerals and the local abundance of pegmatite along the margin of the ore body at some graphite mines.

Graphite genesis in the Highlands undoubtedly had a fairly complex origin in the Middle Proterozoic involving organic and inorganic sources of carbon. More specific details regarding the formation of these interesting deposits remain speculative pending intensive mineralogic, petrologic, and isotopic studies.

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GRAPHITE MINES AND MINING HISTORY IN THE NEW JERSEY HIGHLANDS

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INTRODUCTION

Graphite, also known as plumbago and black lead, is a native element composed of carbon that possesses a metallic luster, a hardness of 1 to 2, and a specific gravity of about 2.23. Graphite occurs as dark gray to black, tabular, hexagonal flakes or plates, in microcrystalline form (often referred to as amorphous), and in fibrous, columnar, or radiating aggregates.

The usefulness of graphite for certain applications has been known for centuries as indicated by the following historical information excerpted from Graffin (1983). The earliest known use of graphite was by primitive man to draw on cave walls. Egyptians later used it to decorate pottery, and in Bavaria, in 1400 A.D., crucibles were fashioned from it. Graphite was first recognized as a distinct mineral species in 1565 by Gessner, and its composition was determined in 1779 by Scheele. It was named in 1789 by Werner from the Greek word "graphien" meaning to write.

Graphite is a good conductor of heat and electricity and if protected from oxidation remains stable to 3,500°C (Weis, 1973). Therefore, its principal uses are as a source of carbon in steelmaking, for high-quality crucibles and refractory ware, carbon brushes in electric motors, foundary facings and, to a lesser extent, in dry-cell batteries, lubricants, paint, and pencils (Evans, 1993). In recent years it has replaced asbestos in the manufacture of brake pads, seen use in the aerospace and nuclear power industries, and also as a component in military stealth technology (Asbury Graphite Mills Inc., 1995).

GEOLOGIC SETTING

Middle Proterozoic rocks of the New Jersey Highlands (fig. 1) consist of a heterogeneous assemblage of metavolcanic, metasedimentary, and metaplutonic lithologies. The occurrence of graphite in New Jersey is restricted principally to the rocks of the Highlands. It is absent elsewhere in the State except as rare detritus in younger rocks and sediments.

Graphite is present in a number of Middle Proterozoic rocks, typically in trace amounts that seldom exceed a few tenths of a volume percent. Of these, biotite-quartz-feldspar gneiss, metaquartzite, and some pegmatite contain graphite in sufficient quantity to warrant mining.

It is noteworthy that practically all of the graphite deposits are confined to the southeast-central Highlands (fig. 1) where they occur along similar trends in the High Bridge, Califon, Gladstone, Chester, Mendham, and Morristown quadrangles. Exceptions are deposits in the Pompton Plains, Wanaque, and Franklin quadrangles.

In Benimoff, A.I., and Puffer, J.H., (editors), The economic geology of northern New Jersey: Field Guide and Proceedings of the fourteenth annual meeting of the Geological Association of New Jersey, 1997, p 21-32.

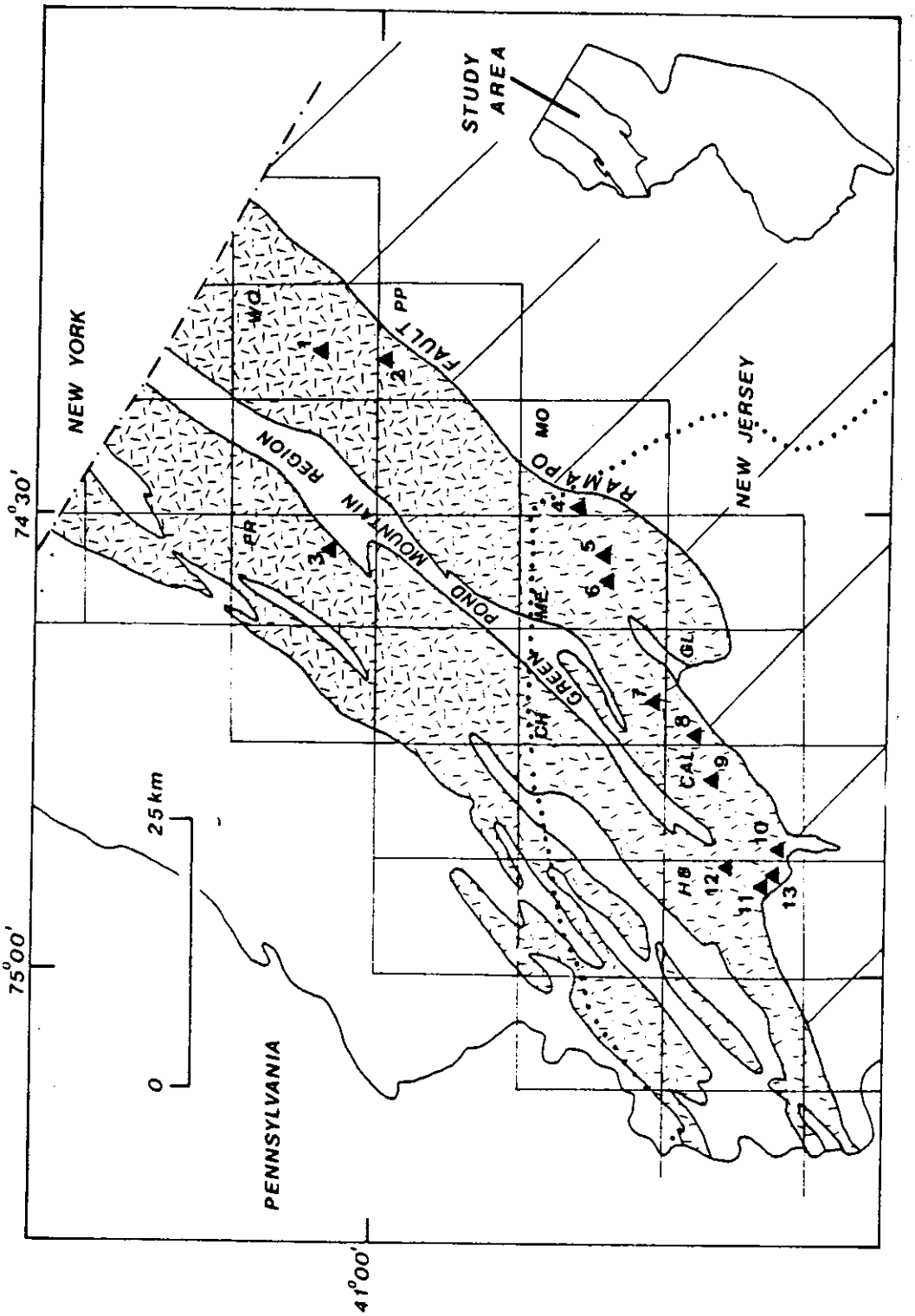

Figure 1. Generalized geologic map of northern New Jersey showing Middle Proterozoic rocks of the Highlands (patterned), Mesozoic rocks of the Newark basin (diagonal ruled), and Paleozoic rocks of the Valley and Ridge (unpatterned). Small inset map locates area shown. Triangles denote known graphite mines numbered as in the text. Abbreviations of 7.5-minute quadrangles are: FR, Franklin; WQ, Wanaque; PP, Pompton Plains; MO, Morristown; ME, Mendham; CH, Chester; HB, High Bridge; CAL, Califon; and GL, Gladstone. Dotted line marks limit of Winsonsinan terminal moraine.

In most deposits of the New Jersey Highlands, graphite occurs as plates up to a few tenths of an inch in diameter, although very locally in pegmatite the plates may reach an inch or more in diameter. Except in pegmatite, graphite is concentrated mainly along foliation surfaces.

MINING HISTORY IN NEW JERSEY

From about 1848 until 1928, various attempts were made to commercially mine graphite from selected Middle Proterozoic lithologies in the New Jersey Highlands. Most of these mines occurred on a relatively small scale and quickly encountered financial problems. However, even the few larger ones had difficulty remaining solvent financially for very long. Part of the problem involved a technology that, at the time, could not easily and inexpensively separate graphite from the host rock. As a result, much ore was discarded with the gangue minerals. Problems arose also through competition from graphite mines in southeastern Pennsylvania, as well as the numerous Adirondack deposits in New York that exploited more extensive graphite-rich ore zones. According to Alling (1918), of the total United States output of graphite in 1916 of about 5,000 tons, the leading producers were Alabama, New York, Pennsylvania, California, and Montana, with New York contributing one fourth of the total.

Figures regarding total extraction of graphite ore from New Jersey are unavailable, and where mentioned in the historical literature, local production figures are sketchy and incomplete. Where known, these data are given under the following individual mine descriptions.

Quite possibly the increased use of less costly imported graphite from such localities as Sri Lanka (Ceylon), or the development of synthetic graphite manufactured from coal in electric furnaces, or as a by-product of blast furnaces, marked the end of graphite exploration and mining activity in New Jersey. According to unpublished documents on file at the New Jersey Geological Survey, between 1928 and 1932 the American Cyanamid Company in Union County produced 1,005 short tons of artificial graphite.

New Jersey Graphite Mines

The numbers preceding each of the following mines are keyed to the mine numbers in figure 1. Most of these mines are on private property and their description here does not convey permission to trespass at these localities.

1. Unnamed mine (Bloomingdale Twp., Wanaque quadrangle)

No historical information is available regarding the name, date of exploration, or the amount of ore extracted here. Nor is this prospect shown on any historical maps. Two small prospect pits or filled shafts occur about 60 feet apart in a wooded area south of Wanaque

Reservoir. A small dump occurs adjacent to each of the openings. Judging by the amount of rock on the dumps, and by the high concentration of graphite that remains in the dump material, it is doubtful that much ore was removed from this location.

Host rock for the deposit is light gray, medium-fine- to medium-grained feldspathic metaquartzite approximately 30 feet thick. A sample of this rock from the mine dump contains 17.2 percent graphite (Volkert, 1997). Exploration here predominantly exploited the locally rusty-weathering sulfidic layers adjacent to the eastern and western contacts with calcitic marble, although the metaquartzite contains abundant graphite throughout. The marble contains accessory serpentine minerals, but lacks graphite. Thin layers of calcitic marble here are conformably interlayered with equally thin layers of pyroxene gneiss that also lack graphite. Foliation trends N24°E and dips 65° SE, steepening to 83° SE from west to east.

2. Bloomingdale mine (Bloomingdale Twp., Pompton Plains quadrangle)

The Bloomingdale mine was one of the larger graphite mines in the Highlands. It was first worked about 1864 or before, but by 1868 was idle due to financial problems. The mine reopened briefly in 1878 under new ownership, closed, and then reopened in 1882 under still different ownership by the Bloomingdale Graphite Company. Mining proceeded for another year or two before ceasing permanently.

The original workings consisted of three openings, a shallow open cut and two shafts 20 feet and 60 feet deep (New Jersey Geological Survey, 1879). In 1882, a drift was driven along strike of the ore from a depth of 30 feet in the deeper shaft. The mineralized zone was reported as being at least 16 feet wide and contained up to 11.2 percent graphite (New Jersey Geological Survey, 1885). No production figures are available prior to 1882, but 300 tons of graphite were subsequently extracted from the drift (New Jersey Geological Survey, 1883).

A graphite mill was constructed near the Bloomingdale mine, one of only four in the Highlands, for the separation of ore utilizing a "dry" process. The ore was crushed under stamps, washed through sieves, and then separated on five different dressing tables into graphite of various grades. When fully operational, the mill at Bloomingdale was capable of separating a capacity of 1,000 pounds of ore per day (Darton and others, 1908). At the present time all of the original mine workings have been developed over and no trace of the mill remains.

Host rocks for the deposit at the Bloomingdale mine are predominantly rusty-weathering, sulfidic, biotite-quartz-feldspar gneiss and pegmatite. The former grades into conformable layers of sparsely graphitic metaquartzite and quartz-poor biotite gneiss that is locally sulfidic and contains up to 30 percent graphite. Thin layers of light green, medium-grained, diopside-rich rock (diopsidite) each several feet thick and containing up to 10 percent graphite are also conformably layered with graphitic biotite-quartz-feldspar gneiss and metaquartzite. Graphite in these rocks is aligned along foliation surfaces and disseminated throughout the rock. Pegmatite is white or light gray, coarse-grained, unfoliated, slightly discordant, and composed of two distinct mineral assemblages: 1) microcline microperthite + plagioclase + quartz + biotite and 2) diopsidic pyroxene + plagioclase ± quartz ± titanite. Graphite is not evenly distributed in pegmatite and is confined to its contact with the adjacent graphitic gneiss. Graphite plates up to

1 inch in diameter occur in the pegmatite where they form around, and are embedded in, feldspar grains and less often quartz. Pegmatite occurs along the contact between biotite-quartz-feldspar gneiss and diopsidite and pegmatite mineral assemblages suggest that they represent local melt from these rocks. Foliation in the vicinity of the Bloomingdale mine trends N30°E to N55°E and dips 45° to 60° SE.

3. Unnamed mine (Jefferson Twp., Franklin quadrangle)

According to the description given in Baker and Buddington (1970), a few prospect pits were dug into graphitic biotite-quartz-feldspar gneiss and graphitic pegmatite seams "near Oak Ridge Reservoir". No additional information is available regarding this exploration. One can assume that it was quite small and probably not much, if any, ore was removed from here. Host rock is rusty-weathering biotite-quartz-feldspar gneiss and the foliation in the area trends N30°E and dips 80° SE (Baker and Buddington, 1970).

4. Betts' exploration (Morris Twp., Morristown quadrangle)

Mining at this location occurred shortly prior to 1868 on the property of the Betts farm, but apparently had ceased by that year. Several shafts of indeterminate depth were sunk along the ore zone which returned rock containing an average of 6 percent graphite (Bayley and others, 1914). Nothing remains of any of the workings. The exact location is speculative, but assumed to be on or near the Betts property shown on historical maps of Morris County.

Host rock for the ore probably is light gray, medium-grained, graphitic, feldspathic metaquartzite. This has been mapped in the vicinity of the Betts farm property (Volkert, 1988b) where it is conformably interlayered with rusty-weathering, sulfidic, graphite-biotite-quartz-feldspar gneiss that trends about N20°E and dips 82° NW to vertical.

5. Unnamed mine (Morris Twp., Mendham quadrangle)

No historical information is available regarding the name, dates of exploration, or the amount of ore extracted here. Nor is this prospect shown on any of the historical maps of Morris County. The remains of a single small shaft and dump occur along the side of a stream south of Route 24.

As with most of the other graphite prospects in the area, the host rock for the deposit is light gray, medium-grained, well-layered, locally schistose, feldspathic metaquartzite that is conformably interlayered with rusty-weathering, sulfidic, garnet-biotite-quartz-feldspar gneiss (Volkert, 1988a). While both units contain graphite, ore was extracted principally from the metaquartzite, which locally contains up to 30.9 percent graphite (Volkert, 1997). Foliation trends N35°W and dips 40° NE.

6. Dickson mine (Mendham Twp., Mendham quadrangle)

No information is available for this mine despite the fact that graphite was present here in sufficient quantity that a mill was constructed. The only reference to this particular operation is a single sentence in Bayley and others (1914). Although the location is reported as being on the road between Brookside and Washington Corners, this mine is not shown on any historical maps of Morris County. If the location is as described, this mine has been completely developed over.

Host rock for the deposit is light gray, medium-grained, well-layered, feldspathic metaquartzite. This is conformably interlayered with rusty-weathering, sulfidic, garnet-sillimanite-biotite-quartz-feldspar gneiss (Volkert, 1988a). This sequence of rocks trends N75°E and dips an average of 51° SE.

7. Englemann mine (Chester Twp., Chester quadrangle)

This mine began operating about 1848 on the farm of Elias Engelmann and likely closed shortly after. Its location is reported as being along the southeast bank of a ravine east of Peapack Brook (Cook, 1868). No additional information is available regarding the extent of the workings, length of operation, or the amount of ore extracted. The width of the mineralized zone varied between 4 feet and 5 feet and was traceable along strike for several hundred feet. Two samples from this locality were reported to contain 13.04 and 14.95 percent graphite (New Jersey Geological Survey, 1879). No trace of the mine workings or dumps could be located.

Host rock for the ore is unknown, although float of medium- grained, graphitic, feldspathic metaquartzite was observed in the approximate area of the mine, and Bayley and others (1914) describe graphitic epidote gneiss from the mine dump. The lithologic sequence here is silicated marble, metaquartzite, epidote gneiss, and quartz- poor pyroxene gneiss (Volkert and others, 1990). Graphite appears to be absent in the marble and pyroxene gneiss. Foliation in the area trends about N70°E and dips gently southeast.

8. Fisher mine (Tewksbury Twp., Gladstone quadrangle)

This operation began about 1880 when "machinery" was set up on the farm of C. Fisher to separate graphite from the country rock (Bayley and others, 1914). The mine was never worked to much of an extent and was abandoned a short time after opening. No information is available regarding the extent of the workings, if any, or the amount of graphite extracted here. No trace of the mine workings could be located, but abundantly graphitic rock was observed in the area.

Host rock is white or buff, medium-fine- to medium-grained, moderately foliated and layered, locally rusty-weathering, sulfidic, feldspathic metaquartzite that contains quartz, feldspar, dark amber to bronze-colored mica (phlogopite?), and graphite. Foliation in the area trends about N60E and dips 45° SE.

9. Sutton mine (Tewksbury Twp., Califon quadrangle)

Operation commenced about 1878 on the farm of George B. Sutton near Fairmount from a single shaft 28 feet deep and two shallow prospect pits. The mineralized zone was reported as being several yards wide in rock that contained an average of 6.87 percent graphite (New Jersey Geological Survey, 1879). No trace of the mine workings remain. However, abundantly graphitic float occurs in fields and on rock fences in the area of the mine.

Host rocks are garnet-sillimanite-biotite-quartz-feldspar gneiss and light gray feldspathic metaquartzite. These are bounded by medium-grained, moderately foliated hornblende granite (Volkert, 1989). Samples of biotite-quartz-feldspar gneiss and metaquartzite collected from the fields for this study contain up to 20 percent graphite. Only a few outcrops of hornblende granite contain graphite, generally less than 2 percent, adjacent to the contact with graphitic meta-sedimentary rocks. Foliation is somewhat variable due to folding, but closest to the area of the mine it trends about N70E and dips gently southeast.

10. Annandale mine (Clinton Twp., Califon quadrangle)

The Annandale graphite mine was the last operating graphite exploration in the Highlands. Unpublished notes on file at the New Jersey Geological Survey dated May, 1921 by H.B. Kummel suggest that exploration occurred about 1920 or 1921 on the property of the State Reformatory. However, according to the same notes, older workings and a dump were observed on the site in 1921. The date of these earlier workings is unknown, unless they are part of the Annandale iron mine which was opened in 1880 (Bayley, 1910).

By 1926, a well equipped and modern mill (fig. 2) had been constructed at the Annandale graphite mine that was capable of processing 4 tons of ore per day. At the time, graphite was still being separated from country rock by the "dry" process described earlier for the Bloomingdale mine. This technique permitted only about 34 percent graphite to be separated after the first run at the Annandale mill and often required two or three additional runs to separate most of the ore. Because this proved uneconomical, the mill was shut down sometime between 1926 and 1928 and modified to utilize a new experimental "wet" process. This involved wet grinding followed by separation of the ore in flotation cells, a technique that reportedly enabled 98 percent of the ore to be recovered on the first run.

Plans for the construction of a larger mill were abandoned and the Annandale mine was closed by 1929, despite the extensive and costly development up to that time by the Annandale Graphite Corporation of Philadelphia. After a change of management, mining resumed in 1931

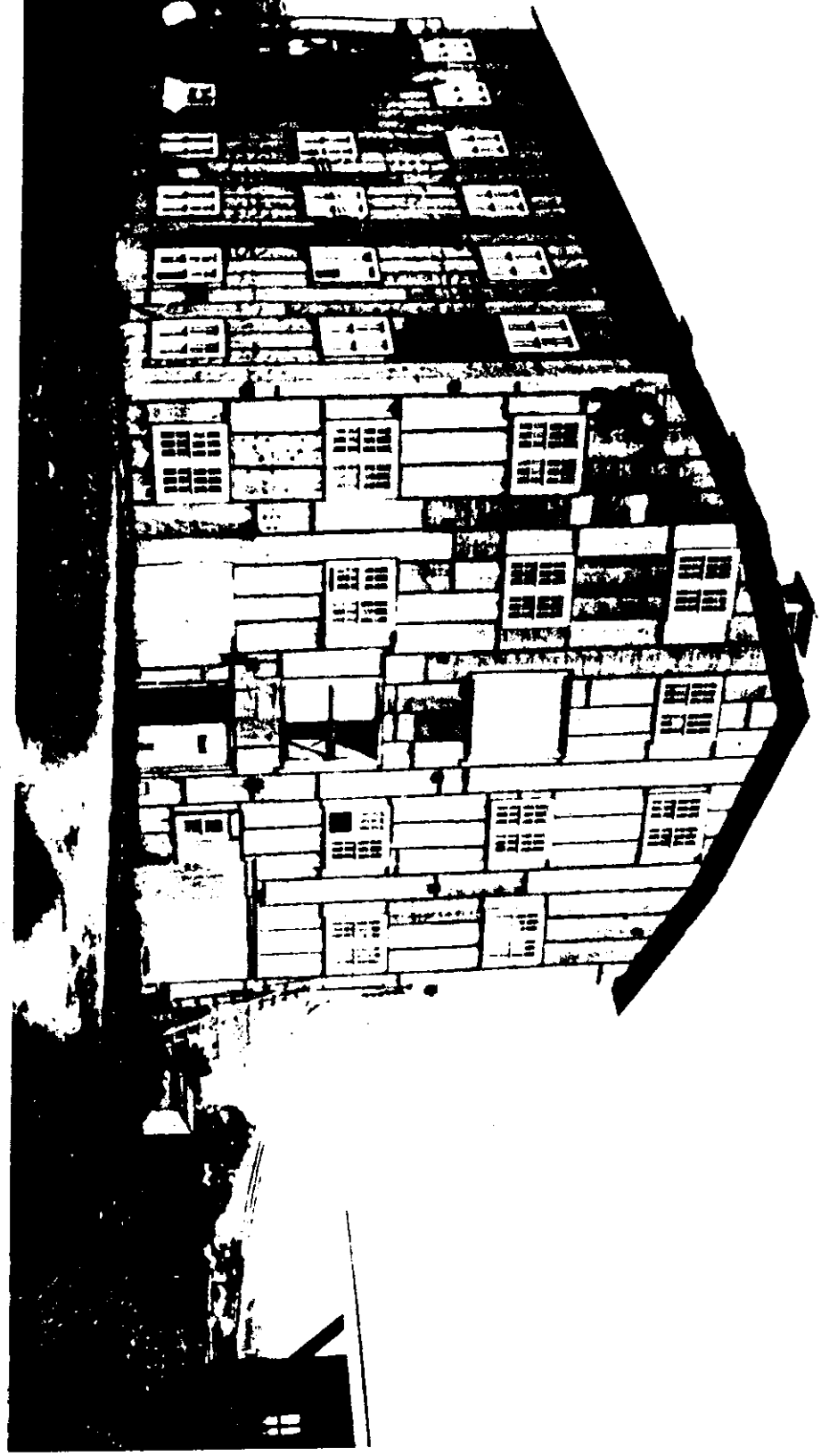

Figure 2. Mill of the Annandale Graphite Corporation in Clinton Township, Hunterdon County. Nothing remains of the original building. Photograph from New Jersey Geological Survey archives taken in 1928 by M.E. Johnson

with the construction of a 600 foot long tunnel into the hillside adjacent to one of the shafts (M.E. Johnson, 1931, unpub. notes). The rock penetrated by this tunnel apparently returned only a small quantity of graphite and the mine again closed, remaining idle from 1931 on.

The principal workings consisted of two shafts 40 to 50 feet deep (H.B. Kummel, 1928, unpub. notes). Graphite was present as disseminated plates up to 3/8 inch in diameter that were concentrated along layers parallel to foliation and adjacent to pegmatite intrusions. The width of the mineralized zone was 18 to 20 feet and the rock within it contained up to 33 percent graphite (H.B. Kummel, 1928, unpub. notes). The total amount of graphite extracted from this mine is unknown, but an unpublished document on file at the New Jersey Geological Survey lists a production figure for 1928 of 50 short tons. The original shaft, although now overgrown and nearly filled in, is still visible along a deep ravine. Several large mine dumps dot the area, but no trace of the mill was observed.

Host rocks for the deposit are locally sulfidic biotite-quartz-feldspar gneiss, feldspathic metaquartzite, and metamorphic pegmatite. A sample of metaquartzite from the mine dump contains 23.1 percent graphite (Volkert, 1997). Foliation near the Annandale mine trends an average of N45W and dips vertically.

11. Beavers' mine (Clinton Twp., High Bridge quadrangle)

This mine began operation about 1878 on the farm of Peter A. Beavers with the development of three prospect pits by the Reading Graphite Company which was leasing the property at the time. The rock reportedly contained an average of 10.09 percent graphite, and locally reached 27.82 percent from a mineralized zone approximately 20 feet wide (New Jersey Geological Survey, 1879). A mill and separator were constructed at High Bridge about 1879, shortly after the one at Annandale, but for some reason it was never fully operational. No trace of the old mine workings could be located.

Host rock for the deposit is moderately layered and foliated, locally schistose, medium-grained, rusty-weathering, sulfidic, biotite-quartz-feldspar gneiss. This is in contact with quartz-poor pyroxene gneiss that contains sparse amounts of graphite and titanite. Foliation trends about N45W and dips steeply toward the northeast.

12. "High Bridge" mine (Clinton Twp., High Bridge quadrangle)

The actual location of this mine is uncertain, but the description appears to match that of abandoned workings near Readingsburg just north of High Bridge. An attempt to mine graphite at the "High Bridge" mine in 1906 or 1907 resulted in the construction of a tunnel 400 feet long along strike of a mineralized zone 30 to 50 feet wide (New Jersey Geological Survey, 1907). Despite the dimension of the ore deposit, the host rock contained only 4 to 8 percent graphite. The mine was abandoned 7 months after opening, following the removal of just "3.5 carloads" of ore (New Jersey Geological Survey, 1907). Presently, a vertical, partially filled opening that extends upslope from the South Branch Raritan River and a few shallow pits rimmed by small dumps are all that remain. The host rock is light gray, medium-grained, feldspathic meta-quartzite. This is spatially associated with sparsely graphitic, rusty-weathering biotite-quartz-

feldspar gneiss that outcrops in the area but is absent on the mine dumps. A sample of the former from one of the dumps contains 23.2 percent graphite (Volkert, 1997).

Conover mine (Clinton Twp., High Bridge quadrangle)

Mining initially occurred about 1878 on the farm of Charles Conover with the development of three prospect pits, the deepest of which was 20 feet (New Jersey Geological Survey, 1879). These were developed in a mineralized zone of unknown width. No figures are available regarding the extent of the workings or the amount of ore removed. A chemical analysis of graphite from this locality was reported as containing 95.79 carbon and 3.6 percent insoluble residue (New Jersey Geological Survey, 1880). Mining operations here had ceased sometime prior to 1885. Host rock for the deposit is unknown owing to the fact that the Conover farm property and mine workings could not be located.

13. Hackett mine (Clinton Twp., High Bridge quadrangle)

Property on the farm of William Hackett was leased by the Reading Graphite Company about 1879 or possibly a few years earlier (New Jersey Geological Survey, 1879). This prospect was never mentioned again in any subsequent literature. Yet mining, or at least limited exploration, apparently did occur as the remains are present of a small water-filled shaft approximately 20 feet long, 6 feet wide, and of indeterminate depth that is rimmed by a small mine dump.

Host rock for the deposit is medium-grained, moderately-foliated, rusty-weathering, sulfidic, biotite-quartz-feldspar gneiss. Foliation in the area trends about N45W and dips steeply northwest.

ACKNOWLEDGMENTS

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THE GEOLOGY AND MINING OF THE STERLING HILL ZINC DEPOSIT OGDENSBURG, SUSSEX COUNTY, NEW JERSEY

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Chief Geologist (Retired)
The New Jersey Zinc Company

The Sterling Hill ore body is one of an unique pair of metamorphosed Proterozoic zinc oxide and silicate deposits of a type unknown elsewhere in the world. With its companion deposit in Franklin, three miles to the North, it made up a mineral district which was a major producer of zinc for over one hundred and fifty years.

The Ore Body

The ore at Sterling Hill is a mixture of franklinite - $(Zn,Mn)O.Fe_2O_3$; willemite - Zn_2SiO_4 ; and zincite - ZnO. These and the gangue minerals, chiefly pyroxenes and olivines, occur in various proportions in a matrix of very coarsely crystalline white marble, the Franklin marble, which has been correlated with the Grenville marble in Canada and the Adirondacks.

Generally speaking, franklinite comprises from forty to sixty percent of the zinc bearing minerals. It typically occurs as black, metallic, rounded to octahedral grains a few millimeters to several centimeters in diameter in intimate association with the other ore and gangue minerals. The metallic ions, zinc, iron and manganese, are present in the mineral in various proportions in different parts of the ore body. As a result, the magnetic properties of the franklinite - and therefore of the ore - vary from strongly magnetic (approaching that of magnetite) to palpably non-magnetic. Microscopically fine particles of the magnetic franklinite are black and opaque (indistinguishable from magnetite) while the non-magnetic particles are ruby red and transparent.

The other principle ore mineral is willemite, comparable in the variety of grain sizes and concentrations in the ore body with the franklinite. It varies in color from deep red to pale pink and from jet black to pale gray or colorless. The colors are due to the presence and relative abundance of micrometer sized inclusions of franklinite which, in various concentrations, also cause differences in the apparent magnetic properties of the willemite from place to place in the ore body. These inclusions do not appear to be genetically related to the macroscopic franklinite occurrences. They are always found associated with similarly minute inclusions of serpentine distributed much like the magnetite inclusions commonly found in serpentinized olivine.

Zincite, an orange to red mineral, comprises about six to ten percent of the ore minerals and is found almost exclusively in the brown willemite parts of the ore body.

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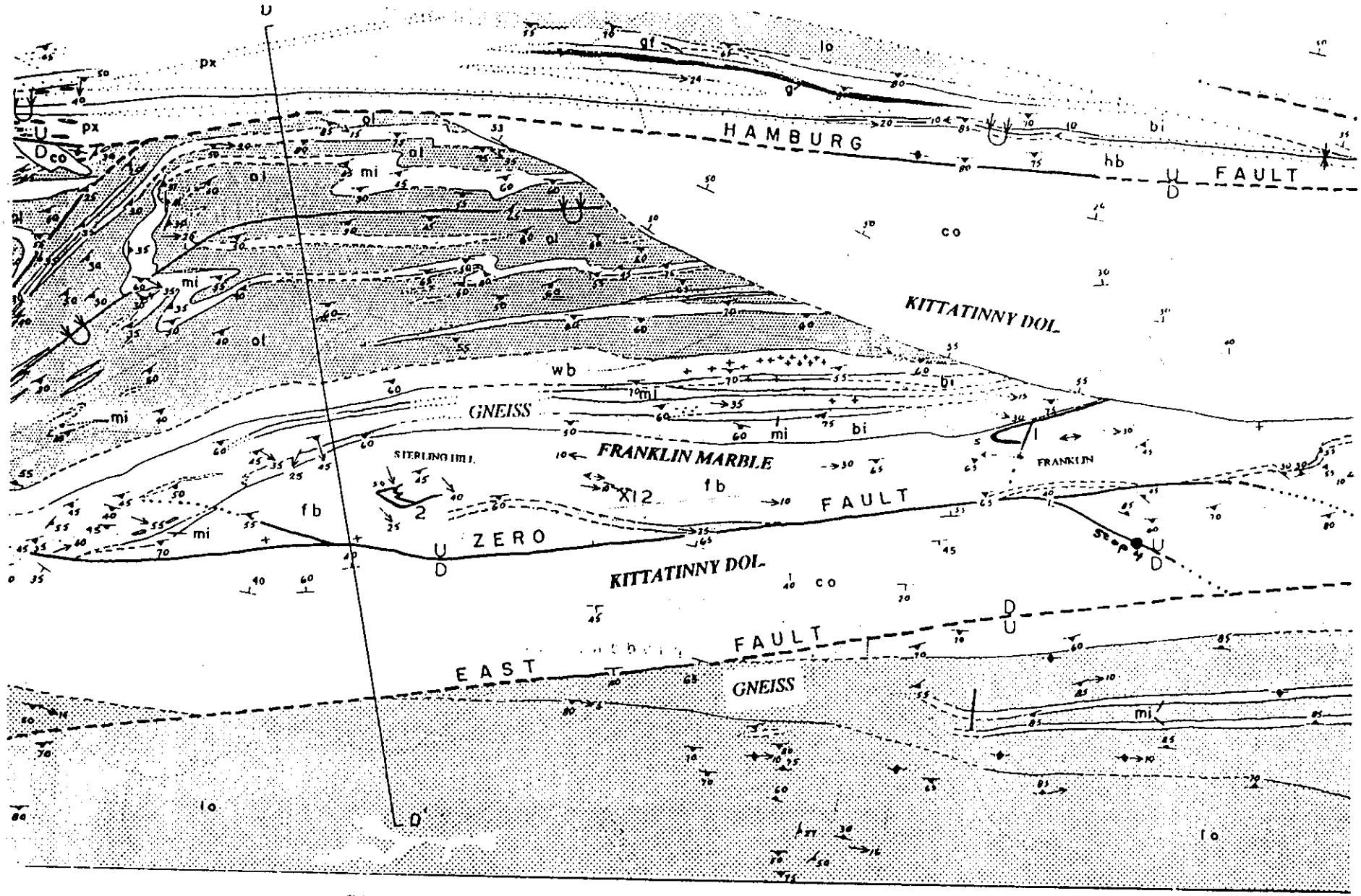
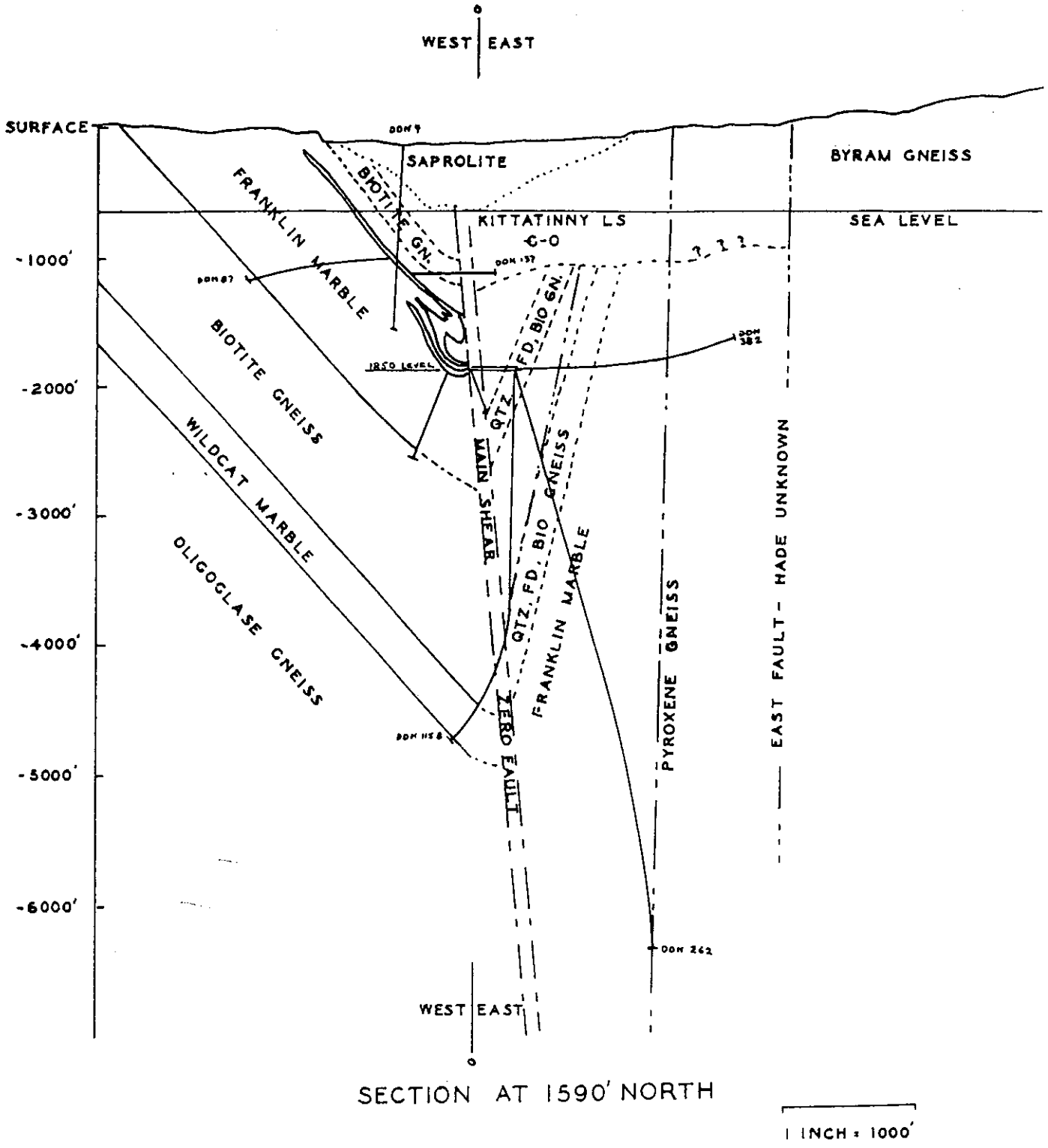


GEOLOGIC MAP OF THE FRANKLIN-STERLING HILL AREA

5,000 Feet

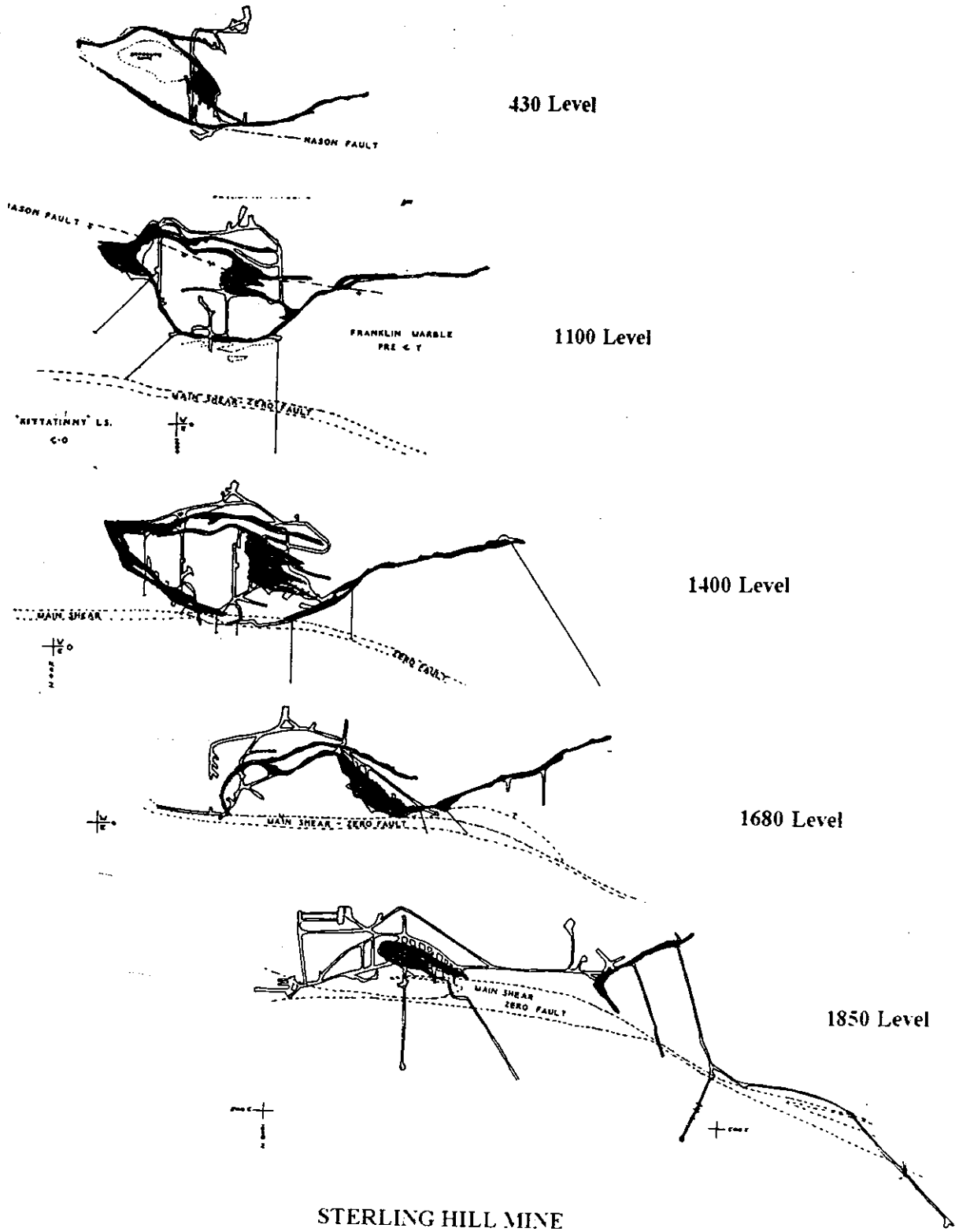
Figure 1.

From Hague et al. (1956)



STERLING HILL MINE
SUSSEX COUNTY, NEW JERSEY

Figure 2.



STERLING HILL MINE
Ogdensburg, Sussex County New Jersey
Scale: 1 inch=600 feet

Figure 3.

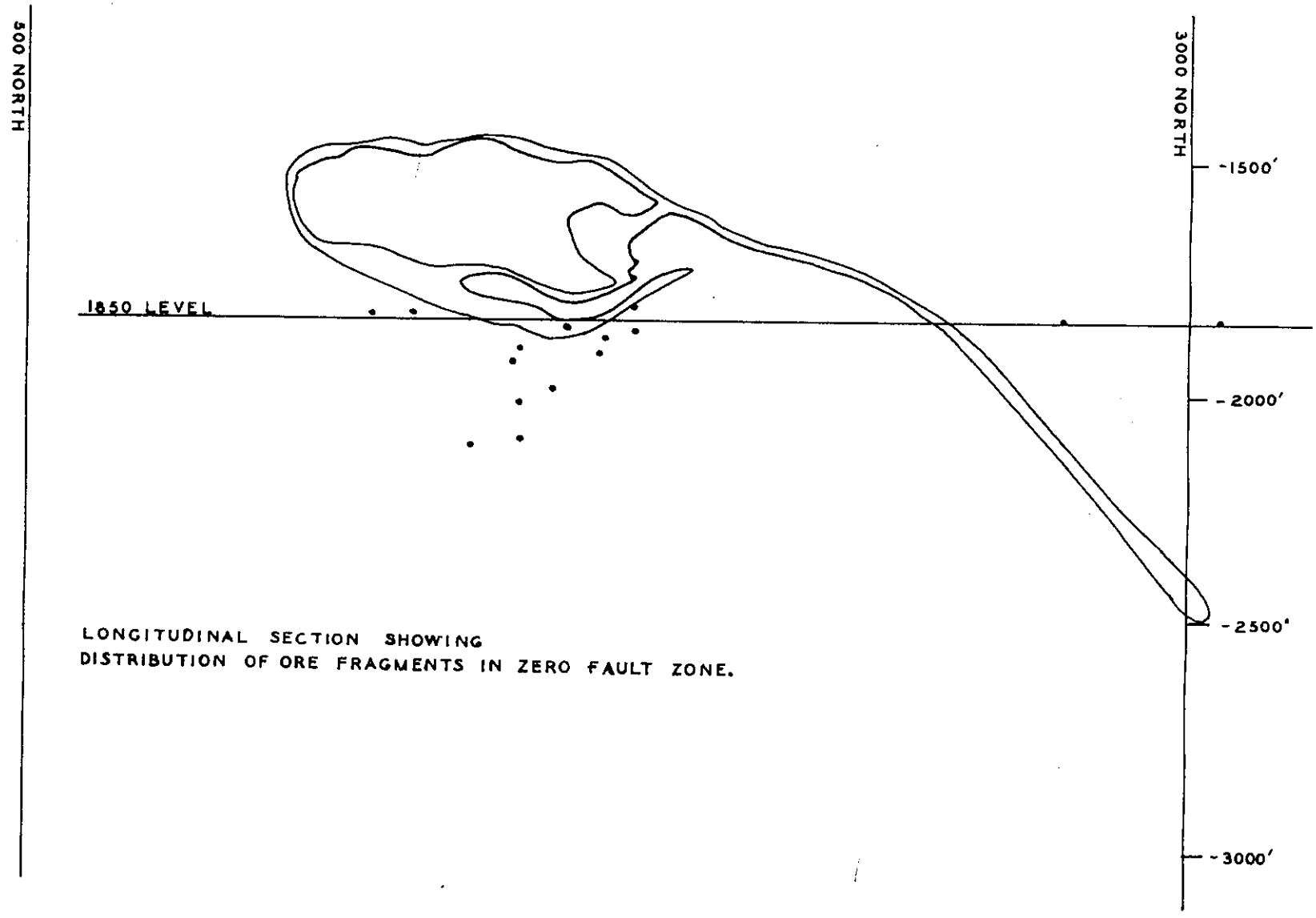

500 NORTH
3000 NORTH
1850 LEVEL
-1500'
-2000'
-2500'
-3000'
LONGITUDINAL SECTION SHOWING
DISTRIBUTION OF ORE FRAGMENTS IN ZERO FAULT ZONE.
STERLING HILL MINE
SUSSEX COUNTY, NEW JERSEY
1 INCH = 300 FEET

Figure 4

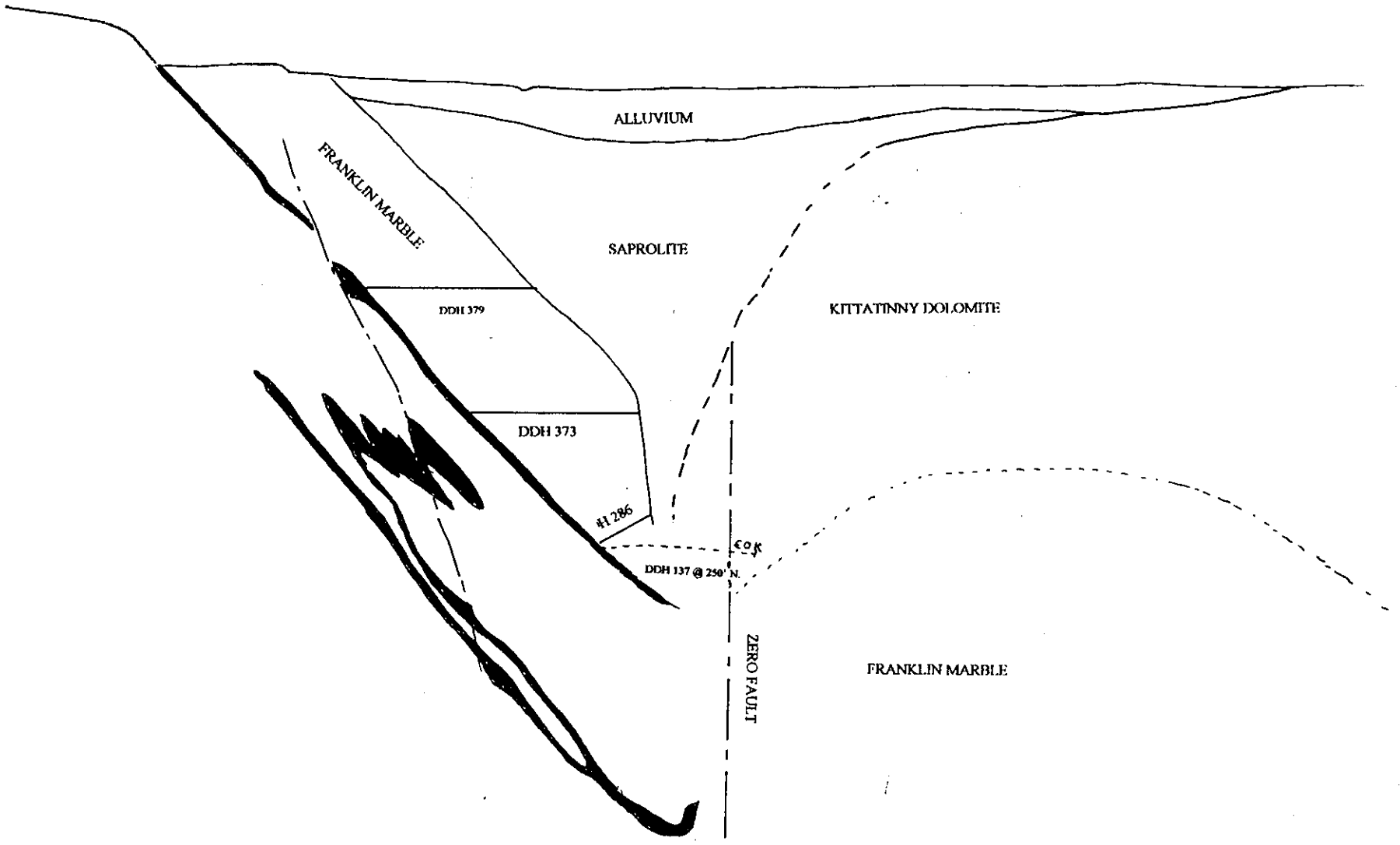
The ore and associated gangue mineral bands occur in complex isoclinal folds which are wrapped around a core of white marble containing xenolith-like blocks of amphibolite. The gross synformal structure plunges 45 degrees almost due east (geographically) while the limbs of the folds strike generally northeastward and dip fifty five degrees to the southeast. The over-all textures and fold structures suggest very strongly that the relatively dense ore body sank through the enclosing marble as an inverted diapir when the carbonate was extremely plastic or almost fluid. Age determinations on galena found in fractures in the marble indicate that the plastic stage of the carbonate, hence the imposition of the fold structures, occurred more than eleven hundred million years ago in pre-Grenville time.

The Zero Fault

Flowing northeastward from Lake Mohawk about seven miles to the Boro of Franklin, the Wallkill River occupies a valley floored by the Cambro-Ordovician dolomitic limestones of the Kittatinny supergroup. These are a minimum of 1,150 feet thick where observed at the Sterling Hill mine and over 1,900 feet thick as determined by drilling in Franklin. The carbonates are bordered on the southeast and northwest by parallel, almost vertical faults (Figs. 1 & 2) which separate them from the Proterozoic rocks on either side. The northwestern border fault is best known because of its impact on the mining operations at Sterling Hill and its contribution to the problems of exploration in the area. It has been investigated extensively in the course of mining and by diamond drilling to as much as seven thousand feet beneath the surface. Because the exposure of the fault in the mine coincides with the zero meridian of the mine survey grid, the fault has been designated as the "Zero Fault" on all modern geologic maps of the area.

As a result of its fifty-five degree easterly dip, the ore body pinches out gradually against the nearly vertical Zero Fault below the 1,500 foot level. Below that elevation, perhaps related to stresses at the time of faulting, a cross-fold which plunges northward at about 25 degrees further complicates the pattern of ore distribution on each level. The ore disappears against the fault at a depth of 2,500 foot from the surface. . While we have no idea how large the severed segment is, it is interesting to note that the dimensions of the ore body increase from the surface downward to its intersection by the fault (Fig.3) beyond which it is missing.

As exposed in the mine, the Zero Fault is marked by a vertical, or nearly vertical zone of intense shearing which ranges up to as much as twenty five feet in thickness. Its schistose layering is lubricated by heavy black smears of graphite gouge which sharply delineate it against the light gray to white rocks on either side. The softness of the shear zone accounts for the lack of natural fault exposures at the surface. Mine openings and drill hole penetrations below the ore body have encountered numerous ore fragments within the fault zone (Fig.4) indicating a substantial vertical downward component to the movement of the block east of the fault. The amount and sense of lateral movement is



STERLING HILL DEPRESSION

1 INCH = 300 FEET

Figure 5.

less clear due, in part, to a lack of data south of the mine. As there are no marker units common to both sides of the fault we can only say that the vertical displacement of the down-faulted block is a minimum of 1,150 feet (the known thickness of the Kittatinny dolomites in the graben.)

A good exposure of the Zero Fault occurs about 3 miles northeast of the Sterling mine. Here the cambro-ordovician sediments are juxtaposed to the Franklin Marble across a steeply dipping strongly mylontonized zone. Evidence of plastic flow in the marble on the northwest (left) side of the very sharply defined gouge-filled break can be seen here. At the same exposure further to the northwest is a graphic granite. How is this related to the enclosing marble?

Saprolite

Directly over the mine, separated from it to the depth of the 600'level by only 300' of marble and by less than 100' of rock at the 1200' level, is a bedrock depression filled with water-saturated mud (Fig. 5) with a north-south surface dimension of 2000' and an east-west dimension of 1000'. The soft, clayey material retains most of the original textures of the rock from which it was altered and is thus classified as a saprolite. The flattened cone-like depression bottoms at about the 1200' level. The great depth of weathering appears to be related to brecciation associated with the Zero Fault.

A second, smaller occurrence of saprolite (Fig.6) replaced the marble and amphibolite core of the ore body from the surface to a depth of 675 feet. This was well explored because it too was hazardous to the mining of the adjacent ore. This saprolite, which also had the consistency of a clayey mud and retained many of the textural and structural features of the rock from which it was derived, had an average assay of about ten percent zinc. Some masses of it assayed as high as forty percent zinc The latter areas were partially indurated with hemimorphite. The "mud" was mined in the late 1800s and early 1900s from two open pits until the water table was encountered.

Development

Although there is some evidence of mining activity in the form of scattered pits as early as 1739, mining in earnest above the 500 foot level did not begin until the latter part of the nineteenth century. In 1913, making use of old workings to a depth of about 340 feet, an inclined (56 degree) shaft was sunk through large masses of ore for development down to the 1850 foot level. To protect the shaft, it was necessary to leave a 200 foot wide pillar. This resulted in tying up a large tonnage of high grade ore.

With the benefit of hindsight it might appear to have been poor planning to locate the main production shaft where so much ore had to be tied up in supporting pillars. However, before 1913 very little was known about the amount and geologic continuity of the ore body below the 500' level. The existence of ore at depth was known only from three widely spaced borings drilled from the surface in the years 1910-1912 The deepest

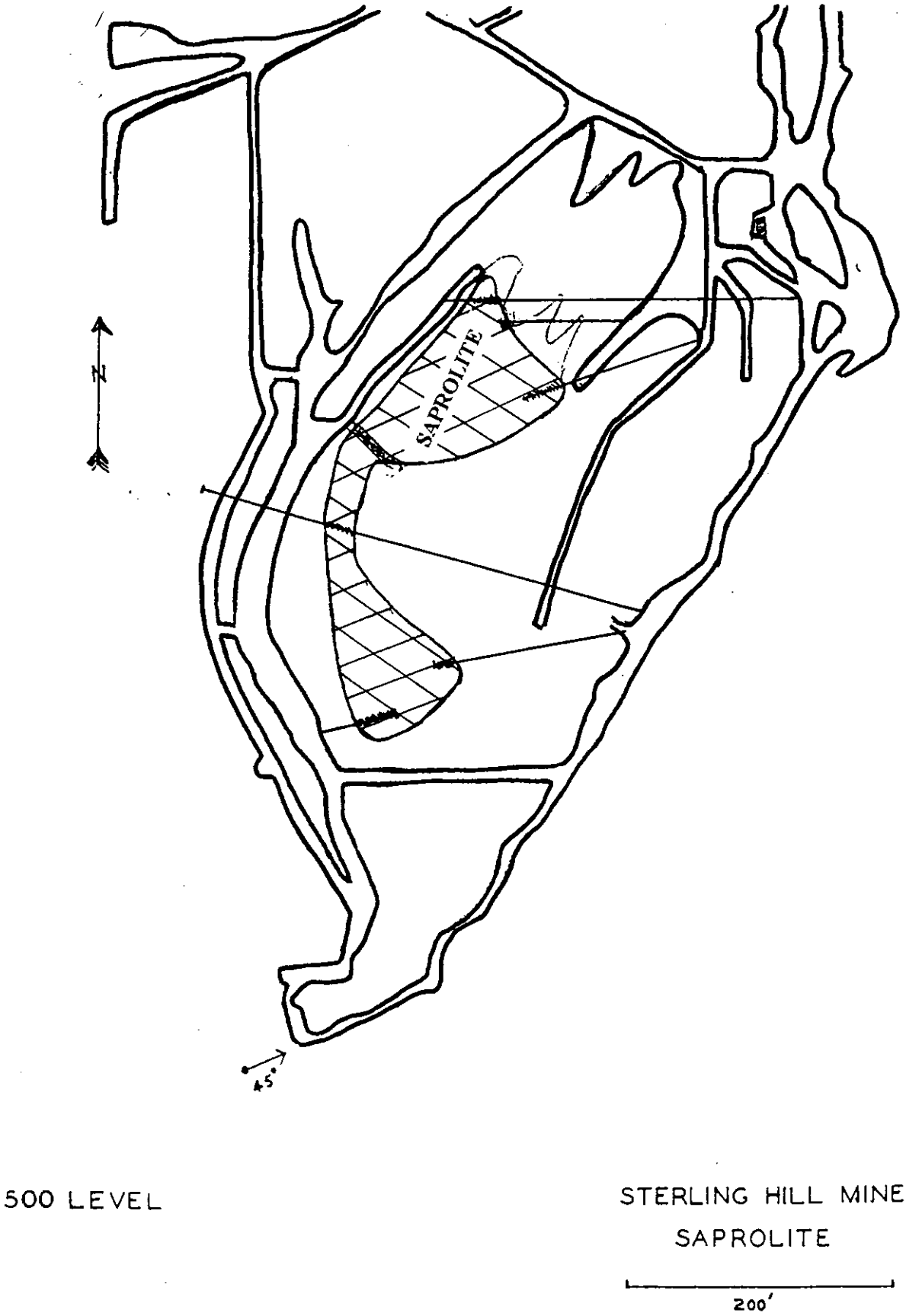


Figure 6.

ore penetrated was less than 1200 feet from the surface and there was no way of knowing how or even whether the occurrences were connected. The decision, therefore, was made to sink the shaft as close as possible to the projected trend of the known ore. From the shaft, levels were established at 100' vertical intervals by driving within the ore to its extremities, working along the ore bands where they were relatively thin (drifts) and across the ore (cross-cuts) where it was thickest in the mid-sections of the folds. This made possible a very accurate estimate of the tonnage and configuration of the deposit with the added advantage of paying for its exploration with the ore thus produced.

As mining progressed during the 1930's in the deeper portions of the ore body above the1850 level it became apparent that, because of the divergence of the trends of the Zero fault and the northern part of the East limb, a substantial tonnage of ore must exist below the that level. An exploration shaft was sunk in ore from the 1850 level to the point where the ore terminated against the fault. Exploration levels at vertical intervals of 100 feet to a depth of 2500 feet, driven in ore, indicated that a sufficient amount was present below the 1850 level to warrant development. As that extension of the ore body was separated from the main shaft by 1,200 feet of barren rock, it was necessary to sink a separate production shaft in its footwall from the 1850 Level and to establish a service level and sumps in rock at the 2550 depth. The bottom of that shaft is at 2,700 feet below the surface and is the deepest point in the Sterling Hill Mine.

Although the ore below the 1850 Level was simply a deeper extension of the East limb that part of the mine became known as the 'North Ore Body' because it had its own shaft and underground hoist. Production from that part of the mine began in 1965

Mining

The East and West limbs and parts of the Cross Member were relatively thin, from place to place ranging in thickness from as little as two feet to more than twenty five feet. In those parts of the deposit the ore was extracted by mining upward and for as much as several hundred feet horizontally along strike from a selected level to the level above (Fig. 7) . The method, known as shrinkage stoping, required that until all the ore was broken between levels only enough be withdrawn to make room for the miners to work at the up-dip limit of the excavation. Not until breaking reached the target level was the stope emptied. This method required the retention of large tonnages of ore in inventory for many months. As the broken ore was withdrawn, unpealed logs were emplaced as stulls to support the overlying marble. When all of the ore in the stope had been removed between levels, a void remained which was from 3 feet to as much as 20 feet thick, 130 feet in the dip direction and as much as 300 or 400 feet along strike. At various times over the years divers materials were used to back-fill the voids.. These included development muck, Franklin Mine tailings, glacial sand and gravel and, for a few years while it was available, electric furnace slag from the smelter in Palmerton, Pennsylvania.

After 1961 cemented fill was used, first mixed by introducing bagged cement to the flow from each carload as it was dumped. As one would expect the resulting low-grade

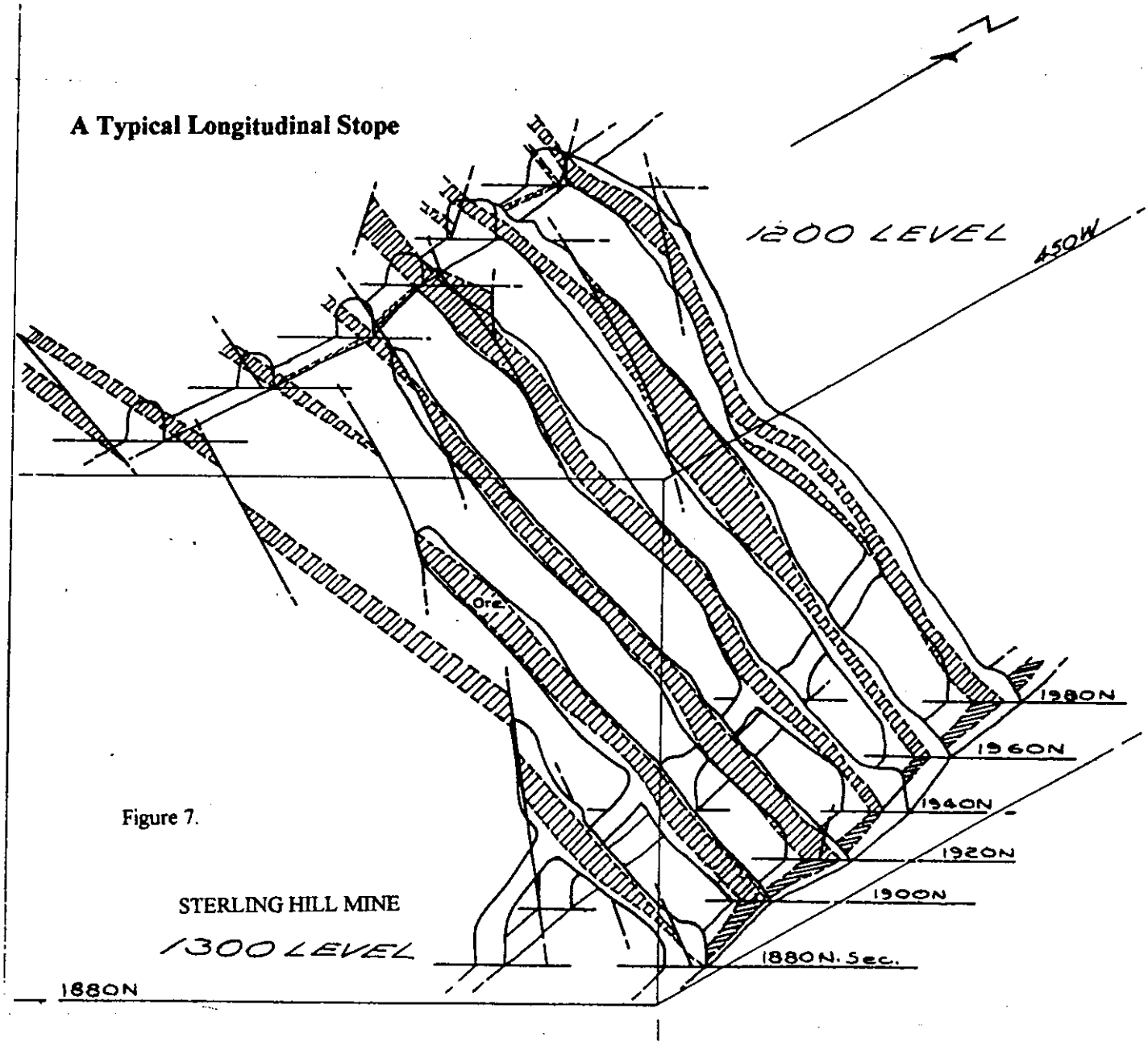
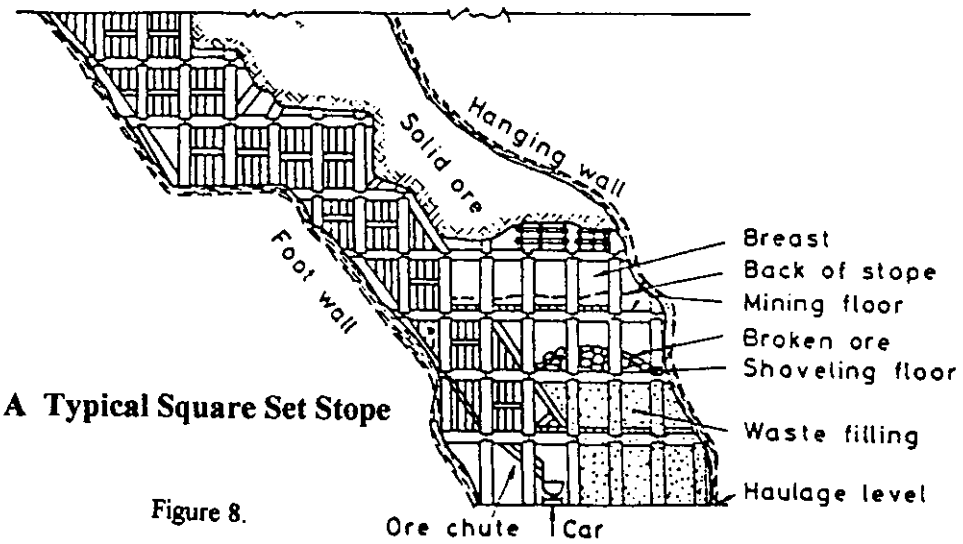


Figure 7.

concrete was anything but uniform in its properties. During the last twenty or so years of mining all fill was introduced hydraulically from a mixing station just under the surface through pipes laid throughout the mine to hand held hoses in the individual working places. This resulted in a much more uniform and controllable grade of concrete.

In the very thick middle part of the ore body a similar shrinkage method was used. However, here the stopes were transverse to the long dimension of the ore body and limited to 18 feet in width with 22 foot wide pillars left between them. Excavation proceeded from foot to hanging wall. On completion, the voids were filled from bridges constructed at the upper level. When all the stopes in a given area had been mined and filled, the pillars were removed. In the early days this was done by excavating in slices from the bottom upward. In the last twenty years or so, mining of the pillars was accomplished using square sets (Fig.8).

In 1961, after a three year shut-down for revising the mining plans, a new shaft entirely within the marble footwall of the ore body was put on line. It had been under construction since 1949 This made it possible to abandon the old shaft and permit the extraction of the huge 200 foot wide shaft pillar.



A Typical Square Set Stope

Figure 8.

Removal of the old shaft pillar had to be approached cautiously as it was located at the center of the arch formed by the eastward dipping hanging wall of East limb of the ore body and was beneath the deepest part of the overlying water-saturated saprolite. Even a very slight shifting of the hanging wall rock would be sufficient to allow large volumes of water and mud to enter the mine through open cracks. That this was a real danger was demonstrated clearly when a drill hole only 1.5 inches in diameter broke through to the mud 150 feet above the 1100 foot level. A slurry of clay and sand gushed from the hole at a rate of about 420 gallons per minute. The water pressure was 410 psi, indicating a continuous column to the surface. The pumping capacity for the entire mine at that time was only 400 gallons per minute. It took almost a week to stop the flow and seal the hole. In the mean-time the mine openings within about 100 feet of the gushing drill hole were filled waist deep in mud. Erosion by the slurry had increased the diameter of the hole appreciably.

Because of the potential danger to the mine from catastrophic flooding, concrete bulkheads with massive hemispherical doors were constructed on each level to protect the shaft and the pumping station at its bottom. Those doors at the 1850 foot level had to withstand water pressures of 800 psi. Pumping capacity was increased to 2,800 gpm although the normal rate to keep the mine dry was only 68 gpm. The centrifugal pumps delivered the water from a sump below the 1850 foot level to the surface in one lift.

At the cessation of operations in 1986 the Sterling Hill Mine had produced about twelve million tons of ore, about half the amount produced from the Franklin mine over its history. In some of the early years the grade as mined had been as low as 12% Zn. This was a result of more primitive and, to some extent, careless methods of extraction. However, subsequent to the re-designing of the of the mining procedures in 1961 the grade produced was consistently between 18.5% and 21% Zinc. This was not due to "high-grading" but rather was the result of more careful attention to the ore-rock contacts and the concomitant reduction of over-breaking.

Because of the unique mineral composition of the ore it was never possible to achieve a beneficiation product at the mine that compared in metal content with the flotation mill products of sphalerite mines. Whereas the over-all grade of ore in sulfide mines such as those in middle Tennessee might be as low as 3% zinc metal, the use of froth flotation methods at the mine makes it possible to separate out a product consisting of almost pure sphalerite. Hence, such mines are able to ship concentrates containing over 60% zinc metal to the distant smelters. At the smelter the sulfides are roasted to obtain zinc oxide.

The ore minerals at Ogdensburg were not amenable to froth flotation. Also, whereas pure sphalerite contains about 67% zinc metal, the combined ore minerals at Sterling Hill, even if it were possible to obtain a perfect separation from the gangue by the gravity and magnetic methods available, would contain only about 40 % Zinc metal. Therefore it was necessary to ship the relatively low grade mill product to the smelter at Palmerton, Pennsylvania, seventy miles distant, for further concentration of the metal. There it was combined with large quantities of anthracite from the nearby coal region. The mixture was charged to a 300 foot long cylindrical rotating kiln (a Waelz kiln) where the zinc was driven off as a metallic fume and immediately oxidized by a current of air flowing over the bed. The resulting low grade zinc oxide was collected in a bag house whence it was drawn for further processing in the manufacture of pure grades of zinc metal and zinc oxide.

Only after the completion of the roasting of the high grade sphalerite concentrates and the more expensive waelzing of the relatively low grade Sterling Hill product could the value of the unique Sterling Hill ore be compared with that of the common sulfide ores. For this reason, although the concentration of metal in the ground at Sterling Hill was much higher than in most other zinc districts in the world, its economic value was not always competitive with that of much lower grade zinc sulfide deposits. The additional energy required for processing and the necessity for transporting a relatively low grade mill product negated much of the economic advantage one might have expected in the mining of such a high grade mineral deposit.

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GENESIS OF THE MARBLE-HOSTED ZINC DEPOSITS AND IRON DEPOSITS AT STERLING HILL AND FRANKLIN, NJ, AND COMPARISONS WITH GNEISS-HOSTED IRON DEPOSITS AT CORNWALL, NY

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Introduction

The zinc+iron+manganese deposits at Sterling Hill and Franklin, New Jersey and the magnetite iron deposits of northern New Jersey-southern New York show obvious similarities. (1) The ores in both deposit types are massive. Sterling Hill and Franklin ores locally reach some 70 wt% zinc+iron+manganese, and at many of the iron deposits grades can locally exceed 70 wt% iron. (2) The ores in both deposit types are predominantly oxide minerals rather than sulfide minerals. (3) The geometry of the deposits can generally be interpreted to have resulted from isoclinal folding of bodies that were originally tabular or stratiform (Skinner and Johnson, 1987), although it is difficult to rule out on structural arguments alone the possibility that the deposits are replacements of already-folded strata.

Despite their similarities, the zinc+iron+manganese deposits and the iron deposits have not traditionally been considered together in formulating genetic hypotheses. The reasons for this are two-fold. First, there is a large chemical and mineralogical gap between the two deposit types. There are no known zincian or manganoan iron deposits that lie between the zinc+iron+manganese deposits on the one hand and the zinc- and manganese-free iron deposits on the other hand. Second, many of the investigations at the zinc deposits have been motivated by an interest in the diversity and quality of mineral specimens found therein. Because the mineralogy of the iron deposits is less interesting, these deposits have received less study generally by different investigators than those that have studied the zinc deposits.

The purpose of this paper is to present a review of selected evidence bearing on the genesis of the marble-hosted zinc+iron+manganese and iron deposits. Comparisons will also be made with gneiss-hosted iron deposits at Cornwall, NY. The Cornwall deposits are representative of many of the gneiss-hosted iron deposits in northern New Jersey and southern New York, but there is sufficient diversity among the iron deposits that the Cornwall occurrences should not be taken as representative of all of them.

For any given deposit, it is possible if not likely that the ores were affected by the Grenville metamorphism and deformation at about 1.0 Ga. As a result, the mineral assemblages that are observed today may well be metamorphic and not be the original assemblages that formed at the time of metal emplacement. This has been shown to be the case at Sterling Hill (Johnson et al., 1990a). The strategy in this review is to focus on features of the deposits which are likely to have been inherited from the protoliths, in particular features which reflect physicochemical parameters such as the oxidation and sulfidation states, or features which reflect the sources of chemical constituents such as the isotopic compositions of carbon, sulfur, and oxygen. At

In Benimoff, A.I., and Puffer, J.H., (editors), The economic geology of northern New Jersey: Field Guide and Proceedings of the fourteenth annual meeting of the Geological Association of New Jersey, 1997, p 47-59

Sterling Hill, it has been shown that some of the ores have gained or lost chemical components during the Grenville metamorphism, mainly along metamorphic veins and fractures that penetrate the deposit (Johnson et al., 1990a, Makovicky and Skinner, 1990). There are sections of the ore zones, however, which show no evidence of metamorphic veins and for which chemical and isotopic data are consistent with the hypothesis that they behaved largely as closed systems on the hand specimen scale during the metamorphic event (Johnson et al., 1990a). These ores are the most reliable indicators of the chemical and isotopic properties of the protoliths and it is observations of these unveined ores that are discussed below.

The Marble-Hosted Zinc Deposits and Marble-Hosted Furnace Magnetite Bed

The Sterling Hill and Franklin zinc+iron+manganese deposits and the Furnace magnetite bed are located within the Franklin Marble, a member of the paragneiss sequence within the Proterozoic Reading Prong. The Sterling Hill deposit comprises a sequence of continuous strata with the form of a north-plunging synform. The ore layers are at the base of the sequence, and they are composed of willemite+franklinite+calcite±zincite. Calc-silicate+calcite strata overlie the ore layers. The Franklin deposit is mineralogically and lithologically similar to Sterling Hill except that there are numerous pegmatites that have crosscut the ore and calc-silicate units creating skarn assemblages at their margins. The Furnace magnetite bed immediately underlies and is similar in form to the Franklin zinc deposit. The Furnace bed is wholly within the Franklin Marble and contains calcite as the major gangue mineral. The proximity of the iron and zinc+iron+manganese ores has long been thought to be evidence of similar origins (Frondel and Baum, 1974); the results of recent work supports this hypothesis (Johnson et al., 1990b). For thorough descriptions of these deposits and their surrounding rocks, the reader is referred to Metsger et al. (1958, 1969), Metsger (1962), and Frondel and Baum (1974).

There are three lines of evidence indicating that the original emplacement of the zinc+iron+manganese at Sterling Hill and, by analogy, at Franklin took place prior to the main Grenville deformation and metamorphism at 1.0 Ga. The first line of evidence is structural and consists of the fact that the orebody is composed of continuous strata that show the same mineral lineations and isoclinal folding as the enclosing host rocks (Metsger, 1962; Metsger et al., 1969). Individual strata can be quite thin and laterally continuous. One distinctive fluorite-bearing layer within the ore body has been traced some 600 meters down the plunge of the synform (Metsger, unpubl. data). The second line of evidence is petrologic and consists of the fact that minerals in the deposits have been shown to have formed at high temperatures indistinguishable from the regional Grenville metamorphic temperatures (Mason, 1947; Carvalho and Sclar, 1988). The third line of evidence is geochronologic and consists of the fact that potassium-argon age dates of Sterling Hill zincian biotites are indistinguishable from the Grenville dates that have been obtained from non-mineralized gneisses both in close proximity to the zinc deposit and elsewhere in the region (Johnson et al., 1990a).

For ores that behaved as rock-dominated systems during the Grenville metamorphism (see discussion in Johnson et al., 1990a), the present-day oxidation and sulfidation states reflect those of the protoliths. The stabilities of relevant minerals and assemblages as functions of oxygen fugacity (fO_2) and sulfur fugacity (fS_2) are shown in Figure 1. The absolute values of fO_2 and fS_2 are not directly applicable to the protolith, but it is clear that the willemite+franklinite± zincite-bearing rocks reflect higher oxidation states and/or lower

sulfidation states than the pyrrhotite- or pyrite-bearing Franklin Marble wall rocks. The Franklin Marble is a metamorphosed marine limestone, so the environment in which the protoliths for the ores were formed was one of a higher oxidation state and/or lower sulfidation state than typical marine limestone.

The Furnace magnetite bed contains minor amounts of pyrite in addition to magnetite. The assemblage magnetite+pyrite is stable over relatively small ranges of fO_2 and fS_2 (Fig. 1). The values of fS_2 are substantially higher than those required to stabilize willemite+franklinite or zincite, so the Furnace bed reflects metal deposition in a more sulfidizing environment than that in which the zinc ores were formed. The relative oxidation states of the two types of ore are not constrained by the mineral equilibria shown in Figure 1.

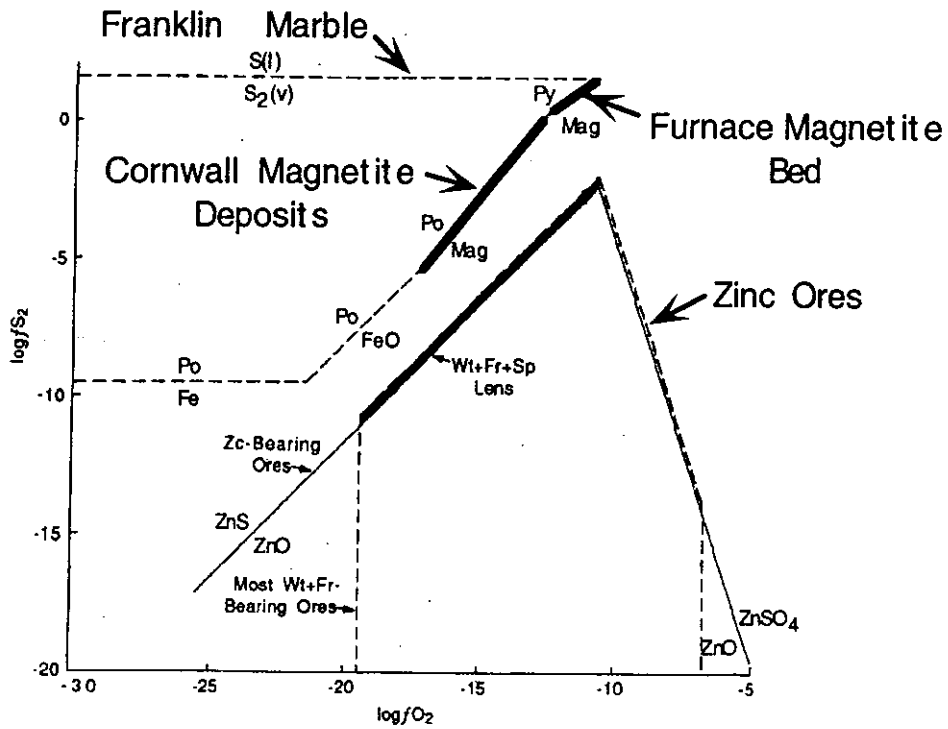

Figure 1. The stability of ore minerals and assemblages as functions of fO_2 and fS_2 at the inferred Grenville metamorphic conditions of 5 kbar and 1000K. Modified from Johnson et al. (1990a).

The fugacities of the halogens fluorine and chlorine are also of interest because they too may serve as a basis for comparing the different deposit types. The common ores at Sterling Hill and Franklin contain neither fluorine- nor chlorine-bearing minerals, so the fugacities of these species cannot be evaluated. The Furnace magnetite bed, however, contains fluorite. Also, tremolitic amphibole with 3 wt% F and less than 0.1 wt% Cl (Fig. 2) has been found coexisting with massive magnetite in drill core from near the keel of the Franklin zinc deposit west of the Parker Shaft. There is some question whether this drill core intersection is actually the Furnace bed and not a separate magnetite layer within the zinc deposit. In either case, because the rock is composed of massive, zinc-free magnetite with minor calcite it is as representative of marble-hosted iron deposits as the Furnace bed. The assemblages in the magnetite layer(s) do not place useful constraints on f_{HF} and f_{HCl}, but the amphibole analyses imply a relatively high f_{HF}/f_{HCl} ratio and provide a basis for comparison with the gneiss-hosted iron deposits at Cornwall (see below).

Another feature expected to have been minimally changed during the Grenville metamorphism is the isotopic composition of sulfur contained in the rocks. Sulfide minerals are not common constituents of the zinc+iron+manganese ores, but there are occurrences within the Sterling Hill deposit that appear from textural evidence to have been part of the metamorphic assemblage and therefore to represent sulfur that was contained in the protolith (Johnson et al., 1990a). The isotopic compositions of representative samples (Ault, 1957) are shown in Figure 3. The existence of $\delta^{34}S_{CDT}$ values as low as -10‰ and the broad range of values spanning some 23‰ are strong evidence that the origin of the sulfur in the protolith was H_2S generated by bacterially-mediated reduction of sulfate. There are no sulfur isotopic data from the Furnace bed with which to compare the Sterling Hill data.

A third feature of interest at the deposits is the isotopic composition of carbon and oxygen contained in calcite, the major gangue mineral in both the zinc ores and the Furnace bed. The carbon isotopic compositions are shown in Figure 4. Sterling Hill ore and calc-silicate calcites lie at 0.8±1‰, a value typical of the Franklin Marble host rock and of marine limestones of late Proterozoic age worldwide. The Furnace bed calcites are systematically lower in $\delta^{13}C$ with a mean value of -7‰. Similarly low values have been observed in carbonate-hosted iron formations of both Algoma- and Superior-types (eg., Becker and Clayton, 1972). The consensus is that the values reflect isotopically light organic-derived carbon, and that carbonate deposition took place in a basin that had restricted communication with the open ocean.

The oxygen isotopic compositions of Sterling Hill and Furnace bed calcite are shown in Figure 5. Both ore types are in the range 11 to 16‰. Of particular interest are the compositions of pure calcite layers that are interbedded with the zinc+iron+manganese-bearing strata. The compositions of these layers at both Sterling Hill and Franklin fall at 15‰, substantially below the 20-25‰ range characteristic of Franklin Marble far from the deposits and of late Proterozoic marine limestones worldwide. If these layers have retained their premetamorphic $\delta^{18}O$ values (see discussion in Johnson et al., 1990a), then they constrain the temperature of their formation and the $\delta^{18}O$ value of the water from which they were deposited. The solution is not unique, but an oxygen isotopic and chemical mass balance model applied to the Sterling Hill ore layers fixes

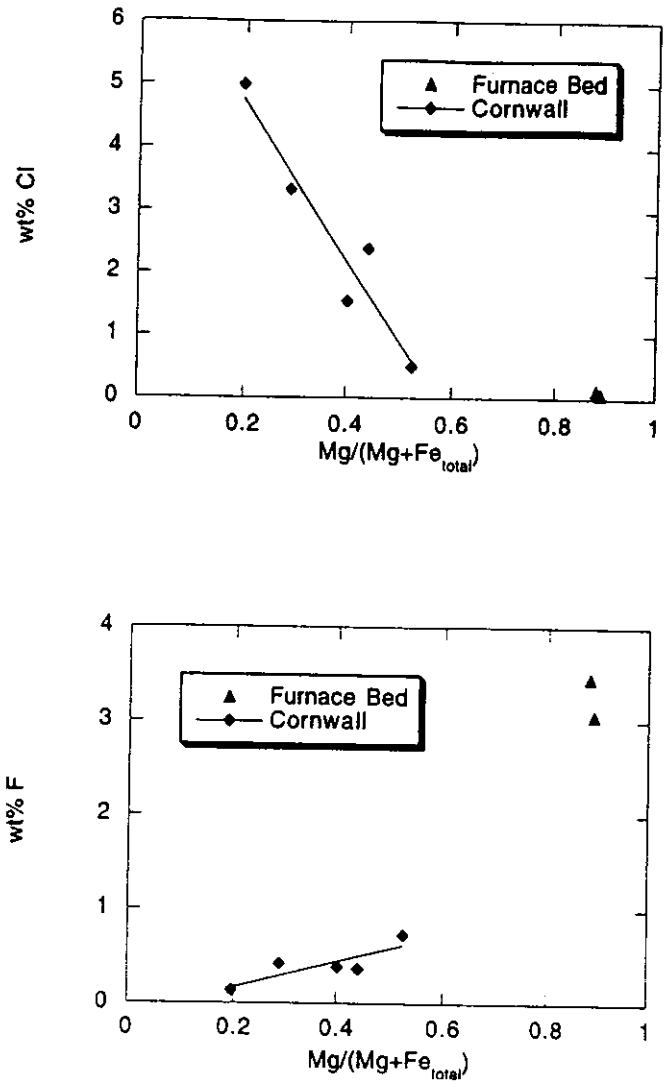

Figure 2. The fluorine and chlorine contents of Furnace magnetite bed (see text) tremolitic amphiboles and Cornwall magnetite deposit pargasitic amphiboles plotted as a function of Mg/(Mg+Fe$_{total}$). Cornwall data are from Leger et al. (1996).

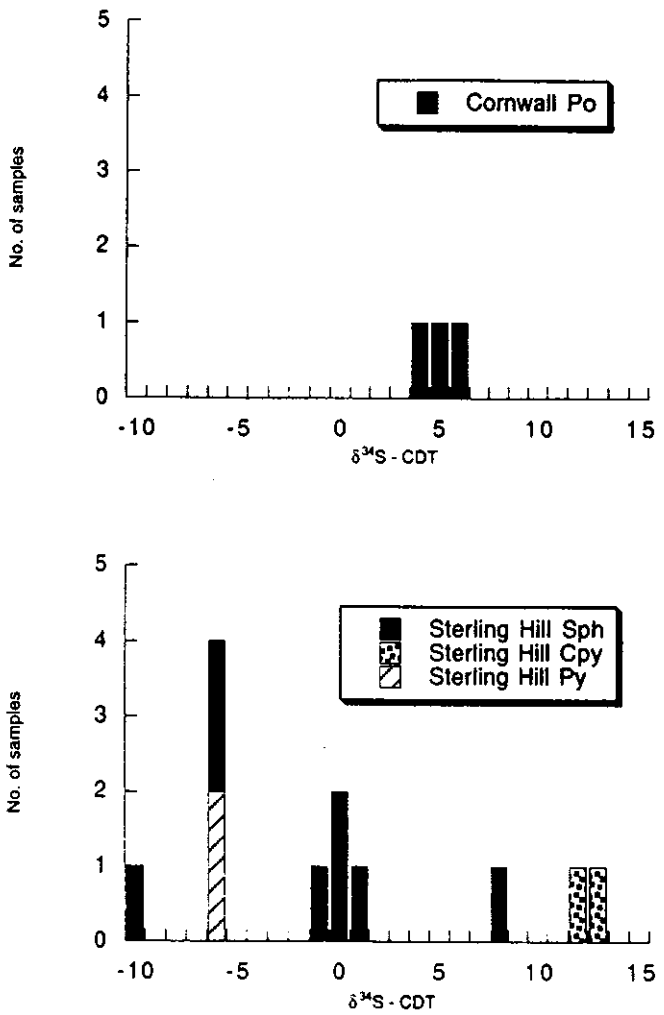

Figure 3. Histogram of sulfur isotopic compositions of Sterling Hill sulfides (from Ault, 1957) and Cornwall pyrrhotite. The compositions are expressed in delta notation where $\delta^{34}S$ (in ‰) = $1000*(R_{SA}-R_{ST})/R_{ST}$, and R_{SA} and R_{ST} are $^{34}S/^{32}S$ for the sample and Cañon Diablo Troilite (CDT), respectively.

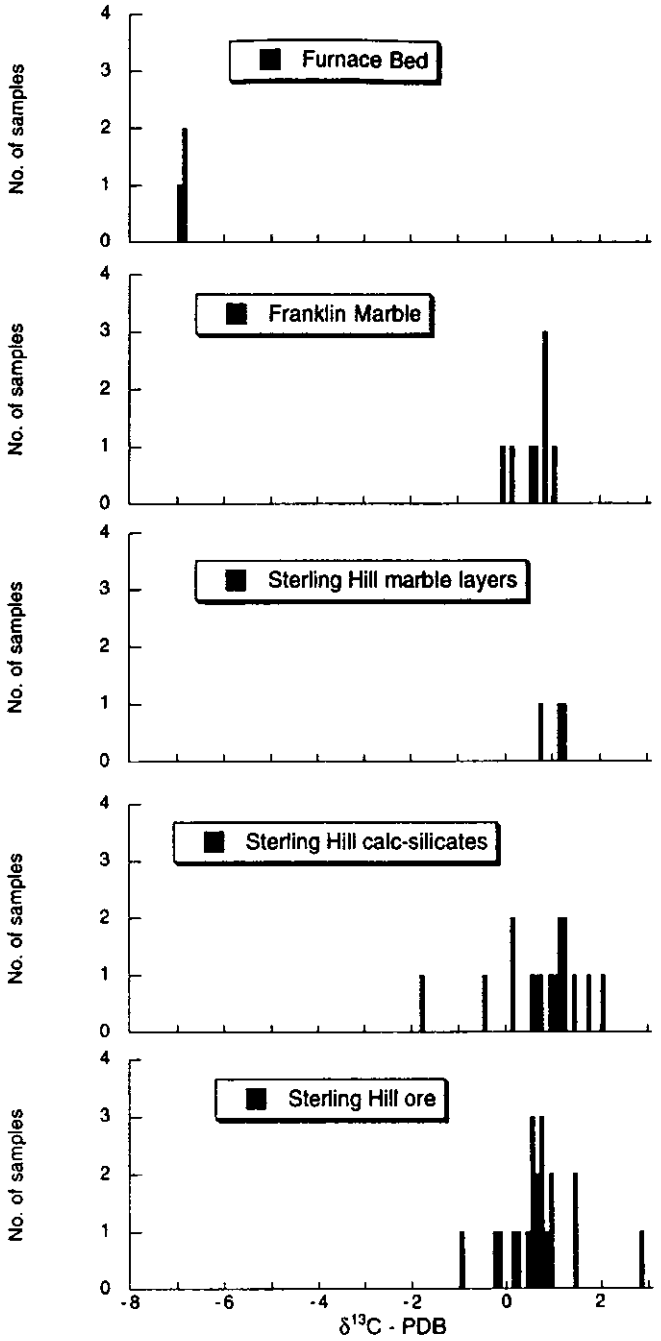

Figure 4. Histogram of carbon isotopic compositions of calcites. The compositions are expressed in delta notation where $\delta^{13}C$ (in ‰) = $1000*(R_{SA}-R_{ST})/R_{ST}$, and R_{SA} and R_{ST} are $^{13}C/^{12}C$ for the sample and Vienna Pee Dee Belemnite (VPDB), respectively. Data are from Johnson (1990).

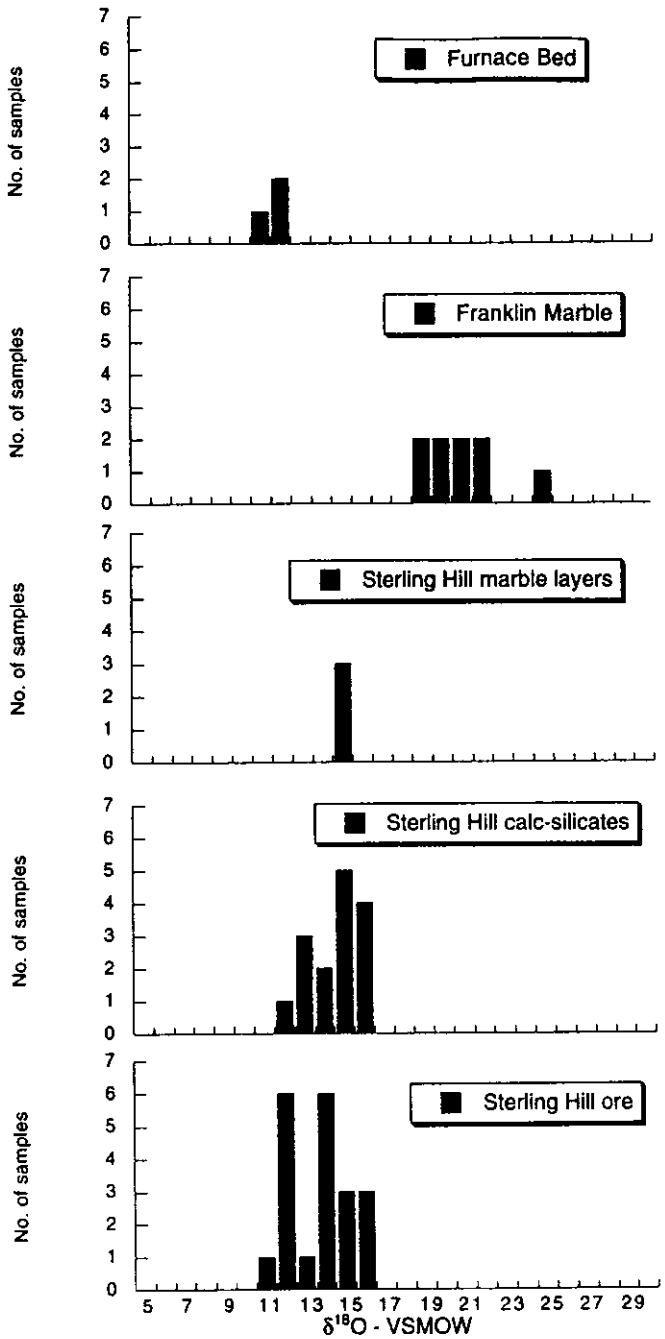

Figure 5. Histogram of oxygen isotopic compositions of calcites. The compositions are expressed in delta notation where $\delta^{18}O$ (in ‰) = 1000*(R_{SA}-R_{ST})/R_{ST}, and R_{SA} and R_{ST} are $^{18}O/^{16}O$ for the sample and Vienna Standard Mean Ocean Water (VSMOW), respectively. Data are from Johnson (1990).

the temperature of zinc+iron+manganese deposition at 150±50°C and the $\delta^{18}O$ value of H_2O in the metal-depositing fluid at a value several permil larger than the value of zero characteristic of pristine seawater. Details and assumptions of the model are given by Johnson et al. (1990a). The same modelling method has yet to be applied to the Furnace bed.

The Gneiss-Hosted Magnetite Deposits at Cornwall, NY

The Cornwall magnetite deposits are located just west of the Hudson River some 10 kilometers south of the city of Newburgh. The deposits are exposed in outcrop and in prospect pits; they are too small to have been of economic interest at the time that other similar deposits in the area were mined. The host rocks for the deposits are strongly banded plagioclase±biotite±clinopyroxene±hornblende+quartz gneisses and kspar+quartz+hornblende granite gneisses. At one locality, well-banded magnetite gneiss occurs at the contact between biotite+plagioclase+hornblende gneiss and a large body of granite. At other localities, the magnetite-rich rocks are pegmatoidal. The pegmatoidal occurrences appear to be less-deformed than the banded occurrences and are surrounded by outcrops of granite. Geologic maps of the occurrences, the results of a ground magnetic survey, and petrographic descriptions of representative rocks are contained in a Hunter College independent study project report that was authored by Askold Chemych and Cheryl Eisenberg in 1991.

At Cornwall, as at many of the magnetite deposits of northern New Jersey and southern New York, it is difficult to determine whether the original emplacement of the iron predated or postdated the Grenville metamorphism. The ground magnetics data show clear evidence of the same isoclinal folds with NNE-trending axial traces that are observed in the outcropping banded gneisses. However, this observation could be interpreted as reflecting either deformation of a preexisting iron formation or metasomatic replacement of an already-deformed horizon or layer. Less-deformed magnetite-bearing pegmatites have been cited elsewhere as evidence of post-folding iron emplacement (Foose and McLelland, 1995), but they cannot be taken as definitive proof because the possibility exists that the pegmatites are partial melts of preexisting iron formations that were formed within the thermal aureoles of the late tectonic granites (Johnson, 1996).

Leaving aside the relative timing question, it is instructive to look at some of the same features that have been described above for the marble-hosted deposits. The oxidation and sulfidation states of the Cornwall magnetite-rich rocks are shown along with the marble-hosted ores in Figure 1. Pyrrhotite is a minor but common mineral in the banded rocks, and its presence indicates more sulfidizing conditions and/or less oxidizing conditions than for the zinc+iron+manganese ores at Sterling Hill and Franklin. The Cornwall deposits, however, lie at lower fO_2 and lower fS_2 than the Furnace magnetite bed (Fig. 1), so the range of conditions at the marble-hosted deposits encompassed those at the gneiss-hosted deposits.

The fluorine- and chlorine-contents of pargasitic amphiboles from the Cornwall rocks have been determined by Leger et al. (1996). Figure 2 illustrates that whereas fluorine contents are low, chlorine contents are quite high. In fact, the coexisting biotite in a sample studied by Leger et al. (1996) has one of the highest chlorine contents ever reported for the mineral. The difference in fluorine contents of the Furnace bed and Cornwall amphiboles can be partly attributed to crystal chemical effects (cf., Morrison, 1991), but the differences are larger than can be accounted for by crystal chemistry alone. The data imply lower f_{HF}/f_{HCl} during iron emplacement at Cornwall than at the Furnace bed.

The isotopic composition of sulfur in typical Cornwall samples is shown in Figure 2. The data are few, but they suggest uniform compositions at about 5‰. The compositions are more suggestive of a magmatic origin for the sulfur than for an origin by in situ bacteriogenic

reduction of sulfate (Ohmoto, 1986). However, a definitive conclusion for the origin of the sulfur will require a more thorough study.

CONCLUSIONS

For the marble-hosted zinc ores, there are several lines of evidence suggesting that metal deposition took place in a marine sedimentary environment. The sulfur isotope data suggest in situ bacterial reduction of sulfate in a diagenetic environment, and the carbon isotope data suggest precipitation from bicarbonate in marine waters. The high oxidation state - low sulfidation state character of the ores is consistent with the hypothesis that seawater controlled these two parameters during ore formation. A modern analog of the protoliths for the Sterling Hill and Franklin ores, first proposed by Callahan (1966), may be the sulfide-poor strata that are forming today beneath ponded brine pools at the bottom of the Red Sea (eg., Pautot et al., 1984). Iron formations have also been recognized in the Red Sea, and these may be analogs for the Furnace magnetite bed. The change from zinc+iron+manganese deposition to iron deposition in the New Jersey occurrences involved a shift toward more reducing conditions, presumably anoxia, and a change in hydrologic conditions so that the overlying basin waters became restricted in their communication with the open oceans.

The gneiss-hosted iron deposits at Cornwall show neither the high oxidation state - low sulfidation state nor sulfur isotopic evidence for a sedimentary environment. The ratio f_{HF}/f_{HCl} is also lower than at the marble-hosted deposits. Despite the fact that the $\delta^{34}S$ values of 5‰ and the apparent high f_{HCl} could both be construed as evidence for a magmatic component in the ore-forming system, it would be premature to conclude that there is no genetic link between the marble-hosted and gneiss-hosted deposits. There remain fundamental unanswered questions for both deposit types. The recognition of a seafloor depositional environment for the marble-hosted deposits sheds little light on the source of the hydrothermal fluid, the source of the metals, and the source of the heat that drove hydrothermal circulation. All three could have been supplied by magmas at shallow levels beneath the Proterozoic seafloor. On the other hand, the lack of obvious evidence for a sedimentary depositional environment at Cornwall may simply reflect an environment of hydrothermal discharge different from that at Sterling Hill and Franklin. Whether the two deposit types shared similar sources for hydrothermal fluids, metals, and heat remains an unanswered and important question.

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A SKETCH OF THE LIME CREST QUARRY, THE FRANKLIN MARBLE AND SOME INTERESTING MINERALS

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The Lime Crest quarry is located at the northwestern edge of the New Jersey Highlands in Sparta Township, Sussex County, NJ, near the center of the Newton East quadrangle, GQ-1707 (Drake and Volkert, 1993). The quarry has been a significant component in the economy of the County and several surrounding municipalities for most of this century. Limestone Products Corp. of America was founded by a group of local businessmen shortly after the First World War and operated the quarry until the mid 1970's. During the past 25 years the operation has had a series of outside Corporate owners. The latest ownership change occurred in early 1997 when Medusa Minerals acquired the property.

The Lime Crest quarry is centered on a small area underlain by a thick, massive Precambrian marble that has customarily been correlated with the Franklin Marble. The Franklin Marble is the only mineable calcite-rich carbonate unit in the region between the mid-Hudson valley of New York and the Lebanon - Lancaster area in Pennsylvania. Until the tide of suburbanization began flowing in earnest in the 1960's Sussex County and adjacent Warren and Orange Counties had large agricultural industries, primarily dairy. The quarry was located in the heart of this agricultural area and was the principal supplier of pulverized limestone, hydrated lime, poultry grits and other products to this large market.

Over the past 40 years there have been major changes in the market for limestone products. Hydrated lime production ceased in the mid-1970's in the wake of the Arab oil embargo and subsequent escalation in fuel cost. Pulverized limestone now goes largely to the lawns of suburbia rather than corn fields. A plant to pelletize pulverized limestone was recently put on line. Pelletization is an attempt of make the application of pulverized limestone to lawns more effective by reducing wind loss and redistribution by rain. The bulk of production goes into crushed stone of various sizes and is consumed by the construction industry. Because of its white color and its coarsely crystalline nature the marble is widely used in a variety of architectural and landscaping applications. It is also processed into a white sand for use in white portland cement concrete most commonly seen in curbs and highway center barrier.

General Geology

The Lime Crest quarry was originally begun in a body of Franklin Marble that outcropped on a hillside overlooking a marsh. The upper slopes of the hill were underlain by gneiss and the marsh by Cambrian age dolomite. Maps by Drake and Volkert (1993) and Herman and Monteverde (1989) show a thrust fault passing just west of the quarry site beneath the marsh and glacial deposits. Cross sections accompanying both maps indicate that the thrust dips at a moderate angle beneath the quarry. The Precambrian rocks and a thin veneer of early

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Cambrian Hardyston and Leithsville Formations have been thrust over the Leithsville and onto the late Cambrian Allentown Formation.

As mapped by Hague, *et al.* (1956) and Drake and Volkert (1993) the marble outcrop area is lens shaped in plan, approximately 6000 ft. x 1300 ft. Part of the marshy area along the northwest side of the marble outcrop was filled and leveled to create a plant site and stockpile yard. The filled area also serves as a barrier between the quarry and the surface water drainage that occupies the adjacent valley. After construction of the plant and yard the mineable outcrop width of the marble was approximately 750 ft. Constrained by noncarbonate rocks to the southeast and abundant surface water to the north, the quarry took on the form of an elongated pit.

At present the quarry is approximately 3000 ft. x 1200 ft. x 300 ft., the long dimension oriented nearly parallel with strike. For most of its history the Lime Crest quarry produced only crushed marble products. However, since the early 1970's, expansion of the pit toward the southeast has required the removal of increasing amounts of gneiss. The gneiss is quarried separately and processed through a separate plant.

The increasing depth of the marble producing part of the main pit has increased both haulage and water pumping costs. Within the past 15 years, since the reconstruction of County Road 669, some work has been done to expand the pit northeastward into unlined areas. However, progress toward this expansion has been slow and most of the mining effort continues to be focused downward. During the spring of 1997 the new owner, Medusa Minerals, diamond drilled at least three holes to 150 feet beneath the lowest level of the quarry. Observation of the core showed that the marble thickness beneath the quarry exceeds that tested by the drill.

The northwestern-most quarry excavation, just north of the access road to the upper benches, has exposed a fault contact between the Franklin Marble and the black dolomite of the Leithsville Formation. At this locality the contact is defined by a northwest dipping zone of brecciation at least 20 feet wide. The breccia zone there is extensive vein filling and replacement of the Franklin Marble by secondary dolomite and minor quartz. The similarities between the secondary dolomite and the Buckwheat dolomite (Peters, et al., 1983; Cummings, 1988), in Franklin, are unmistakable. No evidence of the Hardyston quartzite has been seen.

Toward the southeast the marble is overlain by a metasedimentary sequence very similar to that present in the Franklin-Sterling Hill area. Immediately overlying the marble is a thick sequence of microcline gneiss and amphibolite. The second prominent marble band in the region, the Wildcat Marble, outcrops approximately 2600 feet east of the mining area.

The Franklin Marble

The marble exposed at the Lime Crest quarry is typical of that seen in the main outcrop belt of the Franklin Marble less than 2 miles to the northeast. The marble at Lime Crest is white to light gray, medium to very coarsely crystalline calcite with subordinate dolomite. The analysis of 10 bulk samples of various crushed products done over a period of several years indicates dolomite contents ranging from a few to over 90%. The insoluble component generally ranges between 0.5 and 8%. Higher insoluble residues indicate the incorporation of pegmatite or other silicate rock into the plant feed. The Franklin Marble is the most intensely studied of all the Precambrian units in the Highlands. The descriptions given by Hague, *et al.* (1956), Offield

(1967) and Kearns (1977) of the Franklin Marble in its main outcrop belt apply to the marble at the Lime Crest quarry.

The marble contains a variety of noncarbonate minerals. The most abundant are graphite, chondrodite/norbergite, phlogophite, tremolite, pyrrhotite, scapolite and spinel. Pyroxene is conspicuously absent except in the skarn surrounding pegmatite. Most of the silicate minerals and spinel occur in discontinuous bands that are conformable with both the thin, but continuous, bands of gneiss within the marble and the contact with the overlying gneiss. Long before a quarry was established mineral specimens were collected from the marble outcrops. In 1844 the 5th edition of *Phillips's Mineralogy* listed "Newton" as a locality for corundum, rutile, spinel, sphene and tourmaline. At the time Newton covered a much larger geographical area and although it did not include the quarry site its boundaries came within a few hundred yards. It was also the nearest established settlement. The minerals listed in the 1840's are those that occur most frequently as well formed crystals. As such they are the species that have been most avidly sought by collectors up to the present day.

Pegmatite and gneiss are common in the marble at Lime Crest although they generally account for a relatively minor part of the material mined. Pegmatite usually occurs in lenses up to a few feet thick and a few 10's of feet in maximum dimension. However, at least one large pegmatite in the southwest section of the quarry has been intersected by mining repeatedly for at least 25 years. Pegmatite bodies are rimmed by a pyroxene-rich skarn. The skarn usually forms a thin, irregular band around the outer margins of the pegmatite pods. However, as noted by Kearns (1977) in the Amity, NY area, the reaction rim can be so thick as to consume most of the pegmatite. These skarns may contain numerous euhedral, reddish brown sphene crystals that occasionally exceed 1 inch. In some areas of the Franklin marble the skarns associated with pegmatite and other siliceous gneisses contain large euhedral crystals of diopside and scapolite but such crystals are very rare at Lime Crest.

The Lime Crest quarry contains a thin but prominent layer of hornblende gneiss. This layer is exposed as a series of pods, apparently conformable, that almost certainly are boudins of a formerly continuous bed. In hand specimen the hornblende gneiss is fine grained, massive and is not discernibly layered. In most places the contact between the hornblende gneiss and the marble is sharp and devoid of a reaction skarn assemblage like those that surround more siliceous lithologies. Locally, the contact is marked by pods of margarite [$CaAl_2(Al_2Si_2)O_{10}(OH)_2$], and ferroan dolomite. Margarite, a brittle mica, is normally found in very aluminous environments. It is most often encountered as an alteration product of corumdum. In the Franklin Marble margarite is uncommon but has been found locally in assemblages where corundum is either absent, as in the occurrence in the hornblende gneiss, or is a distinctly separate phase. Dunn and Frondel (1991) interpreted an extensive margarite-bearing horizon in the Sterling Hill area as representing a highly aluminous bed within the marble.

The similarities between the Franklin Marble of the main outcrop belt and the marble at Lime Crest extend to a more detailed level. As in the main outcrop belt of Franklin Marble discontinuous bands and pods of silica deficient silicate minerals are very abundant. In many of these bands chondrodite/norbergite, as yellowish or orange-brown grains thickly dispersed in marble, is the only significant non-carbonate mineral. In many other cases chondrodite/norbergite is accompanied by phlogopite, spinel and, far less commonly, amphibole.

Chondrodite [$Mg_5(SiO_4)_2(F,OH)_2$]/norbergite [$Mg_3(SiO_4)(F,OH)_2$]are two separate species in the humite group. Each has been found as a distinct entity in the Franklin marble but

most often they occur as composite crystals of norbergite over a chondrodite core. They are indistinguishable in hand specimen and so are treated here as one. As abundant as chondrodite/norbergite is at Limecrest and throughout the Franklin marble euhedral crystals are extremely rare. Between 1994 and 1996 a considerable number of well formed crystals up to 1.5" were recovered from a group of boulders. The crystals apparently were limited to one local band of very white, relatively fine grained marble adjacent to a chondrodite/norbergite band of normal texture. The very white marble contained numerous euhedral chondrodite/norbergite crystals, a few discrete flakes of graphite and a few crystals of spinel.

The abundance of chondrodite/norbergite reflects the pervasive presence of significant amounts fluorine throughout the Franklin Marble. Jones, *et al.* (1969) indicated that high fluorine content, relative to hydroxl, was characteristic of chondrodite and norbergite and necessary to stabilize them relative to forsterite. Forsterite occurs in the Franklin Marble but is quite uncommon. Kearns, *et al.* (1980) has shown that relatively high fluorine contents are also characteristic of other hydrous silicates that are abundant in the Franklin Marble. They suggest that fluorine has stabilized some of these minerals and allowed them to persist at severe metamorphic conditions estimated at $836^{o}C+/-40^{o}C$ and 5 to 7 Kbar.

Many of the silicate assemblages are aluminous and contain phlogopite, spinel and, less commonly, corundum. Spinel is locally abundant at Lime Crest where it generally occurs as granular masses and small octahedral crystals. The most common color is dark grayish purple but reddish crystals also occur. Lime Crest is not one of the premier spinel localities in the Franklin Marble. Spinel crystals larger than 1/4" are uncommon and those exceeding 1/2" are rare. The largest individuals found in recent years have been approximately 1" on an edge. Kearns (1977), working around Amity, NY in the most prolific spinel area in the Franklin Marble, noted that phlogophite was a universal associate of spinel. He proposed that the reaction of dolomite and muscovite produced phlogophite, calcite and the alumina that resulted in spinel. Although phlogophite is a common associate of spinel at Lime Crest there are many examples where it does not appear to be sufficiently abundant to account for the spinel present.

Corundum is one of the minerals most sought after by collectors at the Lime Crest quarry and has been found many times. It typically occurs in calcite pods within lenses composed predominantly of phlogopite. The corundum is typically gray with sapphire blue spots and zones. It is almost always associated with small, black, prismatic crystals of rutile. The sapphire blue color is usually intense, but volumes within crystals large enough to produce cut stones are rare. Very rarely, corundum occurs as the ruby variety. In the one example seen by the author the ruby corundum crystallized with red spinel in a small calcite lens in massive, light green amphibole. Unlike the typical corundum-bearing assemblage phlogophite was a very minor component.

Unfortunately, the distribution of corundum occurrences has not been mapped. In the author's experience, covering the past 27 years, most of the corundum discoveries appear to have come from a single conformable horizon extending the length of the quarry, a few tens of feet beneath and conformable with the hornblende gneiss layer. The distribution of lenses of corundum-bearing rock suggest that, like the segmented hornblende gneiss, they may be boudins of a once continuous layer. At present the zone that has produced corundum is not well exposed in active bench faces. Most of its outcrop length is currently in the quarry floor or in debris covered, inactive areas.

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Another characteristic of the Franklin marble are small but widespread concentrations of boron. Because of its distribution and unique set of properties, boron may provide important information concerning the depositional environment of the Franklin Marble (Moore and Swihart, 1990). Tourmaline, mostly uvite [$CaMg_3Al_5Mg(OH)_4(BO_3)_3(Si_6O_{18})$], is distributed throughout the marble as individual crystals and local groups of crystals. Tourmaline is sometimes associated with other silicate minerals, especially scapolite. Frequently, however, tourmaline crystals are found isolated in marble with no significant amount of other silicate minerals nearby. What is frequently seen near these isolated tourmalines are very sparsely scattered arsenopyrite crystals. In spite of being widespread tourmaline is only rarely part of an assemblage distributed more extensively in bands or boudins.

The rare mineral fluoborite [$Mg_3(BO_3)(F,OH)_3$] was first recognized in the 1970's (Kearns, 1975, 1977) as a primary mineral in the marble at 2 localities: Amity, NY and Rudeville, NJ. The Franklin Marble immediately became the World's premier fluoborite locality. Fluoborite is very difficult to recognize in the field because it forms white, vitreous grains that blend with calcite on fresh surfaces and are non-descript on weathered ones. Largely because it is fluorescent moderately bright cream in ultraviolet light, fluoborite has since been recognized at several localities, including Lime Crest. It may be widespread in small amounts throughout the Franklin Marble. At Lime Crest, and at all but one of the other localities where it has been recognized, fluoborite is the overwhelmingly dominant boron species. Unlike uvite, fluoborite is typically found in bands, up to more than three feet thick, where small, thickly scattered grains make up as much as 60% of the rock (Moore and Swihart, 1990). Fluoborite-rich bands may either be isolated or adjacent to and slightly overlap a band of chondrodite/norbergite grains. Other boron minerals have been found with fluoborite. At the Rudeville locality there were minor amounts of kotoite [$Mg_3(BO_3)_2$], sinhalite [$MgAlBO_4$], and warwickite [$(Mg,Fe)_3TiB_2O_8$]. At Amity fluoborite was subordinate to warwickite. To date only fluoborite has been identified at Lime Crest but its mode of occurrence was similar to that seen elsewhere (Steve Misiur, personal communication).

The contact between the Franklin Marble and the overlying microcline gneiss is abrupt in the sense that the initially encountered contact is sharp and there is no extensive transition through calcsilicate gneiss. In a particularly accessible outcrop that existed during 1994-5 least 3 thin, sharply bounded, conformable bands of coarsely crystalline carbonate occurred above the contact within the lower 20 feet of the gneiss. At the initial contact the top one inch of marble contained small, scattered pyroxene grains. Immediately above the marble, and separated from it by a band of pyroxene less than half an inch thick, was a band of granitic rock one to two inches thick. The granitic material was overlain abruptly by more finely crystalline, very well layered, biotite-bearing , microcline gneiss.

The microcline gneiss is generally well layered. Much of the exposed portion of this unit is rich in biotite and layers containing large anhedral to subhedral, maroon crystals of garnet are common. Occasionally, with some imagination, one can make out very crude faces on the garnet crystals.

Microcline gneiss now makes up approximately half of the outcrop area on the pit's southeast side. Because of the moderate southeast dip the gneiss will continue to be exposed on

progressively lower benches as the quarry expands in that direction . Tree cutting, stripping and other preliminary development tasks are currently underway along the crest of the ridge preparing the area for mining.

The gneiss is quarried separately and processed through a separate plant located near the northeast corner of the pit and known as the "granite plant." The "granite plant" is modest in size, approximately 300 tons/hour capacity, but typical of crushed stone producing plants in its operation. It crushes the muck blasted from the upper quarry benches and screens the material into several types of construction aggregate, each having a standard grading. Adjustments in the size and quantity of specific products can be altered by changing screen sizes, crusher settings and recycling circuits. The name given this subsidiary operation is instructive since it emphasizes the strong tendency in the construction industry to call almost all crystalline rocks "granite" or "granite gneiss" regardless of their composition or origin.

Epigenetic Mineral Assemblages

Both the Precambrian rocks of the Highlands and those of the adjacent Great Valley experienced hydrothermal activity during the Paleozoic. In the Precambrian rocks the effects of hydrothermal alteration are limited to fractures and their immediate vicinity. The best examples of secondary hydrothermal mineral development occur along the northwest margin of the Highlands near the contact between the Precambrian and Paleozoic rocks. In most cases deformation within fracture zones created no open space. Alteration is limited to the recrystallization of fault gouge to fine grained chlorite and epidote (Hull *et al.*, 1986). Locally, significant open space does occur in fractures, enough to allow the development of more extensive, interesting and better crystallized mineral assemblages.

The hydrothermal assemblages that occupy open spaces within fractured Precambrian rocks along the northwest margin of the Highlands are of two types: Mississippi Valley-Type (MVT) lead-zinc and Alpine cleft. The mineral assemblages and depositional processes characteristic of each type of mineral deposition are quite different, yet in the region that includes the Lime Crest quarry both types occur in close proximity and in some cases are superimposed on one another.

The mineralogy of MVT deposits is simple: some combination of the common carbonates and sulphides, plus barite and fluorite. Copper is never more than a minor component and is often absent. MVT ore deposits appear to result from the migration of deep sedimentary basin brines to favorable, shallow structures near basin margins. The brines are thought to leach metals from the sediments through which they pass during migration. MVT mineral deposits range from very small curiosities to orebodies containing tens of millions of tons. The chemistry of the deposition site is fluid dominated and the epigenetic mineral assemblage is dominated by introduced components. Because of their economic importance there is a voluminous literature concerning MVT mineral depositing systems. If interested start with Anderson (1991), Anderson and Macqueen (1982), Bethke (1986), Plumlee, *et al.* (1994), and Sverjensky (1984). Although they are among the most intensely studied and best known hydrothermal systems many paradoxes still exist (Spirakis and Heyl, 1996).

Alpine cleft deposits, best known from the central massif in Switzerland, are of interest principally to mineral collectors. Clefts have produced superbly crystallized examples of both common, rock-forming minerals and rarer species. The lack of industrial-scale economic

incentive means that the literature concerning most facets of this type of mineral occurrence is far less focused than that of MVT deposits. A good introductory summary is given by Weible (1966). Briefly, clefts are open spaces that can be of any size. The flux of circulating groundwater is very sluggish, compared with MVT systems, and the fluid chemistry within the cleft is rock dominated. The minerals that crystallize on the walls of the cleft derive all or most of their components from the adjacent countryrock and reflect its composition.

The Highlands probably host many minor Alpine cleft type mineral occurrences. Most contain only small, unexceptional crystals quartz, feldspar, epidote, amphibole and pyrite. However, some clefts in the Highlands, particularly those found in calcsilicate gneisses along its northwestern edge, have contained more interesting species. Several examples have been found, scattered between Bethlehem, PA to Fishkill, NY, that contain ferroaxinite (Cummings, 1983). In 1995 ferroaxinite was found at Lime Crest in very narrow fractures cutting a pegmatite. Most of the best documented and most diverse cleft assemblages have been found in gneisses closely associated with the Franklin and Sterling Hill zinc deposits and the nearby iron mines. Betancourt (1989) described one such example. Another assemblage, rich in sulphides of copper and other metals but with many similarities to Alpine clefts, was recently described from within the Sterling Hill ore body (Jenkins and Misiur, 1994). The metal sulphide-rich nature of this paragenesis, unusual in Alpine cleft environments, may simply reflect the unusual chemistry of the countryrock. The number of cleft localities known in the Franklin-Sterling Hill area probably reflects the intensity of mining and attendant study and collecting.

In the Franklin Marble examples are known in which a later MVT paragenesis is superimposed on an Alpine cleft assemblage. The best known is the Buckwheat dolomite in the Franklin mine (Cummings, 1988). Similarly superimposed assemblages were recognized at the Lime Crest quarry in 1985 and are discussed below. The Buckwheat dolomite was described by Palache (1935) as a vein-like body. It is a secondary dolomite with numerous small vugs containing an extensive suite of minerals including albite, chlorite, microcline, muscovite and rutile (Peters, *et al.*, 1983). Abundant coarsely crystalline calcite and sphalerite were the last minerals to deposit. They fill small fractures and the remaining open spaces in dolomite cavities and, although widespread, are not ubiquitous.

MVT mineral assemblages are widespread in the Franklin Marble including the Lime Crest area. Hague *et al.* (1956) note the occurrence of sphalerite and galena at the Indian Mine, just east of the Lime Crest quarry. The quarry has since expanded past this locality and exposed a fracture zone approximately 20 feet below, and conformable with, the gneiss/marble contact. In exposures that have evolved during the past five years this fracture zone is flanked by an extensive bleached halo and contained one of the best example of MVT mineral deposition that has been recorded from northern New Jersey.

The extent of this fracture is difficult to determine because either much of its outcrop is on inaccessible quarry faces or is obscured by quarry ramps and debris dumps. There is circumstantial evidence, pieced together from fragmental material encountered during mineral collecting by members of the Franklin-Ogdensburg Mineralogical Society (FOMS), that this fault extends the length of the pit and beyond.

During the dozen years prior to 1992 secondary fracture-filling minerals were frequently found in the southern two thirds of the pit. Most of the finds were along the pit's southeast side in large boulders pushed from higher benches onto the quarry floor and appeared to be very limited in extent. In the typical specimen small masses of fluorite or sphalerite were embedded

in secondary carbonate. In some cases the secondary carbonate was calcite that mimicked the marble country rock so closely that the contact between secondary and primary carlcite was difficult to distinguish. Occasionally, the veins were wide enough to contain cavities lined with free standing crystals.

The most extensive and mineralogically diverse of the fracture-filling assemblages that were collected along the pit's southeast side was encountered in 1985. Several hundred tons of rubble were found, near the southern end of the pit, in which the Franklin Marble was fractured and replaced by fine grained yellowish tan dolomite. The size of the boulders suggested that the fault zone source of the material was at least several feet thick. The dolomite contained numerous lenses of vuggy quartz up to several inches long and 1.5 inches thick. There were also numerous flat, very narrow cavities lined with crystalline dolomite. The dolomite cavities also contained crystals of rutile, quartz, and muscovite. Small anhedral crystals and masses of yellow sphalerite were deposited late in the paragenetic sequence and are associated with both dolomite and quartz. Groups of hemimorphite crystals were common in some cavities. The assemblage and paragenetic sequence are similar to that of the Buckwheat dolomite in Franklin. The rubble pile also contained fragments of massive sphalerite, partially oxidized to hemimorphite, in fractured Franklin Marble.

During 1992, quarrying exposed a section of the fault zone approximately 150 feet long and 10 feet thick that contained an anastamosing network of fissures and was flanked by a wide bleached halo. In this section of the fault the fissures contained a well developed MVT mineral assemblage (Cummings, 1993). The MVT assemblage exhibited features including vugs, colloform and cockscomb textures, well layered mineral sequences, and massive sulphide/fluorite that indicate deposition in abundant open space. It was possible to find the entire basic MVT assemblage (dolomite, calcite, pyrite, sphalerite, galena, fluorite and barite) in a single hand specimen. Subsequent quarrying has shown that the concentration of sulphides, fluorite and barite was elongated nearly parallel to strike and had a very limited downdip extension. In outcrops viewed in May, 1997 the fault zone was still marked by bleaching but the fissures contained only secondary dolomite and, locally, very sparce pyrite and fluorite.

Summary

The Lime Crest quarry has been a significant economic force in Sussex County for nearly 80 years. The quarry was once mainly a producer of agricultural lime and pulverized limestone but has evolved into a large producer of landscape, roofing and architectural products. The rare combination of white color and very coarse crystallinity put the quarry in an advantageous position to enter these markets. The operators have taken advantage of the resource available in the microcline gneiss to become one of the area's larger suppliers of construction aggregate and railroad ballast. Huge reserves of both lithologies will sustain production for many years.

The Franklin Marble at Lime Crest has been of interest to mineralogists and mineral collectors since the 1830's. It is possible that the key to understanding the Franklin-Sterling Hill ore deposits lies in the Franklin Marble. The most promising avenue toward deciphering the depositional environment of the Franklin Marble is boron geochemistry. Studies are underway to assess the distribution, mineralogy and isotopic fractionation of boron in the marble (Moore and Swihart, 1990). Because the Lime Crest quarry is the largest exposure of the marble, is

known to contain fluoborite-rich bands and is also collected regularly it will continue to be an important source of mineral material of both scientific and esthetic interest.

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GENESIS OF THE NEW JERSEY HIGHLANDS MAGNETITE DEPOSITS ON THE BASIS OF GEOCHEMICAL AND STRUCTURAL EVIDENCE

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ABSTRACT

Magnetite ore samples from 37 of the approximately 300 abandoned iron mines in the Proterozoic rocks of the New Jersey Highlands were chemically analyzed and the structural relationships to their host rocks were noted. The magnetite of nine of the sampled mines is hosted by coarse grained pyroxenite / amphibolite and occurs as disseminated ilmeno-magnetite (1 to 7.5 % TiO_2) lenses and shoots in shear zones parallel to the regional foliation. Iron saturated fluids were mobilized during prograde regional metamorphism as iron rich amphiboles recrystallized to form iron depleted pyroxenes. The magnetite of twelve of the sampled mines is hosted by biotite bearing potassic schists and gneisses and sodic quartz oligoclase gneiss and also occurs as concordant disseminated ilmeno-magnetite (1 to 5 % TiO_2) lenses and shoots. Iron saturated fluids were mobilized when ferruginous biotite recrystallized to form garnet + K-spar + magnetite at 700 to 740 C^{o} and 5 kb f_{H2O} in equilibrium with concordant magnetite bearing pegmatite emplacement. Eleven of the sampled mines are characterized by undeformed veins of ilmeno-magnetite hosted by quartz oligoclase gneiss and less commonly by pyroxenite. The vein magnetite contains 1 to 3 % TiO_2, but only 0.05 to 0.09 % MnO in contrast to the 0.2 to 0.6 % MnO content of the disseminated magnetite. The magnetite of the undeformed veins precipitated out of hydrothermal fluids from late to post tectonic sources, perhaps mobilized by the intrusion of the Byram granite. The six remaining abandoned mines are hosted by coarse grained marble or calcite, garnet, sulfide granulites and occur as disseminated magnetite (0.01 to 0.2 % TiO_2), or manganiferous limonite lenses. The protolith was a marine carbonate enriched in iron and manganese from submarine volcanic sources.

Only prograde reactions involving biotite and amphibole can release iron and volatiles as they react to form garnet, feldspar, and pyroxene. In contrast, retrograde reactions proposed by others tend to absorb iron and volatiles as ferrugionous epidote and chlorite are generated.

INTRODUCTION

During the 1930s, 40s, and 50s hydrothermal processes dominated the economic geology literature under the influence of Lingren and Bateman. But during the 60s several stratabound syngenetic deposits were recognized and several important ore deposits previously described as hydrothermal were reinterpreted. Then, during the 70s volocanogenic black smoker deposits dominated the literature and again large numbers of deposits were reinterpreted almost to an extreme degree. Now, during the 90s, with the

In Benimoff, A.I., and Puffer, J.H., (editors), The economic geology of northern New Jersey: Field Guide and Proceedings of the fourteenth annual meeting of the Geological Association of New Jersey, 1997, p 71-96.

advent of Fe-oxide (Cu-U-Au-REE) deposits, the literature seems to be going full circle back to granite related hydrothermal solutions. The magnetite ore deposits of the New Jersey Highlands have been variously described as having been generated by each of the above processes and others. But history has shown that over-application of popular geologic models can be dangerous.

A classical hydrothermal precipitation model for the New Jersey magnetite ores was proposed by Sims (1958), and Sims and Leonard (1952). Then Buddington (1966), Buddington and Leonard (1962), and Baker and Buddington (1970) modified the hydrothermal model into a late stage igneous or deuteric release model whereby iron from granite magma locally became converted into the alaskite that is found near some of the iron mines.

Collins (1969a,b), Hagner and Collins (1963), Hagner et al., (1963), and Hagner and Collins (1955), on the basis of host rock and ore zone chemistry, concluded that the iron contained in amphibole bearing rock was mobilized by metamorphic processes and diffused into shear zones. They proposed that iron rich amphiboles recrystallized into magnesium rich pyroxenes plus magnetite. Puffer (1980) and Puffer et al., (1993) suggested a modified version of the Collins model that applies to several of the New Jersey iron mines where biotite was converted during prograde metamorphism into to garnet and feldspar plus iron and potassium saturated water. Magnetite was then precipitated out of the water in a granulite facies environment in shear zones.

A syngenetic model for some of the magnetite deposits involving the accumulation of marine volcanogenic enriched sediments was proposed by Kastelic (1979, 80) and Gunderson, (1986). Johnson (1996), suggests that the marine iron enrichment processes may have been similar to those that concentrated the ore of the Franklin zinc mines. Puffer et al., (1993) apply a similar model to the carbonate hosted magnetite deposits in New Jersey. Pre-Grenville iron accumulation is supported by the foliated texture of the ore and the concordant orientation of the ore lenses to regional Grenville structures.

Foose and McLelland (1995) propose that the hydrothermal processes responsible for the Olympic Dam iron oxide (Cu-U-Au-REE) deposit of South Australia also applies to the magnetite deposits of New York and New Jersey. Meyers (1988) and Grow et al., (1994) suggest that the abundant magnetite and hematite in an Australian calc-silicate-siltstone breccia was precipitated out of hydrothermal solutions at 200 to 400 C^{o} that were generated during the intrusion of post orogenic granites. Foose and McLelland (1995) support their hydrothermal argument with radiometric age dating that indicates a late tectonic intrusion of granite in New York and an association of iron ore occurrences with undeformed veins and pegmatites.

With the exception of Collins et al., (1969a,b), and Puffer et al., (1993) all of the above genetic proposals were made with little regard to the chemical composition of the magnetite that was being mined. The diverse chemical compositions of the magnetites, however, provides data that are essential to an understanding of their genesis.

OBJECTIVE

Since the 1993 GANJ guidebook chapter titled "Precambrian Iron Deposits of the New Jersey Highlands" (Puffer et al., 1993) some important new information bearing on the origin of the iron ore has become available. The 1993 chapter presented new geochemical analyses of a very large collection of New Jersey iron ore samples that I have accumulated over the last 30 years, and offered some new ideas about the origin of the deposits consistent with the geochemical data, particularly titanium content. However, it has recently become clear (Volkert, 1993, 1995) that one of the host rocks and a possible source for some of the deposits (Byram Granite) was emplaced in an anorogenic tectonic setting completely unlike the anatectic, catazonal setting first proposed by Buddington (1959) and accepted by most subsequent authors. In addition, a reexamination of the geochemical data presented in 1993 (particularly MnO content) together with several new analyses has made it clear that there are four distinct geochemical groups of magnetite ore in the New Jersey Highlands. This chapter, therefore is a major revision of the 1993 chapter together with a reassessment of the origin of the deposits in light of the new information.

GEOLOGIC SETTING

The Precambrian rocks of the New Jersey Highlands may be grouped into six types: 1) Granites including the Byram and the Lake Hopatcong granites and related syenites and alaskites; 2) Diorite and related orthogneises; 3) The Losee Gneiss and related sodic meta-volcanic quartz-oligoclase gneisses; 4) Meta-sedimentary schists and potassic gneisses; 5) Marbles including the Franklin and Wildcat Marbles; and 6) amphibolites and pyroxenites (Figure 1). Although most of the larger iron deposits occur within or close to the margins of sodic metavolcanic rocks, potassic metasediments, and amphibolites (Figure 1) at least a few iron deposits are also found within each of these 6 diverse rock types. However, Collins (1969b) notes "the absence of magnetite concentrations in hornblende granite and alaskite".

Almost all of the iron deposits occur as either disseminated or massive magnetite concentrated in lenses oriented subparallel and conformable to the regional north-east strike (Figure. 1) or as networks of closely spaced narrow discordant magnetite veins. Each of the metamorphic rocks were subjected to granulite facies or at least upper Amphibolite facies (>700 C^0) conditions during the Grenville Orogeny that included considerable prograde dehydration, some anatectic melting, and considerable pegmatite development.

The metamorphic rocks of the Highlands were then intruded by A-type granites (Volkert, 1993) during the waning stages of the Grenville orogeny. The initial granitic magma (the Lake Hopatcong Granite) crystallized into an anhydrous assemblage of feldspar, quartz and pyroxene with very minor pegmatite content and presumably carried little if any excess water or other volatiles capable of carrying iron. A secondary or perhaps coexisting intrusion of granite magma crystallized into the hornblende bearing Byram intrusive suite that includes common undeformed hornblende bearing pegmatites with minor magnetite content.

METHODS

Approximately 190 magnetite mines and prospects occur within the central New Jersey Highlands area (Figure 1). During the last 30 years I have attempted to examine each of these mines. Over half of them, however, are no longer exposed and have been completely re-landscaped typically under housing developments or farms. Many of the mines that can be located are poorly exposed as small cuts or are represented by only a few ore samples in a dump or tailings pile.

About 40 mines, however, are still exposed or are represented by a large enough tailings or abandoned drill core pile to gain some insight into the geology of the mine, and about 20 of the mines were well exposed when examined so that underground workings or large open pits could be carefully sampled. Wherever possible a minimum of five three pound ore samples were collected at each mine. The samples judged to be the most representative were then cut, crushed and split into separates to be analyzed for whole rock chemistry, petrology (hand specimen, thin section and polished section analyses), and oxide content. Each ore sample chosen for analysis would be considered high grade on the basis of an oxide content ranging from 25 to 95 volume percent.

Magnetite and in a few cases some ilmenite and biotite were separated from splits of crushed ore using magnetic and heavy-liquid separation methods. Mineral separates were further purified by hand picking under a binocular microscope. Mineral concentrates (Table 1) and high grade ore samples (Table 2) were chemically analyzed with a Rigaku wavelength dispersive x-ray fluorescence spectroscope. Microprobe techniques were rejected as an alternative method for magnetite analysis because of the presence of fine exsolution/oxidation lamellae of ilmenite and other phases that have depleted the titanium content and other elements in the host magnetite below solidus or primary precipitation levels. Microprobe analyses yield erroneous or meaningless subsolvus compositions of magnetites depleted by any elements exsolved into exsolution lamellae.

RESULTS

Four distinct populations of iron deposits within the New Jersey Highlands are identified on the basis of the chemical compositions of the ores, the chemical compositions of the magnetite components, the structural settings of the ores, and the lithologies of the host rocks. One population consists of magnetite disseminated in coarse grained pyroxenite lenses typically contained within larger amphibolite lenses interpreted as meta-basalt. The magnetite of the disseminated to massive pyroxenite hosted deposits contains 1 to 7 % TiO2 and 0.2 to 0.5 % MnO, (Table 1). A second population consists of disseminated magnetite in biotite-sillimanite-garnet schist or K-spar gneiss with common magnetite bearing K-spar biotite pegmatites and in biotite bearing quart oligoclase gneiss.. The magnetite of these disseminated to massive biotite bearing schist and gneiss hosted deposits contains 1 to 5 % TiO2 and 0.3 to 0.5 % MnO, (Table 1).

A third population consists of networks of undeformed magnetite veins injected into several rocks but most commonly in quartz oligoclase gneiss, interpreted as meta-dacite

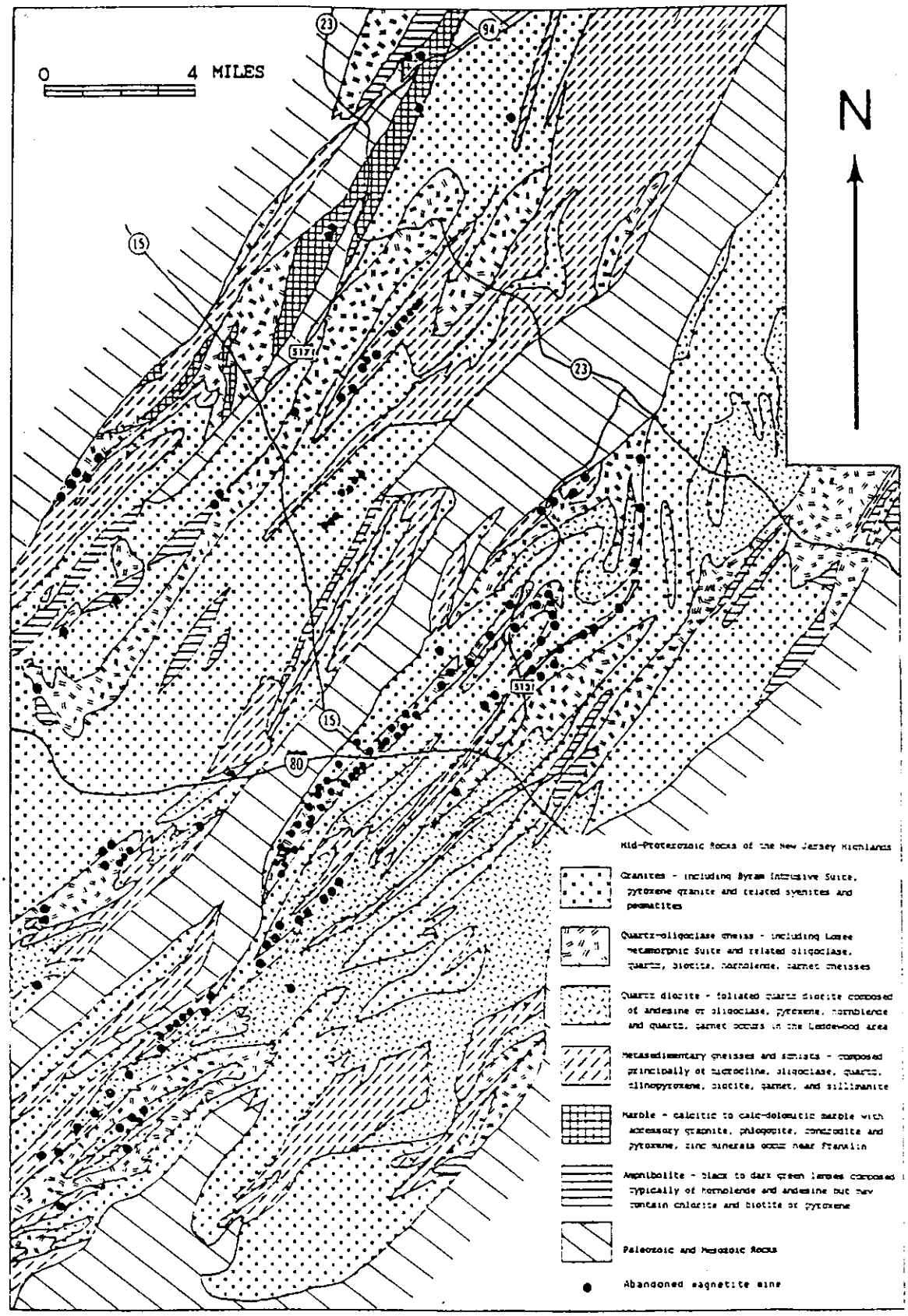


Figure 1. Geologic map of the central New Jersey Highlands with iron mine locations as reported by Baker and Buddington (1971), Sims (1958), Bayley and others (1910), Volkert and others (1989), and Spencer and others (1908).

TABLE 1. COMPOSITION OF MAGNETITE FROM NEW JERSEY HIGHLANDS IRON ORE DEPOSITS

1. Disseminated to Massive

Host rock	Pyroxenite / Amphibolite Hosted								
Mine	Ford	Blue	H.Ledge	Daven	Mt Hope	Dodge	Cogell	DeHart	George
weight %									
SiO2	1.21	1.85	1.39	1.42	2.11	2.01	0.98	1.23	1.38
TiO2	2.52	0.93	7.46	1.35	1.33	3.04	5.32	5.21	3.19
MnO	0.21	0.18	0.31	0.41	0.35	0.33	0.53	0.46	0.24
FeOt	79.38	86.35	91.59	88.95	89.31	92.45	79.72	83.05	82.06
V2O5		0.07	0.1						

2. Disseminated to Massive

Host rock	Meta-sed K-spar Gneiss & Biotite Schist Hosted						Quartz-Oligoclase-Biotite Gneiss Hosted					
Mine	Edison	ShermanB	DavenBV	Bunker	McKean	Beach Gl	Dickers	L.Baker	Wawayu	Byrant	Righter	Gulick
weight %												
SiO2	1.33	1.21	1.21	1.42	1.32	1.72	1.23	0.86	0.98	0.89	0.73	0.86
TiO2	1.31	1.03	4.87	3.99	4.38	1.15	1.59	1.22	0.92	1.97	1.64	1.22
MnO	0.41	0.41	0.55	0.36	0.45	0.26	0.32	0.31	0.35	0.37	0.35	0.36
FeOt	94.01	93.16	86.32	88.93	84.95	94.21	92.32	93.04	91.42	91.46	91.28	87.31
V2O5	0.06											

3. Undeformed Veins

Host rock	Quartz-Oligoclase Gneiss Hosted								Pyroxenite Hosted	
Mine	Baker	Allen	Randall	Richard	Leonard	Elizabeth	Evers	Chester	ScrubOak	Hibernia	Fairview
weight %											
SiO2	1.65	1.59	1.32	1.28	0.99	1.62	0.94	1.27	1.41	0.97	1.24
TiO2	2.7	1.43	2.87	2.11	1.26	1.18	1.76	2.38	0.05	1.19	2.13
MnO	0.08	0.07	0.09	0.05	0.07	0.06	0.09	0.09	0.06	0.09	0.08
FeOt	94.23	95.22	89.39	88.32	93.21	94.22	88.73	89.97	88.86	90.37	90.98
V2O5	0.09	0.07	0.08	0.07	0.09	0.13		0.06	0.01	0.15	0.11

4. Carbonate Hosted

Mine	Roseville	SulfurHill	Andover*	Ahles	Pikes Peak
weight %					
SiO2	2.01	0.87	3.95	0.25	0.41
TiO2	0.27	0.02	0.01	0.08	0.11
MnO	0.45	0.15	2.75	0.73	0.68
FeOt	88.92	87.54	78.53	85.22	86.41

* Some of the MnO may be unseparated fine pyrolusite mixed with a largely fine hematite concentrate.

TABLE 2. CHEMICAL COMPOSITION OF IRON ORES

	New Jersey Highlands Carbonate Hosted Iron Deposits				
mine	Sulfur Hill	Andover	Ahles	PikesPeak	Mine H PA
witght %					
SiO2	29.39	5.98	16.96	1.67	30.45
Al2O3	1.78	3.84	4.06	0.28	0.96
Fe2O3t	32.52	74.83	50.91	45.12	67.35
MnO	0.33	0.45	11.28	2.9	0.1
TiO2	0.04	0.07	0.01	0.01	0.1

	Banded Iron Formations and Black Smoker Sea Mound				
mine	Gogebic	Biwabik	Hamersley	MinasGeris	Pacific
weight %					
SiO2	28.53	46.4	42	0.4	52.42
Al2O3	0.92	0.9	0.3	1.23	0.36
Fe2O3t	68.56	40.58	37.3	51.01	30.26
MnO	0.22	0.63	0.14	0.25	0.49
TiO2	0.06	0.04	0.06	0.01	0.02

	Highlands Gneiss, Schist, and Pyroxenite Hosted Deposits						
mine	Evers	Baker	Allen	Randel	ScrubOak	Richard*	Leonard*
weight %							
SiO2	48.75	8.32	10.96	6.74	30.41	3.77	3.56
Al2O3	1.51	1.08	1.01	0.98	2.38	0.79	0.44
Fe2O3t	41.61	89.55	89.26	90.49	66.15	93.97	93.97
MnO	0.07	0.06	0.06	0.08	0.05	0.06	0.03
TiO2	1.32	2.54	1.29	2.78	0.28	1.3	1.15
mine	Elizabeth*	12 Edison	Taylor	Blue	Hibernia	H.Ledge	Ford
weight %							
SiO2	1.38	56.47	4.21	17.85	8.35	3.4	6.31
Al2O3	0.55	9.64	0.58	1.62	1.86	1.02	0.83
Fe2O3t	93.78	23.42	94.21	64.57	87.57	91.46	86.61
MnO	0.03	0.19	0.03	0.15	0.05	0.29	0.21
TiO2	1.09	1.18	0.82	0.83	1.19	4.39	3.54

* Bayely (1910)

Table 2: Chemical composition of iron ore samples and magnetite concentrates from mines hosted by marbles or calc-silicate lenses within marble from the New Jersey Highlands compared with analyses of BIF (banded iron formation) from South America (VanN.Dorr (1973), the Lake Superior district (Bayley and James, 1973), and Australia (Appel and LaBerge, 1987) and with hydrothermal (black-smoker) sediments from the east Pacific Rise (Barrett and others, 1988).

Table 3 Composition of Co-existing Oxide Pairs from the Edison Iron Mine

	Ore sample 145		Ore Sample 149		Ore Sample 154	
	ilmeno-mag	ilmeno-hem	ilmeno-mag	ilmeno-hem	ilmeno-mag	ilmeno-hem
Fe2O3	70.23	56.61	68.49	46.59	63.94	29.1
FeO	21.77	2.83	25.27	16.32	26.45	12.79
TiO2	1.01	16.16	0.88	23.67	1.39	30.63
Temp Co	748		717		706	
log fo2	-11.46		-12.133		-13.332	

	Microcline Gneiss		Pegmatite	
	ilmeno-mag	hemo-ilmen	ilmeno-mag	ilmeno-hem
Fe2O3	53.72	6.44	70.31	60.43
FeO	31.06	41.45	26.57	14.25
TiO2	2.8	40.44	0.7	20.29
Temp Co	732		707	
log fo2	-13.665		-12.11	

Temperatures and oxygen fugacitites based on Buddington and Lindsley (I964).
Ore sample analyses from Baker and Buddington (I970).
Geniss and Pegmatite analyses from Puffer (I975).

(Volkert and Puffer, 1991). The magnetite of these undeformed vein deposits contains 1 to 3 % TiO2 and 0.05 to 0.09 % MnO, (Table 1).

The fourth iron ore population consists of magnetite or hematite disseminated in carbonate bearing granulites and marble that typically contain garnet and sulfides. The magnetite of these marine carbonate hosted deposits contains 0.01 to 0.25 % TiO2 and commonly coexists with manganiferous limonite.

INTERPRETATIONS

1. Disseminated to Massive Pyroxenite Hosted Deposits

Characteristics:

Amphibolite and pyroxenite lenses and layers are found included within each of the rock types of the New Jersey Highlands and range from a few cm to over 100 m in thickness. The plagioclase of the amphibolites is typically andesine with variable hornblende, pyroxene, and biotite contents. There are several genetic types of amphibolite and related pyroxenites in the New Jersey Highlands (Puffer et al., 1993b) but the amphibolites and pyroxenites associated with iron ore deposits are not typical of the common genetic types (the meta-basalts and meta-shales) and are typically coarser grained and more mafic. Most pyroxenite hosted deposits consist of ilmeno-magnetite disseminated with coarse grained pyroxene, plagioclase, and minor chlorite in green lens shaped zones.

The ilmenite intergrown with magnetite in ilmeno-magnetite occurs as thin blades or lamellae that resemble exolution lamellae but originate from subsolidus oxidation and contemporaneous exsolution of ulvospinel (Buddington and Lindsley, 1964). The ilmenite lamellae are widely spaced and make up less than one percent of typical polished magnetite surfaces but there presence indicates titanium saturation and initial crystallization temperatures higher than the magnetite-ulvospinel solvus. If low temperature oxidation or weathering has effected the deposit, varying degrees of hematite replacement occurs along microcracks in the ilmeno-magnetite host forming a grid-like intergrowth. The ilmeno-magnetite of the pyroxenite hosted ore contains up to 7.5 % TiO_2, (Table 1) which is a level that approaches the TiO_2 content of magnetite from mafic plutonic rock (Figure 3). However, most disseminated magnetite from the New Jersey highlands overlaps the field of magnetite from typical granite and pegmatite suggesting equilibration under similar T/P contitions.

Many of the pyroxenite hosted deposits of the New Jersey Highlands are located in the Dover iron mining district. Collins (1969) mapped the 24 square mile area around Dover, New Jersey and sampled 773 specimens of pyroxenite and amphibolite. Collins compared the unmineralized amphibolites with those containing magnetite deposits and found that hornblende in unmineralized rock is relatively fine grained and aluminous (9.5 to 13.1 percent Al_2O_3) and is typically rich in TiO_2, FeO, MnO, CaO, K_2O, V, Cr, Zn, Pb, Sr, and Ba. In contrast, the amphiboles and pyroxenites of ore deposits is coarse grained containing only 5.1 to 5.5 percent Al_2O_3 but relatively high concentration of SiO_2, MgO, Na_2O, Be and

Bi. Collins found that the magnetite in country rock amphibolites also is relatively rich in MnO, TiO_2, V, Cr, Co, Ni, Cu, Zn, and Pb compared to magnetite in ore zones.

Examples:

Iron deposits hosted by amphibolite and pyroxenite include several described by Sims (1958) as "skarn" deposits although carbonates are rarely found in more than trace amounts in pyroxenite hosted ores and evidence of contact metamorphism is not clear. Some of the larger mines hosted by pyroxenite are the Hibernia Mine and the Mount Hope Mine (Tables 1 and 2).

The Hibernia mine, as of 1958, was the third largest producer in the Dover district with production in excess of 5 million tons of ore averaging 50 percent iron. The deposit strikes N40° to 60°E and dips steeply to the SE. It is tabular and averages about 3 meters thick. The ore is composed of coarse grained massive magnetite with hornblende and pyroxene as the principal gangue minerals together with minor quartz, plagioclase, biotite, and pyrite (or pyrrhotite).

The main shaft at the Mount Hope mine is 2694 feet deep with levels at 200 foot intervals. The ore is in part massive and in part laminated with layers of hornblende, pyroxene, and biotite.

Origin:

The magnetite rich pyroxenite/amphibolite rocks consist primarily of magnetite, clinopyroxene (principally augite with some diopside), and/or orthopyroxene (principally hypersthene), plagioclase, common hornblende, and minor quartz, mica, garnet, sillimanite, epidote, pyrrrhotite, calcite, and chlorite. Prograde metamorphic reactions that involve this assemblage include:

2 epidote => plagioclase + garnet + hematite + quartz + H_2O

Ferrotremolite => Magnetite + quartz + orthopyroxene + H_2O

Hornblende + quartz => hypersthene + anorthite + clinopyroxene + H_2O

Hornblende + 2 biotite + 17 quartz =>15 hypersthene + 4 orthoclase + 3 plagioclase + 5 H_2O

Since the Fe/Mg ratio of each of the amphibole phases decreases during requilibration, the released water is iron saturated. Since chlorine and fluorine are also expelled from hornblende during dehydration, these anions were available to complex with iron and aid in transmitting it to precipitation sites.

The stability of synthetic iron amphibole (ferrotremolite) was determined experimentally by Ernst (1966), and breaks down to yield magnetite bearing assemblages at amphibolite to granulite facies temperatures and oxidizing environments. Support for a prograde metamorphic movement of iron is found in a series of papers authored by Collins and Hagner

(Collins, 1969a,b, Hagner and Collins, 1963, Hagner et al., 1963, Hagner and Collins 1955). They provide abundant and compelling evidence that iron rich ferromagnesian silicates in amphibolites have recrystallized in shear zones as Mg-rich silicates plus magnetite. The process involves a decrease in modal hornblende and an increase in modal clinopyroxene and biotite as magnetite concentrations are approached. They show that this process occurs in an open system but without any introduction of material from a magmatic source. Their calculations measure the chemical changes that would occur if amphibolites composed of various mineral assemblages were recrystalized to form various magnetite enriched pyroxenites. In most cases CaO, TiO_2, MnO, Al_2O_3 is lost, SiO_2 is released to form quartz, K_2O is released to form biotite or potassic feldspar, and iron is released to form magnetite.

Their calculations are similar to conventional mass-balance calculations although they do not balance due to the net loss of CaO, TiO_2, MnO, and Al_2O_3. But since they assume open system conditions, these elements may have been flushed out with the expelled water during prograde recrystallization.

The most compelling evidence in support of hornblende as a source of iron for pyroxenite hosted magnetite deposits is the consistent decrease in the iron content of hornblende as magnetite ore is approached. For example, the FeO content of hornblende from the country rock near the Hibernia magnetite deposit is 15.3 weight percent, considerably higher than the 10.67 percent FeO in hornblende gangue from the ore zone (Collins, 1969). The biotite from country rock adjacent to Hibernia ore is also much richer in iron than biotite in the ore. The biotite in Hibernia ore is extremely depleted in iron and could be characterized as a phlogopite (Collins, 1969).

2. Disseminated to Massive Biotite Bearing Schist and Gneiss Hosted Deposits

Characteristics:

The host rocks of this population consists of : 1.) biotite schist; 2.) K-spar gneiss containing variable quantities of magnetite, K-spar, biotite, quartz, sillimanite, garnet, and plagioclase; and 3.) biotite bearing quartz oligoclase gneiss (Losee Gneiss). These potassic and sodic rocks are generally interpreted as meta-sediments and meta-volcanics. Layers of quartzite found associated with the potassic schists and gneisses support a meta-sedimentary interpretation. The chemical composition of the sodic quartz oligoclase gneiss is consistent with a meta-dacite protolith (Puffer and Volkert, 1991). Magnetite and biotite bearing pegmatite lenses are commonly found emplaced concordant to the foliation of these host rocks. Magnetite disseminated in these schists, gneisses, and pegmatites is concentrated in lenses and shoots that are consistently concordant to the regional foliation.

The iron-titanium oxides of the ore zones include ilmeno-magnetite (magnetite with ilmenite lamellae), hemo-ilmenite (ilmenite with hematite microintergrowths, and ilmeno-hematite (hematite with ilmenite microintergrowths). Ilmeno-magnetite is the most abundant oxide phase making up well over 95 percent of the oxides in most deposits. Wherever any alteration has occurred, particularly weathering, the magnetite surfaces have been partially oxidized to martite.

The ilmeno-magnetite composition of the schist, gneiss, and pegmatites is approximately the same suggesting similar equilibration temperatures. The ilmeno-magnetite contains 1 to

5 % TiO_2 and 0.3 to 0.5 % MnO, (Table 1, Figure 2) and overlap the chemical range of magnetites from granites and pegmatites (Figure 3). However, the TiO_2 content of gneiss hosted deposits greatly exceeds that of any banded iron formation or black smoker rock (Figure 4) and the TiO_2 content of the gneiss hosted magnetites is less than typical of mafic plutonic rock (Figure 3).

Examples:

Some of the larger and relatively well known disseminated lenses and shoot deposits hosted by biotite bearing schist and gneisses include the Sherman-Bunker deposits and the Edison deposits of the Franklin quadrangle. Typical ore, as described in detail by Baker and Buddington (1970) and Puffer et al., (1993), consists of magnetite, quartz, potassium feldspar, biotite, sillimanite, and garnet. The biotite content is particularly variable and is concentrated in schistose layers. Portions of the Edison deposit are mineralized with sulfides (pyrite, chalcopyrite, and molybdenite) that have locally led to saprolitic weathering through sulfuric acid leaching. Pegmatites composed of magnetite, quartz, potassium feldspar, and biotite are common throughout the Edison mine.

Origin:

Some of the balanced prograde metamorphic reactions involving the mineral assemblage of the biotite bearing schist and gneiss hosted deposits that are transitional to or within the granulite facies as presented by Hyndman (1985) include:

Biotite + H_2O => biotite + sanidine + magnetite + H_2O
Biotite + quartz => sanidine + magnetite + hematite + H_2O
Biotite + 3 quartz => 3 hypersthene + 2 anorthite + orthoclase + H_2O
Biotite (annite) + sillimanite + ½ O_2 => muscovite + magnetite + quartz
Muscovite + biotite + 3 quartz => 2 orthoclase + almandine + $2H_20$

Each of these reactions involve the kind of compressional prograde regional metamorphism that occurred during the peak of the Grenville orogeny. The phase boundary between the biotite + H_2O and the biotite + sanidine + magnetite + H_2O fields of the first reaction listed above has been experimentally determined by Wones and Eugster (1965). They show that as temperatures are increased the Fe/Fe+Mg ratio of the biotite is decreased. Iron saturated water expelled from biotite as temperatures are increased is driven out of high pressure zones into any permeable avenue of escape such as shear zones where magnetite would then precipitate. A similar proposed origin for some of the iron deposits of the New Jersey Highlands was proposed by Puffer (1980).

The release of iron from biotite accompanied by any dehydration and precipitation of anhydrous phases such as garnet would set up a complex iron and potassium saturated hydrothermal system. Released water would also contain any other volatile present in the initial biotite such as Cl and F.

Martin and Piwinskii (1969) have experimentally shown that iron under a wide range of pressures fractionates into a vapor phase out of calc-alkaline rock. They found that the

Figure 2. Magnetite from New Jersey Highlands Iron Deposits.
(data of Table 1)

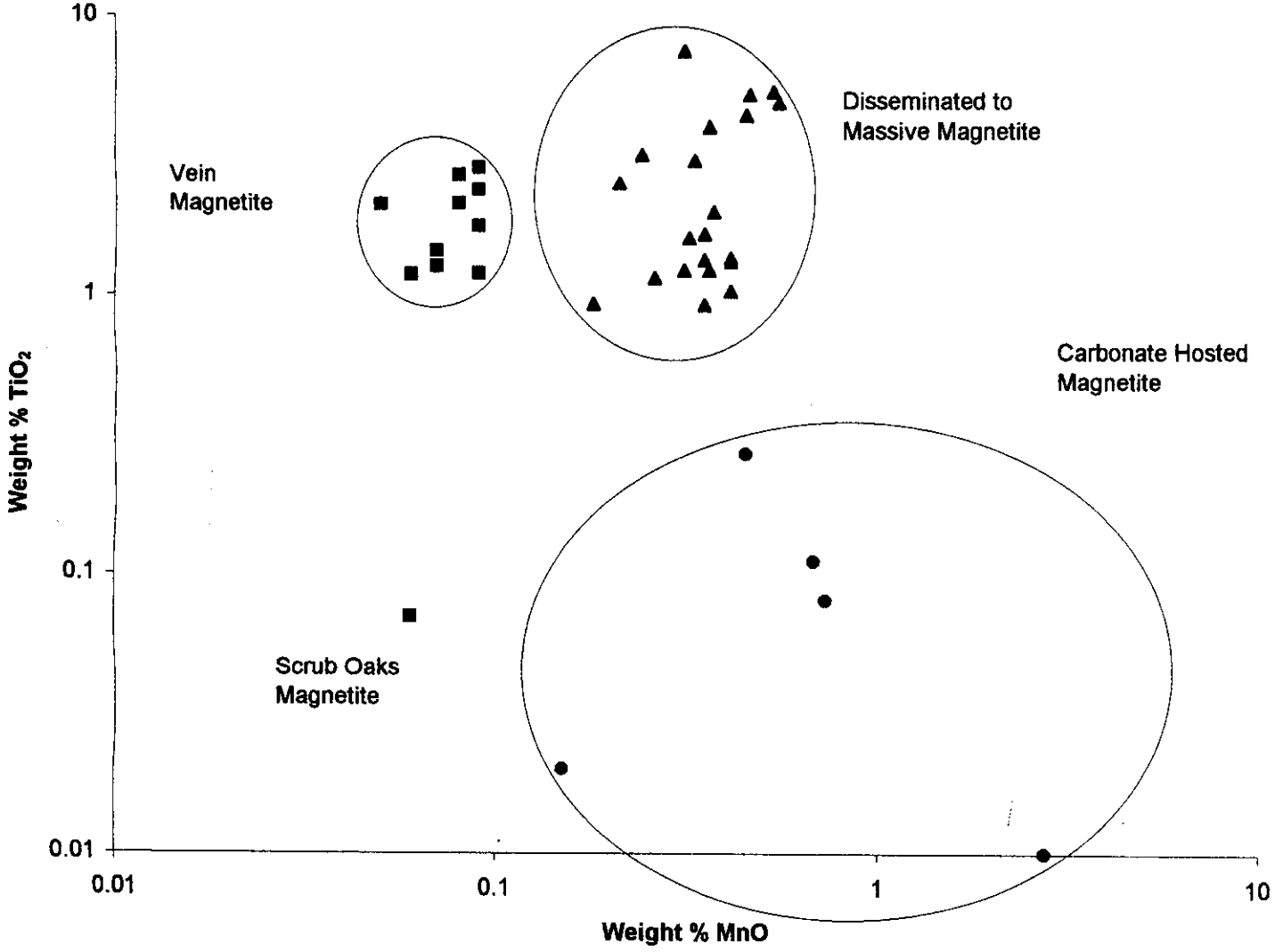


Figure 3. Magnetite from New Jersey Highlands Ore Deposits Compared to Magnetite form Granites, Pegmatites, and Gabbros.

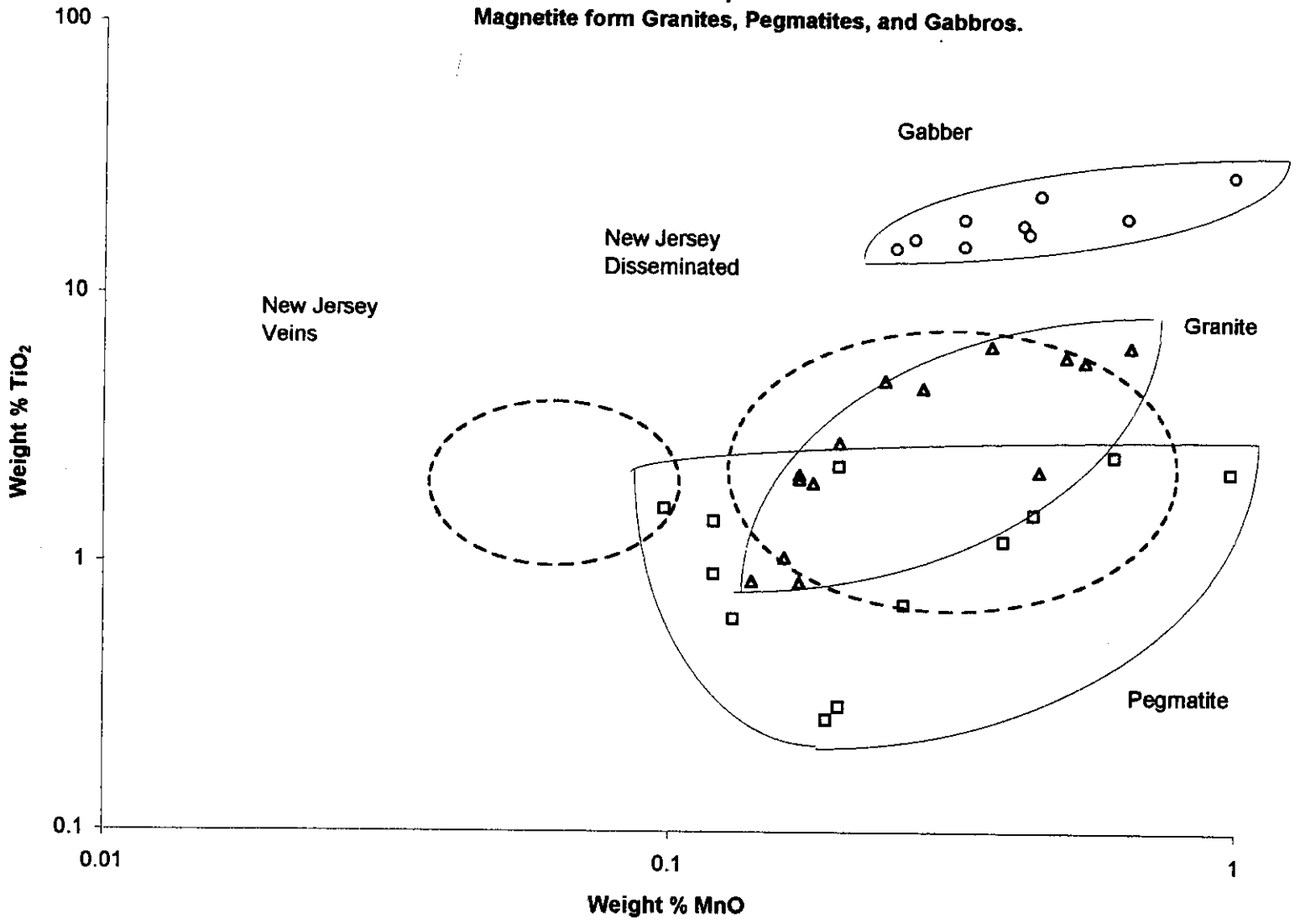
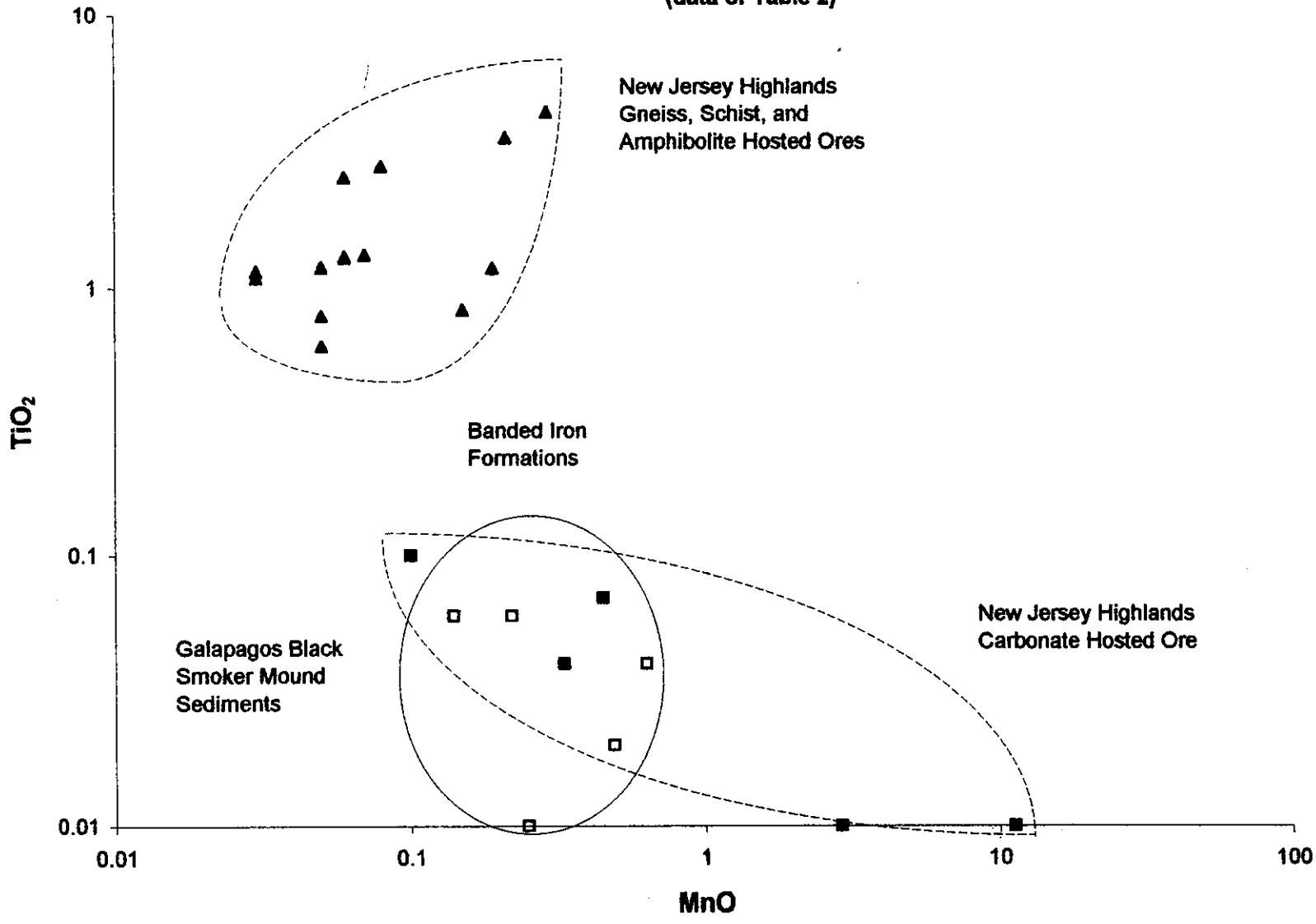


Figure 4. Iron Ore from the New Jersey Highlands, Banded Iron Formations, and Galapagos Black Smoker Mound. (data of Table 2)



presence of a melt, of a specific rock type, or of chloride, fluoride, sulfide, or carbonate anions is not required for the leaching and transport of iron in a vapor phase. Their subsolidus experiments confirm that iron behaves like potassium, and sodium which are leached and then migrate readily from high temperatures (700° C at 5 kb) towards low temperature zones (560° to 450° C) unlike calcium and magnesium which do not.

The upper stability of iron biotite (annite) was experimentally determined by Ernst (1976). The conversion of biotite + quartz into the magnetite + sanadine + quartz + iron saturated fluid field may be forced by an increase in fO^2 or decrease in temperature at constant composition. Increases in fO^2 typically occur in shear zones where hydrothermal fluids have migrated and would be accompanied by a temporary drop in temperature as fluids escape into overlying rocks. These temporary reversals in temperature and fO^2 occur during ruptures or fluid pressure releases and may result in the kind of iron deposition that is found in the Edison Iron mine.

The importance of fO^2 as a control on the phase equilibra of biotite, sanadine, magnetite and ilmenite is demonstrated in the Wones and Eugster (1965) geothermometer and geobarometer:

$$\log f_{H2O} = \frac{3428 - 4212\,(1-x_1)^2}{T} + \log x_1 + 1/2 \log f_{O2} + 8.23 - \log a_{KAlSi3O8} - \log a_{Fe3O4}$$

The f_{H2O} of a biotite + magnetite + ilmenite + sanadine assemblage in the host rock adjacent to the ore zone of the Edison iron mine (Table 3) was calculated using Wones and Eugster's equation. The results of chemical analyses of minerals separated from Edison magnetite ore (Puffer et al., 1993) indicate:

$a_{Fe_3O_4}$ of Edison magnetite = 0.914

$a_{KAlSi_3O_8}$ of Edison host rock Kspar = 0.95

x_1 (mole fraction of annite) in the co-existing biotite = 0.5328

T = 1005 K° (based on magnetite/ilmenite geothermometer of Buddington and Lindsley (1964)

$\log f_{O2} = -13.667$

$\log f_{H2O} = 3.70281$ atm or a $f_{H2O} =$ 5,044 atm.

A f_{H2O} of 5,044 atm. and a temperature of 732 C° is in good agreement with a granulite facies metamorphic environment and supports the interpretation that the biotite + magnetite + ilmenite + sanadine host rock assemblage at Edison is an equilibrium assemblage.

Iron saturated vapors released from the biotite precipitated at slightly lower temperatures and higher oxygen fugacities to form magnetite concentrations in shear zones. The chemical compositions of several coexisting magnetite + ilmenite (and ilmeno-hematite) pairs separated from the ore zone at the Edison mine and from a pegmatite analyzed by Puffer (1975) are listed in Table 3. The oxide pairs yield temperatures in a 706 to 748 C° range only slightly less than the 732 C° of the host rock and the 707 C° of a pegmatite located in the

Edison ore. The oxygen fugacity range (-log fO^2) of the magnetite ore zone is 11.5 to 13.3, slightly higher than that of the host rock; again in good agreement with the experimental studies.

A major portion of the iron contained within the biotite schist and gneiss hosted ore zones, therefore, originated from within the metasedimentary or metavolcanic protoliths and was simply mobilized in response to prograde metamorphic reactions and reprecipitated in ductile shear zones in response to a drop in water pressure.

3. Undeformed Veins Deposits

Characteristics:

Many of the magnetite deposits hosted by the Losee Gneiss (a quartz oligoclase meta-dacite) and a few deposits hosted by pyroxenite consist of an undeformed network of sharp discordant and condordant veins composed of magnesium depleted ilmeno-magnetite (Table 1). Some veins coalesce into dense and massive lenses with quartz and oligoclase as the principal gangue minerals. In addition, a few magnetite veins were also injected into other Highlands rock types but are rare in granite, particularly Lake Hopatcong Granite. Typical low temperature hydrothermal minerals such as sericite, and chlorite are conspicuously rare suggesting that mineralization temperatures exceeded their stability field. Few magnetite bearing pegmatites are associated with the vein networks. Pegmatites contained within the Losee Gneiss are common but are typically small concordant lenses of coarse grained quartz and oligoclase. The magnetite from several mines dominated by magnetite veins (Table 1, Figure 2) contains 1 to 3 % TiO2 and 0.05 to 0.09 % MnO. The titanium and vanadium levels of the vein magnetites overlap those of the disseminated to massive magnetites (Table 1) suggesting an overlapping temperature range of emplacement. However, with few exceptions the MnO content of the vein magnetite is distinctly lower (Figures 2 and 3).

Examples:

Undeformed vein deposits hosted by quartz oligoclase gneiss include several within the Dover mining district as described by Sims (1958) and Bailey (1910). The Dover district included the largest and most productive mines in New Jersey. The geologic map of the Dover district (Sims, 1958) locates most of the 91 listed mines in "oligoclase-quartz-biotite gneiss" and "gneissic albite-oligoclase granite". Sims (1958), however, defines the albite-oligoclase granite as medium grained rock that is composed almost entirely of plagioclase (An8-An15) and quartz with minor muscovite, augite, hornblende, and biotite. Since the Losee Gneiss and the gneissic albite-oligoclase granite of Sims have the same texture and mineral composition I have found them to be indistinguishable and group them together, (Figure 1). Some of the larger Dover area mines are the Richard Mines, the Leonard, and the Elizabeth.

The Richard Mines consists of two deposits: the Mount Pleasant and the Richard. The main shaft is 1,244 feet deep with levels spaced every 200 feet. As of 1958 the Mount Pleasant deposit was being developed on the 1300 and 1500 levels and the Richard deposit was being developed on the 1500 level. The Richard deposit is tabular, strikes N40E and

dips 30° to 50° SE, with an average thickness of 5 meters. It locally contains thin layers of biotite schist. The Mount Pleasant deposit consists of three well-defined shoots that dip 40 to 60° SE. The largest shoot is about 3 meters thick.

The Leonard deposit, up to 8 meters thick and oriented N45°E with a vertical dip, was mined continuously from the surface to the 1700' level of the Mount Hope mine (Sims, 1958). The Elizabeth deposit strikes N45°E but dips 70° SE and is up to 7 meters thick; for structural details see Sims, 1958.

The Scrub Oaks mine consists of six levels 250 feet apart. The deposit is a tabular body that includes several shoots. The ore body and the country rock strike N33°E and dip 55°SE. The Scrub Oaks ore consists primarily of magnetite bands or veins that are parallel to the schistosity of the host rock and veinlets that cut across the schistosity. Smith (1933) describes clear replacement textures including the observation that "The magnetite surrounds corroded remnants of the silicate minerals." and is "... unmistakably later in origin".

The Scrub Oaks ore also includes considerable hematite. At the 1586 and 1587 stopes of the no. 5 level, hematite forms about 50 percent of the ore but is less than 5 percent of the ore elsewhere in the mine. The gangue minerals of the Scrub Oaks ore include the albite and quartz of the hosts Losee Gneiss but also include common apatite and tourmaline, neither of which are found in Losee Gneiss outside of the ore deposit. In addition, their are several rare-earth minerals concentrated in the ore, particularly in coarse grained magnetite ore zones and associated pegmatites (Klemic et al., 1959). Doverite, xenotime, bastnaesite, chevkinite, apatite, zircon and monazite are the principal radioactive rare-earth bearing phases concentrated at Scrub Oaks that are described by Klemic et al., (1959). In addition to the atypical mineralogy of the Scrub Oaks mine, the composition of the Scrub Oaks magnetite is atypical of most vein deposits and contains much less TiO_2 and V_2O_5 than any other analyzed vein magnetite.

Origin:

In contrast to the disseminated concentrations of ilmeno-magnetite in lenses parallel to host rock foliation, the ilmeno-magnetite veins are chemically distinct and commonly cut across foliation planes. The ilmeno-magnetite veins were, therefore, precipitated out of hydrothermal solutions after peak regional metamorphism from a late to post-tectonic source. The titanium and vanadium levels of the vein magnetites overlap those of the disseminated to massive magnetites (Table 1) suggesting an overlapping temperature range of emplacement. However, with few exceptions the MnO content of the vein magnetite, which is independent of precipitation temperature, is distinctly lower (Figures 2 and 3). The contrasting MnO level is evidence of a hydrothermal source unlike the source of the disseminated deposits.

Additional evidence of high temperature precipitation is the absence of low temperature hydrothermal minerals such as sericite, chlorite, calcite, epidote, or sulfides along the margins of most of the veins. The wall rock of most veins was unaltered by retrograde hydrothermal reactions and the antiperthitic textures of some wall rock oligoclase is preserved.

A plausible source of iron enriched hydrothermal vapor is the Byram Granite as suggested by Buddington (1966), Buddington and Leonard (1962), and Baker and Buddington

(1970). These authors proposed a late stage igneous or deuteric release of iron from Byram granite magma as it locally became converted into an iron depleted alaskite facies that is commonly found near some of the iron concentrations. Their proposal is similar to that of Foose and McLelland (1995) who also suggest a granite source of iron rich hydrothermal fluid. Buddington (1959) described the Byram granite as catazonal with most of the characteristics of what later became known as "S-type" granite. However, Volkert (1993) has shown that the Byram granite is an "A-type" or anatectic granite implying post or at least late tectonic emplacement. Recognition of the Byram Granite as A-type is evidence that Foose and McLelland (1995) use to include the Highlands magnetite deposits in the global category of iron deposits that are characterized by an association with such granite. Additional characteristics of Fe-oxide (Cu-U-Au-REE) deposits include an association with Cu, U, Au, and REE mineralization, shallow brittle emplacement in tectonic breccias, low temperature (200 to 400 C^{o}) precipitation temperatures, very low titanium contents in precipitated magnetite, and an association or dominance of hematite mineralization.

None of these additional characteristics apply to any of the Highlands magnetite deposits hosted by potassic meta-sedimentary rocks or pyroxenite which are devoid of Cu, U, Au, and REEs, occur as concordant lenses emplaced in ductile shear zones with pegmatites in equilibrium with granulite facies mineral assemblages, lack low temperature hydrothermal mineral assemblages including hematite, and typically contain 3 percent but as much as 7.5 percent TiO_2. The typically 2 percent TiO_2 content of magnetite from Highlands vein deposits is also much greater than the < 0.1 percent TiO_2 levels of Fe-oxide (Cu-U-Au-REE) type ores such as the Kiruna or southeastern Missouri iron ores. However, the occurrence of A-type granite near the New Jersey Highland vein deposits and their undeformed vein style of emplacement is consistent with an Fe-oxide (Cu-U-Au-REE) type of emplacement.

Fe-oxide (Cu-U-Au-REE) type mineralization may be particularly applicable to the magnetite veins of the Sterling Lake magnetite deposit of the Highlands area described by Hotz (1953) and Puffer (1975). The Sterling Lake veins are associated with hematite and magnetite pegmatites that are completely undeformed. Euhedral crystals of hematite and magnetite containing only 0.1 percent TiO2 are disseminated in a pegmatite consisting of unaltered microcline that could only have been emplaced after compressional regional metamorphism.

Of all the diverse iron mines in the New Jersey, the one that lends itself most clearly to an Fe-oxide (Cu-U-Au-REE) type interpretation is the Scrub Oaks mine located in the Mine Hill area on the western edge of Dover. Most of the characteristics of Fe-oxide (Cu-U-Au-REE) type ores (Hitzman et al., 1992) that <u>do not</u> apply to most Highlands deposits <u>do</u> apply to the highly atypical Scrub Oaks mine:

1. The composition of Scrub Oaks magnetite contains <0.1 percent TiO2 (Figure 2) and closely resembles magnetite from Fe-oxide (Cu-U-Au-REE) type Kiruna and southeastern Missouri deposits.
2. The magnetite commonly occurs in veinletts that crudely resemble the breccia complex of the Olympic Dam deposit.
3. A major portion of the ore is hematite.
4. Some of the wall rock of the magnetite veins is altered to sericite and chlorite.
5. Late veins rich in calcite, pyrite, and chalcopyrite cut through the ore deposit.

6. Uranium and REE minerals (including doverite, xenotime, bastnaesite, chevkinite, apatite, zircon, sphene and monazite) are mixed with the magnetite and commonly occur as aggregates that enclose euhedral crystals of magnetite.

A plausible case, therefore, can be made that the Scrub Oaks mine and possibly some of the other undeformed vein deposits of the New Jersey Highlands are genetically related to the intrusion of A-type granites.

4. Marine Carbonates Hosted Deposits

Characteristics:

This group includes most of the disseminated iron deposits contained within the Franklin Marble and the carbonate enriched meta-sedimentary rocks of the New Jersey Highlands including some but not all of the rocks described as "skarns" by Sims (1958). Such deposits are common in three separate areas of New Jersey, near the towns of Belvidere, Franklin, and Andover. The host rocks of this group typically consist of coarse grained magnetite, carbonates, garnet, sulfides granulites and marble with highly variable concentration of plagioclase, amphibole, and pyroxene. The ore mineral is typically magnetite but a few deposits contain red hematite. In each case the deposits are devoid of ilmenite. The magnetite contains 0.01 to 0.25 % TiO2 and 0.06 to 0.75 % MnO. In the case of the Andover deposit the ore consists of manganiferous hematite containing 2.8 % MnO.

Examples:

Warren County Deposits.

The magnetite deposits of Warren County located east of Belvidere, New Jersey are described by Kastelic (1979) as sharply bounded tabular-shaped bodies composed of magnetite and quartz with minor pyrite. Most of these deposits are hosted by Franklin Marble or calc-silicate lenses adjacent to or surrounded by marble. One of the larger mines in this marble belt is the Ahles mine which was described by Bayley (1910) as a deeply altered soft limonite-pyrolusite rock with minor residual magnetite containing up to 11 percent MnO, and very low concentrations of TiO_2, P_2O_5, and S (Table 2). None of the mines in the marble belt have been preserved and are no longer visible although the Ahles mine dump was a source of ore samples. Some of the marble belt mines, including the Kaiser, Barton, Pequest, and Washington mines, are hosted by calc-silicate rocks that Kastelic (1979) interprets as metamorphosed meta-sediments including siliceous siderite muds intercalated with siliceous dolomite. Kastelic (1979) rejects the suggestion made by earlier workers that the magnetite rich calc-silicate rocks are skarns formed by the intrusion of granitic rock and points out the absence of igneous intrusions near the mines of Warren Co. that could have been the source of iron-bearing fluids.

Andover Deposits.

Another example of metamorphosed marine carbonate hosted magnetite is the Andover mining district (Field Trip Stop #4, this guidebook) described by Sims and Leonard (1952). The district consists of two large mines (the Andover and the Sulfur Hill) and two small mines (the Tar Hill and the Longcore). The ore at the Andover mine consists primarily of hematite, fine grained earthy iron hydroxides, red amorphous silica (jasper) and minor concentrations of relic magnetite. The ore at the Sulfur Hill mine is disseminated magnetite with calcite, dolomite, andradite garnet, pyroxene, and pyrrhotite with minor pyrite, marcasite, sphalerite, chalcopyrite, galena, and molybdenite. The Andover ore is interpreted by Sims and Leonard (1952) as the product of supergene alteration of Sulfur Hill type ore.

Franklin Deposits.

Other examples of deposits hosted by marine carbonates are the magnetite deposits associated with the footwall marble of the Franklin zinc ore body. An approximately one mile long band of disseminated magnetite rich marble (the Furnace magnetite bed) occurs between the Franklin zinc ore layer and the adjacent Cork Hill gneiss.

Origin:

The chemical composition of the carbonate hosted deposits, particularly the very low titanium content, resembles typical Precambrian banded iron formation (BIF), (Table 2 and Figure 4). Although the carbonate hosted group, unlike typical BIFs, is not banded with jasper or chalcedony, it may share a similar depositional environment. The protolith of BIF deposits is usually interpreted as marine sediment that has been enriched in iron by either distal or proximal volcanic processes (Gilbert and Park, 1986). Support for such an interpretation is found by examining iron concentrations currently being precipitated near volcanic vents in the East Pacific Rise. Rock composed largely of iron oxide and silica with very low titanium values comparable to BIFs and the Carbonate Hosted magnetite deposits of New Jersey have been found near volcanic vents from the Galapagos hydrothermal mounds, DSDP Leg 70 (Barrett et al., 1988), along the East Pacific Rise (Table 2)

There are several chemical similarities (Table 2). The very low TiO_2 and Al_2O_3 content contents of carbonate hosted deposits from New Jersey closely resemble BIFs such as the Bewabik Iron Formation, particularly ore from the "Siderite Facies" (Bayley and James, 1973). The TiO_2 content of iron enriched volcanic exhalatives is consistently less than 0.1 percent (Table 2) in distinct contrast to the 2 to 25 percent TiO_2 content of magnetite contained in igneous rock (Buddington and Lindsley, 1964) and the 1 to 7 percent TiO_2 content of magnetite contained in the schist and pyroxenite hosted magnetite deposits of New Jersey.

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THE CLINTON MANGANESE DEPOSIT

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ABSTRACT

The Clinton manganese deposit is a relatively unknown mineral occurrence in Hunterdon County, New Jersey after numerous years of sporadic development. Very little of the mineralized area remains undeveloped. The deposit lies south of Clinton Borough, within Clinton Township, at the intersection of the New Jersey Highlands and the Piedmont Provinces. The Jutland Sequence forms the host rock for the mineralization, and it consists of a thick package of variegated shales and minor interbedded sandstone, chert and limestone of Late Cambrian through early Middle Ordovician age suggested as deposited in an upper slope to outer shelf environment (Lash and Drake, 1984). A north, northeast structural trend of the units around Clinton curves into the more general northwest regional trend. Numerous northwest trending folds and faults permeate the Paleozoic section. Northeast trending, southeast dipping normal faults, related to Mesozoic rifting, also cross cut the local area of the mineralization. Mesozoic fanglomerate deposits crop out just south of the manganese prospect.

Workings on the site show a punctuated history. The first record of the prospect comes from Cook (1865) when the mineralization was mistaken for iron ore. Investigations and limited development on the site continued during the 1860's, 1900's, 1918 and lastly in 1938. All work proved to be short lived, as interest seemed to follow the market value for manganese. Historical reports on manganese oxide content vary from 10 to 45% with the mineral assemblage containing pyrolusite, braunite and/or psilomelane (romanechite). Ore lies within a three- to four-foot thick zone and consists of two to three inch seams along bedding planes. Some investigations listed the ore as forming veins. Workings included several pits yielding a combined length of one thousand feet. The largest pit covered a volume of sixty feet long, twenty-five feet deep and six feet wide at the bottom. One ninety-foot shaft was dug to intercept all the zones at depth. Later work mentioned other mineralized regions within the Jutland region.

Previous investigations performed only limited studies on the ore genesis. Supergene enrichment was the preferred process of emplacement and suggested source horizons included Mesozoic diabase or Proterozoic granites and gneisses. This study supports the previous description of mineralization as several new exposures were found elsewhere within the Jutland. This study proposes that the manganese could have initially formed from primary deposition in the deep-water sediments. Subsequent remobilization of the manganese allowed enrichment at the present locale. The existence of a recently discovered bentonite in the Jutland area along with tuffs and volcanic rocks in the related Hamburg Klippe in Pennsylvania display the existence of volcanic activity as a possible source for the mineralization. The research is on going.

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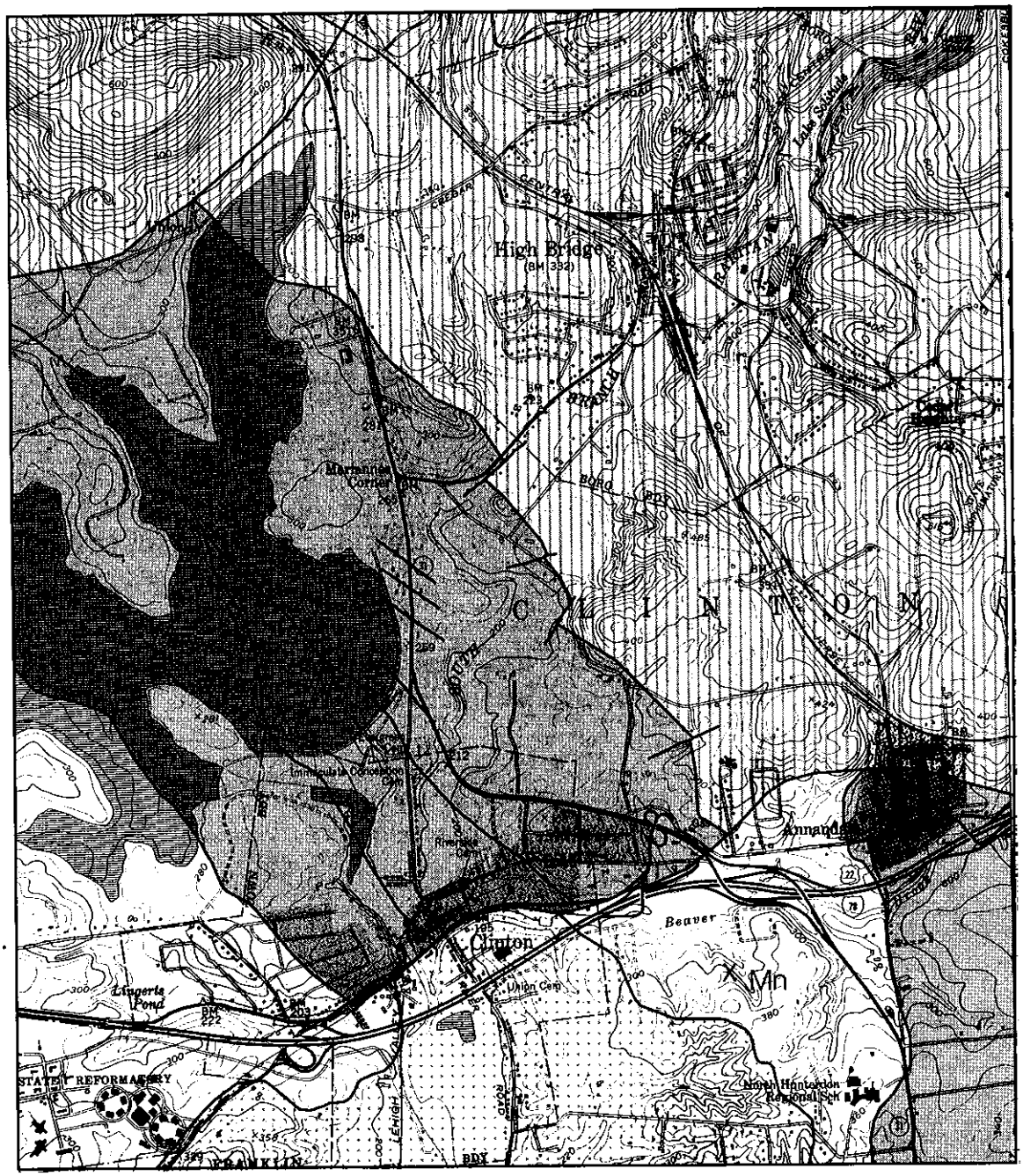

Figure 1. Site geologic map of the Clinton Manganese deposit, High Bridge, New Jersey topotgaphic quadrangle (from Monteverde and others, in press).

Piedmont Province

Mesozoic rocks

X Mn - Location of Clinton Mn deposit

New Jersey Highlands Province

Jutland Sequence

Lower Paleozoic Kittatinny Supergroup

Proterozoic granites and gneisses

INTRODUCTION

New Jersey has a long detailed history of mining activities. Development of this mineral potential drove a prosperous industry, thereby aiding the early growth of the state. The mineral wealth of the state was exploited long before the landing of the first Europeans (La Porta, 1996). Early man quarried and mined the chert found in the carbonate rocks of the Valley and Ridge for various tools necessary for their survival. La Porta (1996) showed that these early "geologists" followed select beds of chert for exploitation. Studies of these early mining attempts suggest a highly organized technology using locally derived quartzite material in quarries more than 200 feet long.

The arrival of colonists from Europe began a new era of mineral exploration. One of the oldest works includes the Pahaquarry Copper Mine. In the 1600's, the Dutch colonists prospected and developed this site along the northwest slope of Kittatinny Mountain within the Bloomsburg Red Beds (Woodward, 1944). Later, more copper mines, located in the Mesozoic Basin rocks, opened in the 1700 through 1900's (Woodward, 1944).

A second important resource in New Jersey was iron ore. Many old iron mines dot the countryside (Bayley, 1910). These mines occur predominantly in granites and gneisses and rarely in carbonate inliers, all within the New Jersey Highlands. They formed a significant part of the industrial development of New Jersey.

One of the last active mines to close was the Sterling Hill Mine, last worked by the New Jersey Zinc Company (Frondel, 1990; Metsger, this volume). The zinc ore comes from the Franklin Marble in the Ogdensburg and Franklin region. This area has one of the longest histories of workings within the state. Spencer (1908) suggested the possibility that the colonists prospected in this deposit as early as 1650. The first authenticated removal of ore is in 1774 (Spencer, 1908).

One commodity rarely associated with the mining industry of New Jersey is manganese. Although it did form a byproduct of other mining activities it was not the primary resource desired. A history exists of prospecting and limited working of a manganese deposit in the Clinton area of Hunterdon County. This prospect crops out within the Jutland Sequence (Perissoratis and others, 1979; Drake and others, 1996; Monteverde and others, in press). Original workers characterized the host formation as Mesozoic rocks (Cook, 1865, 1868) and later within the Martinsburg Formation (Markewicz, 1967; Lyttle and Epstein, 1987). Market fluctuations of the price of manganese drove the speculation and development of this site. The following will outline the geology and history of the Clinton manganese deposit and speculate on possible genesis.

REGIONAL GEOLOGY

The Clinton manganese prospect developed in a structurally complex area in north central New Jersey. A boundary between two physiographic provinces, the New Jersey Highlands and the Piedmont Province, lies immediately south of the prospect(figure 1). It divides the older Paleozoic and Proterozoic units of the Highlands from the younger Mesozoic rocks in the Piedmont to the south.

Proterozoic-aged gneiss and granitoid rocks along with minor amounts of marble comprise the dominant units of the Highlands (Kummel, 1940, Drake, 1984). They record the signature of the Grenville Orogeny (Drake and Volkert, 1991; Drake and others, 1991). Younger rocks of the New Jersey Highlands include early Paleozoic-aged basal sandstones and passive margin

carbonates of the Kittatinny Supergroup (Kummel, 1940; Drake and Lyttle, 1985). The Kittatinny consists of, from oldest to youngest, the Leithsville Formation, Allentown Dolomite, and the Beekmantown Group. The approximately 3000 foot thick Kittatinny forms the New Jersey portion of an extensive carbonate platform developed throughout the Appalachians during this time period. Dolomite and rare limestone, siltstone and shale comprise its lithology. Deposition of this carbonate dominated package in New Jersey was almost entirely restricted to the near shore environment. Age of the Kittatinny ranges from Early Cambrian to Early Ordovician.

Time equivalent to the upper part of the Kittatinny Supergroup is the Jutland Sequence (Drake and others, 1996). It forms an interbedded package of red, green and gray shale, siltstone and minor limestone and sandstone. Monteverde and others (in press) estimated the thickness of the Jutland at 3,800 feet. Markewicz (1967) proposed a near shore environment with intermittent deep water cycles for the middle part of the Jutland and deeper marine conditions for the lower to upper sections. He described it as a member of the Middle and Late Ordovician Martinsburg Formation and stated that the fossil evidence was inconclusive for the older age given by Dodge (1952). Perissoratis and others (1979) suggested deep-water conditions for the Jutland and gave a Lower Ordovician age. Lash and Drake (1984) correlated the Jutland to the Windsor Township Formation of the Greenwich Slice of the Hamburg Klippe, which formed in a continental slope environment. Lyttle and Epstein (1987) mapped the eastern half of the Jutland as Martinsburg Formation while differentiating the western half as Jutland Klippe rocks. Drake and others (1996) and Monteverde and others (in press) maintained the Jutland as deeper water deposits and time equivalent to the Beekmantown Group of the Kittatinny Supergroup. Fossil evidence, consisting of graptolites and conodonts, yields an age ranging from Late Cambrian through Early Ordovician (Perissoratis and others, 1979; Karlins and Repetski, 1989; Repetski and others, 1995; Parris and others, 1995, Parris, written communication, 1996, Parris and Cruikshank, 1986).

The Middle Ordovician Jacksonburg Limestone lies unconformably above the Kittatinny Supergroup. Generally this contact is a disconformity with variations up to 10 degrees. The Jacksonburg consists of a lower fossiliferous, fine-grained limestone that grades upward to argillaceous limestone. It marks a marine transgression that continues to deepen through a carbonate platform margin type into the deep-water flysch of the overlying Martinsburg Formation. Plate convergence from the Taconic Orogeny drove this Middle to Late Ordovician transgression. The Martinsburg does not crop out in the vicinity of the Clinton manganese deposit, but occurs approximately 7 miles to the north in Musconetcong Valley.

Lying unconformably above all older rocks is the Triassic and Jurassic Newark Basin sediments. The major stratigraphic units of the Newark basin include, from older to younger, the Stockton, Lockatong and Passaic Formations. Rift clastic units including red brown siltstone, sandstone and cobble and pebble conglomerates comprise the Stockton and Passaic. Along strike to the southwest, dark gray to black argillite to mudstone from a lacustrine environment dominate the Lockatong and develop thin traceable basin wide cycles in the Passaic Formations. Jurassic-aged diabase intruded the sedimentary units of the Mesozoic basin and lies to the east, upholding Cushetunk Mountain.

All the rock units have undergone some degree of deformation. Mesozoic rocks in the Clinton region formed in a relay ramp between two major fault segments (Schlische, 1992, 1993). These fault segments funneled fluvially derived sediments from the eroding Highlands. Localized rider blocks also exist to the west (Schlische, 1992, 1993). Dominant joint trends formed parallel to the bounding listric fault traces, which trend northeast, and the north trending intrabasinal Flemington Fault. Paleozoic formations portray a variable deformational signature containing both

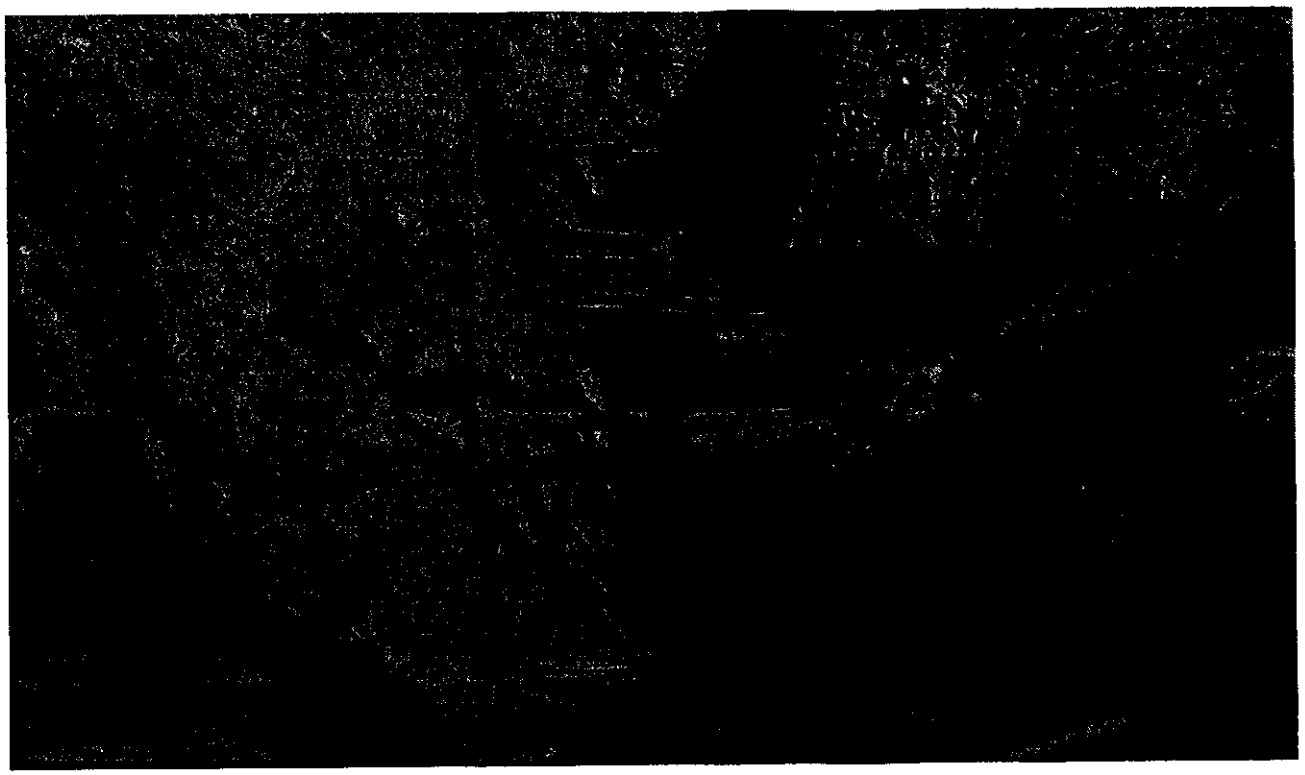

Figure 2. Photograph of one of the trenches dug during the development by the New Jersey Manganese Company in 1918.

reverse and normal faults and folds. Reverse faults, folds and corresponding cleavage and joints trend northwest in the Clinton area, parallel to local strike of bedding. Mesozoic normal faults, trending northeast, related to rifting also cut these units. Ratcliffe and Costain (1985), Burton and Ratcliffe (1985), Ratcliffe and Burton (1985), and Schlische (1992) suggested that the Newark basin bounding normal faults are reactivated Paleozoic-aged reverse faults.

HISTORICAL BACKGROUND OF MANGANESE EXPLORATION AND DEVELOPMENT

The Clinton manganese mine has a punctuated history with frequent reinvestigations proceeding and during times of armed conflict. First note of any investigations in this region appears in Cook (1865). The occurrence, on the land of JT Leigh, future mayor of Clinton, lies between Clinton and Clinton Station. Cook (1868) suggested the original search was for iron ore as "It was said to be hematite." He noted a strong resemblance to manganese and analysis recorded 45 per cent manganese oxide. The material worked at that time consisted of a three or four foot wide vertical vein without apparent gangue. Red sandstone, believed to be part of the Newark Basin, was described as the host rock.

In a later publication, Cook (1868) went into greater detail in the description of this mineral occurrence. He listed the location as near Clinton, on the land of John T. Leigh and the estate of General George Taylor, a Union General killed at Manasses in September 1862. At this time the workings were several discontinuous shallow excavations approximately five feet deep and trending southeast for one hundred and fifty feet in length. He described the vein as approximately ten feet

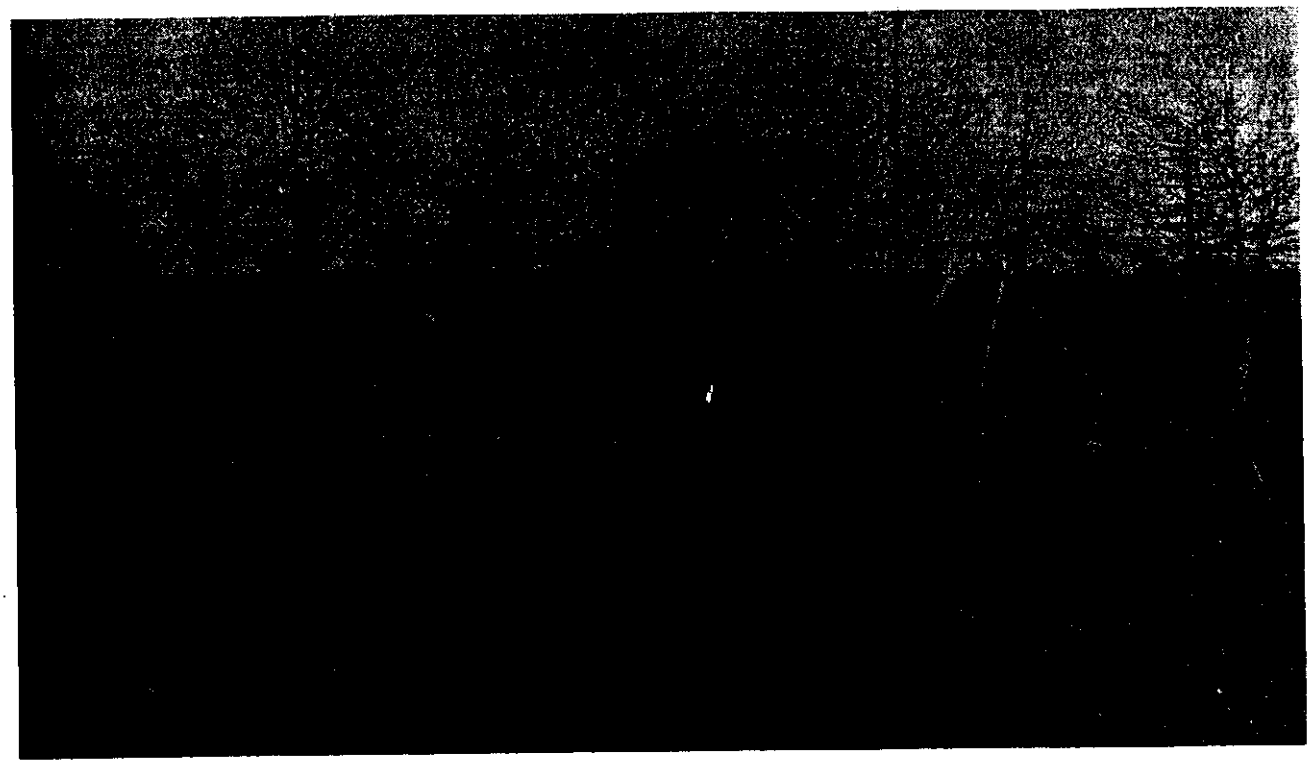

Figure 3. Photograph showing the limited amount of workings during the development in 1918 by the New Jersey Manganese Company. Photograph appears to be looking northwest and the workings would be down the slope towards the north(right corner of picture).

wide and cross cutting the host rock stratification. Cook (1868) stated the ore as 47 per cent metallic manganese in the forms of pyrolusite and braunite. Accessory material included 7 per cent iron and 25 per cent insoluble material. The Newark Basin rocks remained as the host assemblage as Cook (1868) stated "red sandstone and fine quartzose conglomerate" formed the hill containing the workings. Twenty tons were removed, including one railroad car sent to the blast furnaces in nearby Bethlehem, Pennsylvania.

Weeks (1886) noted the initial speculation of the Clinton manganese deposit as iron ore. The manganese oxide mineralogy consisted of pyrolusite and braunite with a chemical analysis of 47 per cent metallic manganese, 7 per cent iron and 25 per cent of insoluble matter. The host rock is again listed as "Triassic rock." The source of Weeks' (1886) information appears to be Cook (1868) as no new information was given. Subsequent information of the ore sent to blast furnaces (Cook, 1868), listed the investigation of the material for use in the production of spiegel iron (Weeks, 1886). Results of the study at the steel works in Bethlehem proved unsatisfactory and production apparently halted.

Smock (1894) used Cook (1868) as a source in noting the manganese occurrence. Smock restated the mineralogy as pyrolusite and braunite and described the host as "Triassic sandstone."

Letters on file at the New Jersey Geological Survey (NJGS), Trenton, New Jersey record a rebirth in the exploration of the manganese mineralization in 1902. J.S. Stewart of Phillipsburg, NJ, as president of a newly founded company with capital of $300,000, initiated prospecting into the

Figure 4. Photograph of one of the excavations of the New Jersey Manganese Company in 1918. Bedding is dipping from the upper left to the lower right. Other parting planes are evident in a near vertical orientation.

extent of the manganese occurrence. Surface digging increased to one thousand feet in length in nine excavations including the original trench from earlier prospecting. Manganese occurred throughout the entire length beginning at six inches and increasing downward to a seven feet wide zone. Water filled the trenches and caused the company to develop an adit ninety feet long with the intent of intercepting all the ore outlined in the trenches at depth. Chemical analysis yielded 30 to 45 per cent manganese with local high contents of iron and phosphorus. U.S. Steel Corporation received samples and recommended the ore quantity be proven. Mr. Stewart's company intended to sell stock to finance the exploration development of the property. NJGS geologists visited the locale, informally named the Crammer site, and collected samples for analysis. No further information of this attempt was encountered.

Interest in the manganese prospect resurfaced during the latter years of World War 1 due to the increased price of manganese from $15.00 to $68.50 for domestic ore. A mining engineer from New York City investigated the site on May 1917 and suggested the deposit represented a "true vein" (Kirby Thomas, written communication to NJ State Geologist, May 16, 1917). There is some question if this inquiry concerns the Clinton manganese deposit or a different one in the Annandale area. Mr. Thomas listed the site as "near Annandale, New Jersey, on the old Fox place, now owned by Mr. Sharp". He further stated the host rock as "schist". This information does not agree with previous descriptions of the Clinton site. Also, a record exists of a psilomelane locality in Annandale but it lacks a complete site description (W.S. Valiant, written list of mineral localities of New Jersey dated 1914, on file at NJGS). Research on maps of the Annandale area (Hunterdon

County Atlas of 1873) outlined several lots in the northern part of the town under the name of Fox. This region falls within a belt of Proterozoic metasedimentary rocks (Monteverde and others, in press). Several graphite mines occur in this area (Volkert, this volume). The lack of a complete site description of the "Annandale manganese occurrence" makes it impossible to definitely know which site Thomas queried. The response from the State Geologist (the copy in the NJGS files was unsigned) to Mr. Thomas only gave a description of the Clinton site. No further correspondence was found on this inquiry.

Activity did commence late in World War I. By February 1918, a company under the title of the Annandale Mining Company actively worked the deposit. At that time Twitchell (written note on file at the NJGS) noted 50 tons of ore removed with manganite as the probably ore mineral. Twitchell described thin layers of ore, two to three inches thick, continuing downward thirty to forty feet and pinching out at folds in the shale with the total ore zone three to four feet thick. Ore locally reoccurred in new layers one or two feet below the last mineralization. The total ore zone measured three to four feet thick. Kummel (written note on file at NJGS) revisited the site in March of that same year. He described the trend of the openings as near north south in an uphill direction. The largest dig continued fifty- to sixty-feet long, twenty to twenty-five feet deep and three to five feet at the bottom (figure 2). The workings remained limited (figure 3). Kummel further described the host rock as "shales, much cleaved and slickensided" (figure 4). He suggested that the rock resembled the regional Triassic rocks but due to the cleaved nature of the rock is better associated with the "graptolite shale west of Clinton." This is the first note of the host rock correlated with the Jutland Klippe Sequence (Drake and others, in press). During this year the company name changed to New Jersey Manganese Company. By November of 1918, an Annandale train agent informed a NJGS geologist that only two carloads of ore were shipped and a further two carloads remained along the tracks (note on file at NJGS). Everything at the mine was abandoned. Hewitt (1921) listed the total output of the mine in 1918 as 64 tons. No value of ore was given.

The last public inquiries on the manganese mineralization occurred in the later 1930's. Correspondence between AR Archer of Somerville, NJ, EP Earle of New York City and Meredith E. Johnson, the New Jersey State Geologist, shows the desire for further information on the geology of the manganese. Background information on Archer and Earle is lacking. Johnson's response, dated May 24, 1938, to Archer's original letter (currently missing from the NJGS files) explains the location of previous published descriptions of the site along with a recent field visit by Johnson. He listed the exposures as poor with no mineralization found greater than 3 inches. Johnson described the host rock as "highly compressed shale beds of the Martinsburg formation" and had no opinion on whether the mineralization is primary or secondary. He further stipulated "From a commercial standpoint the possibility of mining the ore does not at present appear very hopeful. The manganese ore does not constitute more than 5 to 10 percent of the rock that would have to be mined in order to extract it and, of course, the percentage of manganese would be less than one-half, or say a maximum of 5 percent of the material to be mined." He did suggest a thorough program of diamond drilling before abandoning the site. The missive ends with a mention of a second mineralized zone in a "borrow pit adjacent to the main highway about 1-1/2 miles west of Clinton." The Clinton Block Quarry appears to be the location of the second zone alluded to by Johnson. No other information is given.

Archer responded to Johnson's comments in a letter to Earle dated June10, 1938. He agreed with Johnson concerning the discouraging circumstances surrounding the initial site but added the possibility of mineralization in the graphitic rock to the northeast (possibly that queried by Thomas) and on the grounds of the Clinton Reformatory (presently the Edna Mahan Correctional Facility for

Women). Speculation on manganese mineralization according to Archer rested on "soil showing the distinctive purple coloration... which points to a wide distribution of the impregnation." Archer did not recommend the use of private funds in manganese exploration of this area but did "consider the favorable chances quite sufficient to warrant thorough examination with the use of Federal Government funds as part of our national defense economy." Johnson agreed with this assessment. No further communication exists of this inquiry.

On August 29, 1940 Meredith Johnson released a "special feature story" to an unknown source entitled "Occurrence of manganese ore in New Jersey". In the manuscript he described the Clinton site and speculated on it's origin. He delineated manganese occurrences on two hilltops (Clinton site and the borrow pit to the west?) and suggested secondary enrichment as the mode of origin. The process, according to Johnson, included initial dissolving by ground water of an originally diffuse manganese mineralization from overlying sediments. Concentration and redeposition of the manganese under proper conditions followed. Johnson theorized the presence of other, possibly richer deposits "beneath the barren capping of younger Triassic rocks which covers the Ordovician shale south of a curving line which runs through Allerton, Clinton and the former railroad station at Grandin, halfway between Lansdowne and Jutland." No other data were found on this manuscript.

Thurston (1951) produced the first and only detailed study to date of the manganese in Clinton. He placed the deposit, which he called Clinton Point, within the Martinsburg Formation. An atypical assemblage of rock at the site as compared to the Martinsburg of the Kittatinny Valley area was noted. He described the ore as thin discontinuous stringers following fracture planes within the shale and as thicker veins showing a microscopic boxwork appearance. The main thrust of the study was to identify the manganese mineralogy and describe limited examples of the ore habit. Pyrolusite and psilomelane comprised the ore minerals (Thurston, 1951). The deposit was suggested as formed through supergene enrichment though no detailed analysis was given. Thurston (1951) considered the overlying Paleozoic shales (Jutland Sequence) as the most likely source rock for the manganese. Other secondary sources included the Highlands and even less likely, the Mesozoic igneous units.

More recent studies of the region have attempted to elucidate the geological setting of the Clinton site. Markewicz (1967) mapped the geology of the High Bridge quadrangle, including the location of the manganese deposit at a scale of 1:24,000. He agreed with the supergene enrichment theory proposed by Thurston (1951) and showed possible source rocks for the manganese. Markewicz placed the host rock of the Clinton deposit within the Jutland Member of the Martinsburg, a new member he designated. Analysis of weathered limestone layers and lenses showed 5 to 15% manganese oxides present. Reference was made to the limestone undergoing manganese enrichment through percolating meteoric waters. Markewicz (1967) indicated a shallow shelf as the depositional environment for these limestone layers occurring in the middle of his Jutland Member. A direct correlation exists between the Jutland Member as outlined by Markewicz (1967) and the Hamburg Klippe of Pennsylvania (Markewicz, personnel communication 1997).

Lyttle and Epstein (1987) differentiated between the Martinsburg Formation and the Jutland Klippe within the Clinton area. They stated the Jutland and Martinsburg as two unrelated units without any overlap in time, but still mapped the Clinton manganese site as within the Martinsburg. Jutland was only mapped farther to the west. The most recent mapping of Drake and others (1996) and Monteverde and others (in press) placed the host rocks of the manganese mineralization within the Jutland Klippe Sequence(figure 1).

The preservation of the mining activity at Clinton remains poor. Several of the original

trenches have been relocated on the Beaver Brook Golf Course with the aid of a groundskeeper. The largest of the relocated workings forms a filled elongated shallow depression approximately 15 feet long and 4 feet wide. It shows some subsidence at its southern end with a depression approximately 4 feet in diameter and 3 feet deep. Loose blocks fill the cavity and show the possibility of further subsidence. The second excavation has almost totally been closed. Only a very shallow scar remains of the old mine activity. A small dump pile exists alongside the larger hole that has been recently raided for fill material by the golf course groundskeepers. Material collected from the site is red shale that portrays a great deal of fracturing. Manganese mineralization occurs along both bedding planes and fractures. Thicker accumulations of mineralization, up to 0.5 inches, occur on the bedding planes.

Recently other examples of manganese mineralization have been discovered within the Jutland. During the early 1990's construction of a gas pipeline trench exposed continuous outcrop over hundreds of yards in a northwest-southeast direction. It bisected parts of the sequence that are younger than the host rock for the Clinton deposit. The mineralization is strata-bound and occurs as thin lenses of a black manganese ore, generally following bedding planes. It is associated with medium- to dark-gray, micritic limestone, and black and green shale and to a lesser extent red shale. The manganese can reside along foresets in cross-laminated sandstones (K. Muessig, written communication) and also as lenses on bedding. Unfortunately samples were not collected while the trench was open so no further analysis could be performed.

GENESIS OF MANGANESE MINERALIZATION

Detailed study of the genesis of the manganese mineralization in the Jutland is lacking. Previous workers suggested that supergene enrichment played the dominant role in the placement of the manganese (Thurston, 1951; Markewicz, 1967; Meredith Johnson, unpublished notes). They listed mixed gneisses and granitoid rocks of the Highlands and sediments within the Jutland as the most likely sources for the manganese. Localized diabase intrusives of Mesozoic age were also considered as a secondary possibility. Of these reports only Markewicz (1967) gave direct evidence for elevated manganese concentrations as a possible source horizon. He stated that 5 to 15% manganese occurred within the "brown leached porous rock surface" in the thin-bedded limestones within the middle section of the Jutland. The exact mineral species of the manganese was not given. Markewicz (1967) suggested that the manganese percolated downward via meteoric waters and solidified along fractures within the shales. He stated that some of the manganese also enriched still lower limestone layers. It is unclear in this model whether the manganese originated as a primary accumulation during limestone deposition of the Jutland and later remobilized to the present location by supergene enrichment, or the manganese originated from an unidentified unit and moved through the Jutland by a several stage supergene process. An analysis of recent literature on manganese deposition in marine sediments suggests a primary depositional sequence of the manganese in either the limestone facies or the red green and black shales of the Jutland. Limited subsequent remobilization of the manganese could have occurred through a supergene process as professed by earlier workers.

Much work has been done characterizing the mode of formation of sedimentary-hosted manganese deposits (Clavert and Pedersen, 1996; Frakes and Bolton, 1992; Roy, 1992; Okita, 1992; Dagupta and other, 1992; Delian and other, 1992; Force and Cannon, 1988). Manganese deposits generally form in sediments laid down in shallow coastal basins offset from the open ocean by topographic obstructions that hinder the replenishment of dissolved oxygen from deep-water

currents. Therefore they are generally thought to be shallow water deposits (Roy, 1992) though examples are known in turbidites and greywackes from deeper-water environments (Calvert and Pedersen, 1996; Gross, 1990; Xu and others, 1990, for example).

Roy (1997, and papers cited therein) reviewed the mechanism for deposition of manganese in sedimentary rocks within a stratified ocean basin. Two general processes exist to deliver manganese to a marine setting. From a continental source, fluvial channels carry the eroded products from weathered bedrock and dump these manganese bearing materials into the near shore environment. Manganese is also directly input to marine surroundings through hydrothermal mid-ocean ridge systems, mid-plate seamounts and subduction related island-arc settings (Iijima and others, 1990; Liu, 1990; Nath and others, 1997; Roy, 1997; Varentsov and others, 1990). Manganese introduced hydrothermally may disperse thousands of kilometers from its source (Roy, 1977). Once in the marine environment the manganese needs to intensify greater than 200 times the average crustal concentration of 0.15 percent (Force and Cannon, 1988). Average concentrations in marine sediments range between 0.17-0.38 percent manganese (Baturin, 1988). A stratified ocean aids in enrichment of manganese. Dissolved manganese consolidates in the deep anoxic water layer due to Eh and pH conditions (Force and Cannon, 1988). The metallic ions move to the interface of the anoxic-oxygenated water by vertical advection-diffusion. Above this interface, the manganese oxidizes to a manganese oxyhydroxide solid. This particulate matter settles back through the interface where it again dissolves. The degree of precipitation of the manganese compounds relates to the amount of mixing in the water column (Force and Cannon, 1988). Actual deposition of the manganese only occurs where the redox front intersects the sloping basin margin. Preservation of the precipitate requires the newly formed mineral to intersect the basin bottom before crossing the redox interface. Force and Cannon (1988) suggested the manganese crystallized in an oxide habit or within carbonates depending on the pH and Eh conditions of the environment. Oxide facies develop on margins under oxygenated conditions and at the oxic-anoxic interface (Maynard and other, 1990) in shallow water while manganese bearing carbonates form on either oxic (Frakes and Bolton, 1992; Force and Cannon, 1988) or reduced substrates in deeper water. Roy (1997) stated that manganese-rich limestones are the most susceptible for remobilization and subsequent supergene enrichment in the weathering zone.

Others suggest a different interplay of the redox front not necessarily within an initially stratified ocean. Calvert and Pedersen (1996) stated that oxyhydroxides of manganese form in oxic environments while manganese carbonates precipitate in anoxic conditions within the sediment pile directly beneath the oxygenated waters. They suggested that the deposits originally formed in oxygenated conditions. Due to rapid burial of organic material and slow resupply of oxygen by either diffusion or irrigation, conditions become anoxic at relatively shallow depths in the sedimentary layers (Calvert and Pedersen, 1996). Microbes utilize the oxygen in the sediment in breaking down the organic material thereby creating the oxygen deficient conditions within the sediment pile. Sediments formed under these conditions display a unique reddish to dark brown outer layer (oxygenated) which grades down to olive-green sediments (reducing). Manganese and iron oxyhydroxides are present in the reddish surface layers. Roy (1997), in a review of Phanerozoic manganese deposits, stated that manganese oxyhydroxide might evolve early before undergoing reduction during the process of organic oxidation that leads to manganese carbonate, similar to the theory stated by Calvert and Pedersen (1996).

Ground water has also been shown to influence the formation of a manganese deposit during diagenesis. Force and others (1986) proposed a model whereby the mixing front between fresh and saline ground water influenced the type of manganese mineral phase to crystallize. The model

consisted of initial deposition in an oxygenated environment that recrystallized during diagenesis under an anoxic front governed by ground water. They gave an alternative hypothesis where the ground water front could be responsible for the first introduction of the manganese to the region.

Roy (1997) further associated the stratified ocean and thereby the manganese deposition to a globally warmer paleoclimate concomitant with marine transgressions. Xu and others (1990) correlated a marine transgression caused by glacial melting as responsible for deposition of the late Proterozoic deposit in the Datangpo Formation of south China. They proposed a generally cooler climate than that of Roy (1997). According to the warmer climate model, ocean water mixing stops due to the shutting down of deep-water currents such as the North Atlantic Deep Water. This lack of mixing of deep ocean currents with the ocean margins creates the development of stratified ocean basins. During subsequent transgressions, waters flood across shallow continental regions. These ideal circumstances cultivate increased biologic activity in the newly flooded areas. An augmentation in organic activity enhances the supply of dead matter to the deep-water environment and the basin floor. Break down of the carbonaceous material utilizes all the available oxygen and strengthens and enlarges the anoxic environment (Frakes and Bolton, 1992). Decomposition of the organic rich sediment in an anoxic environment leads to the development of black shales. Roy (1997) reviews the association between black shale basins and manganese mineralization in the various models of manganese deposits. It creates a vertically stratified sequence of manganese oxyhydroxide in the oxygenated layer, followed deeper to manganese-rich carbonates at the oxic-anoxic front and finally black shales in the anoxic layer. Others suggested that deposition of black shale does not mandate anoxic conditions. Initial formation could occur in an oxygenated environment so that the anoxic event happening within the sediment layers themselves (Calvert and Pedersen, 1998).

DISCUSSION

Did the manganese in the Clinton site originate as a primary feature of the Jutland Sequence? Although a detailed study needs to be completed, several characteristics of the Jutland allude to a possible answer. A potential source of the manganese exists within the Jutland as both submarine volcanics and tuffs have been described in the Hamburg Klippe in eastern Pennsylvania (Lash and others, 1984; Ganis, 1997), a unit often correlated to the Jutland (Lash and Drake, 1984; Perissoratis and others, 1979; Markewicz, 1967). An exposure of bentonite has been discovered within the Jutland. The bentonite could be a second location for one described by previous workers (M.E. Johnson and H. Herpers, permanent notes on file at the NJGS). This places igneous activity, and correlative elevated manganese content in the region during deposition of the Jutland. Equivalent deep-water sedimentary deposits in Newfoundland also contain increased manganese and other metals. Submarine igneous activity is responsible for dispersing the metal enriched hydrothermal fluids into the ocean as recorded in the Humber Arm Allochthon (Botsford and Sangster, 1990).

The depositional environment of the Jutland fulfills the criteria proposed for the manganese models discussed above. Sedimentary structures including turbidites and disorganized limestone conglomerate and glauconite, suggest an upper slope to possibly outer shelf depositional environment for the Jutland (Flugel, 1982; Lash and Drake, 1984; Perissoratis and others, 1979; Markewicz, 1967; Dodge, 1952). Dominant fossil assemblages include graptolites and conodonts (Parris and others, 1995; Parris and Cruikshank, 1986; Repetski and others, 1995). Conodonts found to date fall within the North Atlantic Realm, which consists of a colder, deeper water affinity

then the North American Mid-continent Realm developed in the time equivalent platform carbonates of the Kittatinny Supergroup (Karlins and Repetski, 1989; Repetski and others, 1995).

Both recent and historical study of the region encountered manganese in several locations within the sedimentary assemblage (Johnson, written note; H. Herpers, written note; Markewicz, 1976; Muessig, written communication). It arises as a strata bound deposit, both along fracture planes and bedding planes, in thinly interbedded, red, green and black shales and brown altered limestone. Alternation of red and green shales supposes the changing of oxygenated and reduced conditions. To date the manganese mineral assemblage has not been worked out in both the limestone and shales. Samples of the shale bearing manganese exist but so far have eluded characterization. Locations of elevated manganiferous limestone, as described by Markewicz (1967) have not been refound. These sites must be located and mineralogy described to validate the models proposed. Although supergene enrichment may have altered the location of some of the mineralization, the stratabound nature of the deposit insinuates more of a primary origin. So far the construction of interbedded red, green and black shale and limestone meets the models as proposed by Calvert and Pedersen (1996) and in the review papers of Roy (1997, 1992). It is unsure which of the oxidation front models, stratified ocean of Roy (1997, 1992) or within the sediment pile of Calvert and Pedersen (1996) supplies the best explanation for the development of the deposit.

Most of the models propose a stratified ocean to allow the deposition of the manganese in the sediments (Roy, 1992, 1997; Force and Cannon, 1988; Frakes and Bolton, 1992). Botsford and Sangster (1990) in their study of correlative deposits in western Newfoundland found changing conditions of ocean circulation. A stratified ocean developed during the late Cambrian and the Arenig. The Tremadoc had more ventilated conditions from better ocean current conditions.). Botsford and Sangster (1990) suggested that the alternating conditions of a stratified ocean controlled the deposition of manganese and other metals in the deep-water sediments. Cambrian and Ordovician carbonates portray many cycles of transgression and regression (Koerschner and Read, 1989; Read, 1989; Taylor and others, 1992). Warmer conditions prevailed during this time period. This suggest that all the necessary criteria for manganese deposition as proposed by Roy (1992, 1997) prevailed during the time period of Jutland sediment deposition.

CONCLUSIONS

The Clinton manganese deposit is a relatively unknown segment of the mining history of New Jersey. Although it never arrived as an economic deposit, it did elicit interest during years of armed conflict. It was initially thought to be hematite by the early explorationist. Geologists later identified manganese as the ore. Previous workers identified pyrolusite, braunite and psilomelane as ore minerals. Manganese ore developed in fractures and along bedding planes. Mining of the ore in highly fractured red shale consisted of individual pits and a single adit. Historical documents state that two train carloads were sent to the Bethlehem steel mill for analysis possible use in steel production. Results proved unsatisfactory. The period of greatest development was in the years 1917 and 1918 when the price of manganese rose from the need for armaments in the later part of World War I. During the 1930's the final interest in the deposit peaked. Previous workers stated supergene enrichment as the method for metal emplacement.

Cambrian and Ordovician Jutland Sequence constitutes the host formation of the Clinton deposit. It contains an interbedded package of red, green, and black shale and siltstone and subordinate amounts of limestone and sandstone. Depositional environment ranges from upper slope to outer shelf for the sediments. Manganese has recently been found in several other locations

in the Jutland as stratabound, generally parallel to bedding.

A preliminary assessment of the deposits suggests that the mineralization resulted from primary deposition of manganese ore in to the Jutland sediments. Igneous activity supplied the manganese to the marine column of water. Deposition formed under an oxygenated and reduced interface. This could have developed in a stratified ocean as suggested by Roy (1997, 1992) or in the sediments themselves as described by Calvert and Pedersen (1996). Clearly, further analysis needs completion before a definitive mode of origin can be selected for the Clinton manganese deposit.

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EIGHTEENTH AND NINETEENTH CENTURY COPPER MINING IN NEW JERSEY

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Copper extraction in New Jersey almost certainly began in prehistoric times. Early setters found copper nuggets near Somerville and New Brunswick, the largest weighing 74 pounds. Copper carbonate ore-minerals, common at New Jersey mines, are bright green in color and are easily spotted by lay people. Aboriginal copper tools have been found in New Jersey, but aboriginal mining sites are not yet known. Popular historical accounts credit Dutch settlers from Kingston, New York with the discovery of the Pahaquarry Copper Mine near the Delaware Water Gap. Although a road called Old Mine Road was built from Kingston to the Delaware Water Gap by 1700, no physical evidence of Dutch miners has been found at the Pahaquarry site. A Hollander, Arent Schuyler, found a copper mine on his estate around 1712. Many New Jersey copper mines were exploited by British colonists using Welsh miners. The Griggstown Mine was known to have employed 160 Welsh miners during 1765. Ore was discovered during farming activity, digging water wells, actual prospecting, and by excavation at New Brunswick after the mysterious appearance "of a blue flame, about the height of a man" in a field. British Law originally limited the processing of ores to concentrating by hand-cobbing; therefore, the product was then shipped to Swansea, Wales. Later concentrating techniques included stamp mill and roll crushing, followed by washing and, at Flemington during 1835, the use of buddies after a table wash. Smelting at or near the mine became common during the nineteenth century. A 12-stamp mill at New Brunswick was well-suited to native copper recovery, and employed a 1 -mile long tunnel as a head-race for water power. The British smelting tradition employed roasting followed by reduction in reverberatory furnaces. As early as 1784, a German smelter worker produced copper near Somerville, most likely using blast furnace technology. The technology employed underground by copper miners in New Jersey depended upon the mine-site, the geometry of the mineralization and the local underground conditions. Initially open-cut mining started a mine; however, at major mines tunneling, shaft sinking, drifting, crosscutting, raising and stoping was necessary to extract the ore. Excavation went on by pick and shovel at some mines, with hand-steel drill and blast techniques used as needed. Mine props were often constructed from local timber. Stoping techniques included undercut, room and pillar, breasting and gallery. Water control and ventilation both presented problems. Water was initially handled by drain tunnels and air supply was achieved by ventilation shafts. The first steam engine in America was a Neucomen atmospheric engine, used in 1753 to power Cornish pumps underground at the Schuyler Mine. Steam-powered pumps, fans and compressors were widely used during the 19th century. From 1830 to 1870, a burst of reports, investigation, promotion and mining activity correlated with the discovery of the Michigan Keweenaw Peninsula copper deposits in Michigan. The geologic setting of copper in basalt flows and in close association with diabase dikes makes many New Jersey copper mines similar to those in the Michigan Copper Range, an observation fully exploited by nineteenth- and twentieth-century stock

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promoters. In Michigan the initial investments exceeded profits by four to one, in New Jersey no records exist, but this ratio is likely much higher.

ROAD LOG OF THE ECONOMIC GEOLOGY OF NORTHERN NEW JERSEY FIELD TRIP

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INCR.	CUM.	REMARKS
	0.0	Parking lot of Marriott Hotel
0.0	0.0	Leave parking lot of Marriott Hotel on Route 10 East
0.6	0.6	Enter I-287 North
2.6	3.2	Boonton Reservoir on right. From here north we repeatedly cross the trace of the Ramapo fault that forms the structural boundary between the Highlands on the west and Mesozoic rocks of the Newark basin on the east.
3.8	7.0	Low outcrop in median of conglomerate of Upper Jurassic Boonton Formation.
1.5	8.5	Gabions here north on right contain local bedrock from quarries to the north

In Benimoff, A.I., and Puffer, J.H., (editors), The economic geology of northern New Jersey: Field Guide and Proceedings of the fourteenth annual meeting of the Geological Association of New Jersey, 1997, p 117-128.

Figure 1. Field trip route with stop locations plotted onto a portion of the Bedrock Geologic Map of Northern New Jersey (Drake et al., 1996).

1.1	9.6	Steep roadcuts on left, and for next several miles, of mylonitic Middle Proterozoic rocks along the footwall of the Ramapo fault. Note the strong brittle overprints visible as slickensides and extensive fracturing of the bedrock. Observe the anchor bolts, shotcrete, ditch below road level and rock barriers for slope stabilization.
3.8	13.4	Exit I-287 North
1.7	15.1	Right onto Mathews Ave.
0.7	15.8	Bear left
0.1	15.9	Turn Left into Rock Creek Crossing
0.1	16.0	STOP 1

STOP 1: GRAPHITE NEAR BLOOMINGDALE

The residential development here is situated in part over the abandoned Bloomingdale graphite mine which was active in the late 19th century. At the present time all of the original workings have been developed over. See Volkert (p. 24, this volume) for a more complete description of the Bloomingdale mine.

Artificial cuts expose graphite layers trending N30° E to N55°E that vary in thickness from a few inches to seven feet and occasionally contain locally abundant pyrite. The Bloomingdale graphite deposit is hosted mainly by rusty-weathering biotite-quartz-feldspar gneiss and pegmatite. Graphite occurs in the former as plates disseminated throughout the rock and in somewhat more massive form in layers that are conformable to foliation and layering. Graphite in the pegmatite occurs as large plates up to 1 inch in diameter that form around, and are embedded in, feldspar grains mainly along the contact with graphitic gneiss. Pegmatite lacks graphite away from this contact.

Lenses of light green, medium-grained diopside-rich rock (diopsidite) that grade into diopside + plagioclase ± quartz ± titanite ± pyrite gneiss are conformably layered with graphitic biotite-quartz-feldspar gneiss and metaquartzite. Diopsidite contains local disseminated graphite plates that coarsen along the contact with pegmatite. Thin veins <1 foot to about 2 feet thick of buff to light green, coarse-grained pegmatite occur along the contact between graphitic biotite-quartz-feldspar gneiss and diopsidite. Pegmatite is composed of two distinct mineral assemblages: 1) perthitic microcline + plagioclase + quartz + biotite ± graphite and 2) diopside + plagioclase quartz ± titanite ± graphite. X-ray analysis shows the diopside in the pegmatite and in the diopsidite to be nearly identical. These pegmatites represent local melt from biotite-quartz-feldspar gneiss and diopsidite respectively.

Graphite layers in biotite-quartz-feldspar gneiss and diopsidite formed largely through the metamorphism of carbonaceous material in the precursor sediments. Pegmatite subsequently obtained carbon from these rocks and precipitated graphite along its contacts and in other permeable zones.

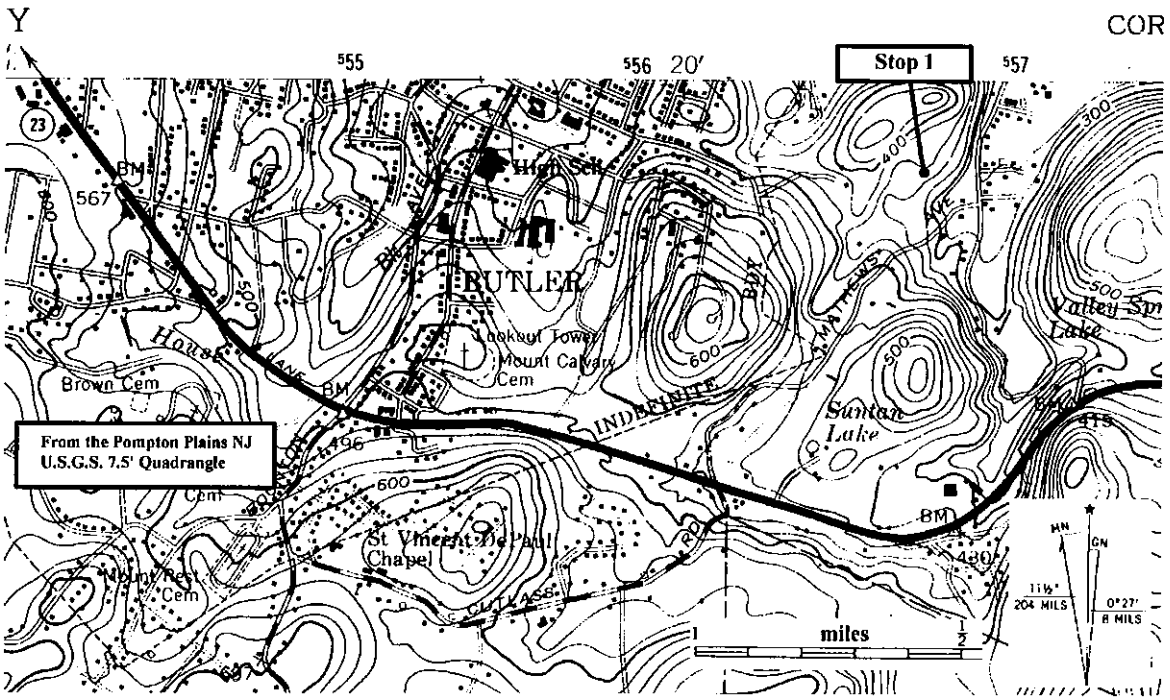

Figure 2. Site map showing Stop 1. From U.S.G.S. Pompton Plains, NJ 7.5' quadrangle.

0.9	16.9	Retrace route to Route 23 North
0.3	17.2	Outcrop of biotite-quartz-plagioclase gneiss on left
3.3	20.5	Diorite outcrops on right and for the next several miles
5.3	25.8	Outcrop of hornblende granite of the Byram Intrusive suite
6.4	32.2	Charlotteburg Reservoir on left. Prominent topography on left(Copperas Mountain) and on right(Kanouse Mountain) is underlain by Silurian Green Pond Conglomerate that unconformably overlies Middle Proterozoic rocks midway upslope. For the next several miles we are traversing a major down-faulted synclinorium of predominantly Silurian and Devonian sedimentary rocks of the Green Pond Mountain region.
0.2	32.4	Syncline-anticline fold pair on right. and anticline hinge in center median in quartzite facies of Green Pond Conglomerate.
3.7	36.1	Oak Ridge Reservoir on left. Crossing the trace of the Reservoir fault that forms the structural boundary between Paleozoic rocks of the Green Pond Mountain region on the east and Middle Proterozoic rocks.
0.2	36.3	Deformed outcrop of amphibolite on footwall of Reservoir fault.
0.2	36.5	Outcrop of pyroxene granite of the Lake Hopatcong Intrusive Suite on both sides of road for next mile. Note the thick, rusty, sulfide-rich seams on right.
0.7	37.2	More pyroxene granite
0.9	38.1	Outcrop of pyroxene gneiss on left
2.4	40.5	Outcrops of quartz-oligoclase gneiss of the Losee Metamorphic Suite in core of major northwest-overturned antiform on both sides of road for next couple of miles.

2.6	43.1	Left turn onto Route 517 South
1.6	44.7	Right turn onto Passaic Ave.
0.6	45.3	Left turn into mine entrance
0.1	45.4	STOP 2

STOP 2 - STERLING HILL ZINC MINE

The tour through the Sterling Hill Zinc Mine will be conducted by Robert Metsger who was the chief geologist during the mines most productive years and during its final closure in 1986. Robert Metsger is generally recognized as New Jersey's most distinguished mining geologist and we (the editors of this guidebook) are particularly pleased that he has agreed to participant in this conference.

The mine, since its closure has been rescued from abandonment by Richard and Robert Hauck who have converted it into a museum under their supervision. Although the museum includes an excellent mineral collection in the "Exhibition Hall" together with very interesting mill artifacts, most of our tour will focus on geologic aspects of the marble hosted ore body.

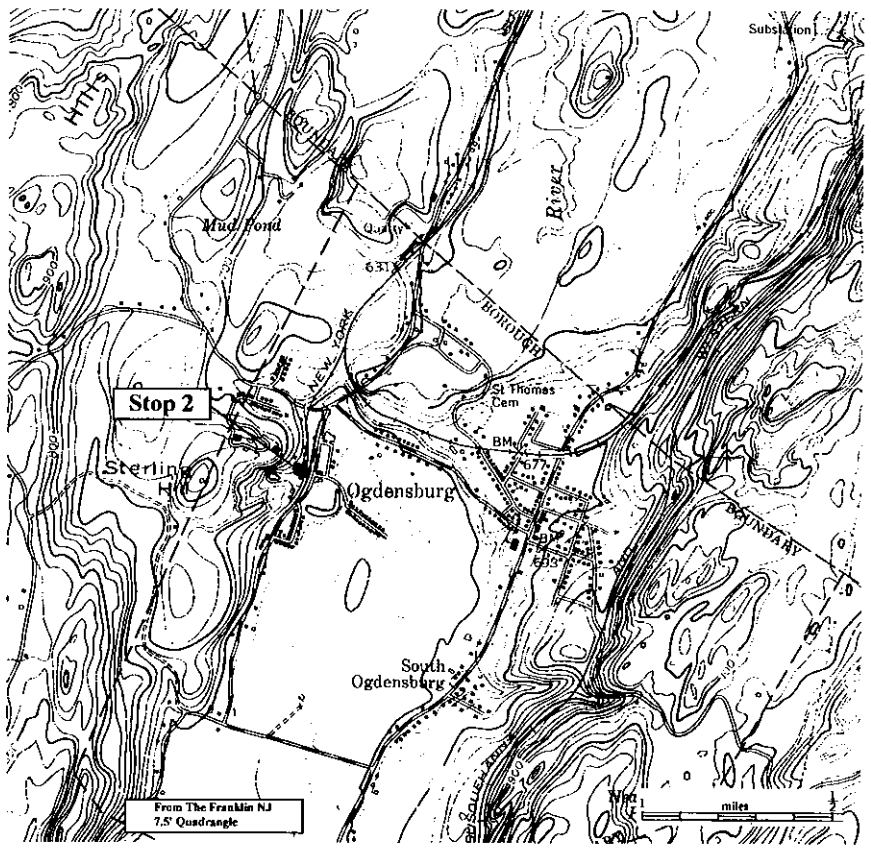

Figure 3. Site map showing Stop 2. From U.S.G.S. Franklin, NJ 7.5' quadrangle.

0.1	45.5	Right turn onto Passaic Ave.
0.7	46.2	Right turn onto Route 517 South
2.9	49.1	Continue straight through traffic light
0.4	49.5	Crossing over NJ Route 15
6.1	55.6	Outcrops of Allentown Dolomite on both sides of road
0.5	56.1	Continue straight through light onto Route 517 South
1.3	57.4	Continue straight through light onto Route 616 West
0.3	57.7	Turn right onto Sussex Mills Road
1.1	58.8	Outcrop of Franklin Marble on right
0.5	59.3	Turn right onto Route 669 North
1.4	60.7	STOP 3 Plant Entrance - Medusa Minerals

STOP 3 LIME CREST QUARRY

We are grateful to Medusa Minerals Inc ,the new owner of the quarry, for granting GANJ permission for today's visit (November 1, 1997) to the Lime Crest Quarry. Permission must be obtained from Medusa Minerals Inc., P.O. Box 217, Sparta, NJ 07871 for any future visits.

Exposed here at the Lime Crest quarry is the Franklin Marble. The quarry has been in operation for more than 75 years, and the Franklin Marble that occurs here is generally similar in all respects to that found at the type locality. The quarry has expanded to include part of the microcline (Cork Hill) gneiss, that overlies the marble, and the Cambrian Liethsville Formation. Other features of interest include faults, related to the fold and thrust belt recently described by Herman et al. (1997), and the secondary mineral assemblages that some of these faults contain. The Lime Crest quarry produces a wider variety of crushed stone products than the standard group of construction aggregates produced by most quarries. The white color and unusually coarse crystallinity of the marble makes possible the fabrication of many additional materials with architectural and landscaping applications.

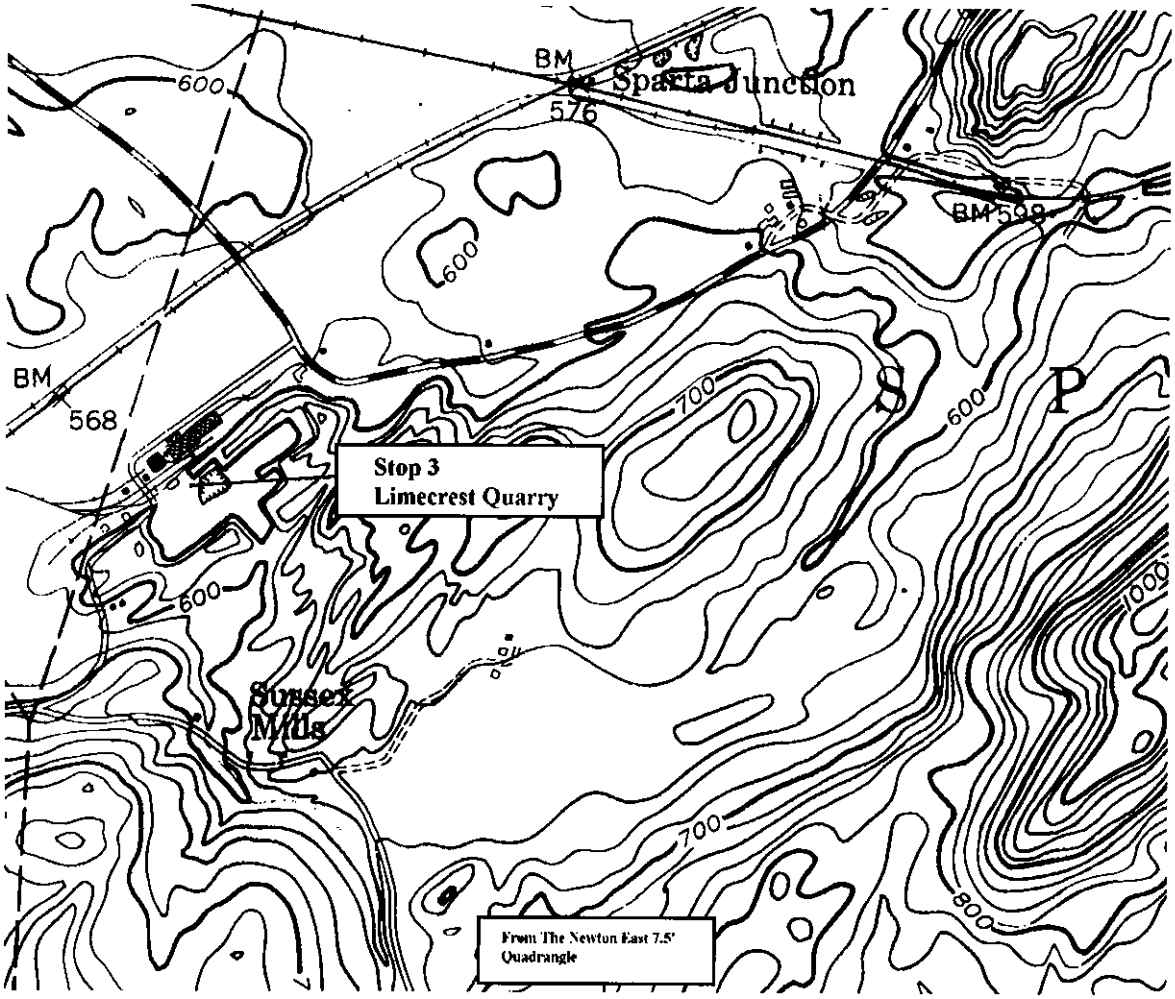

Figure 4. Site map showing Stop 3. From U.S.G.S. Newton East, NJ 7.5' quadrangle.

0.0 60.7 Continue on Route 669 South

3.0 63.7 STOP 4 The Sulfur Hill and Andover Iron Mines

STOP 4 THE SULFUR HILL AND ANDOVER IRON MINES

Proceed east down the mine road about 300' to the open trench to the south. This is the Andover Mine (Figures 6 and 7). The Sulfur Hill Mine is located about 500 ft. northeast of the Andover Mine. Excellent samples of several different sulfides and garnet in a carbonate matrix can be found in the mine dump located between the two mines. The Andover Iron Mine was worked from some time before 1763 until 1863 while the Sulfur Hill mine was worked from 1855 until 1880 (Sims and Leonard, 1952). The Andover Mine was abandoned when most removable ore was mined out, while the Sulfur Hill Mine was abandoned because of problems with the high sulfur content of the ore.

More ore was taken out of the Andover Mine during its 100 years of operation than the Sulfur Hill Mine during its 25 years of operation and has left larger excavations. The

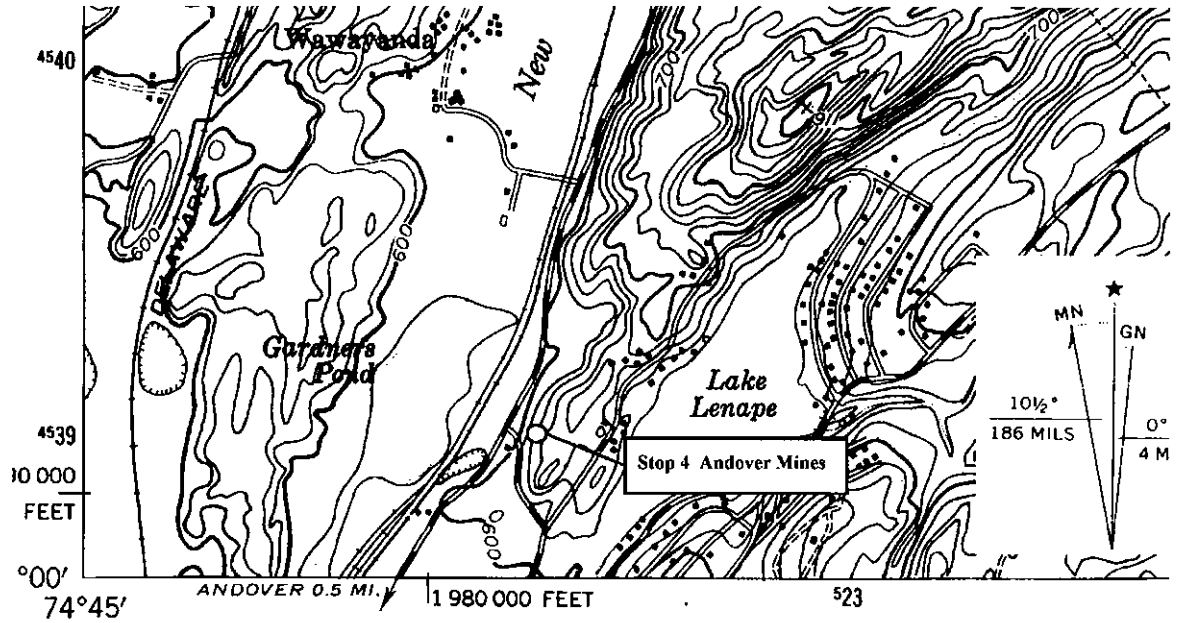

Figure 5. Site map showing Stop 4. From U.S.G.S. Newton East, NJ 7.5' quadrangle.

Andover mine consists of an open trench about 750' long, 70' deep, and 50' wide that is partially filled in plus some underground workings beneath the hill at the NE end of the trench (Figure 6).

The SE wall of the trench is a fault plain with common slickensides that have generated a chloritized mylonite zone veined with quartz. The NW wall is also sheared and highly altered. The foliation of the metamorphic host rocks strikes N35°E and dips almost vertically.

The Andover ore is largely a red hematite ore with residual magnetite in a chloritized, highly altered matrix. Sims and Leonard (1952) suggest that the hematite was formed by supergene alteration of magnetite. The descending supergene solutions were presumably the kind of low temperature and oxidizing solutions. The protolith of the Andover hematite ore is probably Sulfur Hill type ore which consists of undeformed coarse grained granoblastic rock. The ore resembles impure marble or skarn and is composed of clacite, garnet, magnetite, pyroxene, and pyrrhotite in widely varying proportions with accessory black sphalerite, galena, chalcopyrite, and molybdenite. Several similar iron oxide deposits occur in the Franklin Marble some of which have been described by Kastelic (1979) although most are not as richly mineralized with sulfides as the Sulfur Hill Mine.

The geology of the Andover area has been mapped several times with widely differing interpretations. The Franklin Furnace Quadrangle mapped by Kummel and others (1908) locates the Andover and Sulfur Hill mines in a north-east trending one to one half wide band of hornblende, pyroxene gneiss (Pochuck gneiss) that is parallel to a

band of quartz oligoclase gneiss (Losee gneiss) located less than one half mile to the east. Kummel and others map several thick lenses of Franklin Marble within the band of Pochuck gneiss.

A much more detailed map by Sims and Leonard (1952) locates the Andover mine in a narrow (200' wide) band of quartz oligoclase gneiss (Losee gneiss ?) and the Sulfur Hill mine in a lens of carbonate "skarn" (Figure 6). Simms and Leonard map a band of Microcline Granite Gneiss along Limecrest Road just west of the mines and bands of pyroxene feldspar gneiss and amphibolite just east of the mines (Figure 6). Sims and Leonard (1952) also interpret some of the highly chloritized rock exposed along the Andover mine as a diabase dike and suggest that it intruded during the Triassic. Additional evidence of diabase, however, has not been found during any of my approximately 20 searches through the mine area.

Recent mapping (The Newton East Quadrangle by Drake and Volkert, 1993) locates the mines in Microcline gneiss between two branching bands of Amphibolite located about 300' to the north west and south east of the mine (Figure 7). They locate a band of Marble about 1/4 mile east of the mines and biotite quartz oligoclase gneiss (Losee gneiss) over 1/4 mile north of the mine. The Microcline gneiss is described by Drake and Volkert (1993) as a meta-sedimentary, medium grained, grayish-orange to pale pink, well-layered, and moderately well foliated gneiss composed largely of quartz and microcline, and lesser oligoclase with accessory biotite, garnet, magnetite, and locally sillimanite.

Map detail at the scale published by Sims and Leonard (1952) is not possible at the scale published by Drake and Volkert (1993) and the absence of map conformity may not be as extreme at implied by comparing Figure 6 with the Newton East Quad. The Drake and Volkert (1993) description of Microcline gneiss allows for variable oligoclase content and each of the three maps agree that amphibolites are part of the mine area. The locations of a few samples that I have recently collected from the mine area are plotted onto Figure 6. At least four lithologies are clear on the basis of careful mineralogical identifications including sodium cobaltinitrite staining of K-spar:

1. Pyroxenite (mine dump samples)
2. Carbonate, magnetite, garnet, sulfide granulite (mine dump samples)
3. K-spar gneiss (locations 1, 2, 3)
4. Quartz oligoclase gneiss (locations 4, 5)

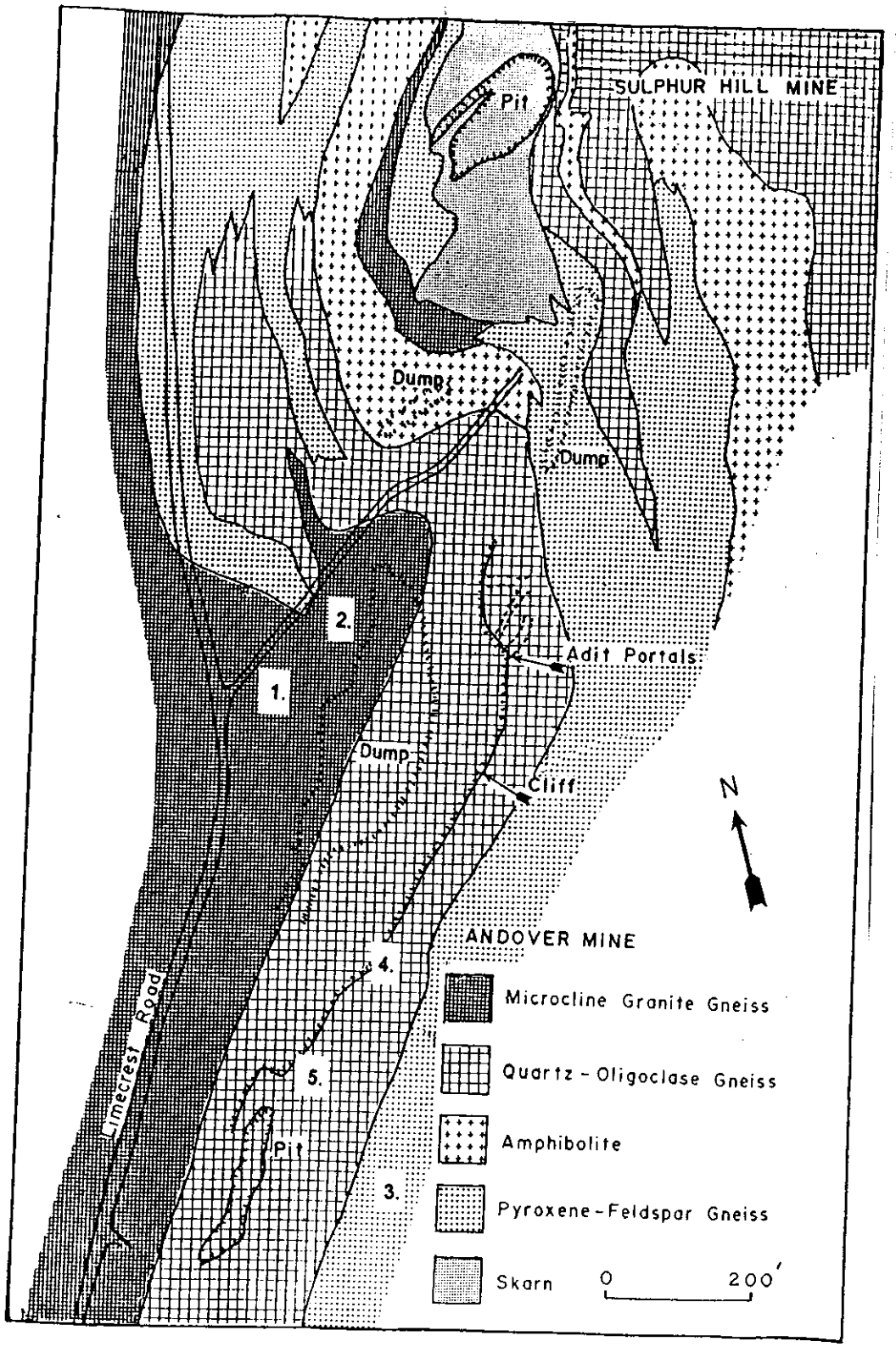

Figure 6: Geologic map of the Andover mining district after Sims and Leonard (1952). The skarn unit here is a carbonate-magnetite-garnet-sulfide rock. On the basis of recent mapping by Drake and Volkert (1993) there is probably less quartz-oligoclase gneiss and more microcline gneiss than indicated. Map locations 1, 2, and 3 are bedrock exposures of microcline gneiss; locations 4 and 5 are bedrock exposures of quartz oligoclase gneiss.

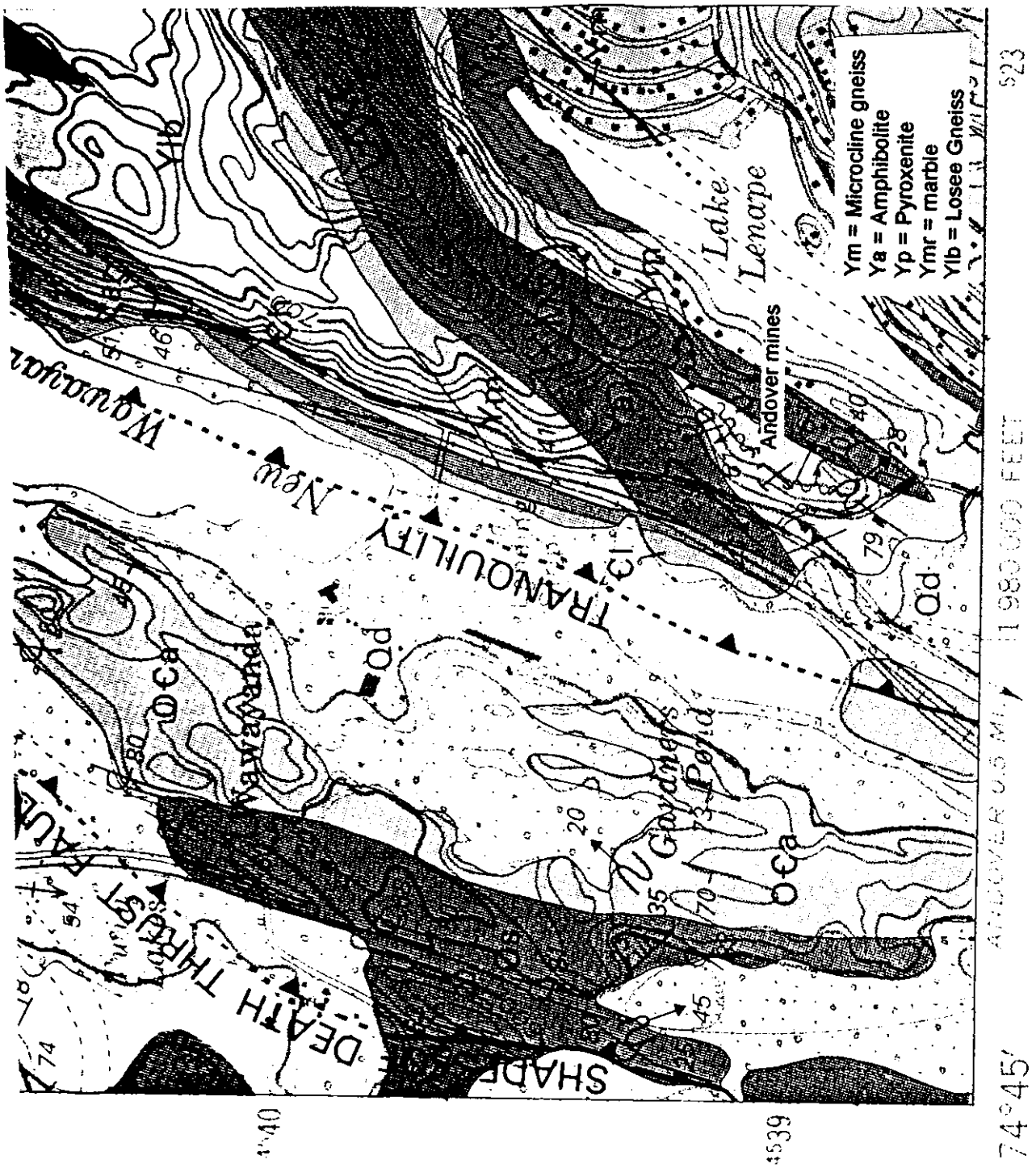

Figure 7: Geologic map of the Andover and Sulfur Hill mine area, after Drake and Volkert(1993)

0.0	63.7	END OF FIELD TRIP - RETURN TO HOTEL
0.8	64.5	Turn left onto US Route 206 South
5.8	70.3	Bear right and follow signs for I-80 East
1.2	71.5	Enter I-80 East
17.4	88.9	Exit onto I-287 South
2.2	91.1	Exit onto NJ Route 10 West
0.9	92.0	U-turn
0.7	92.7	Parking lot of Marriott Hotel

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