Hydrothermal Alteration, Mass Transfer and Magnetite Mineralization in Dextral Shear Zones, Western Hudson Highlands, NY

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INTRODUCTION

The Reading Prong is a Grenville basement massif that forms part of the spine of the Appalachians, connecting the Blue Ridge and Green Mountain provinces, in Pennsylvania, New Jersey, New York, and into Connecticut (Figure 1). Known as the New Jersey Highlands and the Hudson Highlands in New York, it hosts thousands of iron deposits, the first discovery of which was as early as 1730 (Lenik, 1996). These deposits were extensively mined throughout the 18th and 19th centuries (Lupulescu and Gates, 2006).

Earliest investigations of the iron districts in the Highlands focused on the ore deposits (*e.g.* Kitchell, 1857; Cook, 1868; Wendt, 1885; Ruttman, 1887), and generally accepted an origin from metamorphosed sediments. Rodgers (1840), however, proposed that the magnetite deposits were of igneous genesis. Colony (1923) later proposed that the ore deposits were the result of magmatic replacement of country rocks. More recent investigators found that the deposits were either the result of magma driven replacement or hydrothermal processes (*e.g.* Sims and Leonard, 1952; Hotz, 1953, 1954; Sims, 1958; Hagner *et al.*, 1963; Buddington, 1966; Collins, 1969; Baker and Buddington, 1970; Foose and McLelland, 1995; Martinko and Gates, 2000). Gundersen (1984, 1985, 2000) proposed that they formed in a back arc setting, but Volkert (2001) proposed that the magnetite deposits were formed in an extensional setting, prior to metamorphism. It is evident that there are multiple modes of emplacement for the magnetite bodies of the Hudson Highlands (Lupulescu and Gates, 2006), as there are multiple types. Gundersen (2000) and Puffer (2001) acknowledge several types, including vein deposits created by remobilization of magnetite into faults and fractures, as well as deposits that are related to plutonic rocks and hydrothermal activity (Gundersen, 2004).

This guide will focus on two, kilometer scale shear zones that host several hydrothermally mineralized veins containing massive magnetite bodies. The NE trending dextral shear zones formed in crystalline rocks of the western Hudson Highlands late in the Grenville orogenic cycle. The veins are exposed at several abandoned magnetite mines within Harriman State Park. New modeling for the mode of formation of these deposits is discussed, following an overview of the bedrock geology, and brief description of each of the veins. New geochemical evidence from the wall rock-vein contact will also be explored. This guide will then conclude with an overview of the current working model for the formation of these iron oxide deposits.

BEDROCK GEOLOGY

The origin of the bedrock of the Hudson and New Jersey Highlands has been the subject of much controversy over the past two centuries. Earliest interpretations agreed that the gneisses were of meta-sedimentary origin (*e.g.* Rodgers, 1840; Kitchell, 1857; and Cook, 1868), but later Spencer (1904, 1905, 1909) and Berkley (1907) proposed a plutonic origin. Bayley (1910) proposed that they were derived from both plutons and metamorphosed sediments. Later workers also concurred with this hypothesis (*e.g.* Sims and Leonard, 1952; Hotz, 1953, 1954; Sims, 1958; Hagner *et al*., 1963; Buddington, 1966; Collins, 1969; Baker and Buddington, 1970; Foose and McLelland, 1995). More recent geochemical investigations found that some of the rocks are of plutonic origin (Gundersen, 1984, 1985, 1986, 2001; Volkert and Drake, 1999; Gates *et al.*, 2001, 2003, 2004; Puffer and Gorring, 2005). However, others have recognized that many of the quartzofeldspathic and amphibole-pyroxene gneisses show striking similarities in major-element chemistry to volcanic rocks (Helenak, 1971; Jaffe and Jaffe, 1973; Drake, 1970, 1984; Murray, 1976; Grauch, 1978; Kastellic, 1979; Gundersen, 1984, 1985; Gates *et al.*, 2001, 2003, 2004; Puffer and Gorring, 2005).

Currently, several models exist for the formation of these rocks. Gundersen (1984, 1985, 2001, 2004) proposed that many of these gneisses formed in an extensional backarc marginal basin whereas others have a bimodal, volcanic origin. In contrast, Volkert and Drake (1999) interpret many of the igneous gneisses a result of fractionation of a single parent diorite and the metasedimentary units as unconformable sequences. Gates *et al.* (2001) propose formation of a volcanic pile with a volcaniclastic apron in an island arc or marine magmatic arc setting, characterized by layered intermediate and mafic gneisses and associated plutons, for the genesis of these rocks.

Prior geologic mapping in this region sub-divided the abundant quartzofeldspathic gneisses based upon the individual ferromagnesian minerals (Dodd, 1965; Dallmeyer, 1974). However, considering that 80% of these rocks are quartz-feldspar gneisses (Gates, 2003), this guide will use the system first conceived by Gundersen (1986), and later adapted by Gates *et al.* (2001), using a type of sequence stratigraphy

Figure 1. Map of the Northeastern United States showing distribution of Grenville rocks, the Reading Prong, Hudson Highlands, and study location (Figure 2).

for metamorphic rocks. Units are grouped into lithofacies based on various rock types, to define quartzofeldspathic, metasedimentary (calc-silicate, pelite, and psammite), and metavolcanic (plagioclase-amphibole-pyroxene) assemblages (Figure 2).
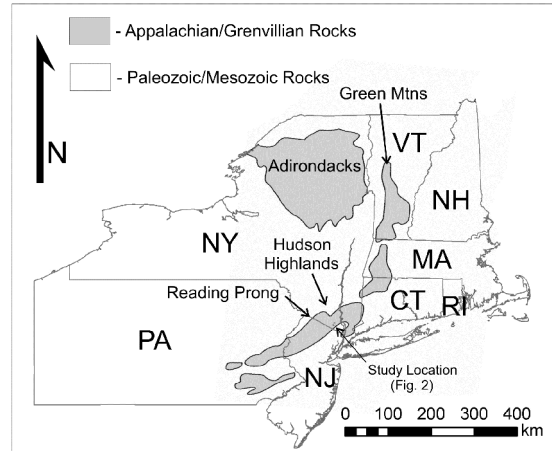
Metavolcanic Gneiss

The metavolcanic unit consists of strongly banded sequences of inter-layered dark mafic and gray intermediate gneisses, interpreted to represent rocks with volcanic protoliths (Gates *et al.,* 2001). Compositional banding ranges in thickness from 5cm to 5m with varying quantities of each rock type. Mafic assemblages are composed primarily of medium to coarse grained amphibole, plagioclase, clinopyroxene and hypersthene, and minor magnetite locally. Intermediate bands contain medium to coarse grained plagioclase and quartz, with minor amounts of amphibole, biotite, clinopyroxene and hypersthene. This unit also contains localized interlayers of quartzite, marble, and calc-silicate gneiss, as well as migmatites. The contacts with the quartzofeldspathic unit are interstratal gradational.

Quartzofeldspathic Gneiss

The quartzofeldspathic gneiss ranges from massive to layered quartz-plagioclase gneiss to quartz-K-feldspar-plagioclase gneiss with lesser amounts of clinopyroxene, hypersthene, amphibole, and/or phlogopite, layer pending. Minor amounts of magnetite and garnet can also be observed locally. Compositional layering is defined by the proportions and species of ferromagnesian minerals present. Closest to the contact with the metavolcanic unit, it is also locally interlayered with quartzite and mafic gneiss. Locally, this unit also contains apparent fining upwards sequences defined by an increase in the amount of mica and decrease in the layer spacing, showing sharp contacts between sequences (Gates *et al.,* 2001). However, it is difficult to interpret such relict sequences in granulite terranes. Gradational contacts with the metavolcanic lithofacies, composition and mineralogy, and internal compositional layering suggest that this unit represents a volcaniclastic sequence (Gates *et al.,* 2001).

Metasedimentary Gneiss

Throughout the western Hudson Highlands there are belts of rock considered to have sedimentary protoliths including meta-pelitic, meta-psammitic, calcsilicate gneisses, quartzite and marble. Belts of rock may contain all or some of these lithologies interlayered at the scale of meters to 100's of meters (Gates *et al*., 2001). The calc-silicate gneiss is quartzofeldspathic containing salite, K-feldspar, apatite, sphene, scapolite, and amphibole. Centimeter-scale quartzite layers and discontinuous layers of diopside marble also appear in this unit. The meta-pelite consists of interlayered biotite-garnet gneiss with medium to coarse quartz, plagioclase, K-feldspar, with cordierite and sillimanite locally (Gates *et al.,* 2001).
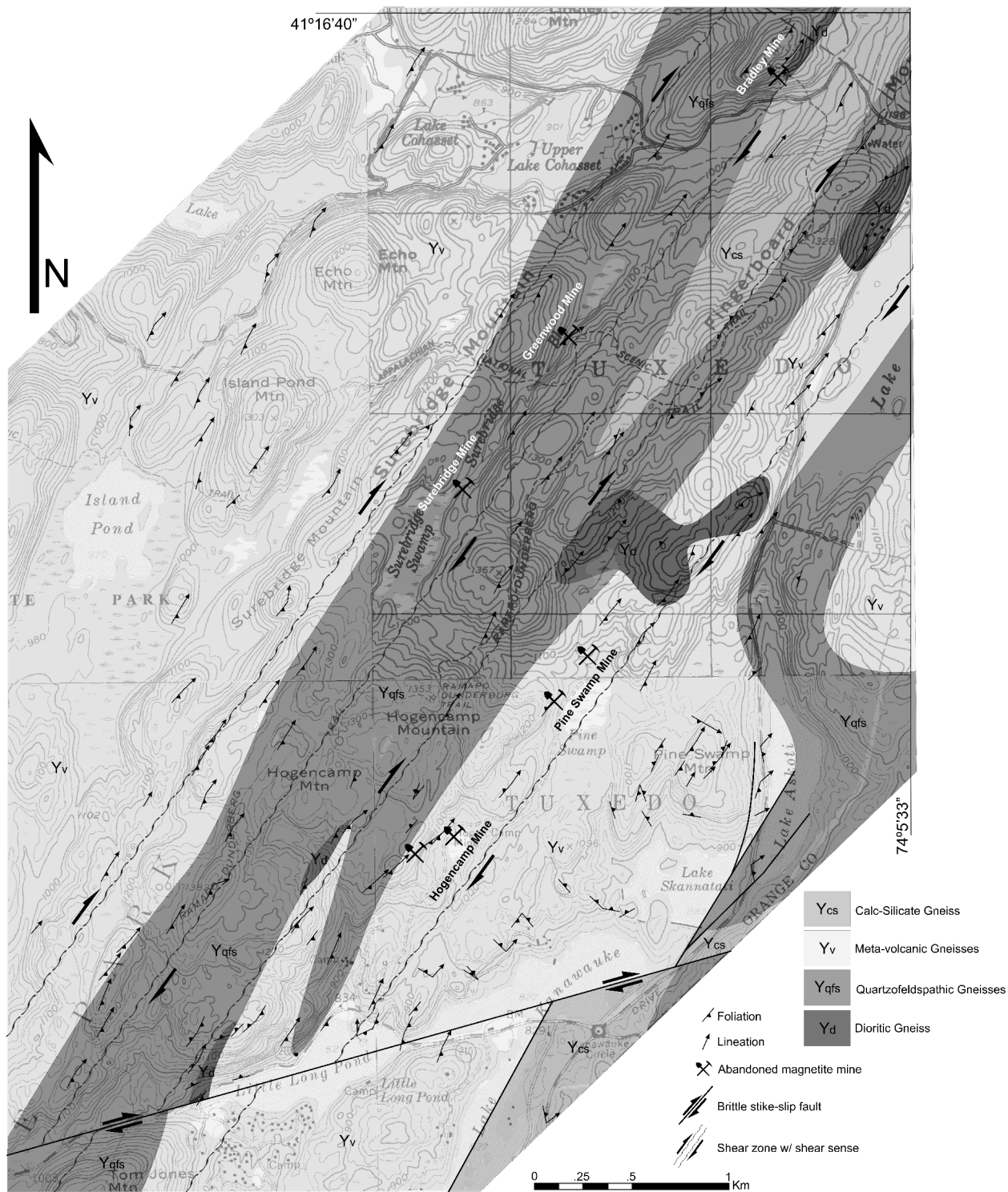
Figure 2. Geologic map of the field trip area, located within Harriman State Park, NY.

Lake Tiorati Diorite

Coarse to very-coarse grained black and white diorite dikes and bodies containing plagioclase, pyroxene, amphibole, and minor biotite locally, are found throughout the field area. Some small bodies appear concentrated in certain areas, possibly indicating larger bodies at depth (Gates *et al.,* 2001). The diorite also grades to pyroxene-poor, anorthositic compositions locally. Textures vary from coarse granoblastic to foliated and mylonitic with type II S-C fabrics (Lister and Snoke, 1984), exhibiting dextral shear sense (Figure 3). Locally, the diorite contains country rock xenoliths, showing intimate, ductile contacts which are partially melted, forming a rind of coarse pegmatite granite around them. Granite also fills fractures in the diorite that opened after crystallization, but while the granite was still liquid, which implies emplacement at depth (Gates *et al.,* 2001).

Pegmatites

Two generations of pegmatites occur throughout the field area. The earliest dikes are white and contain K-feldspar, quartz, muscovite and minor garnet locally. They are concordant to semi-concordant to the gneissic foliation, commonly boudinaged and containing internal fabrics and deformed grains. Thickness of these dikes ranges from 1cm to 1m. The later pegmatitic dikes are pink and very coarse grained, containing K-feldspar and quartz with muscovite, amphibole, magnetite, pyroxene, titanite, and/or garnet locally, depending on the rock intruded. They are highly discordant, commonly within brittle faults and contain xenoliths of faulted country rocks. They show minor to no deformational fabrics, and thickness ranges from 1m to 10m. They are also associated with small granite bodies (Gates *et al.,* 2001).

MINERALIZED ZONES

Two small, sub-kilometer wide shear zones formed within a 35-km-wide anastomosing dextral strike-slip shear system in the western Hudson Highlands (Gates, 1995). Each one contains a concordant to slightly discordant, late stage dilational, brittle fracture zone where hydrothermal fluids interacted with local country rock and deposited mineralized veins (Figure 4). Both NE-trending zones are defined by steeply dipping foliations, penetrative mineral lineations, and type II S-C mylonites with other dextral kinematic indicators (Figure 3). The vein-wall rock contact is sharp and semi-concordant to the mylonitic foliation. However, on the small-scale it appears slightly discordant, crosses foliation, and erodes into the wall rock.

The geometry of the vein deposits is characterized by three distinct zones sub-parallel to the wall rock boundary (Figure 5). The unaltered wall rock, grades into a 1-2cm "bleached zone" that is lighter in color and marked by alteration of the original minerals and formation of new minerals. The bleached zone is in direct contact with the vein deposit. The vein proper is characterized by two distinct zones, a layered sequence along the wall rock and a core of massive mineral assemblages. The layered sequence is characterized by distinct, dark colored bands of fine-coarse grained, pyroxene, amphibole, and/or biotite-rich assemblages, which range in thickness from 2-10 centimeters and have semi-gradational to sharp contacts. The core of the veins consists of massive deposits, characterized by very coarse, randomly oriented Fe/Mg-rich assemblages including magnetite and also containing late stage interstitial cementing minerals. The magnetite deposits in the core are characterized dominantly by massive magnetite, with minor am-
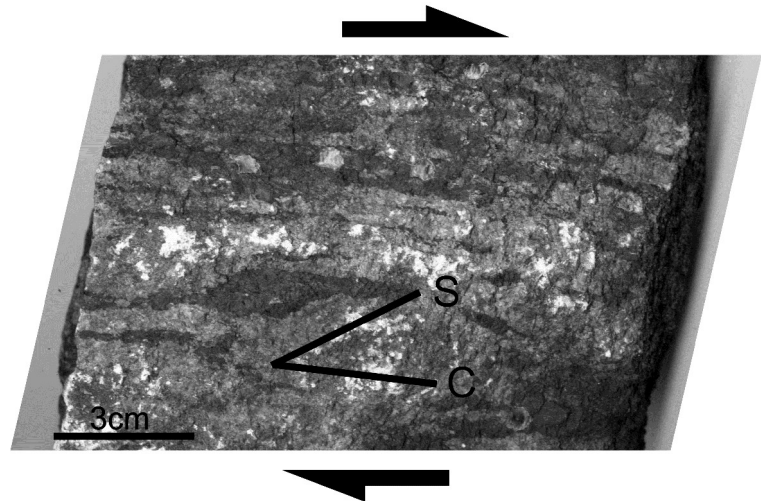

Figure 3. Hand sample of the meta-diorite, with type II S-C fabrics and σ porphyroclasts showing dextral shear sense.

phibole and pyroxene gangue minerals. The thickness of the core ranges from 1-10 meters. Thickness of the entire zones ranges from 2-15 meters and from tens of meters, to one kilometer in length. The much narrower zones that connect the magnetite deposits are thinner and typically composed of randomly oriented to aligned clinopyroxene with minor magnetite, phlogopite and/or quartz. The zones are commonly intruded by late pegmatite dikes that contain the mineralized rock as xenoliths.

Bleached zone mineral assemblages, vein material, and late-stage cementing minerals vary with location within each mineralized zone. In areas of Ca-rich country rock, clinopyroxene/calcite rich mineral assemblages dominate the bleached zone and throughout the vein. In areas of quartzofeldspathic country rock, amphibole/quartz assemblages dominate layered and massive vein material, whereas localities with sulfide-rich country rock contain orthopyroxene/sulfide rich vein assemblages.

Southeastern Shear Zone

Hogencamp Mine. The Hogencamp Mine lies in the southern part of the southeastern mineralized zone, where the sheared wall rock is dominated by metavolcanic gneisses, with calcsilicate gneiss and marble locally.

Figure 4. Map showing location and orientation of the vein deposits.

It is characterized by a series of 1–10 meter horizontal and vertical mine shafts and open pit mines where the magnetite ore was extracted from the vein deposit. These mines can be traced along strike for up to one kilometer. The mineralized zone that hosts Hogencamp Mine is roughly 6 kilometers long and ranges in thickness from 2-10 meters at the mine locations to as little as one meter in the narrow zones connecting the massive deposits. The bleached zone is characterized by calcite and scapolite, and retrogression of pyroxene to amphibole, with phlogopite, and minor apatite locally. The layered portion of the vein is composed of hornblende- and orthopyroxene-rich layers, also containing calcite and phlogopite closest to the semi-gradational bleached zone contact where observed (Figure 5). The central massive portion of the vein deposit is characterized by substantial magnetite ore and gangue minerals of euhedral crystals of clinopyroxene cemented by late-stage, interstitial calcite (Figure 6 A).
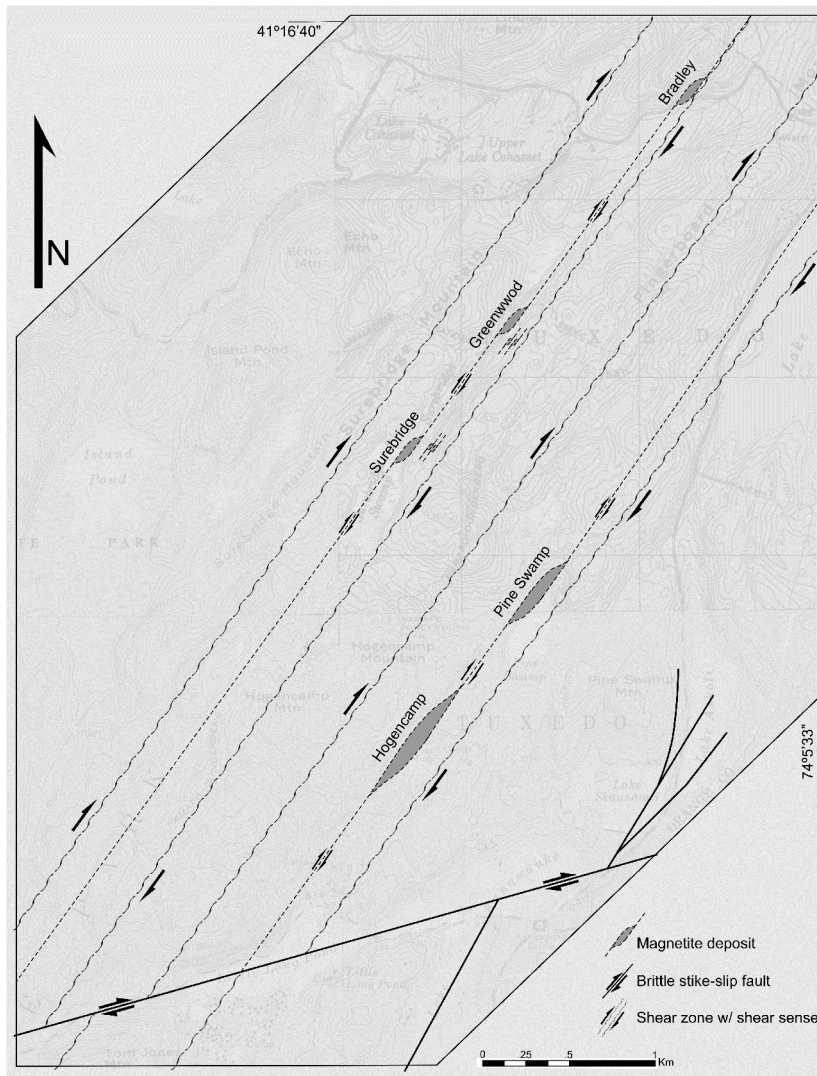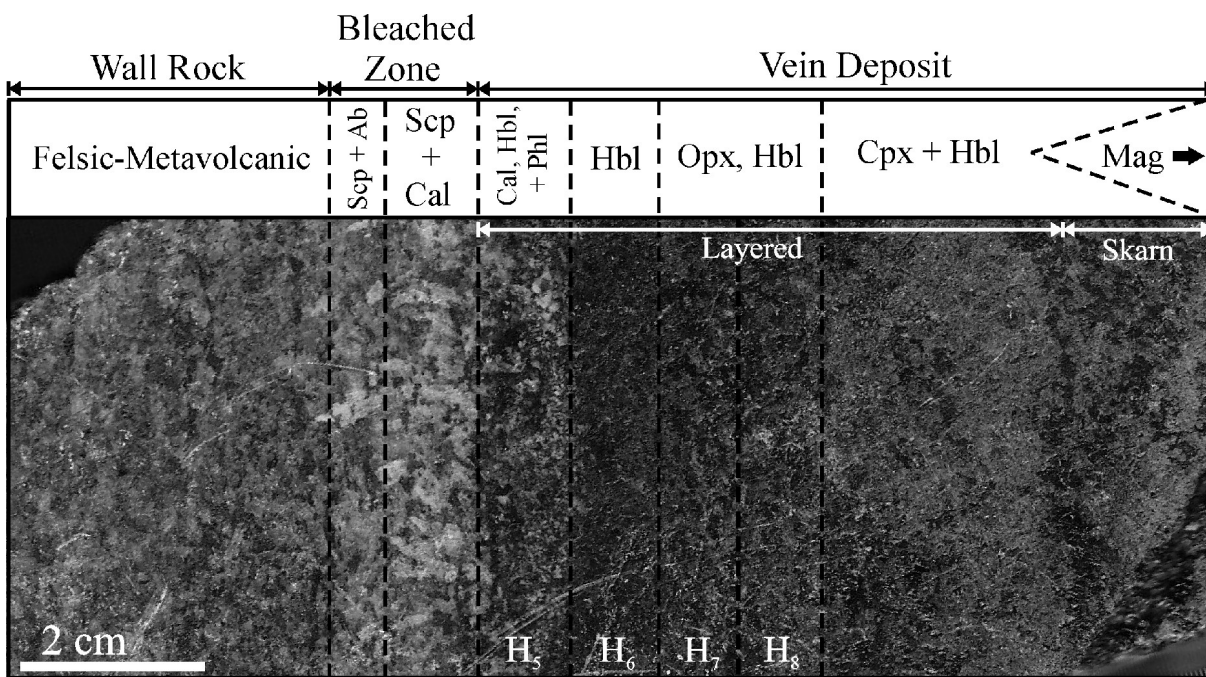
Figure 5. Hand sample from Hogencamp mine, showing banded mineral assemblages across the wall rock, bleached zone, vein deposit contact. Also shown are the layered and massive vein zones and layers H5-H8.

Pine Swamp Mine. Pine Swamp Mine, lies along strike, about one kilometer to the NE, of the Hogencamp Mine deposit. Pine Swamp is also characterized by a 5-meter wide horizontal mine shaft and meter scale open pit mines. However, unlike Hogencamp, the vein which hosts Pine Swamp Mine is only several hundred meters long, with minor semi-concordant veins in the northern extent of the deposit. This portion of the vein lies primarily in sulfide-bearing quartzofeldspathic gneiss and metavolcanic gneiss country rock, which varies among interlayered mafic and intermediate gneisses. The bleached zone is primarily defined by retrogression of pyroxene to amphibole, but also contains minor amounts of scapolite, and apatite locally. Orthopyroxene and amphibole dominate the thin layered vein and thick massive sequences including the magnetite deposits. Massive minerals are cemented by late stage sulfide minerals, mainly pyrite and pyrrhotite.
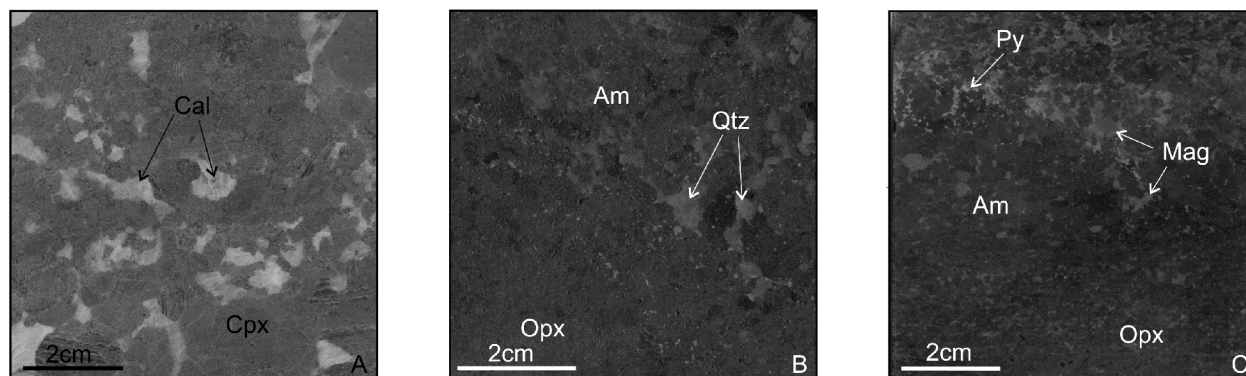

Figure 6. Hand samples of massive vein material from Hogencamp (A) and Greenwood Mines (B & C), showing late-stage interstitial cementing minerals; (A) calcite in massive clinopyroxene, (B) quartz in amphibole, orthopyroxene, and minor magnetite, and (C) pyrite in amphibole, magnetite, and layered orthopyroxene.

Northwestern Shear Zone

Bradley Mine. Bradley Mine is the northern-most deposit in the northwestern shear zone. Bradley Mine hosts a several meter discontinuous body of diopside marble. This marble is composed of medium-coarse grained calcite, fine-medium diopside, which occurs in cm-scale irregular aggregates, and it includes minor amounts of garnet locally. The vein deposit is a few hundred meters in length and varies in thickness from 2 to 10 meters. The bleached zone here is characterized by scapolite, diopside, and sericite, followed by a narrow zone of layered clinopyroxene and very fine grained micas. Massive intergrowths of clinopyroxene cemented by calcite, and magnetite appear in the core of the deposit.

Greenwood Mine. Greenwood Mine lies in the middle of the northwestern shear zone, between the Bradley mine to the NE and the Surebridge mine to the SW. It occurs within the quartzofeldspathic gneiss unit and the deposit is characterized by a series of sub-parallel veins. The main deposit ranges in thickness from 2-10 meters and several tens of meters in length. The two sub-parallel veins are a few hundred meters to the SE of the mine and are 2-4 meters thick and 10-20 meters long. The bleached zone is absent in all cases with the vein in direct contact with the wall rock. Most wall rock-vein contacts are relatively sharp, and lead directly into a layered orthopyroxene-rich zone, with minor amphibole. The massive zone contains magnetite and amphibole/orthopyroxene-rich deposits cemented by interstitial quartz, or less commonly pyrite and, pyrrhotite (Figure 6, B & C).

Surebridge Mine. The Surebridge Mine is located at the southern end of the northwestern shear zone. Country rock is sulfide-bearing quartzofeldspathic gneiss, but also contains mafic metavolcanic gneiss. The Surebridge Mine is characterized by a roughly 5 meter wide by 30 meter long central vein deposit, and two, narrower, sub-parallel secondary vein deposits. These secondary vein deposits are about 100 meters to the east of the main vein. The much narrower veins are about 2-3 meters thick and up to 20-30 meters in length. They can currently be observed as two, 5 meter deep pits, from which the iron ore was extracted. The narrow bleached zone at Surebridge is composed primarily of amphibole with only minor scapolite and sericite locally. The layered vein sequence in this locality is only a few centimeters in thickness and dominantly composed of amphibole and magnetite with minor pyroxene and quartz. Massive minerals are pyroxenes and amphibole, followed by magnetite in the core, with sulfides and some quartz as a late cementing agent.

GEOCHEMISTRY

Samples were collected at each of the mines, across the wall rock-bleached zone boundary, for chemical analysis. They include one each from the northern and southern Hogencamp and Pine Swamp deposits, and one each from the Bradley and Surebridge deposits. Small, cm-scale samples were removed from the unaltered wall rock and from the bleached zone at each locality and pulverized for chemical analysis. These samples were analyzed for bulk oxides

| | Hogencamp | | | | Pine Swamp | | | | Bradley | | Surebridge | |
| | North | | South | | North | | South | | | | | |
	Wall	Bleach	Wall	Bleach	Wall	Bleach	Wall	Bleach	Wall	Bleach	Wall	Bleach
SiO_2	68.30	55.96	57.48	53.06	56.06	53.29	66.18	63.48	43.88	43.01	69.46	59.39
TiO_2	0.43	0.31	1.10	2.02	0.95	1.40	0.72	0.78	0.13	0.11	0.36	1.00
Al_2O_3	12.98	14.05	16.04	14.89	15.92	14.59	13.80	15.56	17.23	19.09	13.84	13.97
Fe_2O_3	4.51	9.69	8.46	10.86	8.29	10.02	6.64	8.09	12.66	11.54	6.82	11.28
MnO	0.16	0.23	0.05	0.05	0.05	0.05	0.15	0.17	0.15	0.17	0.09	0.16
MgO	1.73	3.80	4.75	6.06	4.83	6.19	2.36	2.37	3.85	3.27	2.07	4.20
CaO	3.86	11.93	7.90	9.40	8.13	9.24	4.89	5.48	19.79	21.89	4.39	7.13
Na_2O	3.11	4.23	5.08	4.41	5.01	4.35	4.23	4.67	0.83	0.15	4.36	3.89
K_2O	5.34	0.57	1.63	1.12	1.34	1.17	1.61	2.47	0.18	0.07	0.79	0.95
P_2O_5	0.11	0.24	0.22	0.18	0.20	0.19	0.16	0.17	1.49	0.68	0.06	0.29
Total	100.53	101.03	102.71	102.06	100.78	100.50	100.74	103.24	100.19	99.98	102.24	102.25

Table 1. Bulk oxide geochemical results (in wt.%) for samples from select mines. Wall rock compositions are an average from several analyses.

and minor elements using a JY ULTIMA-C, ICP-OES at the ICP OES & MS Laboratory at Montclair State University, New Jersey. Wall rock chemistries were averaged at each sampling location to reduce variations due to compositional heterogeneity in the cm-scale samples of gneiss (Table 1) (Grant, 2005). Bulk rock chemistries were also acquired in cm-scale bands across layered vein material from Hogencamp Mine (Table 2, Figure 7). The locations of the bands are shown in Figure 5.

Bulk oxide chemical analyses results are shown in Table 1. Wall rock samples from Hogencamp Mine varies from felsic to intermediate in composition, containing about 58-68% silica, moderate amounts of alumina (13-16%) and iron (4.5-8.5%), and varying amounts of potash (1.6-5.3%), soda (3-5%), and calcium (3.8-7.9%). Pine Swamp wall rock samples show similar bulk chemistries to Hogencamp Mine. Silica ranges from 56-66%, alumina varies from about 14-16%, and iron ranges from 6.6-8.2%, whereas soda and potash appear relatively consistent between sample locations. Larger differences in calcium (4.8-8.1%) and magnesium (2.3-4.8%) are also observed. Calc-silicate wall rock at Bradley is very high in calcium (19.8%), alumina (17.2%), and iron (12.6%), and very low in silica (43.9%), soda (0.8%), and potash (0.2%). Surebridge Mine samples are from sulfide-bearing quartzofeldspathic gneiss, characterized by almost 70% silica, with significant amounts of alumina (13.8%), iron (6.8%), calcium (4.4%), and soda (4.4%), and minor quantities of magnesium (2%) and potash (0.8%).

Bulk chemical composition of the layered vein material from Hogencamp Mine resembles that of a mafic to ultramafic igneous rock (Table 2). Silica progressively decreases from 48-44.8% into the vein, whereas magnesium increases from 6.1-13.3%. Iron shows a small net loss from 13.1-11.6%, inward whereas alumina and soda decrease progressively away from the wall and increase into the innermost analyzed layer, from 11.2-7.5% and 1.9-0.7% respectively. Potash decreases close to the wall rock from 0.7-.3%, but increases to 1.1% in the innermost layer. Similarly, calcium and phosphorus progressively increase but then decrease in the innermost layer from 17.6-21.4% down to 18.6%, and 0.5-1.3% down to 1.0%, respectively. Titanium remains relatively stable around 0.2-0.3% across all layers (Figure 7).

The geochemistry of the wall rock and bleached zone samples were compared to constrain elemental gains and losses into the metasomatic fluids within the zone. However, comparison of two sets of geochemical data may lead to misinterpretations without the knowledge of the relationship between composition and volume changes that accompany the processes (Gresens, 1967). To that end, Grant's Isocon analysis (1987), after Gresens (1967) equation for metasomatic alteration was applied to the bulk oxide geochemistry results and density calculations. In doing so, constant alumina was generally assumed.

Hogencamp Vein				
	H_5	H_6	H_7	H_8
SiO_2	48.00	47.13	45.98	44.76
TiO_2	0.22	0.32	0.32	0.32
Al_2O_3	11.18	7.93	7.49	8.62
Fe_2O_3	13.06	13.95	12.10	11.59
MnO	0.25	0.27	0.20	0.17
MgO	6.08	8.59	11.73	13.25
CaO	17.58	20.45	21.42	18.58
Na_2O	1.86	0.91	0.65	0.91
K_2O	0.67	0.28	0.50	1.06
P_2O_5	0.49	0.79	1.30	1.02
Total	99.42	100.61	101.69	100.27

Table 2. Bulk oxide geochemical results (in wt.%) from layered vein material.
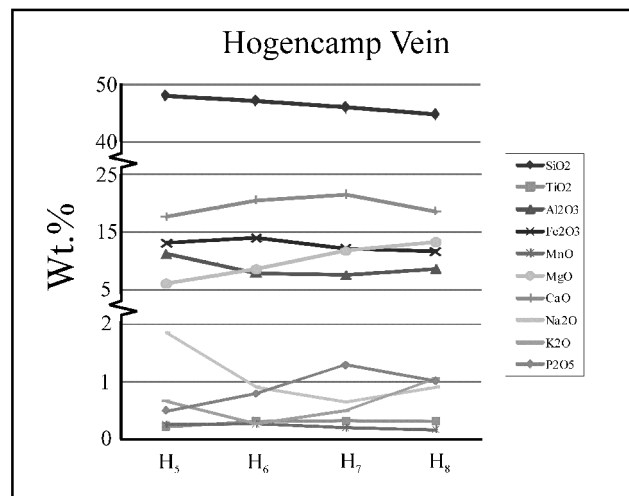

Figure 7. Plot of bulk oxides (wt%) for data in Table 2, Figure 5, showing layered vein material progression.

The results are shown in Figure 8. Samples from both northern and southern Hogencamp and Pine Swamp deposits reveal significant gains in CaO, MgO, and Fe_2O_3, with considerable losses in K_2O, and SiO_2 to a lesser extent (Fig-
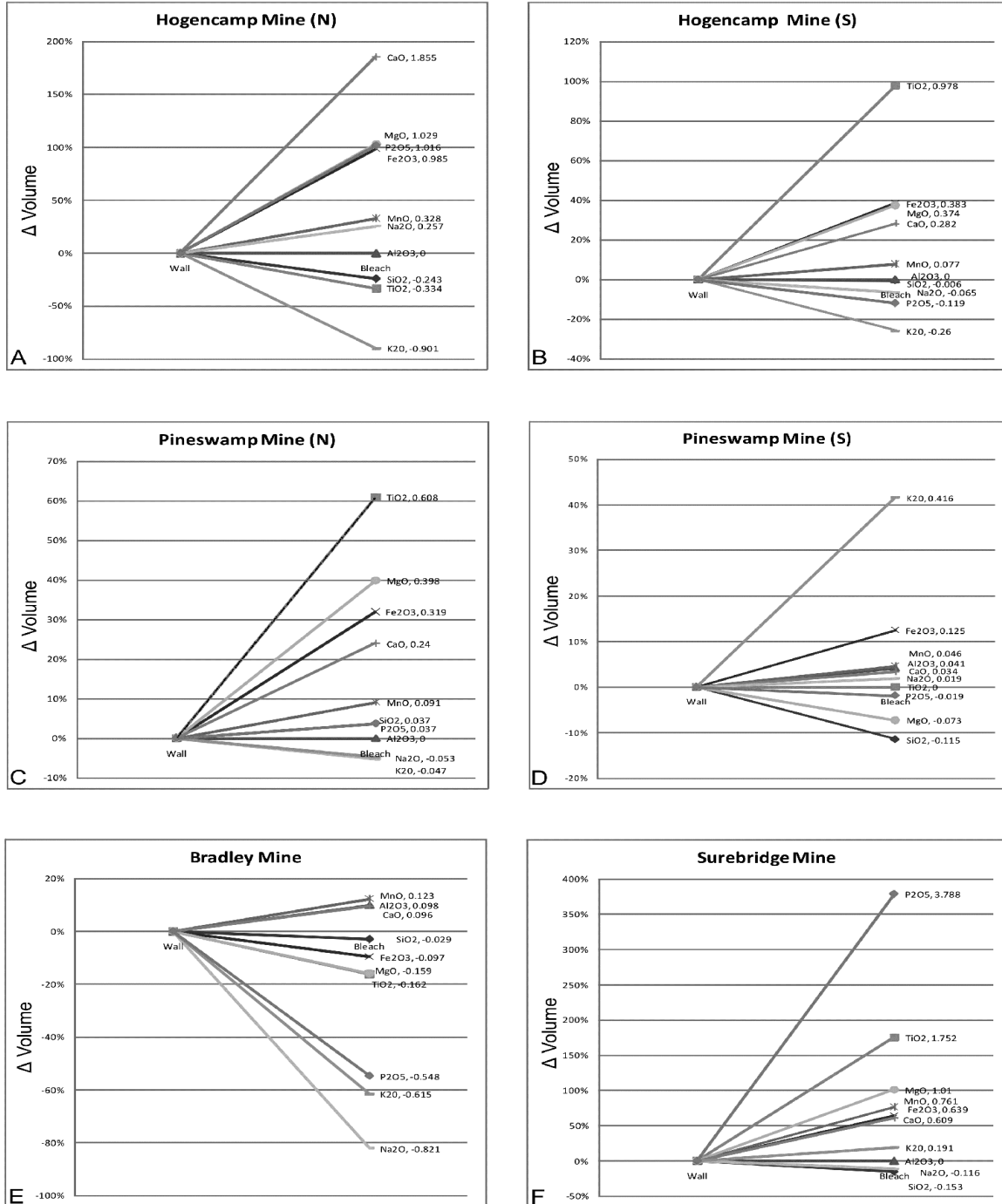

Figure 8. Volume percentage change in oxides for samples from select mines, across the wall rock-bleached zone boundary, as determined by Grant's isocon analysis (1987).

ure 8, A-D), into the bleached zone. Mass transfer analysis from the Bradley deposit shows small gains in CaO and Al_2O_3, and large losses in Na_2O, and K_2O, and lesser amounts of MgO and Fe_2O_3 (Figure 8, E). Geochemical modeling from the Surebridge deposit reveals significant gains in MgO, Fe_2O_3, and CaO, with only minor losses of SiO_2 and Na_2O (Figure 8F).

DISCUSSION

A volcanic pile formed about 1.2 Ga in an island arc or marine magmatic arc setting characterized by layered intermediate and mafic rocks, and associated plutons and volcaniclastic sediments (Gates *et al.*, 2001). This sequence underwent granulite facies metamorphism, about 1,050 Ma associated with the Ottawan phase of the Grenville orogeny (Gates *et al.*, 2003). Locally, anatexis produced migmatites, granite sheets, and the early pegmatites (Gates *et al.*, 2003). Subsequent diorite intrusions occurred around 1,008 Ma, either the result of delamination at the end of the first event, or the early dilational stages of the next event (Gates *et al.*, 2003). This second event is characterized by dextral strike-slip movement during a period of rapid uplift and unroofing at approximately 1,008 Ma to 924 Ma (Gates and Krol, 1998). Thick anastomosing zones of mylonite formed, overprinting previous features. Offset reached upwards of 100's of kilometers (Gates *et al.*, 2003). Strike-slip shearing continued during rapidly decreasing temperatures, resulting in the shear zones crossing the brittle-ductile transition and becoming dilational (Gates *et al.*, 2003).

These dilational structures began to open at small angles to the shear zone boundaries, resulting in sub-parallel, pull-apart structures (McCaig, 1987), connected by brittle faults. Hydrothermal fluids flushed into the faults, and chemically equilibrated with the composition of the local wall rock. These reactions produced the bleached zones in the wall rocks. In this way, the fluids were chemically buffered locally, and varied in composition along strike. During continued deformation, these now buffered metamorphic fluids were transported along strike likely through such mechanisms as seismic pumping (Sibson *et al.*, 1975). Alteration in the bleached zone and precipitation in the vein was driven by chemical or physical changes encountered as the fluids were flushed along strike. This was controlled by the stability or instability of individual minerals in response to changes in P, T, pH, or solution composition (Eugster, 1986).

Acid-base reactions are important in metamorphic and ore-forming environments, because in many mineral alteration reactions, Al is conserved in the solid, and HCl is produced (Eugster, 1986). This is the case for the alteration of K- and Na-feldspars to mica (Hemley, 1959). The addition of SiO_2 and S_2 into solution may also drive pH levels of the fluid to higher acidity, whereas introduction of Ca drives fluids to more basic values. Such positive feedback, of varying pH solutions from addition and subtraction of certain chemical species into and out of solution, may have driven the earliest stages of the bleached zone formation as fluids fluxed along the zone. Furthermore, redox, acid-base, and exchange reactions are the principal mechanisms by which metals are acquired by hydrothermal fluids (Eugster, 1986). Acids released during early wall rock reactions may have aided in the mobilization of metals which contribute to the later massive deposits, where Fe is preferentially mobilized during more oxidizing conditions (Eugster, 1986). Ultimately, chemical interaction primarily drove much of the early precipitation and alteration seen in the bleached zone and layered vein sequence, whereas decreasing pressure gradients primarily caused precipitation of the later massive mineral assemblages.

Higher pressures and acidity permitted higher levels of iron and magnesium to remain in solution in the narrow faults connecting the mineralized zones. Wall rock in these areas likely exhibited lithostatic pressure, and buffered the pH of fluids to more acidic values (Eugster, 1986). Dropping from lithostatic pressure, to hydrostatic pressure at the extensional segments of the fault, where fluids were also buffered to higher pH, led certain chemical species to become supersaturated and prompted precipitation.

Influx of Fe/Mg-rich fluids into the dilational fractures resulted in the deposition of the early pyroxene/amphibole-rich assemblages. Locally buffered fluids may have interacted in some localities, resulting in early clinopyroxene-rich layered and massive mineral assemblages in areas of CaO enrichment, amphibole-rich in areas of SiO_2 enrichment, and orthopyroxene-rich in localities dominantly increased only in Fe_2O_3 and MgO. Layered vein sequences were likely formed in the earlier volume confined conditions, whereas the conditions favored formation of massive

assemblages as dilation continued, volume increased and pressures began to drop. In this way, the main driving force switched from a chemical to pressure controlled precipitation mechanism, forming the massive ferromagnesian and magnetite assemblages, central to the deposit.

Inflowing fluids progressively became extremely Fe-rich, magnetite ore forming fluids, as these dilational structures continuously opened, and the fluids continued depositing the massive magnetite ore bodies. Residual, depleted fluids or fluids later buffered from local country rock, played a role in the late-stage, interstitial, cementing minerals found throughout the vein deposits. Calcite, quartz, and sulfides cement occurred in areas of calcium, silica, and sulfide-rich country rock, respectively. Granitic pegmatites intruded as fault activity waned, concentrated along the faults, suggesting a genetic relationship (Gates *et al.*, 2003). Fe-rich, dark pink potassium feldspars are evidence of intrusion along the same iron-oxide forming fluid conduits.
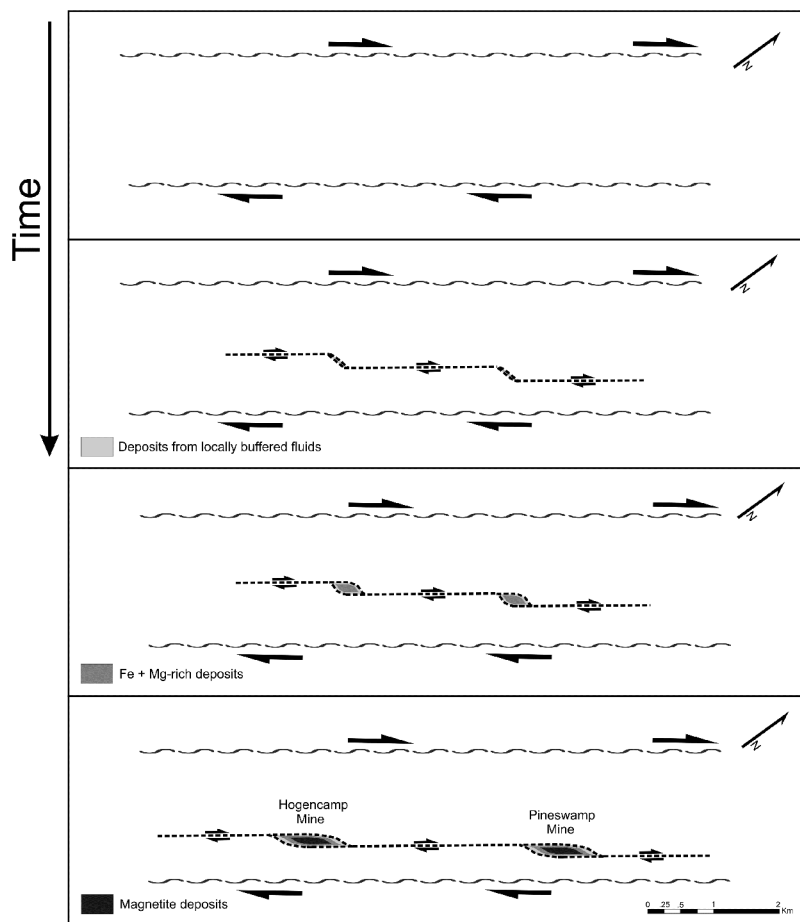

Figure 9. Schematic model of the successive stages of the vein deposits formation.

CONCLUSION

New geochemical and structural analysis of the mineralized faults leads to the interpretation of a mode of formation for one type of massive magnetite ore bodies in Harriman State Park (Figure 9). During the latest stages of dextral shearing, rapid temperature decreases caused the shear zones to cross the brittle-ductile transition and dilational structures formed a variably closed and open fracture system. Metamorphic fluids flushed through fractures and reacted with wall rocks. Changes in chemistries into the altered bleached zone, at the wall of the vein, relative to unaltered country rock reflect the exchange of various chemical species from the buffered fluids. Buffered to the composition of local country rocks, these fluids were transported along strike, eventually encountering favorable physical and/or chemical conditions for precipitation.

As deformation continued, the dilational structures continued to pull-apart, and fluids turned to ferromagnesian-rich solutions, interacted with locally buffered fluids, and deposited early layered deposits, primarily pyroxenes and amphiboles. With continued dilation, the dominant mode of precipitation switched from chemical to pressure driven, favoring the deposition of massive over layered deposits. The zones continuously opened, as fluids turned extremely iron-rich. As the fluids flushed through the dilational zones, they encountered lower pressures, less acidic conditions, and oxidized dissolved Fe, deposited as the massive mineralized magnetite bodies. Late interstitial minerals were precipitated from residual fluids that were locally buffered. Latest pegmatites intruded along some of the same pathways as the vein forming fluids.
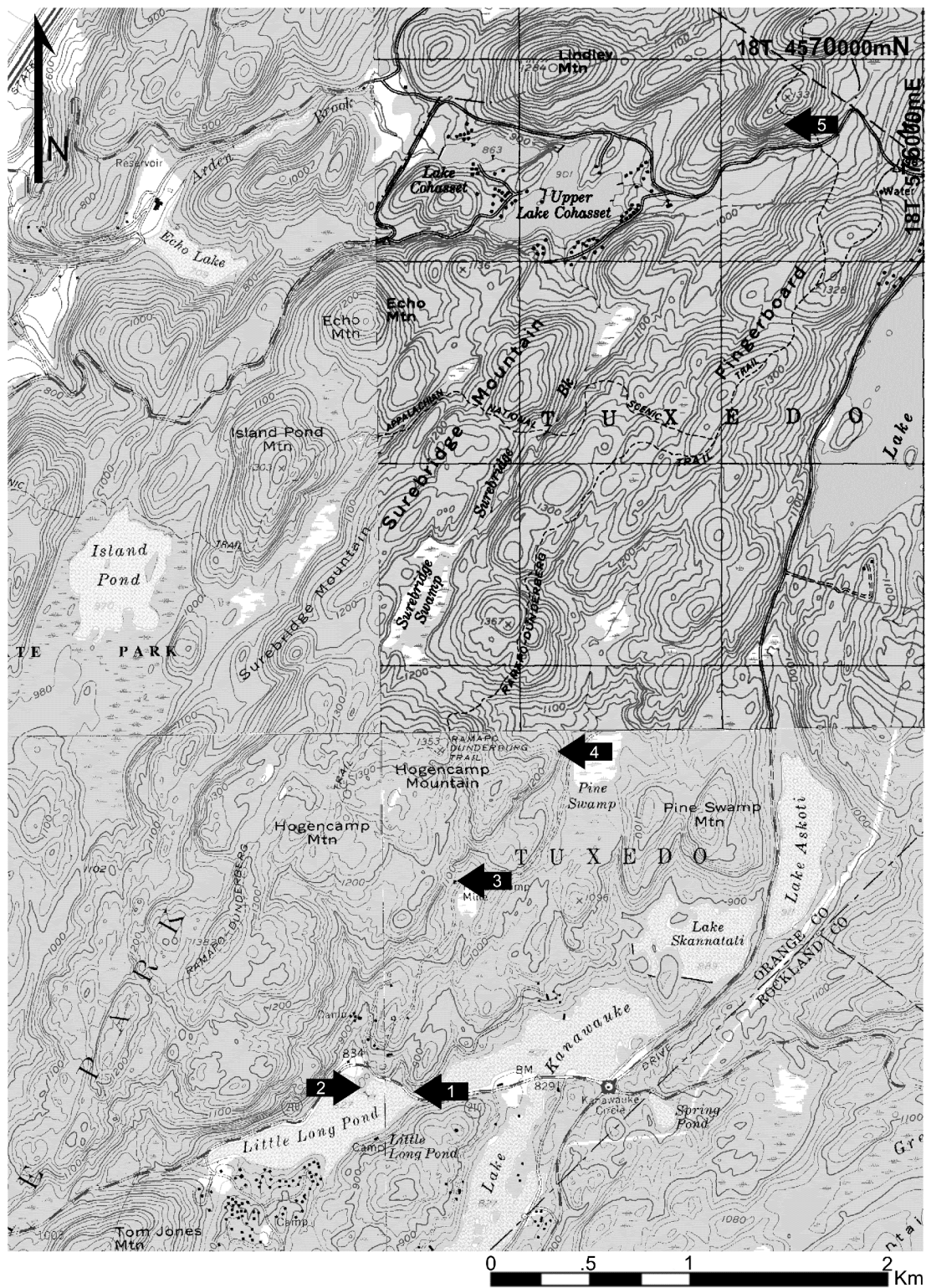
Figure 10. Map of field area showing locations of each stop.

ROAD LOG

As we are in a State Park, we would ask that no hammer be used and only pictures taken. Be prepared for two and a half miles of trail hiking.

Mileage
0.0   Start at SUNY New Paltz, NY.
0.3   Turn right at County Route 17/Jansen Rd.
1.8   Turn right at NY-208/State Route 208.
3.1   Turn right at Main St/NY-299/NY-32.
34.5   Take the ramp onto I-87 South.
35.1   Take exit 16 for NY-17/US-6 toward Harriman.
35.4   Keep right at the fork; follow signs for Harriman/US-6/NY-17/West Point.
39.5   Turn left at NY-17/NY-32/State Route 32.
42.0   Turn left at Arden Valley Road.
44.6   Turn left to stay on Arden Valley Road.
47.9   At the Tiorati traffic circle, take the 1st exit onto Seven Lakes Parkway, heading south.
51.2   At the Kanawauke traffic circle, take the third exit onto County Route 106, heading west.
51.8   Continue on County Route 106 for 0.6 miles until reaching the bridge over the eastern-most extent of Little Long Pond. If needed, parking can be found a quarter mile east on route 106 at the picnic area (lavatory facilities). Walk 100 feet to the north-west on route 106 to the first road-cut (Stop 1, Figure 11).

Stop 1. Unsheared Metavolcanic Gneiss (30 MINUTES) (UTM: 18 T 0573484 4565104). Strongly interlayered intermediate and mafic gneisses with migmatitic bodies, just outside of the SE shear zone (Figure 11). Mafic layers are characterized by assemblages of clinopyroxene and amphibole with minor plagioclase, magnetite, sphene, and apatite. Intermediate layers are mainly plagioclase with minor quartz, apatite, amphibole, and biotite. The felsic leucosome is composed of coarse plagioclase, quartz, and K-feldspar, which form veins and clots including classic "net veining". Minerals exhibit preferred orientations in the gneiss and appear granular in the leucosome. Late stage K-feldspar pegmatites can also be observed in this outcrop, containing mafic gneiss xenoliths.

Gneisses exhibit a strongly banded, intermediately dipping, foliation that strikes N-NE. Isoclinal intrafolial folds can also be observed, with axis following similar orientation. This deformation is indicative of main stage Grenville tec-

Figure 11. Stop 1, Road-cut of migmatitic metavolcanic country rock.

tonism, and was unaffected during the later tectonic event, associated with the formation of the ore deposits. Contrast these rocks with Stop 2.

51.9   Head west on route 106 for roughly 600 feet to the large peninsula protruding into Little Long Pond. Proceed 250 feet to the south to reach the tip of the peninsula (Stop 2, Figure 12).

Stop 2. Southeastern Shear Zone Boundary (30 MINUTES) (UTM: 18 T 0573272 4565124). Several meter scale lozenge and cigar shaped boudins of mafic gneiss are contained within mylonitic quartzofeldspathic gneiss, with folded biotite and local amphibole-rich layers (Figure 12). The layers appear contorted and wrap around the mafic bodies. The encased mafic gneiss is similar to that of Stop 1, however it also contains contorted folds and veins of magnetite. The long axis of the bodies and fold axis appear sub-parallel, and shallowly plunge to the northeast.

Both Stops 1 and 2 are similar in composition and therefore grouped within the same metavolcanic sequence. Long axis and fold axes roughly parallel shear zone boundaries and fabrics within. This location is at the edge of the southeastern dextral, strike-slip shear zone. Deformation steadily increases to the northwest, into the central shear zone, characterized by a steepening of planar fabric and increase in intensity of linear fabric. The contrast of the features at Stop 1 with Stop 2 shows the difference between the first main Grenville and second strike-slip events.

51.9   Locate the gated path directly on the opposite side of route 106. Walk this path heading north for a quarter mile, to the intersection with Dunning Trail (E-W trending). Head roughly northeast on Dunning Trail for about a half mile until you reach several large holes in the ground, and the path crosses a small stream. Hike upstream for roughly 100 feet until you are almost cliff side. A linear open pit mine should be visible to the southwest and a large mine shaft to the northeast into the cliff, beneath Cape Horn (Stop 3, Figure 13).

Stop 3. Hogencamp Mine (50 MINUTES) (UTM: 18 T 0573790 4566280). Hogencamp Mine lies in the southern part of the SE mineralized zone. Here the sheared wall rock is dominated by quartzofeldspathic and amphibole-pyroxene (metavolcanic) gneisses with interlayered calc-silicate gneiss and marble locally. Hogencamp Mine was active from the earliest to latest 18th century. It is characterized by a series of meter to several meter scale horizontal and vertical mine shafts and open pit mines. The mines can be traced along strike for up to a kilometer. The mineralized zone that hosts Hogencamp Mine is roughly 6 kilometers long and extends into Pine Swamp Mine (Stop 4). The vein ranges in thickness from 3-15 meters at the mine locations to as little as one meter in the narrow zones connecting the deposits. The Hogencamp Mines can be followed to the southwest from this location for up to a kilometer.
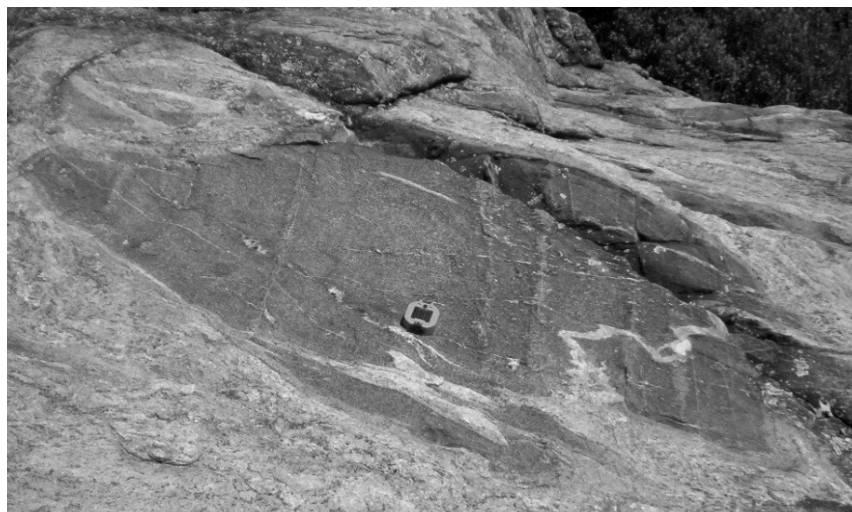

Figure 12. Stop 2, Outcrop on tip of peninsula, northeastern Little Long Pond.

Figure 13. Hogencamp (Stop 3), Pine Swamp (Stop 4), and Bradley Mines (Stop 5). From left to right.

The vein-wall rock contact is sharp and semi-concordant to mylonitic foliation. On the small-scale it appears slightly discordant, crosses foliation and erodes into the wall rock. This is best observed in the open pit mine, directly in front of the northeastern most mine shaft, below Cape Horn (Figure 13). Here, the bleached zone is characterized by the deposition of calcite and scapolite, and retrogression pyroxene to amphibole, also containing phlogopite, calcite, and minor apatite locally. Earliest vein deposit is characterized by layered amphibole, orthopyroxene, and clinopyroxene, later by massive clinopyroxene and magnetite, cemented by late stage, interstitial calcite in the ore zone. Clinopyroxene and localized magnetite are euhedral, forming doubly terminated crystals, thought to have crystallized in cavities. The veins are also intruded by very coarse grained pegmatites which contain xenoliths of the mafic vein material.

51.9   Back on Dunning Trail, walk east than north-northeast, another half mile until you reach a large swamp to the east, and a large, steep hillside to the west, littered with dark-colored mine tailings. Locate the makeshift path on the hillside. Take this path uphill for about two hundred feet, until your reach the entrance to the mine (Stop 4, Figure 13).

Stop 4. Pine Swamp Mine (50 MINUTES) (UTM: 18 T 0574253 4566795). Pine Swamp Mine lies in the northern part of the SE shear zone within the same mineralized vein which hosts Hogencamp Mine, roughly one kilometer alone strike (NE). Pine Swamp is also characterized by a 3-12 meter horizontal mine openings and several, meter scale open pit mines (Figure 13). The mines that compromise the Pine Swamp deposit can be traced along strike for several hundred meters, with minor semi-concordant offshoots in the northern extent of the deposit. This portion of the hydrothermal vein lies dominantly within sulfide-bearing quartzofeldspathic gneiss country rock, which contains interlayered metavolcanic gneisses. The bleached zone is primarily defined by retrogression of pyroxene to amphibole, also containing minor amounts of scapolite, and apatite locally. Orthopyroxene and amphibole dominate the narrow layered vein and thick massive sequences followed by the magnetite deposits. Massive minerals are locally cemented by late stage sulfide minerals, mainly pyrite and pyrrhotite. The flat wall adjacent to the mine proper is yellow to rust colored, due to the weathering of the sulfide-rich country rock. Massive minerals orthopyroxene, amphibole, and substantial magnetite can also be observed.

51.9   Take Dunning Trail south and west back to the first intersection of paths. Head south on the first path, to exit back at the gated entrance.
52.7   Drive east on route 106 to Kanawauke Circle, and take the third exit to head north on Seven Lakes Drive.
56.0   Head north on Seven Lakes Drive until reaching Tiorati Circle (lavatory facilities and parking if needed). Take the third exit to head west on Arden Valley Road. ***Lunch & Lavatory Stop***

56.8   Drive west on Arden Valley Road for 0.8 miles until a large, cliff-face is visible to the north, roughly thirty feet off of the road. Walk uphill, along the path directly adjacent to the cliff (west-side) for about 200 feet (Stop 5, Figure 13).

Stop 5. Bradley Mine (50 MINUTES) (UTM: 18 T 0575371 4569883). Mine is the northern deposit in the north-western shear zone, located a few hundred meters north of Arden Valley Road (Figure 13). Bradley Mine was active through the latest 19th century, when it closed permanently in 1874. Wall rock is primarily calc-silicate gneiss, containing salite, K-feldspar, apatite, sphene, scapolite, and amphibole. Bradley Mine also hosts a several meter thick discontinuous body of diopside marble, composed of medium-coarse grained calcite, fine-medium diopside, and minor amounts of garnet locally. Diopside is generally dispersed, though locally forms cm-scale aggregates.

The vein deposit at this location appears only a few hundred meters in length and varies in thickness from 2-10 meters. The bleached zone here is characterized by scapolite, diopside, and sericite, followed by narrow zone of layered clinopyroxene, calcite and very fine grained micas. Massive intergrowths of clinopyroxene cemented by calcite, than magnetite, appear in the center of the deposit. Both types of pegmatites can also be observed here. The earliest, white-colored pegmatite appears slightly deformed; containing minor internal fabrics, whereas the latest stage, pink pegmatites appear very coarse grained, and shows no deformational fabrics. A much more recent brittle fault also cross cuts the entrance to the mine proper.

53.7   Head west on Arden Valley Road.
56.2   Turn right to stay on Arden Valley Road.
60.1   Turn right at NY-17/Rte-17, heading north.
60.5   Turn right onto the ramp.
61.2   Take the I-87 exit toward Albany.
92.0   Continue toward and merge onto I-87 North.
92.8   Take exit 18 toward New Paltz.
94.2   Turn left at Main St/NY-299.
95.6   Turn left at S Chestnut St/NY-208.
95.9   Turn left at County Rte-17/Jansen Rd.

END OF TRIP

REFERENCES CITED

Baker, D.R. and Buddington, A.F., 1970, Geology and magnetite deposits of the Franklin Quadrangle and part of the Hamburg Quadrangle, New Jersey: US Geological Survey Professional Paper 638.

Bayley, W.S., 1910, Iron mines and mining in New Jersey: Geological Survey of New Jersey Final Report Series VII.

Berkey, C.P., 1907, Structural and stratigraphic features of the basal gneisses of the highlands: New York State Museum Bulletin 107.

Buddington, A.F., 1966, The Precambrian magnetite deposits of New York and New Jersey: Economic Geology, v. 16, p. 484-510.

Collins, L.G., 1969, Regional recrystallization and the formation of magnetite concentrations, Dover Magnetite District, New Jersey: Economic Geology, v. 64, p. 17-33.

Colony, R.J., 1923, The magnetite iron deposits of southeastern New York: New York State Bulletin, No 249-250.

Cook, G.H., 1868, Geology of New Jersey: Geological Survey of New Jersey.

Dallmeyer, R.D., 1974, Metamorphic history of the northeastern Reading Prong, New York and Northern New Jersey: Journal of Structural Geology, v. 15, p. 325-359.

Drake, A.A. Jr., 1970, Structural geology of the Reading Prong, In Fisher, G.W. ed., Studies in Appalachian Geology-Central and Southern, John Wiley, New York, p. 271-291.

Drake, A.A. Jr., 1984, The Reading Prong of New Jersey and eastern Pennsylvania, An appraisal of rock relations and chemistry of a major Proterozoic terrane in the Appalachians, In Bartholomew, M.J. ed., The Grenville Event in the Appalachians and Related Topics, Geological Society of America Special Paper 194, pp. 94-109.

Dodd, R.T. Jr., 1965, Precambrian geology of the Popolopen Lake quadrangle, southeastern New York: New York State Museum and Science Service Map and Chart Series, No. 6, 39 p.

Eugster, H.P., 1986, Minerals in hot water: American Mineralogist, v. 71, p. 655-673.

Foose, M.P. and McLelland, J.M., 1995, Proterozoic low-Ti iron oxide deposits in New York and New Jersey, Relation to Fe-oxide-Cu-U-Au-rare earth element deposits and tectonic implications: Geology, v. 23, p. 665-668.

Gates, A.E., 1995, Middle Proterozoic dextral strike-slip event in the central Appalachians- Evidence from the Reservoir fault, NJ: Journal of Geodynamics, v. 19, p. 195-212.

Gates, A.E. and Krol, M.A., 1998, Kinematics and thermochronology of late Grenville escape tectonics from the central Appalachians: Geological Society of America, Abstracts with Programs, v. 30.

Gates, A.E., Valentino, D.W., Gorring, M.L., Chiarenzelli, J.R., and Hamilton, M.A., 2001, Bedrock geology, geochemistry and geochronology of the Grenville Province in the western Hudson Highlands, New York, In Gates, A.E., ed., Geology of the Lower Hudson Valley: New York State Geological Association Guidebook, p. 177–204.

Gates, A.E., Valentino, D.W., Chiarenzelli, J., Gorring, M., and Hamilton, M., 2003, Field Trip to the Western Hudson Highlands: 2003 Long Island Geologists Conference, 30 p.

Gates, A.E., Valentino, D.W., Chiarenzelli, J.R., Solar, G.S., and Hamilton, M.A., 2004, Exhumed Himalayan-type syntaxis in the Grenville Orogen, northeastern Laurentia: Journal of Geodynamics, v. 37, p. 337–359.

Gates, A.E., Valentino, D.W., Gorring, M., Thern, E.R., and Chiarenzelli, J.R., 2006, Rodinian collisional and escape tectonics in the Hudson Highlands, New York, In Pazzaglia, F.J., ed., Excursions in Geology and History: Field Trips in the Middle Atlantic States, Geological Society of America Field Guide 8, p. 65-82.

Grant, J.A., 1986, The isocon diagram—a simple solution to Gresens' equation for metasomatic alteration: Economic Geology, v. 81, p. 1976–1982.

Grant, J.A., 2005, Isocon analysis: A brief review of the method and applications: Physics and Chemistry of the Earth, v. 30, p. 997-1004.

Grauch, R.I., 1978, Geology of the uranium prospect at Camp Smith, New York, with a new model for the formation of uranium deposits in the metamorphosed submarine volcanogenic rocks: US Geological Survey Open-File Report, 76-582.

Gresens, R.L., 1967, Composition–volume relationships of metasomatism: Chemical Geology, v. 2, p. 47–55.

Gundersen, L.C.S., 1984, Geochemistry and geology of Precambrian rocks of from the Greenwood Lake area, New York and New Jersey, Implications for the genesis of uranium-magnetite deposits: Geological Society of America, Abstracts with Programs, v. 16.

Gundersen, L.C.S., 1985, Stratigraphy of the northern Reading Prong and its tectonic implications: Geological Society of America, Abstracts with Programs, v. 17, p. 9.

Gundersen, L.C.S., 1986, Geology and geochemistry of the Precambrian rocks of the Reading Prong, New York and New Jersey - Implications for the genesis of iron-uranium-rare earth deposits, In Carter, L.M.H. ed., USGS Research on Energy Resources -1986 Programs and Abstracts, US Geological Survey Circular, v. 974, p. 19.

Gundersen, L.C.S., 2000, A backarc basin setting for a portion of the Proterozoic Reading Prong: Geological Society of America, Abstracts with Programs, v. 32, p. 21-22.

Gundersen, L.C.S., 2004, Tectonics and metallogenesis of Proterozoic rocks of the Reading Prong: Journal of Geodynamics, v. 37, p. 361-379.

Hanger, A.F. and Collins, L.G., 1963, Host rock as a source of magnetite ore, Scott Mine, Sterling Lake, New York: Economic Geology, v. 58, p. 730-768.

Helenek, H.L., 1971, An investigation of the origin, structure and metamorphic evolution of major rock units in the Hudson Highlands: PhD Thesis, Brown University.

Hemley, J.J., 1959, Some mineralogical equalibria in the system K2O-Al2O3-SiO2-H2O: American Journal of Science, v. 57, p. 241-270.

Hotz, P.E., 1953, Magnetite deposits of the Sterling Lake-Ringwood, N.J. area: US Geological Survey Bulletin, 982-F.

Hotz, P.E., 1954, Some magnetite deposits in New Jersey: US Geological Survey Bulletin, 995-F, p. 201-254.

Jaffe, H.W. and Jaffe, E.B., 1973, Bedrock geology of the Monroe Quadrangle, Orange County, New York: New York State Museum and Science Service Map and Chart Series, 20.

Kastellic, R.L., 1979, Precambrian Geology and magnetite deposits of the New Jersey Highlands in Warren County, New Jersey, Master Thesis, Lehigh University.

Kitchell, W., 1857, Annual Report of the Superintendent and State Geologist for 1856: Geological Survey of New Jersey, p. 5-38.

Lister, G.S. and Snoke, A.W., 1984, S-C mylonites: Journal of Structural Geology, v. 6, p. 617-638.

Lenik, E.J., 1996, Iron mine trails: New York – New Jersey Trail Conference.

Lupulescu, M. and Gates, A.E., 2006, Iron deposits from Hudson Highlands, NY: Systematics, mineralogy, mineral chemistry and tectonic setting: Geological Association of New Jersey Field Guide and Proceedings, v. 23, p. 46-59.

Martinko, G.S. and Gates, A.E., 2000, Hydrothermal mineralization and chemical changes associated with magnetite ore in Hogencamp iron mine, Hudson Highlands, NY: Geological Society of America, Abstracts with Programs, v. 32, p. 57.

McCaig, A.M., 1987, Deformation and fluid-rock interaction in metasomatic dilatant shear bands: Tectonophysics, v. 135, p. 121-132.

Murray, D.P., 1976, Chemical equilibrium in epidote-bearing calc-silicates and basic gneisses, Reading Prong, New York: PhD Thesis, Brown University.

Puffer, J.H., 2001, Five types of Proterozoic magnetite deposits in the New Jersey Highlands: Society of Economic Geology Guidebook Series, v. 35, p. 103-110.

Puffer, J.H. and Gorring, M.L., 2005, The Edison magnetite deposits in the context of pre-, syn-, and post-orogenic metallogenesis in the Grenville Highlands of New Jersey: Canadian Journal of Earth Sciences, v. 42, p. 1735-1748.

Rodgers, H.D., 1840, Final report on the geology of the State of New Jersey: Geological Survey of New Jersey, Report of the State Geologist, v. 36, p. 12-22.

Ruttman, F.S., 1887, Notes on the geology of the Tilly Foster ore body, Putnam County, New York: Trans. A. Inst. Min. Eng., v. 15.

Sibson, R.H., Moore, J.M., and Rankin, A.H., 1975, Seismic pumping-a hydrothermal transport mechanism: J Geol. Soc. London, v. 131, p. 653-659.

Sims, P.K. and Leonard, B.F., 1952, Geology of the Andover mining district, Sussex County, N. J.: New Jersey Department of Conservation and Development Bulletin 62.

Sims, P.K., 1958, Geology of the Dover magnetite district, Morris County, New Jersey: US Geological Survey Bulletin, 982-G, p. 245-305.

Spencer, A.C., 1904, Genesis of the magnetite deposits of Sussex County, New Jersey: Mining Magazine, v. 10, p. 377-381.

Spencer, A.C., 1905, Progress of work in the Precambrian rocks: Geological Survey of New Jersey Annual Report of the State Geologist, p. 161-186, p. 246-253.

Spencer, A.C., 1909, The Mine Hill and Sterling hill zinc deposits of Sussex County, New Jersey: Geological Survey of New Jersey Annual Report of the State Geologist, p. 23-52.

Volkert, R.A. and Drake, A.A. Jr., 1999, Geochemistry and stratigraphic relations of Middle Proterozoic rocks of the New Jersey Highlands: U.S. Geological Survey Professional Paper 1565-C, 77 p.

Volkert, R.A., 2001, Geologic setting of Proterozoic Iron, Zinc, and Graphite Deposits, New Jersey Highlands: Society of Economic Geology Guidebook Series, v. 35, p. 59-73.

Wendt, A.F., 1885, The iron mines of Putnam County, New York: Trans. A. Inst. Min. Eng., v. 13.