To the Graduate Council:

I am submitting herewith a thesis written by James Barry Hersch entitled "Petrographic Relationships among the Host Dolostone, Gangue Dolomite, and Ore Mineralization, in the Austinville-Ivanhoe Lead-Zinc District." I recommend that it be accepted in partial fulfillment of the requirements for the degree of Master of Science, with a major in Geology.

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PETROGRAPHIC RELATIONSHIPS AMONG THE HOST DOLOSTONE, GANGUE DOLOMITE, AND ORE MINERALIZATION, IN THE AUSTINVILLE-IVANHOE LEAD-ZINC DISTRICT

A Thesis

Presented for the

Master of Science

Degree

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James Barry Hersch

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ABSTRACT

The purpose of this investigation was to study the relationships between the host carbonates (Shady Formation) and primary sulfide mineralization in the Austinville-Ivanhoe lead-zinc district using standard petrographic and cathodoluminescence (CL) techniques. Hand specimens representing variations in host rock lithology and sulfide mineralization, were obtained from four ore horizons in the Austinville mine and one ore horizon in the Ivanhoe mine.

Petrographically, the Shady Formation carbonates may be divided into 3 main groups: (1) xenotopic, decimicron-sized, host dolostones which were deposited in a hypersaline, probably sulfate-rich, sedimentary environment; (2) microsparites which appear to be calcitized dolostones (dedolomites), and probably formed by replacement of preexisting dolomites due to Ca^{++} ions released from dissolution of sulfates; (3) white gangue dolomite ("met") comprising coarse-grained, hypidiotopic, baroque dolomite crystals which are often interlaminated with the fine grained host dolostone. The gangue dolomite exhibits microbrecciation under CL and apparently formed from a previous carbonate rock which was later subjected to solution collapse brecciation and neomorphism.

The primary ore in the district consists predominantly of pyrite, sphalerite and galena. Stratigraphically the ore zone is distributed over 364 meters in the Ribbon and Austinville Members of the Shady Formation. Favorable beds were mineralized as open space filling during one main period of mineralization. The ore appears to have been

emplaced during or after orogenic movements with migration of the ore fluids along zones of greatest permeability.

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CHAPTER I

INTRODUCTION

I. SCOPE OF THE PRESENT INVESTIGATION

The Austinville-Ivanhoe lead-zinc deposit of southwest Virginia (Figure 1) has been classified by various authors as a Mississippi Valley-type deposit (Lindgren, 1933; Callahan, 1964; Brown and Weinberg, 1968). Although features such as the host dolostones, ore mineralogy, low-temperature and predominantly open-space filling mineralization support such a classification, in some respects the Austinville-Ivanhoe deposit is dissimilar to many Mississippi Valley-type deposits. This is especially true in comparing these to other Mississippi Valley deposits in East and Central Tennessee. The Austinville-Ivanhoe district is the only major Cambrian base metal deposit in the eastern U.S. (Brown and Weinberg, 1968). The ore occurs in the Lower Cambrian Shady Formation (Dolostone) and it is not associated with an unconformity. Despite the absence of an unconformity, there is still an abundance of brecciation similar to that found in other Mississippi Valley-type zinc deposits of the Appalachian region. Many writers (Brown, 1935; Currier, 1935a; Brown and Weinberg, 1968; Weinberg, 1971) are of the opinion that some of the brecciation is tectonic in origin. A nontectonic origin for the brecciation has also been proposed. Callahan (1964) suggests that the brecciation may be the result of dissolution of favorable stratigraphic horizons vertically stacked along a facies

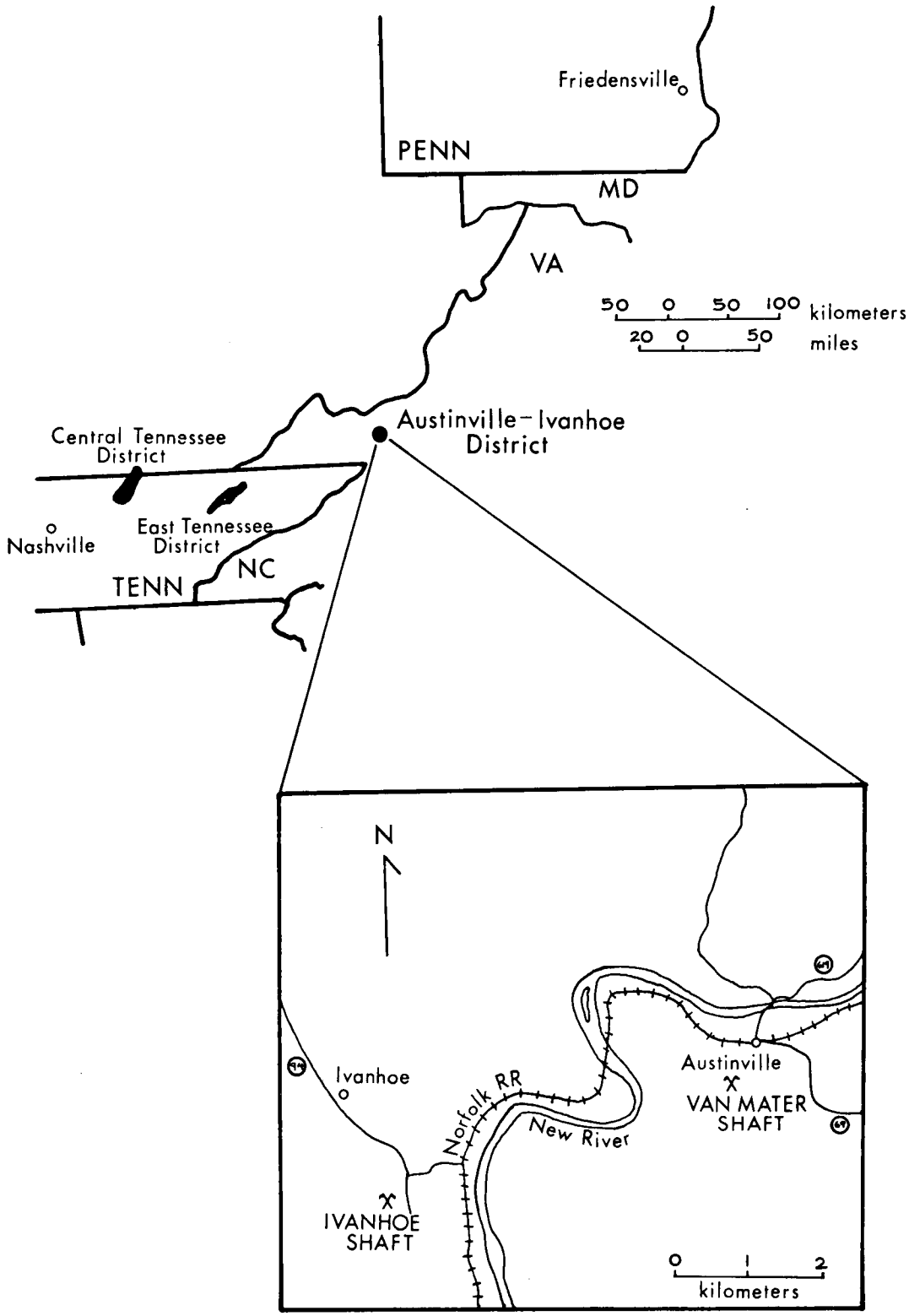


Figure 1. Map of the Austinville-Ivanhoe lead-zinc district, southwest Virginia.

change. Hoagland (1971) contends that the brecciation represents a fore-reef facies which has been selectively mineralized.

The facies relationships within the Shady Formation are difficult to decipher because of pervasive thrust faulting. Currier (1935b) and Stose and Jonas (1939) postulate that the pronounced facies changes at Austinville are simply the effect of juxtaposition of significantly different sedimentary environments by thrust faulting.

The problems concerning the sedimentary setting, reason(s) for brecciation, origin of the gangue dolomite ("met") and relationship of the sulfide mineralization to orogenic movements are by no means settled. This paper investigates the relationships between the host dolostone, gangue dolomite, and ore minerals, by selective sampling of four of the ore horizons exposed in the subsurface workings of the Austinville-Ivanhoe mines.

II. SAMPLES AND STUDY METHODS

To cover the range in variation of the host rock lithologies in the major ore horizons in the Austinville-Ivanhoe mining interval, a total of 170 hand specimens representative of 4 different ore horizons, were obtained from 33 stopes in the south-limb facies (Austinville) and 5 stopes in the north-limb facies (Ivanhoe). The relative stratigraphic positions of these ore horizons (Malden, Mottled Marker, 40 Orebody and 100 Orebody) in the south-limb facies are shown in Figure 2. The ore horizon represented by the north-limb facies specimens is roughly equivalent to the 40 Orebody horizon of the south-limb facies.

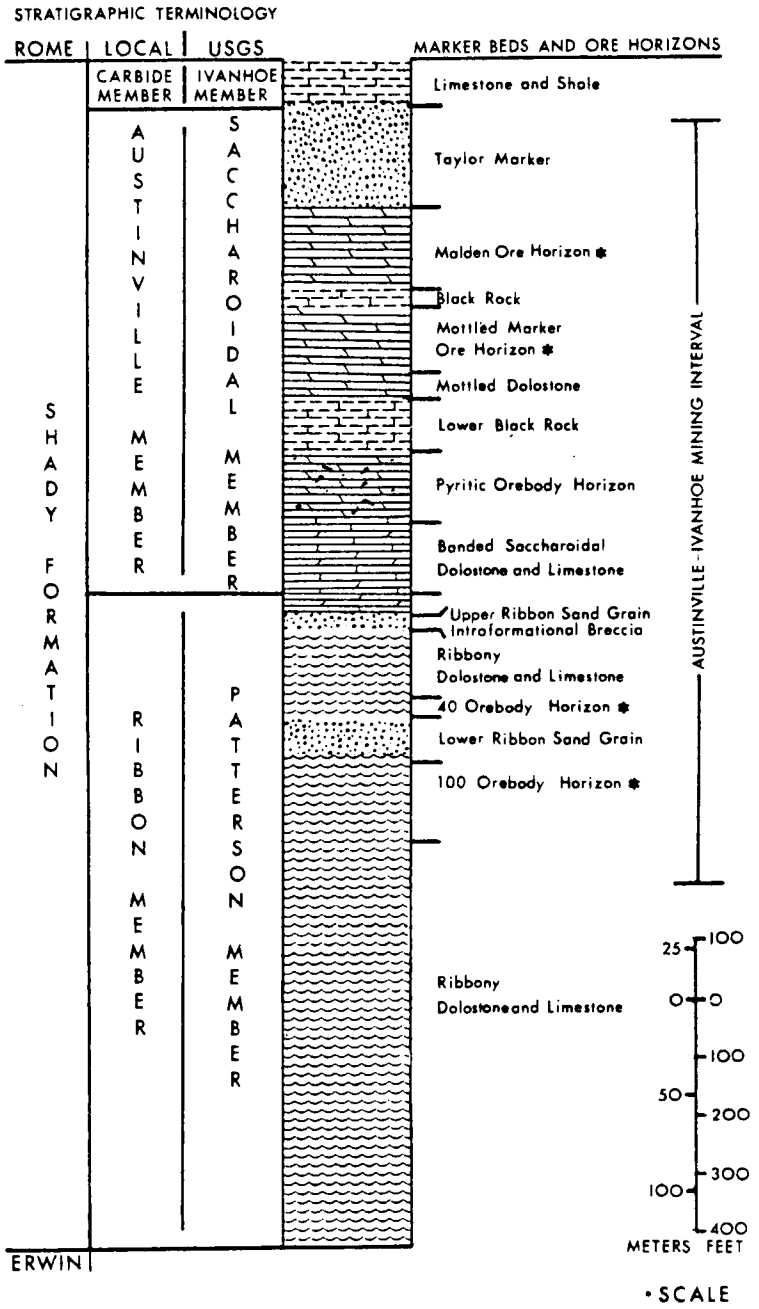


Figure 2. Generalized stratigraphic column of the south limb facies at Austinville, after Brown and Weinberg (1968)

*Indicates ore horizons from which specimens were obtained for the present study.

All the samples from the Malden and Mottled Marker ore horizons and a few of the specimens from the 40 and 100 ore horizons were provided by Michael L. Ebers (Mine Geologist, Austinville). The remainder of the samples were collected by the author under the guidance of Edgar L. Weinberg (Regional Geologist) and Michael L. Ebers. Further details of specimen locations and sampling procedures are included in the Appendix. Laboratory study included examination of polished slabs, transmitted-light microscopy of stained thin sections, reflected-light microscopy of polished sections, and study of thin sections with cathodoluminescence (CL). Several of the above samples were acid leached and the resulting insoluble residues were examined under the petrographic microscope. Carbonate rocks were classified according to the crystal form and size (Friedman, 1965).

III. PREVIOUS WORK

Results of stratigraphic studies on the Shady Formation in the vicinity of the Austinville-Ivanhoe district have been reported by Currier (1935a,b), Stose and Jonas (1939), Stose and Stose (1957), Brown and Weinberg (1968), Weinberg (1971), and Byrd (1973). Two distinct nomenclatures for the members of the Shady Formation are in common usage. According to the U.S.G.S. nomenclature, the Shady Formation is subdivided into Patterson, Saccaroidal, and Ivanhoe Members, whereas mine personnel and local workers prefer a subdivision into Ribbon, Austinville and Carbide Members (Figure 2, p. 4). In this study the older, "local" terms Ribbon, Austinville, and Carbide Members will be used.

Published discussions specifically related to the relationships between the host dolostones, mineralization, and gangue dolomite are limited, the papers by Brown (1935) and Brown and Weinberg (1968) being notable exceptions. These workers drew a direct, quantitative relationship between the intensity of mineralization and the amount of coarsely crystalline, white gangue dolomite ("met"). From field relations and megascopic petrography, these authors concluded that the ore fluids migrated along faults and mineralized favorable horizons which had been brecciated.

Byrd (1973) provided a regional investigation of the Shady Formation based on examination of measured sections and petrographic study. He described briefly the brecciation, evaporite pseudomorphs, and facies relationships southeast of Austinville. Byrd concluded that the lithofacies he observed could be best explained by an environmental pattern analogous to the Bahama Banks.

Biostratigraphic and paleontological studies of the area have been reported by Resser (1938), Yochelson (1970), Byrd, Weinberg, and Yochelson (1973), and Balsam (1974). The above workers are in general agreement on the stratigraphic divisions of the Shady Formation and the identification of the fossil allochems which it contains. Resser (1938) reported the occurrence of the reef former Archeocyathus in the Carbide Member; however, as reported by Balsam (1974) some workers believe that the so-called "reef" simply represents a mud bank with abundant shrimp burrows.

CHAPTER II

GEOLOGIC SETTING

I. STRATIGRAPHY

In the Austinville-Ivanhoe district, the Lower Cambrian Shady Formation contains all of the sulfide mineralization. The stratigraphic sequence of the study area is given in Figure 2 (p. 4).

As indicated in Figure 2 (p. 4), the vertical extent of mineralization is approximately 365 meters thick. The total thickness of the Shady Formation in the vicinity of Austinville is estimated to be approximately 608 meters (Brown and Weinberg, 1968).

In the Austinville-Ivanhoe area, the Shady Formation gradationally overlies the Erwin Quartzite which is a member of the Chilhowie Group (Lower Cambrian). The Erwin formation is a thick bedded, locally shaly, protoquartzite, with an approximate thickness of 456-608 meters (Cooper, 1961).

Gradationally overlying the Erwin Quartzite is the lowermost member of the Shady Formation, the Ribbon Member (Currier, 1935b). Characteristically, it is a ribbon-textured limestone and dolostone with wavy carbonaceous partings (Currier, 1935b). The ribbon texture arises from various combinations of carbonaceous material, shale, and dolostone interlaminated with coarse white gangue dolomite (Currier, 1935b).

Conformably overlying the Ribbon Member is the Austinville Member. The Austinville Member has been described as a saccharoidal, locally

shaly, dolostone with intercalated limestone lenses. Mine geologists for the New Jersey Zinc Company have divided the Austinville Member into nine lithologic subunits as shown in Figure 2 (p. 4).

Conformably overlying the Austinville Member is the Carbide Member which is a shaly, locally fossiliferous, calcilutite of limited areal extent. The occurrence of the Carbide Member is restricted almost exclusively to the immediate vicinity of Austinville-Ivanhoe district (Resser, 1938). Exposed Carbide units southeast of Austinville include the post-Taylor-Marker (PTM of Figure 3) beds that are possibly time-equivalent to the basal Rome Formation.

The Rome Formation (Lower Cambrian) which conformably overlies the Shady Formation is approximately 547-820 meters thick in the Austinville-Ivanhoe area (Butts, 1940). It consists of red sand and siltstones interbedded with light grey dolostones and limestones. The unit is characterized by abundant primary sedimentary features, such as mudcracks, ripple laminations, raindrop impressions, and evaporite pseudomorphs.

The Carbide, Austinville, and Ribbon Members of the Shady Formation show facies variations in the Austinville area along and across strike (Figure 3). The most significant of these facies variations occurs along dip and across the prominent structure in the area, the Austinville Anticline. Three different facies of the Shady Formation have been recognized. These are referred to as the Laswell facies, the north-limb facies, and the south-limb facies. The rocks of the Shady Formation northwest of the Laswell Fault are referred to as the Laswell facies.

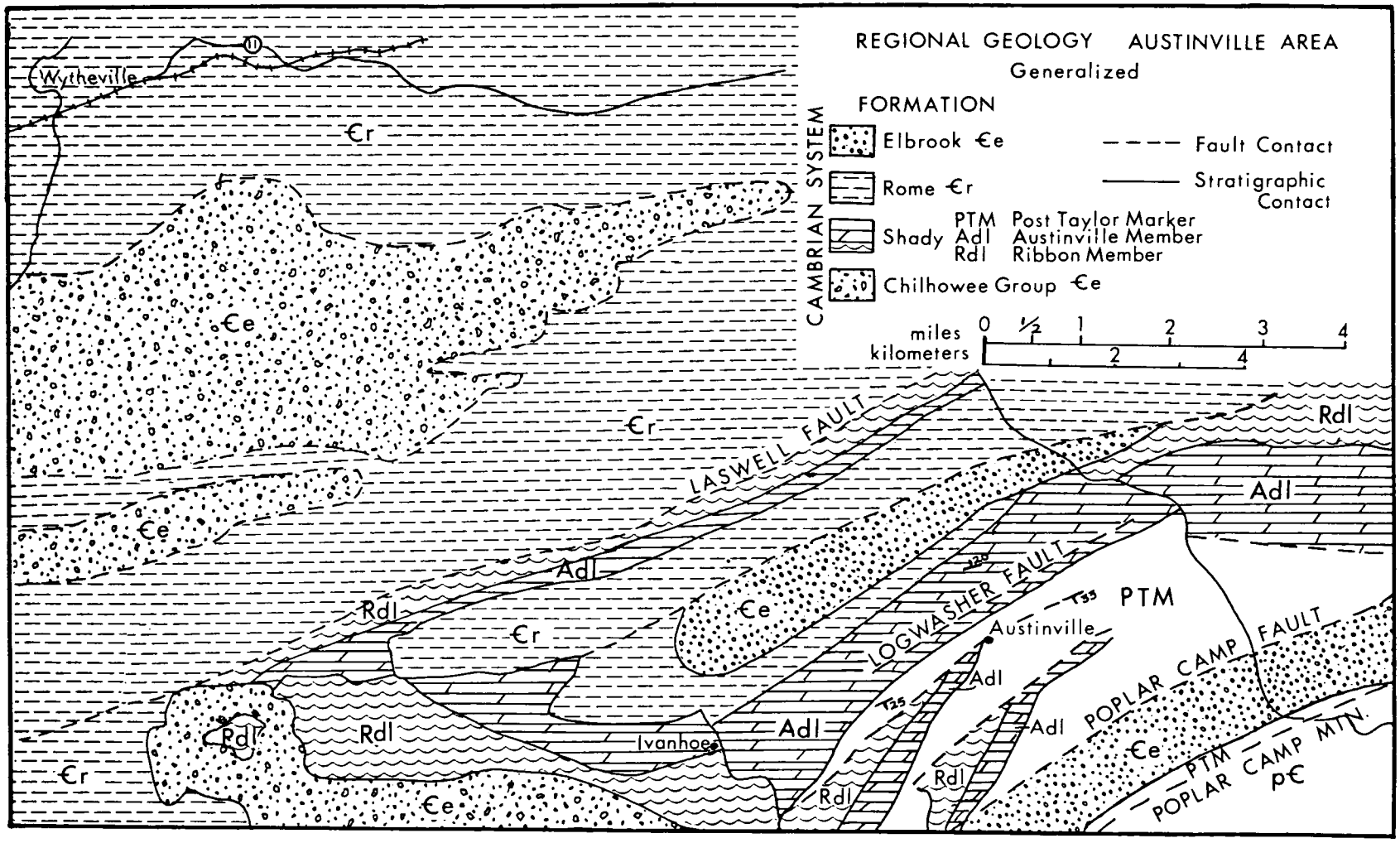


Figure 3. Regional geology of the Austinville Area, modified from Brown and Weinberg (1968).

The north-limb facies is limited to the area between the Laswell and Logwasher Faults on the north-limb of the Austinville Anticline. South of the Logwasher Fault and north of Poplar Camp Mountain is the south-limb facies (Figure 3).

The north-limb facies of the Austinville Member differs from the south-limb facies in four general ways: (1) the shale content of the dolostone increases to the southeast; (2) the south-limb facies has distinctive marker beds which are absent in the north-limb facies; (3) Lower Cambrian fossils present in the south-limb facies are not found in the north-limb facies; and (4) the detrital content of the Austinville Member increases in the north-limb facies (Stose and Jonas, 1957; Brown and Weinberg, 1968).

The Laswell facies is similar to the Austinville Member, the former being composed of saccharoidal grey and white dolostones (Currier, 1935b).

The ore deposits are localized in the transition zone between the north-limb facies and south-limb facies. This suggests that the facies change might have been the major control (Callahan, 1964) or at least one of the significant controls of ore localization (Brown and Weinberg, 1968). The origin of the facies change itself is controversial. The transition between the north and south-limb facies has been interpreted variously as fore-reef breccia (Hoagland, 1971), a true sedimentary facies change of laterally contiguous environments (Callahan, 1964), or lithologic changes created by pervasive imbricate thrust faults (Stose and Jonas, 1939).

II. STRUCTURE

The Austinville-Ivanhoe district is located near the boundary between the Appalachian Valley and Ridge Province and the Blue Ridge Province. The gently folded sedimentary rocks of the Appalachian Valley and Ridge Province have been structurally complicated by numerous imbricate faults. In the study area, the Shady Formation has been overthrust to the northwest onto the Rome Formation. To the southeast, the exposed Shady Formation beds are overlain by Cambrian and Precambrian rocks of the Blue Ridge thrust sheet.

According to some authors the region is characterized by a series of broad folds and associated thrust faults which moved strata over large distances (Currier, 1935a; 1935b; Stose and Jonas, 1939). However, Callahan (1964) suggested that the faulting actually took place along high angle reverse faults (on the order of 45°) which caused very little lateral displacement of strata.

Figure 4 is a map of the structural features of the Austinville area as interpreted by Currier (1935b). The major structural features of the area are the Laswell Overthrust, the mineralized Austinville Anticline, Hematite Mountain Overthrust, Gray's School Fault, and the New Castleton Fault.

Stose and Jonas (1939), as illustrated in Figure 5, disagreed with the map by Currier (Figure 4) and indicated that the Gray's School Fault and the New Castleton Fault are actually one fault, the Little Mountain Overthrust. The implication of these two interpretations will become evident in the following discussion.

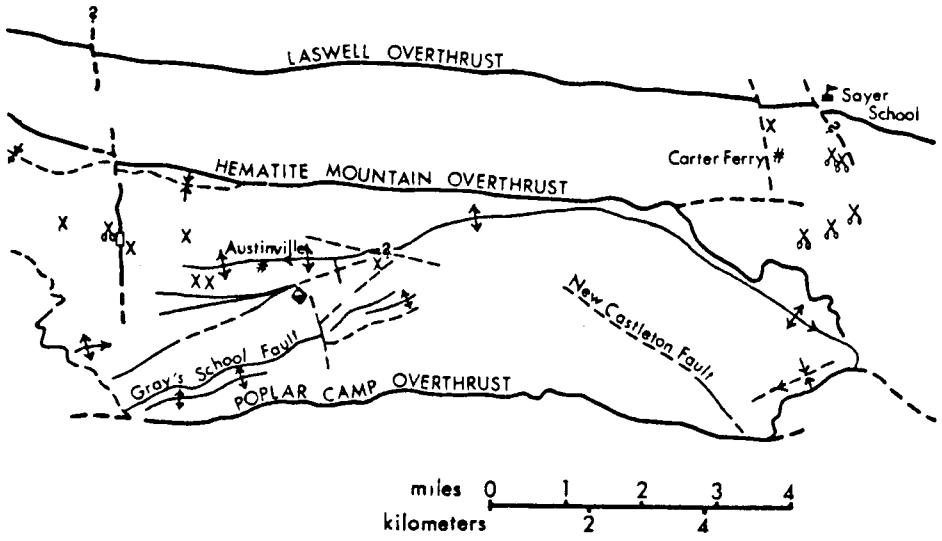


Figure 4. Structural features and occurrence of lead-zinc ores, from Currier (1935b).

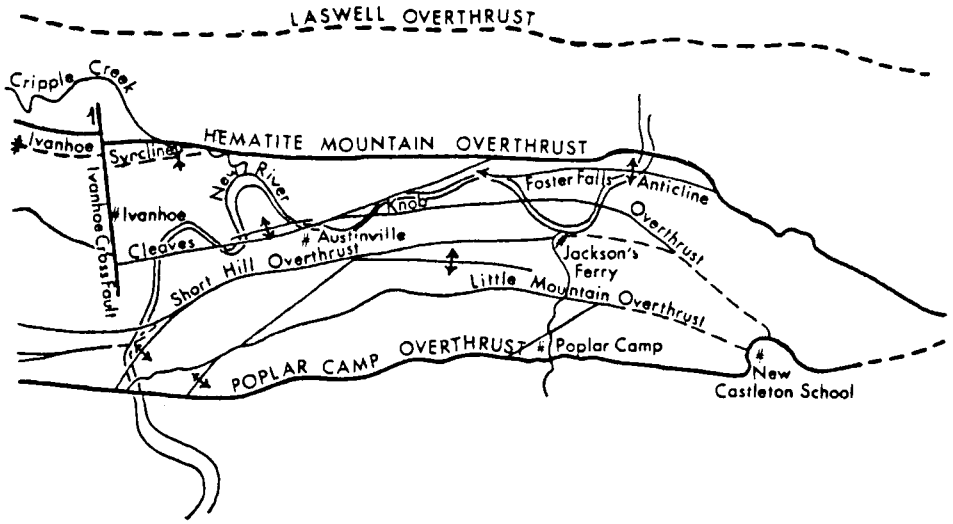


Figure 5. Map of the Austinville fault blocks within the overthrust mass, from Stose and Jonas (1939).

Cutting across the strike of the thrust faults and the crest of the Austinville Anticline is a series of cross faults (Figures 4 and 5). These include the Ivanhoe Fault (due west of Austinville), and several small unnamed cross-faults described by Brown (1935) and Currier (1935b).

The cross faults offset the major thrust belts in both Currier's (Figure 4) and Stose and Jonas' interpretation (Figure 5). Currier (1935a) stated that "there is no conclusive evidence as to the direction of movement along the fault planes." He continued to say that the cross-faults "are practically contemporaneous with the major deformation, following it very slightly in point of time and in essential continuity."

As mentioned earlier, the mineralized rocks of the district are primarily in the zone where the south-limb facies grades into the north-limb facies (Callahan, 1964; Brown and Weinberg, 1968). Callahan (1964) indicated that a sedimentary facies change is the controlling factor for the localization of the ore and rejected the idea of thrust faulting proposed by Stose and Jonas (Figure 5). Callahan discounted these thrust faults as being capable of moving strata over long distances by stating,

> Thrust faulting is present in the area but our data indicates that the dip thereof is of the order of 45°, at the least 10° steeper than dip of bedding, and the stratigraphic displacement is small. Consequently, it appears that location of the facies change is proximate to its present position rather than originally remote therefrom.

Stose and Jonas (1939) explained the mineralized facies change in a different manner. According to them, the south-limb facies was developed a considerable distance to the southeast and was thrust to its present position (Short Hill Fault Contact, Figure 5). If this interpretation

is correct there may not be a drastic sedimentary facies change as proposed by Callahan (1964). The observation that reverse faults in the Shady are on the order of 45° dip should not negate the Stose and Jonas model. Thrust faults, capable of displacing strata many kilometers often travel at low angles through incompetent beds such as shale, and then ramp through more competent beds (e.g., carbonates), at a much higher angle (King, 1960).

CHAPTER III

PETROLOGY

I. HOST ROCK CARBONATES

Examination of thin sections with the petrographic microscope and CL demonstrated that three distinct lithologies are present in the ore horizons sampled for the current study (Figure 2, p. 4): (1) Host Dolostone, (2) Microsparite, (3) Coarsely Crystalline White Gangue Dolostone.

The host lithology for sulfide mineralization in the district is a demicron-sized, xenotopic, dolostone. The grain size of the host dolostone is not constant. In many of the thin sections studied, the grain size varies systematically across the sections. Textural fabrics of this type have been interpreted by Bathurst (1975) to be indicative of neomorphism. The crystal size of the host dolostone may not change uniformly and the variability in grain size often has an erratic distribution. The variability in crystal size of the host dolostone is typically observed in the vicinity of stylolites. Textures of this type are possibly related to pressure solution or hydrothermal solutions accompanying sulfide mineralization. The host dolostone is of secondary (penecontemporaneous) origin. This is suggested by the replacement of earlier organic structures such as burrows, pellets and fossil allochems by dolomite crystals, and by the extreme fine-grained texture of the dolostone (Blatt et al., 1972).

Accessory lithologic constituents include pellets, invertebrate fossils (trilobites, conulariids, and pelmatozoans), evaporite pseudomorphs, carbonaceous matter, silt-sized euhedral microcline, sand-sized quartz, clays, chalcedony spherules, and occasionally, vug-filling euhedral calcite.

Sedimentary fabric(s) observed in these rocks may be classified as primary or secondary. Primary features, formed during sedimentation or early surface exposure, include burrows, laminations, geopetal structures, and solution breccias. Secondary features formed after burial include stylolites, tectonic breccias and fracture patterns.

Accessory Lithologic Constituents

Fossils. Individual fossil specimens or fragments are scarce in the Shady Formation. In the study interval (Figure 2, p. 4) where found, they mostly occur as fragmental accumulations along stylolite seams (Figure 6). Extensive neomorphism has obscured the primary skeletal microstructures of all the recognizable fossils in the ore zones. Most of the Shady fossils are tiny allochems which are difficult to identify; however, some of the fossils can be identified by external morphology where whole or almost whole individuals are present.

Salterella conulata, a Lower Cambrian elongate conical fossil of uncertain affinities, was found in samples from 100 Orebody, Austinville mine and the 8th level drift (40 Orebody Horizon), Ivanhoe mine (Figure 2, p. 4). The samples containing these Salterella conulata fossils were collected stratigraphically 230-260 meters above the contact of the



Figure 6. Photomicrograph showing Salterella conulata in a stylolite seam observed with normal light. Note that the opaque area on the right portion of the slide is insoluble residue of stylolite. Bar scale is 1 mm long (Sample Number 8-29).

Ribbon Member with the underlying Erwin Quartzite. Occurrence of Salterella conulata at roughly equivalent stratigraphic position has been reported from surface exposures in two areas in Virginia as well as in subsurface drill cores at the Austinville mining district (Byrd et al., 1973). These occurrences of S. conulata in the subsurface workings of the 100 Orebody (Austinville) and 8th level drift (Ivanhoe) is important because of its potential usefulness as an index fossil for the Lower Cambrian Carbonate in the Appalachian region (Yochelson, 1970; Byrd et al., 1973).

Trilobite fragments ranging from 0.5 to 5 mm in size are common in the Shady Formation. Most are tiny allochems recognizable as trilobites by their typical "button hook" appearance and diagnostic sectional views (Figure 7). Specific and generic names could not be determined because of fragmentation and poor preservation.

A kind of circular fossil fragment occurs throughout the section studied. These enigmatic fossils consist of plates of dolomites or cryptocrystalline quartz with centers of recrystallized dolomite (Figure 8). The exact nature of these fossils is not known although their morphology is suggestive of pelmatozoan columnals. Byrd (1973) contends that these may represent the primative echinoderm Helicoplacus, which is believed to have been an unattached echinoderm.

Peloids—pellets. These two allochems are discussed together because the distinction between them is usually a matter of their definition. Peloid is a general term first proposed by McKee and Gutschick (1969) to embrace spherical to elliptical aggregate grains of



Figure 7. Photomicrograph of host dolostone showing a trilobite fragment when observed with normal light. Bar scale is 1 mm long (Sample Number 8-24).

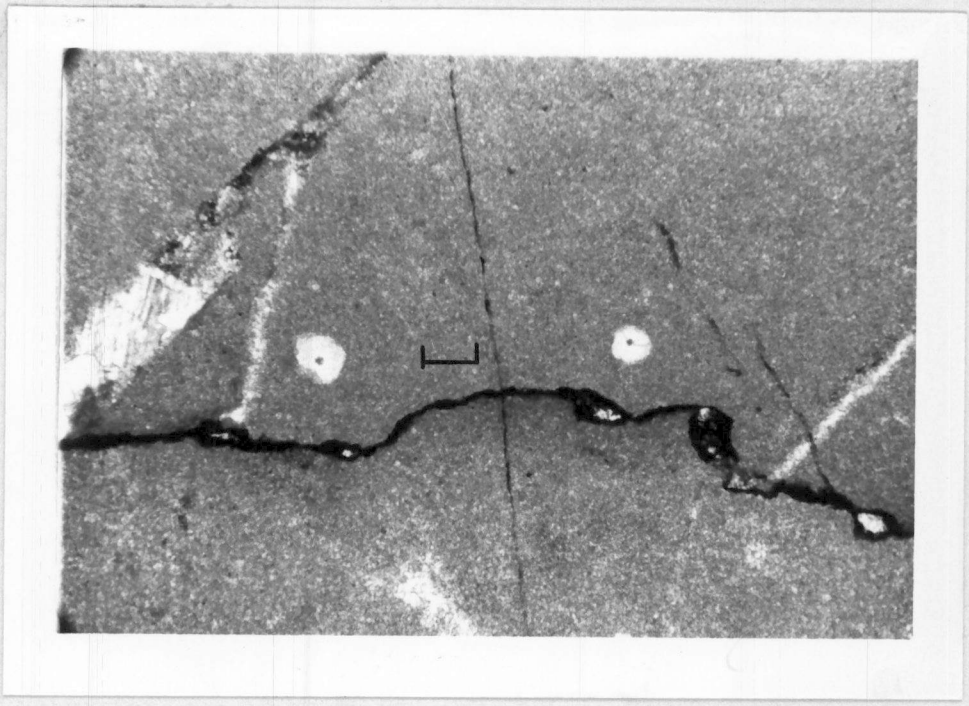


Figure 8. Photomicrograph showing enigmatic fossils resembling pelmatozoan grains observed under normal light. Bar scale is 1 mm long (Sample Number 8-27).

carbonate material when the mode of origin is unknown. Pellet (Folk, 1962) is a more restricted term than peloid, and its usage is limited to describe rounded, commonly ovoid aggregate accretionary carbonate material of fecal origin. Some authors, notable Bathurst (1971) and Brown (1965), contend that most ellipsoidal peloids observed in ancient sediments are probably of fecal origin.

The peloids observed in the Shady Formation have been neomorphosed into xenotopic dolomite. However, ghosts of peloids are common in host dolostone; their elliptical shape being attributable to flattening during neomorphism or possibly to fecal origin (Figure 9).

Other minerals. Much of the host rock, especially the Ribbon Member, has a dark coloration which is caused by laminations of what appears to be organic residue (Figure 10). Minute pyrite cubes are associated with the dark laminae of the Ribbon and Austinville Members. The black residue is probably a mixture of carbonaceous material and pyrite whose origin may be related to the locally reducing conditions caused by organic decay. This carbon and sulfide may have originated from the decomposition of algae, as suggested by Gebelein and Hoffman (1971).

Chert nodules and chalcedony concretions are common in certain horizons of the ore zones. The cryptocrystalline quartz seems to be of two varieties: chalcedony or a fibrous variety which is optically length-slow (Figure 11), and a cryptocrystalline variety of chert, occasionally fibrous, which is optically length-fast. According to Folk and Pittman (1971), length-slow chalcedony characterizes silicification in sulfate-rich environments and is often the only

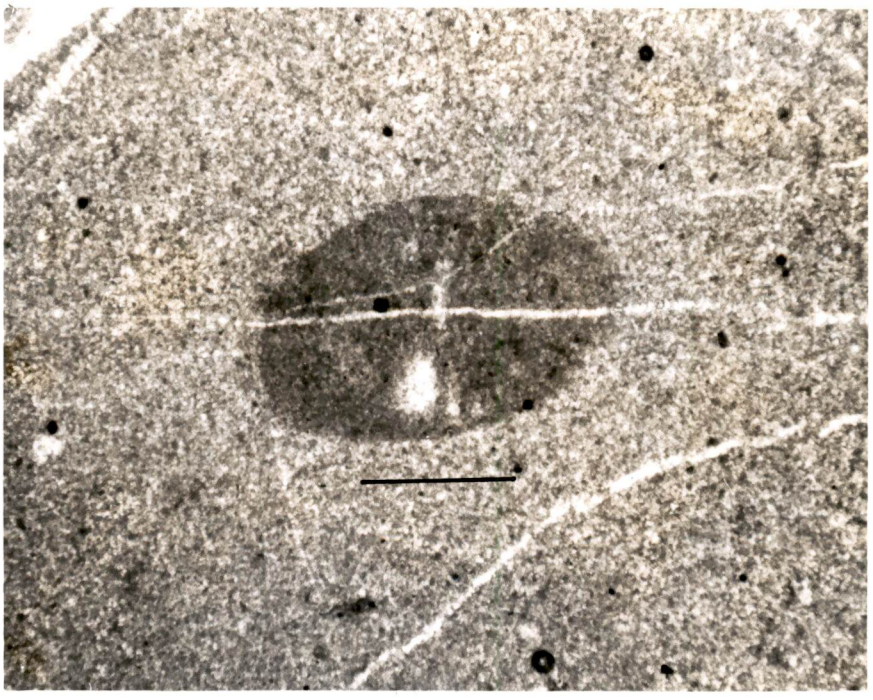


Figure 9. Photomicrograph showing flattened peloid when observed under normal light. Bar scale is 1 mm long (Sample Number 26).



Figure 10. Photomicrograph showing apparent organic-rich laminations alternating with white gangue dolomite. This is the characteristic texture of the Ribbon Member in the ore zone when observed under normal light. Bar scale is 10 mm long.

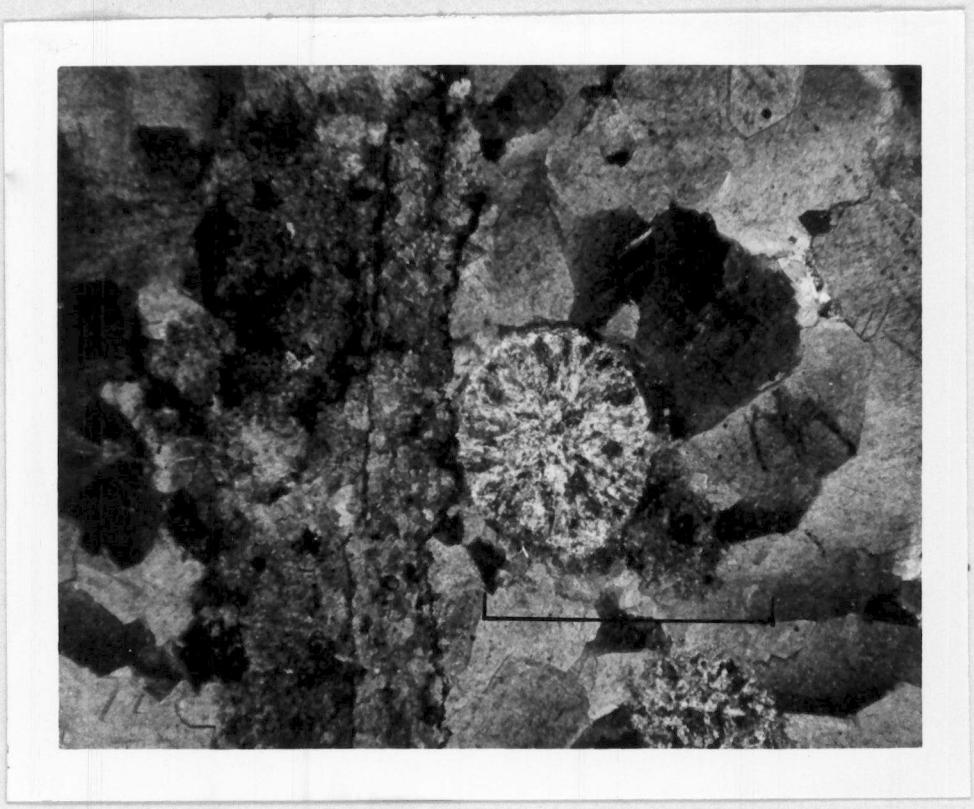


Figure 11. Photomicrograph showing chalcedony spherules observed under crossed-nicols. The chalcedony is optically length-slow, suggestive of formation by replacement of evaporite minerals. Bar scale is 1 mm long (Sample Number 31).

available indicator of "lost" evaporites in a particular stratigraphic section; the length-fast chalcedony forms later, after complete flushing of the evaporitic fluids.

Silt, clay, and quartz are often associated with the dark laminae discussed above. The silt is composed of well-sorted, angular quartz and feldspar grains ranging from 0.02 mm to 0.1 mm in size. The clay content of the host dolostone is quite variable and ranges from less than 1 percent up to approximately 50 percent of the rock.

Insoluble residues of samples in the vicinity of specimens containing abundant evaporite pseudomorphs (discussed later) were found to contain euhedral, small (30-40 microns in size) microcline crystals. These were identified by their diagnostic twinning.

The detrital versus authigenic origin for these microcline crystals is problematical. The evidence for authigenic growth lies mainly in the observation that the microcline crystals are euhedral and of uniform size. Similar lines of evidence for authigenic growth of albite crystals (Lowville, Limestone, Pa.) have been reported by Honess and Jefferies (1940). The observation of crystal habit, uniformity in size, and the apparent absence of detrital cores make a detrital origin for the feldspar unlikely.

Authigenic feldspar in Lower Paleozoic carbonates have been reported to be indicators of possible hypersaline conditions (Buyce and Friedman, 1975). Alberstadt and Reesman (1978) concur with this conclusion and suggest that evaporation on a tidal flat may increase the K ion concentration in interstitial pore water sufficiently to form feldspar crystals provided aluminum and silicon are also present. Honess and

Jeffries (1940) and Buyce and Friedman (1975) suggest that alteration of pyroclastics present in carbonate muds by interaction with sea water may result in the formation of authigenic feldspar; another source for the ions necessary for authigenic feldspar formation could be detrital clay minerals. The possibility that authigenic feldspar is an indicator of hypersalinity is at present debatable. However, sufficient evidence shall be discussed later in this paper which supports the hypersalinity hypothesis for authigenetic feldspar growth.

Primary Sedimentary Features

Silt-sized quartz laminations. A ubiquitus feature of the host dolostone is the presence of light grey, undulatory laminations which are essentially parallel to bedding and are composed chiefly of silt-sized quartz (Figure 12). The silt is well sorted and angular. These laminations are most abundant in the Ribbon Member of the Shady Formation, but they actually occur throughout the study interval. At present, there is no conclusive evidence to suggest an organic (algally trapped) sedimentary origin for this primary feature.

Pseudomorphs of dolomite. Isolated dolomite crystals ranging in size from 0.1 mm to 0.2 mm in width by 0.1 mm to 0.5 mm in length are common in unfractured areas of the Shady Formation. The majority of these crystals appear to be in form with rectangular and rhombohedral variations in shape caused by oblique orientation of thin sections through the crystals (Figure 13). From their external forms the

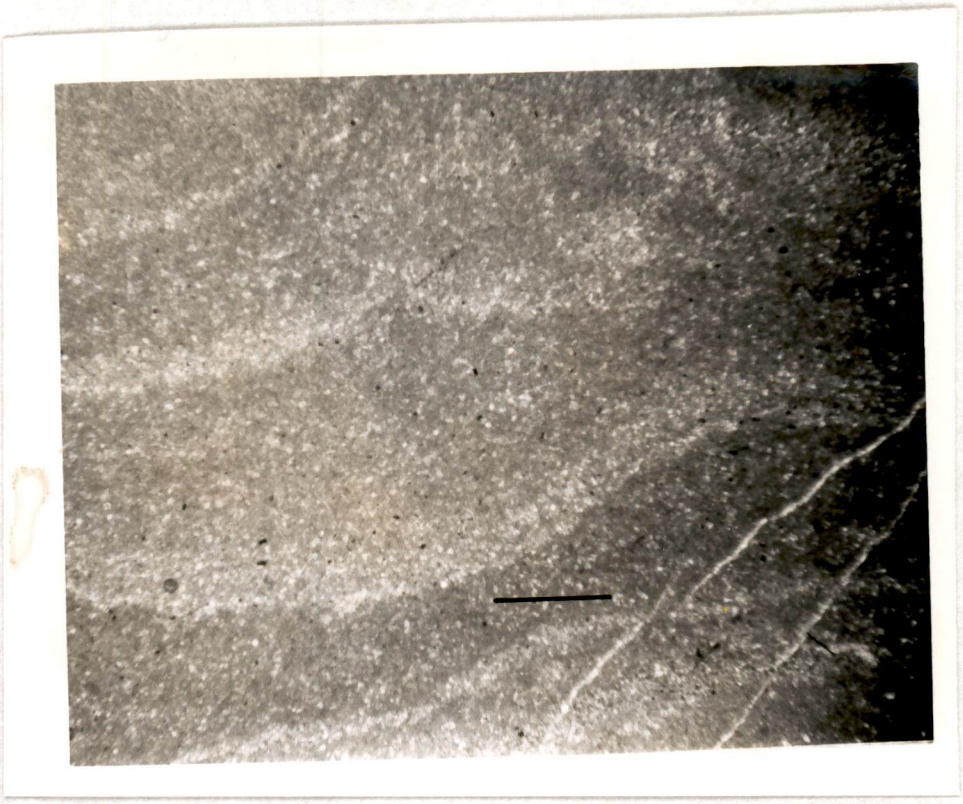


Figure 12. Photomicrograph showing three positively graded quartz-silt laminations when observed under normal light. The laminations are indicated by numbers 1, 2, 3, to the left of the figure. Bar scale is 1 mm long (Sample Number 13).

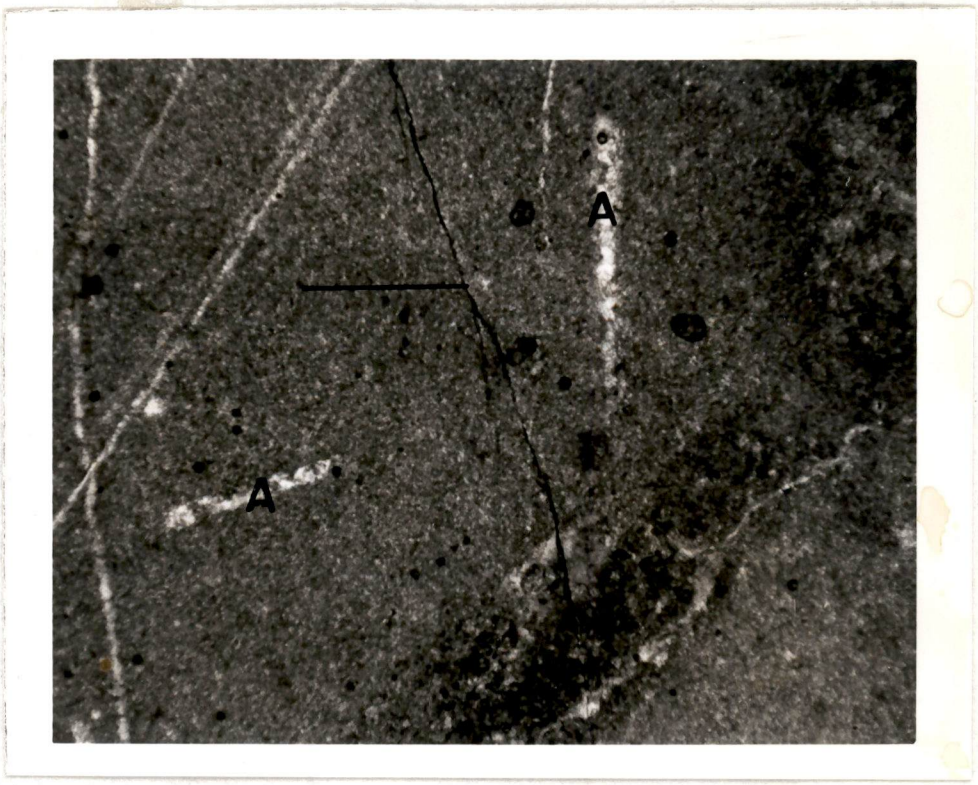


Figure 13. Photomicrograph showing dolomite (A), pseudomorphous after sulfate minerals when observed under normal light. Bar scale is 1 mm long (Sample Number 14).

dolomite crystals appear to be pseudomorphs after sulfate minerals (probably gypsum) as reported by Lucia (1972) and Gill (1977). Byrd (1973) has described similarly shaped dolomite crystals in the Ribbon and Austinville Members of the Shady Formation. Byrd has suggested that these dolomite crystals were originally gypsum, probably deposited in a supratidal environment.

Geopetal fabric. Geopetal structures or "geopetal fabric" are internal features which indicate the original orientation of stratified rock (Sander, 1951). The Shady Formation, especially the Ribbon Member, has abundant geopetal structures. The features are generally limited to oblate spheres and discontinuous pod-like features in the host dolostone which are filled with "baroque" dolomite (dolomite with undulatory extinction) and silt-sized dolomite (Figure 14). The features shown in Figure 14 resemble structures interpreted by Lucia (1972) as dolomitized anhydrite nodules. The nodules may have originally been filled with an aggregate of dolomite and anhydrite crystals. The more soluble anhydrite was leached away causing the dolomite crystal to collapse and fill the bottom of the void. Hypidiotopic, baroque dolomite spar then filled the remainder of the void (Figure 14). White baroque dolomite which filled fractures and voids of Triassic rocks in Italy has been interpreted by Folk and Asserto (1974) to be indicative of a hyposaline (meteorically-diluted) environment with Ma/Ca ratio near 1:1 and possible presence of sulfate or sulfide ions. This is in agreement

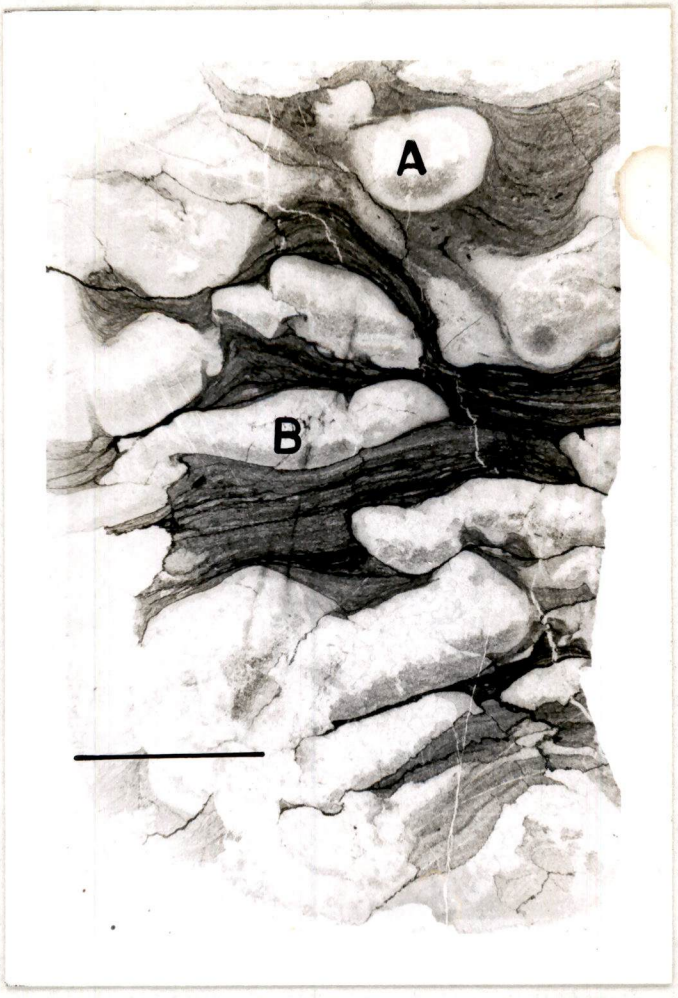


Figure 14. Photomicrograph showing numerous geopetal structures when observed under normal light. Note the nodular variety of geopetal structure (A) and a podiform variety (B). Bar scale is 10 mm long (Sample Number 24).

with the conclusion presented earlier (Figure 13, p. 28) that the isolated crystals of dolomite represent pseudomorphous replacement of sulfate minerals.

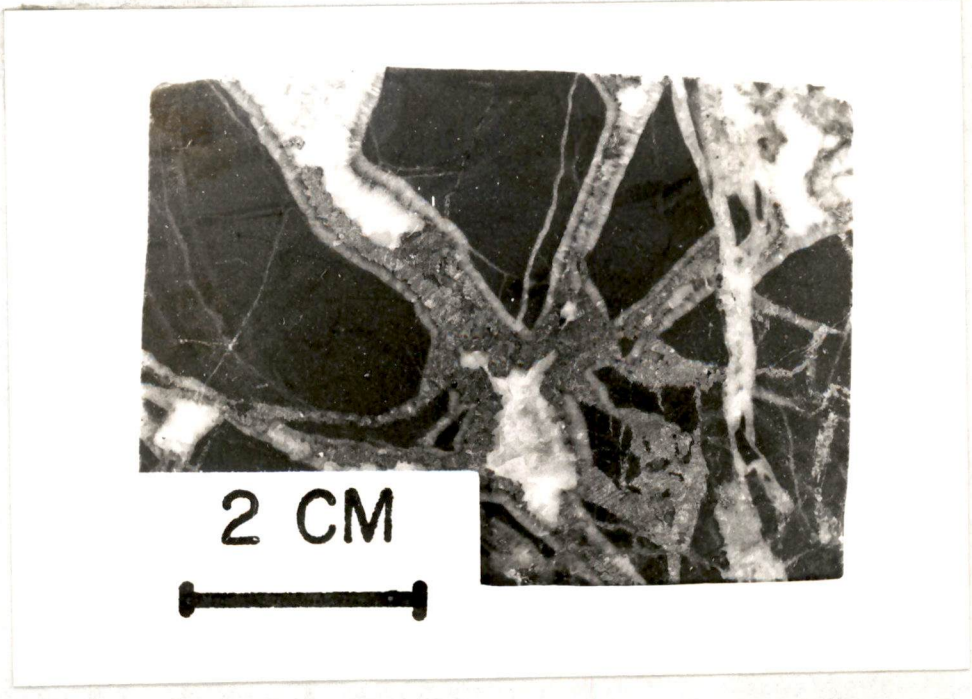
Another possible explanation for these structures could be the effects of pressure solution or the dissolution and replacement of fossils. However, this appears unlikely in view of the relative scarcity of fossils in the Shady Formation.

Brecciation. Brecciation in the Shady Formation is similar to that found in the East Tennessee zinc district, although there is no evidence of the brecciation being related to a post-Shady unconformity. Crackle and rubble brecciation structures (Norton, 1917) are present, ranging in size from small slump breccias to rubble breccias hundreds of meters in extent (Figure 15A, B). The host rock breccia clasts vary in size from 0.1 mm up to 60 mm. Sulfide mineralization with calcite, dolomite, or chert are often associated with the brecciated zones as typical open-space fillings (Figure 15B). The origin of these breccia bodies is problematical with sedimentary (Callahan, 1964), tectonic, and fore-reef talus origins having been proposed.

Borings and burrows. Circular openings, now filled with neomorphic dolomite spar or internal sediments are scarce in the decimicron-sized, xenotopic, host dolostone. Often the burrows occur as ghosts in the host dolostone and are recognizable only by changes in crystallinity or color.



(A)



(B)

Figure 15. Photomicrograph showing typical crackle and rubble brecciation structures when observed under normal light. Note the unmineralized crackle breccia (A, Sample Number 5-34-59) and the mineralized rubble breccia (B, Sample Number 7-40-65). Bar scale is 10 mm long (A).

Secondary Features

Stylolites. Stylolites are the most common structure present in the host dolostone (Figure 16). They are common in every stratigraphic unit examined and many laminae are actually microstylolitic seams. The stylolites vary in form from sutured to undulatory or hummocky in cross sectional view. They may be either parallel or oblique to bedding, with relief along the stylolite surface varying from less than 1 mm to tens of mm. Concentrations of dark black organic residue with disseminated pyrite occur in many of the stylolite seams.

Stylolites are caused by stress-induced solution as documented by Sorby (1879, 1908), Stockdale (1943), Dunnington (1954), Bathurst (1971), and Logan and Semeniuk (1976). The sutured variety of stylolites are similar to stylolites described by Trurnit (1968), which are thought to be indicative of an advanced stage of solution by grains of equivalent relative pressure solubility. Sutured solution seams that are parallel to bedding seem to result from overburden stress (Wanless, 1977). The rock units of the Shady Formation have been subjected to tectonic stress and this might have resulted in the stylolites that occur at sharp angles to bedding. Deeply sutured stylolites only occur in host rocks that have low contents of platy minerals (Sorby, 1879; Barret, 1964). The nonsutured, undulatory or hummocky stylolites may be related to areas of host rock which contain fine, insoluble, platy particles. Similar features are especially common in limestones and dolostones which have been subjected to tectonic stress, and have been attributed to pressure solution effects (Wanless, 1977). By analogy with the work done by

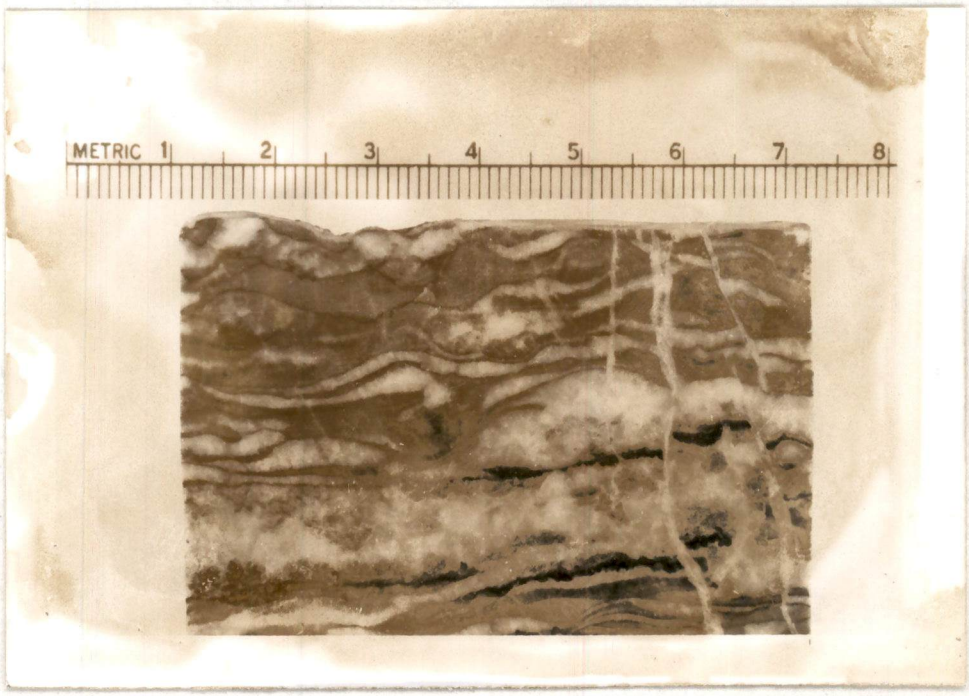


Figure 16. Polished slab of speciman 11-19-80 showing undulatory stylolites.

the above authors, it appears that two periods of stylolite development took place in the Shady Formation. The first episode of stylolite development was related to overburden pressure and formed stylolites which are generally parallel to bedding. The second generation of stylolites crosscut the first generation of stylolites, are oblique to bedding, and probably are related to tectonic events. A similar situation was observed by the author from Shady Formation exposures in Bumpass-Cove (Embreeville), Tennessee.

Tectonic breccia and fractures. Breccias of apparently tectonic origin occur in areas of the mine where thrust faulting is prominent. Sulfide mineralization often accompanies the brecciation, with the breccia fragments consisting of host dolostone cut by numerous fractures filled with coarsely crystalline calcite, dolomite, or silica.

Significant structures in the host rock are sets of subparallel fractures which form diagonal intersections with stylolites (Figure 17). These fractures cut across all the primary and secondary features observed and are interpreted to represent tectonically-controlled fractures related to orogenic movements. Several fractures of this type are mineralized (Figure 17) indicating that at least some of the mineralization may be syn- or post-tectonic.

II. MICROSPARITE

The least extensive of the three lithologies is a microsparite, interpreted to represent a dolostone which has been extensively

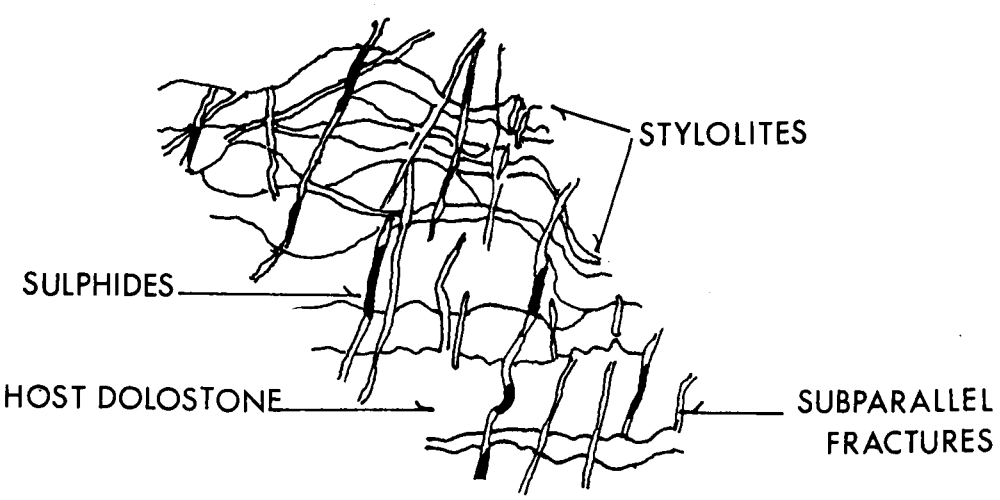


Figure 17. Drawing of mineralized subparallel fractures observed in the 100 Orebody (actual size).

replaced by calcite (dedolomite). This rock type, previously considered to be limestone because of its apparent calcite-rich constitution, is sparsely but erratically distributed in the mining interval. The microsparite exhibits sharp contacts with the host dolostone and relict structures may be traced from one unit to the other across the contacts (Figure 18). The dedolomites examined in this study were collected directly along strike 500 feet S 45° W from the 100 Orebody and from the 8th level drift Ivanhoe (Figure 2, p. 4).

The microsparite is composed of equant, 4 to 20 micron-sized calcite grains with polygonal grain boundaries. The origin of limestone with a fabric of this type is difficult to decipher and is subject to varying interpretations. There are three basic processes which may generate a fabric of this type: (1) coalescive or porphyroid neomorphism (Folk, 1965), (2) mechanically deposited carbonate silt (Folk, 1965), and (3) dedolomitization (Shearman et al., 1965). An origin from mechanically deposited carbonate silt is not applicable in the present case because cross-bedding, laminae, and other primary depositional features are absent. The choice between two remaining possible origins, neomorphism and dedolomitization is a difficult one since the end product of either process may be very similar.

A microspar of neomorphis origin must have clear evidence of originating from micrite (Folk, 1965). Evidence suggested for this include (1) gradational to sharp contacts with micrites, (2) aureoles

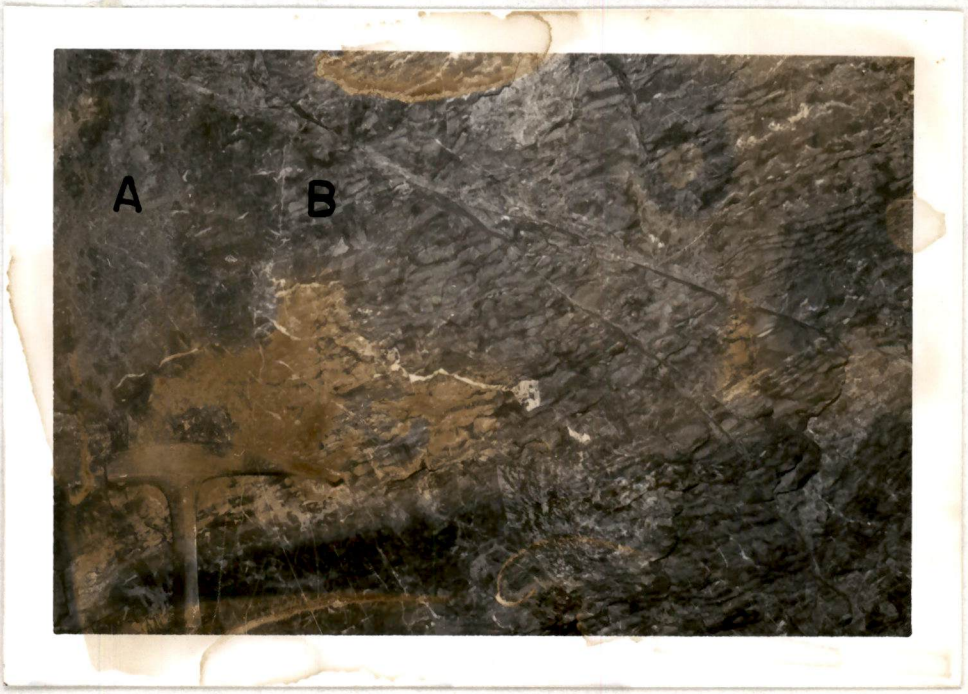


Figure 18. Photograph showing the contact between the microsparite (A) and the host dolostone (B) exposed in a stope face at the Austinville Mine. Exposed part of hammer is approximately 20 cm tall.

of microspar around allochems or grains, and (3) microspar cutting across fecal pellets. These criteria are of little aid in evaluating the origin of the microspar in question because: (1) the contacts of the microspar in the Shady Formation are in sharp contact with xenotopic host dolostone, (2) there is a lack of allochems and there are no microspar aureoles, and (3) the rare, scattered peloids show no evidence of microspar cutting across them. The allochems which do occur in the microspar are restricted to insoluble accumulations along stylolite seams. No fossils with the exception of scattered peloids are found within the microspar away from stylolites. It is difficult to imagine an original micrite with such an abiotic character.

The remaining possibility for the origin of the microspar is dedolomitization. Dedolomitization is the reverse process of dolomitization and is effectively brought about by solutions with high Ca^{++}/Mg^{++} ratios reacting with dolomite to form calcium carbonate (Evamy, 1967). The evidence for dedolomitization is determined by interpreting the texture of the rock as observed in a stained thin section. The utilization of Alizarine red-S stain or some other Ca^{++}-seeking stain enables calcite to be detected and the resulting textures caused by dedolomitization to be observed.

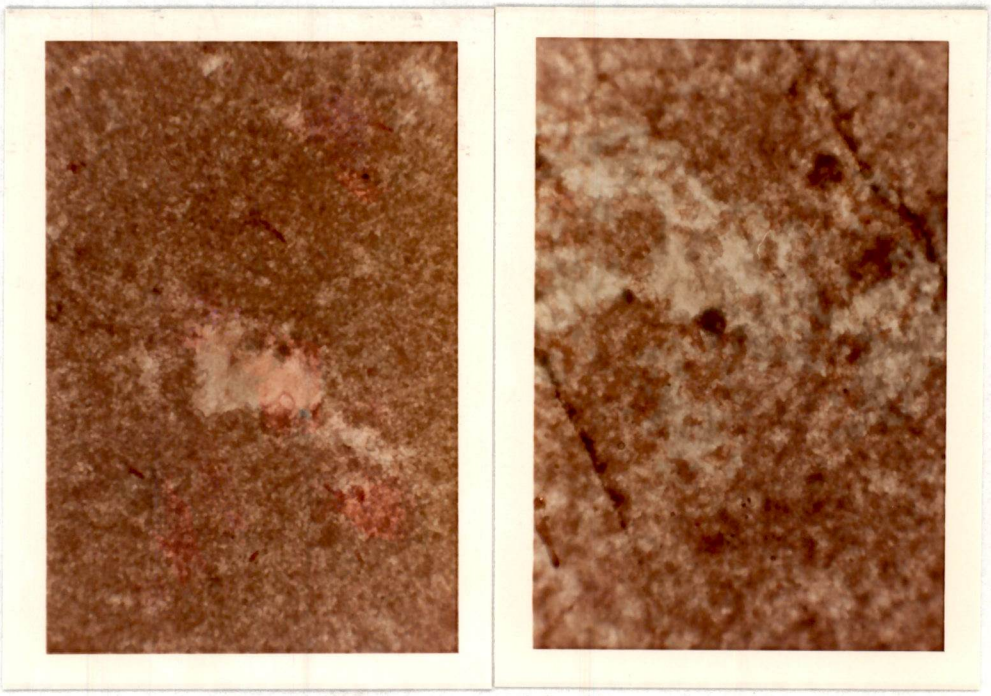
Textures resulting from dedolomitization have been described by Shearman et al. (1961), Evamy (1967), and Chilingar (1956). The textural evidence cited by the above workers for dedolomitization include (1) relics of incompletely replaced dolomite crystals, (2) the

presence of pseudomorphs of calcite after dolomite, and (3) generation of equicrystalline calcite mosaics, and development of a clotted or "grumeleuse" texture.

The textures of the stained sections show a marked resemblance to the equigranular "limestone after dolomite" texture first described by Shearman et al. (1961). Shearman et al. concluded that the final product of dedolomitization would result in a limestone (dedolomite) consisting of equigranular calcite grains. The new calcite mosaic would be equigranular with complex grain boundaries. Their description appears to describe the basic texture of the microsparite observed in the Shady Formation. A similar explanation for the origin of the microsparite in the Shady may also explain the lack of typical limestone features such as allochems or cements.

The "grumeleuse" or clotted texture reported by Evamy (1967) is also present in the microsparites in the dolostone. The "grumeleuse" texture is believed to have resulted from dolomitization and subsequent dedolomitization. The photomicrographs in Figure 19 show the anhedral, equigranular, calcite crystals and the unstained poorly crystalline dolomite which are typical features of dedolomites (Evamy, 1967).

Another feature common to the dedolomites is the prevalence of unaltered subhedral dolomite spar in many of the stylolite seams, especially those stylolite seams which are rich in insoluble residue (Figure 20). It is proposed that the fluids which generated the anhedral, equigranular, calcite crystals from previously-existing host dolostone were unable to penetrate the insoluble material (Figure 20).



(A) (B)

Figure 19. Photomicrographs showing typical features of dedolomites observed under normal light. Bar scale is 1 mm long.

A. Typical microsparite (red) stained with Alizarine red S and unaffected white, poorly crystalline dolomite (clotted texture, Sample Number H-10).
B. Calcite microspar (red) replacing dolomite (white) yielding a grumeleuse texture (Sample Number H-30).



Figure 20. Photomicrograph showing unaltered (white) subhedral dolomite and grey microsparite observed under normal light. Bar scale is 1 mm long (Sample Number H-5).

The dedolomite (microspar) contains many of the lithologic constituents, primary and secondary features discussed in the sections concerning the host dolostone. The major lithologic constituents are anhedral calcite and hypidiotopic dolomite spar. Accessory lithologic constituents include pellets, invertebrate fossils (usually fragments), evaporite pseudomorphs, carbonaceous matter, and chert. The explanation and possible significance of these features have been explained in the sections describing the host rock.

In summary, the absence of limestone fabrics, the presence of equigranular calcite mosaics, the development of "grumeleuse" texture strongly suggest that the microsparite in the Shady samples, examined by the author, formed by dedolomitization. It may be noted that similar textures observed in the Cambro-Ordovician Knox Group carbonates of Tennessee have been interpreted to have originated by dedolomitization (Alberstadt and Reesman, 1978; Alberstadt, personal communication, 1978).

Experiments on dedolomitization by De Groot (1967) and Yanat'eva (1955) have demonstrated that dedolomitization is a near-surface phenomenon enhanced by the presence of calcium sulfate. Early workers such as von Morlot (1948) suggested that dissolution of evaporites in the presence of dolomite could result in a dedolomite. A similar explanation for the origin of the microsparite observed in the Shady Formation appears likely, especially when other evidence of hypersaline (evaporitic) conditions are present in the host rock lithologies.

III. WHITE GANGUE DOLOMITE ("MET")

This material is composed of coarse, hypidiotopic, baroque, dolomite crystals and is often interlaminated with the host dolostone (Figure 21A, B). This lithology had previously been described as "met" by Brown and Weinberg (1968). Statistical studies done by the New Jersey Zinc Company have shown that the distribution of ore is related to the proportion of "met" present in a particular stratigraphic horizon (Brown and Weinberg, 1968).

The term baroque was first used by Folk and Assereto (1974) to describe white dolomite crystals which exhibit undulatory extinction (Figure 21B). They believe that baroque dolomite crystals are formed in the presence of sulfates or sulfides. In the Austinville-Ivanhoe ore horizons studied, the presence of baroque dolomite is restricted to the white gangue dolomite which occurs predominantly as cavity-fillings in the host dolostone and in association with sulfide ores. A small amount of baroque white gangue dolomite also occurs as geopetal-fill in the host dolostone (see p. 30). It appears that the baroque dolomite of the first category formed in the presence of sulfides whereas that of the second category formed in the presence of sulfates.

The gangue dolomite has been previously described as originating from metamorphism or recrystallization of the wall rock. In hand specimens, the gangue dolomite appears as coarsely crystalline white dolomite with no discernible variation between specimens (Figure 21A). Examination of approximately 180 thin sections utilizing CL has shown



(A)



(B)

Figure 21. Typical features of white gangue dolomite ("met").

A. Photograph of high "met" area exposed in a stope face at the Austinville mine. Rock hammer is approximately 35 cm tall.

B. Photomicrograph of baroque dolomite showing characteristic bent crystals when observed under normal light. Bar scale is 1 mm long (Sample Number 10).

some of the white gangue dolomites to be microbrecciated on a crystal-size scale (Figure 22). The relic breccia blocks luminesce a distinctly different color than the later stage dolomite which fills the open spaces between the microbreccia blocks. The microbreccia blocks appear to be crystal fragments of an earlier carbonate rock as evidenced by the minor CL zoning of the breccia clasts. The microbreccia exhibits both rubble and crackle breccia structures under the luminoscope. It is interesting to note that neither the relic microbreccia blocks nor the later open-space fill dolomites exhibit any appreciable CL zoning. This is in contrast to cathodoluminescence work done by Ebers (1976) on Mississippi Valley-type deposits in East Tennessee. The micro-stratigraphy and subsequent relationships to ore deposition Ebers established in East Tennessee, may not be applicable to the Austinville-Ivanhoe area.

Staining techniques described by Dickson (1966) and Choquette (1971) to determine the genesis of carbonates proved of little value in this study. Calcite is so rare in this lithology that Alizarine red-S stain has no practical application in studying the white gangue dolomite. Lack of staining by potassium ferricyanide suggests that the dolomite does not contain significant Fe^{++}.

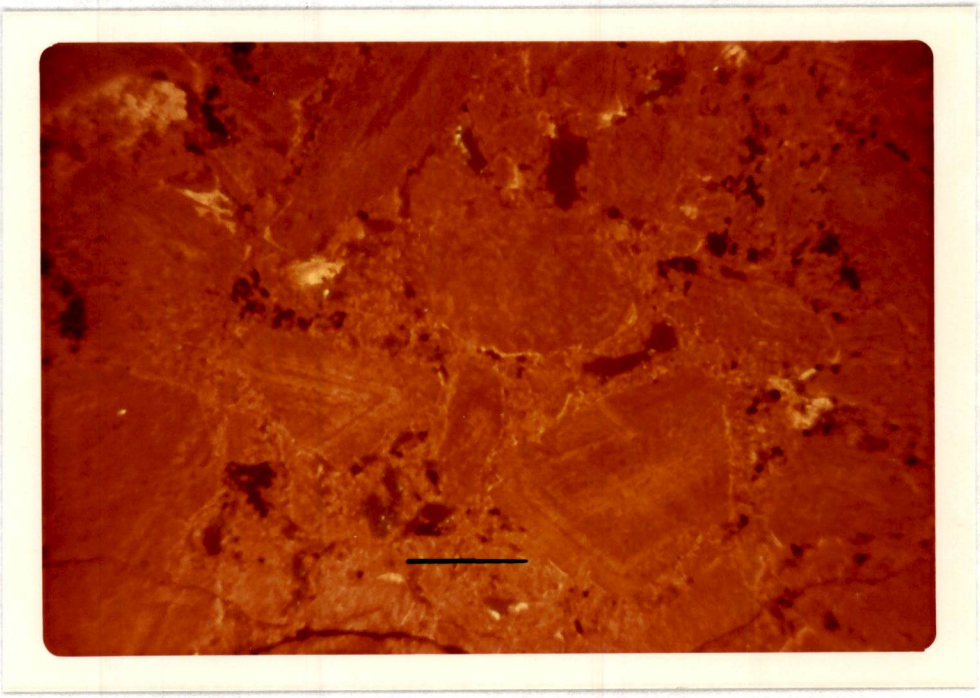


Figure 22. Photomicrograph showing zoned rubble breccia and subsequent breccia-fill dolomite each luminescing a different color. The black or nonluminescing mineral dispersed throughout the photomicrograph is pyrite. Bar scale is 1 mm long (Sample Number 5-02-16 A).

CHAPTER IV

ORE PETROLOGY

I. NATURE OF MINERALIZATION

The bulk of the mineralization in the Austinville-Ivanhoe lead-zinc district occurs in association with the white gangue dolomite. The main ore minerals of the district comprise sphalerite, pyrite, and galena. These primary sulfide minerals were epigenetically emplaced into the host Shady Formation, mostly as typical open-space fills but some also as replacements. At the Austinville-Ivanhoe mines, much of the ore is in breccias, the sulfides occurring as open-space fill between the brecciated fragments of the Shady Formation and associated with the white gangue dolomite. In occurrences of this type, the ore minerals are generally crustiform with colloform textures also being common. Well-formed crystals of sulfides, dolomite, quartz and calcite often coat box-work vugs and fractures throughout the orebodies and are most common in the north-limb facies (Ivanhoe). In particular a lemon-yellow variety of euhedral sphalerite found in the vugs of the Ivanhoe mine is a much sought collector's item. Replacement ores, which constitute a very minor proportion of the sulfide mineralization, occur as disseminations and stringers in organic-rich layers of the host dolostone. Although quantitatively unimportant, replacement of white gangue dolomite by sphalerite has also been observed in all the ore horizons studied.

Crustiform Ore

Crustiform ore may be defined as sulfide mineralization which occurs as layers lining previously open cavities (Figure 23). Figure 15 (p. 32) illustrates a mineralized breccia similar to the rubble breccias of the East Tennessee zinc district. The sulfides, pyrite and galena, can clearly be seen to be occurring as fracture fillings and encrusting breccia fragments of host dolostone. The specimens corresponding to Figure 15 (p. 32) and Figure 23 contain white gangue dolomite which consists of baroque dolomite in direct association with the ore.

Colloform Ore

Colloform ore is used here as a textural term to describe sulfide aggregates with rounded external form and concentric banding, without any suggestion of a colloidal origin of these aggregates. The colloform ore "floats" in the baroque, white gangue dolomite as discrete spherical forms when viewed in three dimensions. In the specimens studied, only sphalerite and pyrite show well-developed colloform textures (Figure 24A, B). The colloform sphalerite consists of concentric bands of honey-yellow sphalerite which in many cases alternate with bands of dark brown sphalerite. The banding of honey-yellow sphalerite may be attributed to difference in grain size or crystallinity; compositional differences between the bands, if any, have not been ascertained. Often, an irregular mass of fine-grained pyrite (framboidal?) forms the core of the collorform sphalerite, in other cases the core is composed of fine-grained sphalerite. The colloform pyrite is devoid of sphalerite.

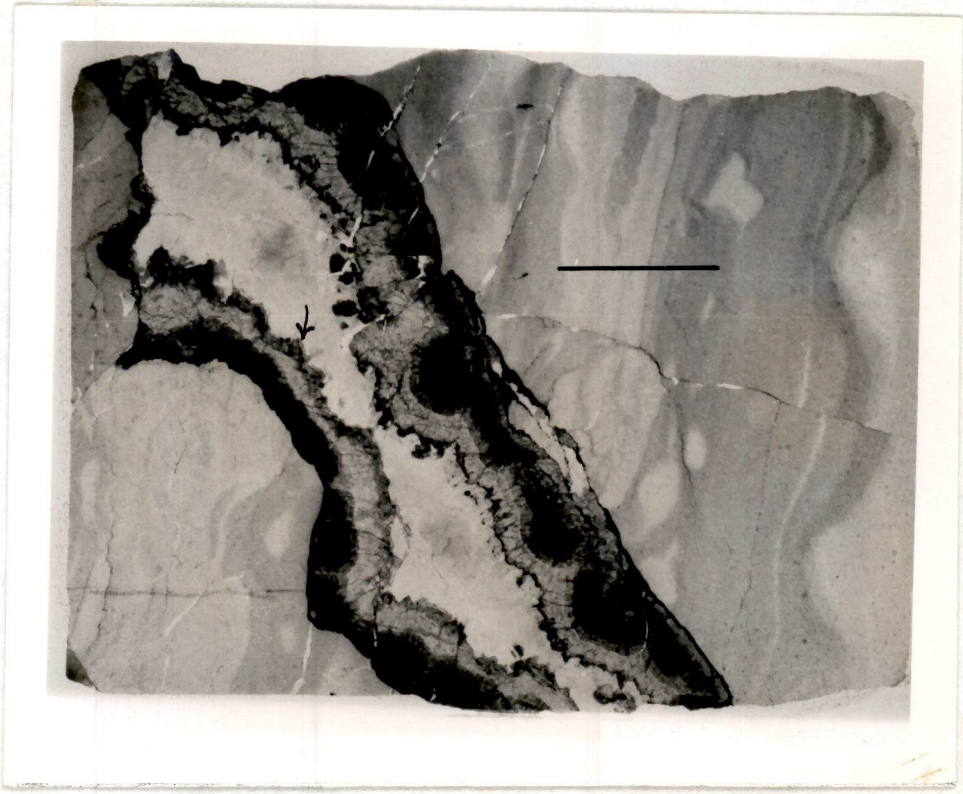
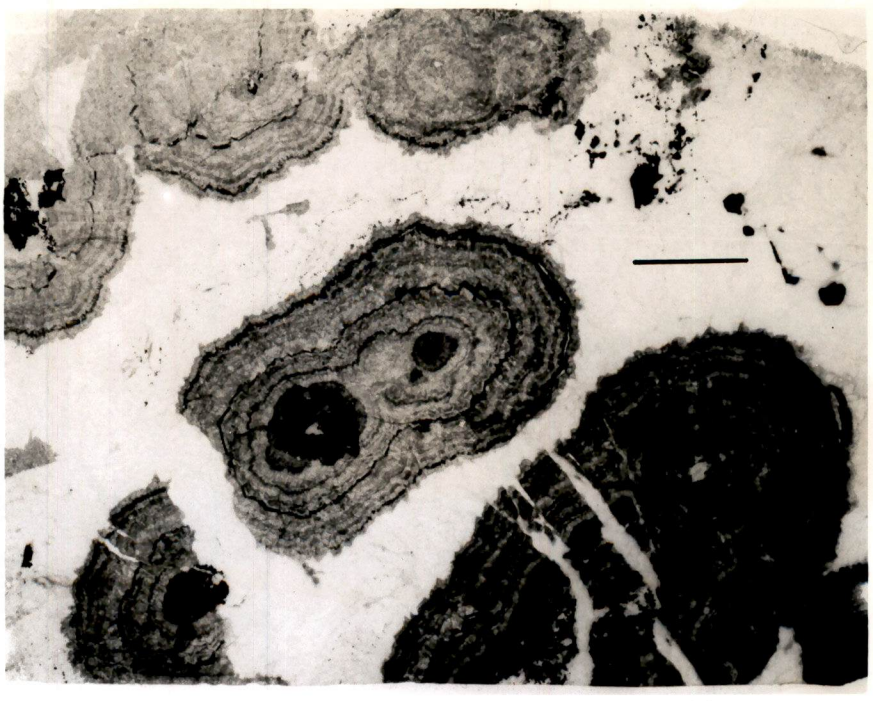
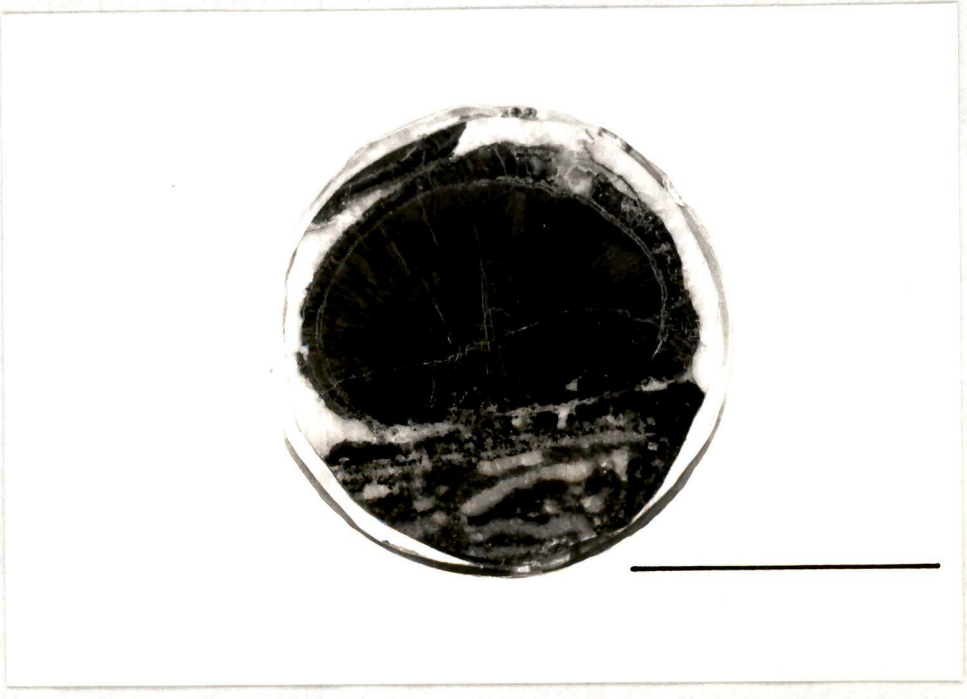


Figure 23. Photomicrograph showing crustiform sphalerite ore (arrow) observed under normal light. Bar scale is 10 mm long (Sample Number 11-48-89).



(A)



(B)

Figure 24. Typical colloform ore textures.

A. Photomicrograph of brecciated colloform sphalerite when observed under normal light. Bar scale is 10 mm long (Sample Number 5-02-09).
B. Megascopic view of a polished section showing colloform pyrite. Bar scale is 10 mm long (Sample Number 7-04-05).

Typically, it contains a core of globular (framboidal) pyrite and a shell of euhedral pyrite growing outward from the core. The core may show various stages of disaggregation and recrystallization, and in some cases concentric banding. Both the colloform sphalerite and colloform pyrite show interruption of concentric banding due to juxtaposition of adjacent colloform rosettes and occasional brecciation of the rosettes.

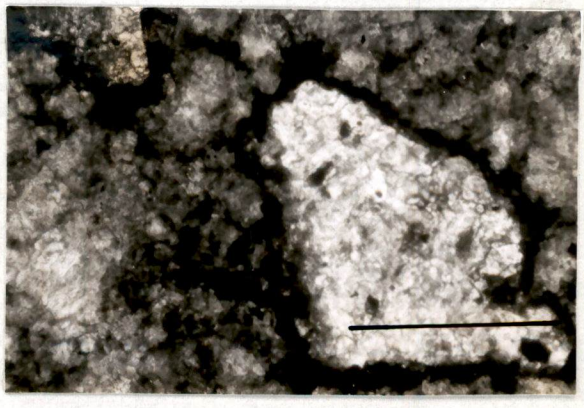
Replacement Ore

As mentioned earlier, much of the ore mineralization is restricted to the white gangue dolomite. Minor amounts of disseminated sphalerite, galena, and pyrite, interpreted to represent replacement of organic-rich, silty, host dolostone, is also present. Of these, the sphalerite shows the best evidence for a replacement origin. Some dolomite crystals are only partially replaced by sphalerite at the edges; in other cases, almost complete replacement is suggested by preservation of the outline and islands of the original dolomite crystal (Figure 25B). In the Ribbon Member, stringers of sphalerite and galena have replaced portions of the host dolostone without destroying the ribbony structure (Figure 25A). The minor amount of sulfide mineralization in the host dolostone may be attributed to replacement in locally reducing environments.

All of the ore horizons studied exhibit replacement textures between sphalerite and the white gangue dolomite. Figure 26 illustrates one of these textures which is interpreted to be indicative of sphalerite replacing gangue dolomite along grain boundaries. This texture grades into several types of replacement textures, one of which is illustrated in Figure 27.



(A)



(B)

Figure 25. Typical replacement textures.

A. Photomicrograph of replacement sphalerite in organic-rich layers of host dolostone when observed under normal light. Bar scale is 10 mm long (Sample Number 39).
B. Photomicrograph of sphalerite replacing dolomite while retaining the external morphology of the dolomite (normal light). Bar scale is 1 mm long (Sample Number 39).

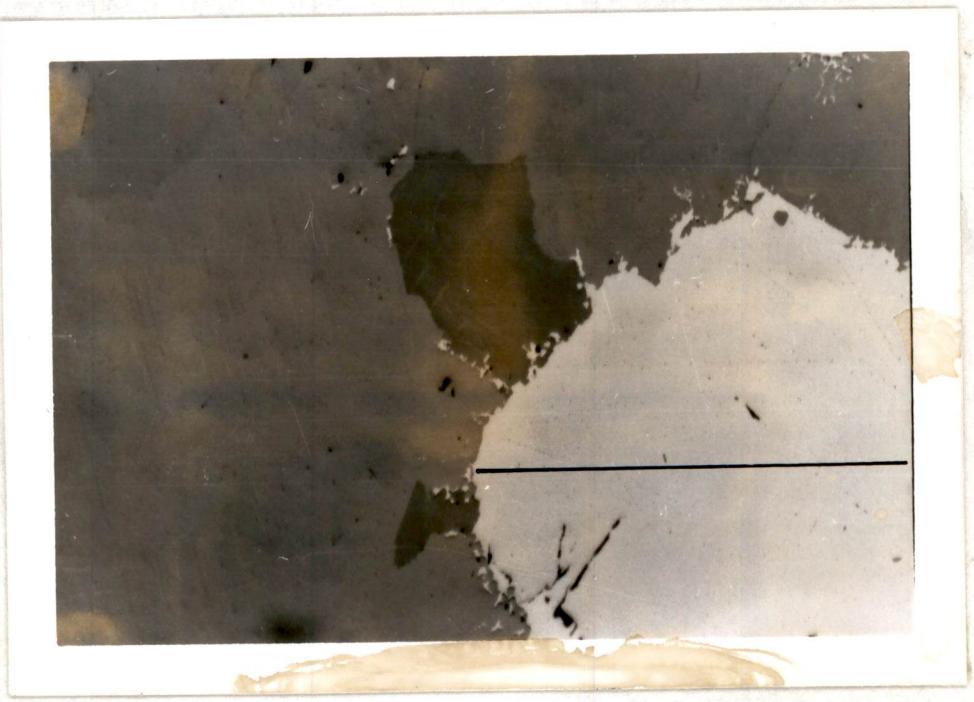


Figure 26. Photomicrograph of sphalerite (light grey) replacing along grain boundaries of gangue dolomite (dark greys, reflected light). Bar scale is 1 mm long (Sample Number 4-03-46).

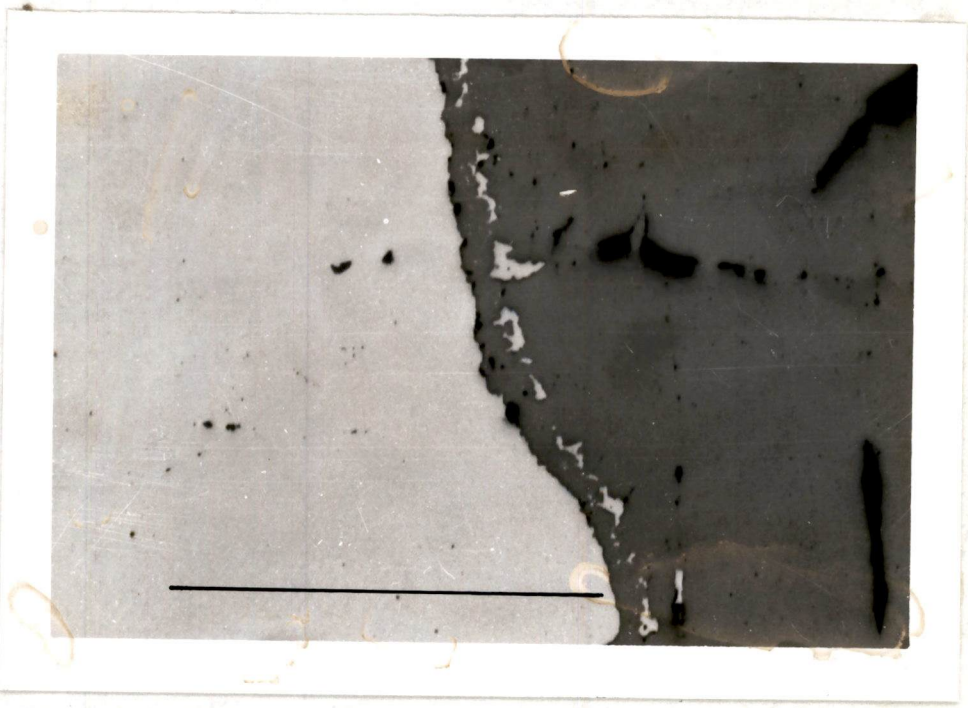


Figure 27. Photomicrograph showing typical replacement texture between sphalerite (light grey) and gangue dolomite (dark grey, reflected light). Bar scale is 1 mm long (Sample Number 5-03-30).

Replacement of sulfides by sulfides is very limited, but does occur between sphalerite and galena. This replacement is shown by veinlets of galena which cut across the sphalerite leaving remnant islands of sphalerite enclosed within the galena. The replacement of sphalerite by galena probably represents the very last phase of the mineralization event.

II. PARAGENESIS

Paragenesis has been discussed briefly in two previous papers devoted to ore zoning (Brown, 1935) and the general geology of the Austinville-Ivanhoe district (Brown and Weinberg, 1968). Based on a study of megascopic and microscopic textural features, Brown and Weinberg (1968) suggested the following general sequence (from earliest to latest) of sulfide deposition:

(1) pyrite

(2) medium to dark brown sphalerite

(3) galena

(4) light yellow sphalerite

Twenty-four polished sections, 90 thin sections and 50 polished slabs from 90 hand specimens were examined in the course of the present study. The crustiform and colloform nature of much of the mineralization and extreme paucity of cross-cutting relationship among the sulfides indicate that the ores were deposited sequentially (pyrite $\longrightarrow$ sphalerite $\longrightarrow$ galena) in a single period of mineralization. Erratically distributed throughout the mining interval is a lemon-yellow variety of

euhedral sphalerite filling box-work vugs. The vugs which contained these crystals lack other sulfides. Apparently, the sphalerite crystals grew in available open spaces from the residual ore fluids. In no instance were textural relationships observed which would suggest multiple periods of mineralization. It appears that the colloform, crustiform, and disseminated ores formed during one main period of mineralization. Thus, the field and microscopic observations in the present study support the paragenetic sequence suggested by Brown and Weinberg (1968). As noted earlier, the white gangue dolomite is intimately associated with the ore. Megascopic mineral zoning in mineralized fractures and box-work vugs show that the white gangue dolomite formed both before and after the main period of sulfide mineralization. The post-ore white gangue dolomite exhibits micro-brecciation (Figure 23, p. 50), suggesting that fluids capable of dissolution of dolomite were active after the main period of sulfide mineralization.

III. DISCUSSION

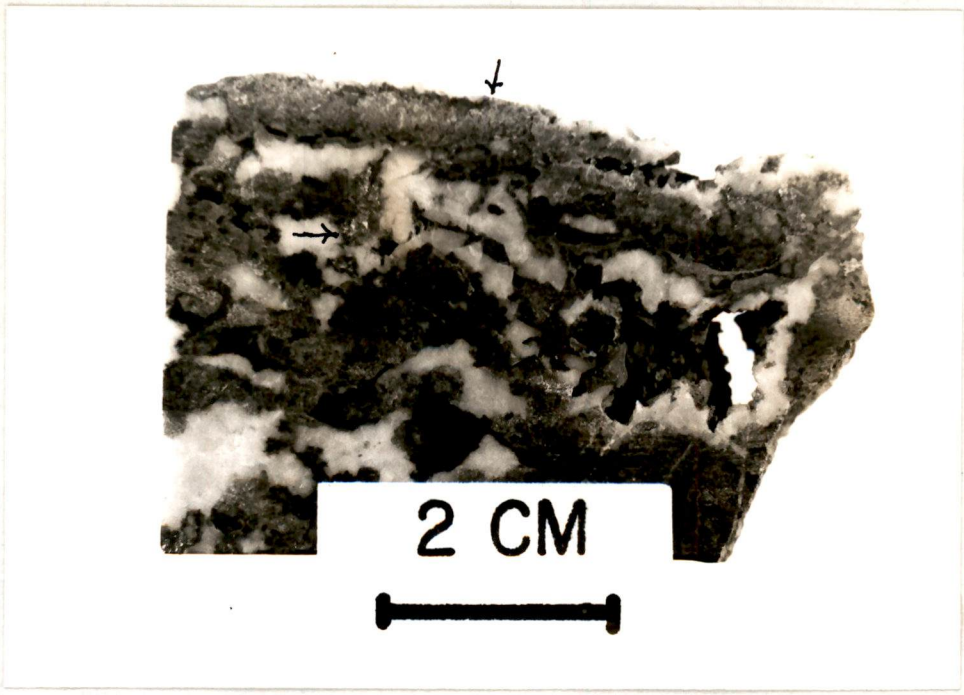
There is not much textural evidence for stress-induced metamorphism of the sulfides. The colloform pyrite and sphalerite ores, some of which are nearly perfect spheres, show no evidence of plastic or cataclastic deformation. Pyrite however, frequently occurs as aggregates showing foam texture (commonly indicative of annealing) and fracturing.

The presence of subparallel tectonic fractures gives some support that at least some of the mineralization was syn- or post-orogenic

movement. Studies of the Zn/Pb, Fe/Pb, and Fe/Zn ratios have been interpreted by Brown (1935) to indicate that the ore fluids migrated along faults and mineralized favorable horizons.

The question of the age of mineralization in the Austinville-Ivanhoe district is debatable. The argument for post-orogenic mineralization is weak, however, the available data for mineralization when the beds were horizontal is even weaker. The author believes that at least some of the ore is post-orogeny on the basis of mineralized tectonic fractures and the probability that some of the brecciation may also be tectonic.

The favorable horizons apparently were sites of mineralization because of their abundant open spaces (Figure 28A, B). The specimens illustrated (A) mineralized and (B) unmineralized, demonstrate that the ore and white gangue dolomite were open-space fillings of previously vuggy host dolostone. Specimens of this type have open vugs which show bedding in both outcrop and hand specimens (Figure 29). The vugs cut across laminations and are obviously secondary in origin. Megascopic and microscopic observations such as the paucity of fossils, length-slow chalcedony spherules, pseudomorphs of dolomite after sulfate, the fine-grained nature of the host rock, dolomitized anhydrite nodules with geopetal structures, and baroque dolomites (all discussed in previous sections) indicate that evaporites were probably present in the host rock prior to mineralization. Selective dissolution of the evaporites when the host rock was semilithified could have created the open-spaces and subsequent brecciation. It appears that dolomite dissolution may also have contributed to creating porosity in the



(A)



(B)

Figure 28. Typical secondary porosity textures.

A. Slab showing sulfide mineralization and white gangue dolomite filling previous vugs (Sample Number 11-10-17).
B. Slab showing unmineralized open vugs (Sample Number 11-10-09).



Figure 29. Photograph showing the bedded vugs exposed in a stope face at the Austinville Mine (Ribbon Member). Pencil is approximately 14 cm long.

ore horizons. Analytical studies by Roedder (1977b) on fluid inclusions of Mississippi Valley-type deposits indicate that the ores formed from dense, hypersaline brines with temperatures of formation from 75° to 200°C.

The petrologic evidence suggests that the host dolostone for the Austinville-Ivanhoe ores were deposited in a hypersaline environment. Periodic influxes of less saline waters caused the dissolution of evaporite minerals and subsequent geopetal structures filled with baroque dolomite. A second period of baroque dolomite formation occurred with the primary ore mineralization and is probably related to sulfide ions being present in the ore fluids. It is suggested here, that the ore fluids migrated along permeability zones and mineralized stratigraphic horizons which were made favorable by dissolution of previous evaporites and/or dolomite.

CHAPTER V

DIAGENESIS

I. INTRODUCTION

Diagenesis as used herein refers to all changes that occurred to the sediment after deposition, excluding metamorphism. For the purposes of this report, mineralization shall be included in the discussion of diagenesis under the section referring to anadiagenesis.

Three stages of diagenesis are recognized: syndiagenesis, anadiagenesis, and epidiagenesis. Syndiagenesis (Bissell, 1959) is defined as near-surface phenomenoma which take effect immediately after sedimentation. Syndiagenesis is characterized by the presence of large amounts of interstitial or connate waters which are expelled slowly during burial. The process of syndiagenesis has been divided into two major divisions (Dapples, 1962). These divisions are called: (1) initial stage, controlled by the chemistry of superjacent water and (2) early burial stage controlled by connate water and the modifying effect of organisms.

Anadiagenesis (Fairbridge, 1967) is a term which includes the compaction and maturation phases of diagenesis. It may include tectonic effects and may be effective to depths of approximately 10,000 meters (Fairbridge, 1962). It is characterized by expulsion of connate waters and hypogene mineralizing fluids either of a magmatic or nonmagmatic origin.

Epidiagenesis is a term proposed by Fairbridge (1967) to describe the emergent or final stage of diagenesis. It occurs during and after orogeny but before erosion. It is characterized by oxidation and the effects of meteoric ground water.

Syndiagenetic processes include bioturbation, penecontemporaneous dolomitization, pseudomorphous replacement of sulfates by dolomite, dedolomitization, and solution collapse brecciation. Anadiagenesis processes include tectonic fracturing and brecciation, stylolitic solution, emplacement of "met," and ore mineralization. Epidiagenetic processes include solution and oxidation with subsequent precipitation of oxide-enriched ore minerals. The order of occurrence of these processes at Austinville is given in Table 1.

II. SYNDIAGENETIC PROCESSES

Bioturbation

The host dolostone and microsparite (dedolomite) are rarely burrowed. Where present, the burrows are indistinct and filled with blocky dolomite spar. Some of the burrows are filled with internal sediments and were recognized by changes in crystallinity or color.

The organism(s) which created these features is unknown since there is no trace of the organism(s) in or near the burrows. Apparently, the burrowing organisms were soft bodied and are not preserved.

Penecontemporaneous Dolomitization

The fine-grained nature of the host dolostone is considered as an indicator or hypersaline conditions (Fold and Siedlecka, 1974). The

TABLE I

SEQUENCE OF DIAGENESIS IN THE SHADY DOLOMITE

Diagenetic Event	Syndiagenetic	Anadiagenetic	Epidiagenetic
Bioturbation	——— ---		
Penecontemporaneous Dolomitization (baroque dolomite in geopetal fabric)	—— ---		
Pseudomorphs	——— ----		
Dedolomitization	——?——		
Solution Collapse Brecciation	——— ---		
Tectonic Fracturing Brecciation		——?——	
Stylolitic Solution	———	———	
Ore Mineralization		——?——	
Solution and Oxidation			———
Oxide-enriched Ores			———
White Gangue Dolomite (Baroque)		———	

presence of length-slow chalcedony, pseudomorphs of dolomite after sulfate, dolomitized anhydrite nodules with geopetal structure, baroque dolomite, paucity of fossils, and the euryhaline nature of the fossils present in the host rock support this conclusion. It appears from the petrologic evidence cited above that the mineralized beds of the Shady Formation formed in a sabkha or tidal flat environment. A tidal flat depositional environment has previously been suggested for many of the ribboned dolomites in the Ribbon Member of the Shady Formation (Byrd et al., 1973). Byrd suggested that light-grey laminae observed in the Ribbon Member are representative of a marine influx of carbonate mud across a tidal flat. The pervasive dolomitization of the original lime mud created the decimicron-sized, xenotopic, penecontemporaneous dolostone without completely destroying primary sedimentary structures and lithologic constituents such as laminations, pellets, and geopetal structures. Possibly, a future study concentrating on bore holes in the vicinity of the mine area would be able to place the various features of the host dolostone into a supratidal, intertidal, and subtidal environmental framework.

Pseudomorphs of Dolomite after Sulfate Minerals

Replacement of dolomite after surface minerals are common in the host dolostone and microsparite lithologies of the Shady Formation. The prismatic monoclinic-appearing crystals (probably gypsum) grew in the interstitial voids of the then unlithified host lithologies. The sulfates were then dissolved and selectively replaced during penecontemporaceous dolomitization. The dissolution of evaporites

in the host lithologies could have resulted in the vuggy, often brecciated structures in the host dolostone.

Dedolomitization

The dedolomite (or microspar) probably resulted from a process very similar to the one first proposed by Morlot (1848). The dissolution of evaporites and the subsequent release of Mg^{++} and Ca^{++} combined with the presence of dolostone resulted in formation of a calcitized dolostone (dedolomite). The dedolomitization took place after penecontemporaneous dolomitization and before deep burial. The transformation of an original limestone mud to dolostone and subsequent replacement by calcite could explain the lack of typical limestone fabric in the microspar (dedolomite).

Brecciation

Solution collapse brecciation is common to the host dolostone and microspar of the Shady Formation. Small slump breccias and breccia bodies limited to specific beds appear to be solution related. Some of the larger mineralized breccia bodies are very similar to the mineralized breccia bodies of the East Tennessee zinc district. It is geologically reasonable to expect that some of these breccias are related to dissolution of evaporite minerals and subsequent collapse. Many megascopic features of the host dolostone such as discontinuous folds, bedded vugs, and localized slump features are difficult to explain without dissolution of evaporites. The absence of a post-Shady unconformity precludes the application of the paleo-karst hypothesis of breccia formation to the mineralized breccias of the Shady Formation.

The abiotic nature of the host dolostone also makes the fore-reef hypothesis of breccia formation in the Shady Formation (Hoagland, 1971) as highly unlikely. Many of the textural features observed both microscopically and megascopically are subject to various interpretations. It is however, difficult to visualize a reasonable geologic set of circumstances which can explain the gross features of the lithologies of the mineralized zone without the role of evaporite dissolution.

III. ANADIAGENESIS

Stylolitic Solution

Stylolites are the most common feature observed in the host dolostone. The dark, argillaceous laminae of the host dolostone contain numerous undulatory stylolites which are parallel to bedding. Stylolites of this type are thought to have been formed by overburden-induced pressure-solution. Other stylolites, which are steeply inclined to bedding, are believed to have been generated by pressure solution effects related to tectonism. The difference in external form of stylolites (sutured-nonsutured) are probably related to the platy mineral content of the original sediments as proposed by Sorby (1879), Barret (1964), Roehl (1967), Wanless (1977).

Tectonic Fracturing and Brecciation

Numerous subparallel fractures oblique to bedding cut across the host dolostone and microsparite. These features are usually filled with coarsely crystalline, subhedral dolomite. Calcite spar filling the fractures is rare. Frequently, the fractures form diagonal

intersections with stylolites that are parallel to bedding. The predominantly dolomite-filled veins probably originated during tectonic activity and subsequent mineralization.

Emplacement of White Gangue Dolomite and Ore Mineralization

The main ore minerals of the Austinville-Ivanhoe lead zinc district comprise sphalerite, pyrite, and galena. These primary ore minerals were apparently emplaced during or after orogenic movements. The ore textures and cross-cutting relationships indicate that the ore predates the majority of the coarsely crystalline, hypidiotopic, baroque dolomite (met). The white gangue dolomite which is always associated with the ore is often microbrecciated and exhibits dissolution features with CL.

The published literature concerning the mode of formation of baroque dolomite crystals is limited. There appears to be general agreement that the formation of baroque dolomite is indicative of evaporitic conditions (Folk and Asserto, 1975; Dunsmore, 1973). It is apparent from the bent saddle-shaped crystals that the baroque dolomite grew in an open void. Dunsmore (1973) suggested that the "zebra-rock" fabric of the Pine Point orebodies resulted from dissolution of preexisting carbonate rock by corrosive fluids generated by a reaction between petroleum and sulfate ions. The fabric zebra-rock described by him (Dunsmore, 1973, Figure 3, p. 9) is similar to the fabric of the Ribbon Member (Figure 10, p. 23). Dunsmore (1973) contends that zebra-rock (dolostones which are banded parallel to bedding with baroque dolomite) originated by corrosive fluids which replaced carbonate along planes of preferred

permeability parallel to bedding. His conclusions are based primarily on analogies drawn from oil fields and evaporite complexes which are presently being commercially exploited in the same formation as the Pine Point orebodies.

The petrologic evidence presented elsewhere in this report suggests that evaporites were present in the Shady Formation. As yet there is no evidence suggesting the former presence of hydrocarbons. Analysis of fluid inclusions from the Austinville-Ivanhoe district may yield evidence of hydrocarbons which would support Dunsmore's corrosive fluids model.

From the available petrologic evidence observed in the host lithologies of the Shady Formation, it appears that baroque dolomite is related to dissolution of evaporite minerals. The development of the crystals is probably related to sulfate ions and/or their reduction to sulfide. The baroque crystals are apparently limited to open-space growth in which the voids may have been created by corrosive fluids. The precise origins and nature of these fluids is speculative.

IV. EPIDIAGENESIS

Solution and Oxide Enrichment

Surface exposure and solution along joints and permeable zones by meteoric water has caused pinnacle weathering attaining a maximum relief of about 60 meters. The oxidized zone affects rocks 700 feet below the present day surface. The known oxidized zinc minerals from the

Austinville-Ivanhoe district include smithsonite, hydrozincite, and hemimorphite. Oxidized lead minerals include anglesite, cerussite, and very rare massicot and minium (Brown and Weinberg, 1968). The secondary ore minerals are concentrated in crests and troughs between the pinnacles. Their distribution influenced the early economic development of the district.

CHAPTER VI

SUMMARY AND RECOMMENDATIONS

I. SUMMARY

The conclusions summarized below are based on 181 thin sections, 24 polished sections, and numerous polished slabs. The samples were obtained from the Malden, Mottled Marker, 100 and 40 Orebody horizons (Figure 2, p. 4). The majority of the specimens were from orebodies in the Ribbon Member (Figure 2, p. 4). These conclusions are:

1. The Lower Cambrian Shady Dolomite ore horizons are subdivided into 3 lithologies based on lithologic constituents and relict sedimentary structures. Lithologies recognized are: host dolostone (decimicron-sized, xenotopic), microsparite (dedolomite), and white gangue dolomite ("met").

2. Lithologic and mineralogic constituents recognized are pellets, invertebrate fossils, dolomite pseudomorphs after sulfate minerals, carbonaceous matter, silt-sized euhedral microcline, sand and silt-sized quartz, clays, chalcedony concretions, calcite spar, dolomite spar, and baroque dolomite.

3. Relict sedimentary features are divided into primary and secondary types. Primary features include: burrows, laminations and solution breccias. Secondary features include: stylolites, tectonic breccias and fracture patterns.

4. Processes of diagenesis are: syndiagenetic, which include bioturbation, penecontemporaneous dolomitization, pseudomorphs of dolomite after sulfate, dedolomitization, solution collapse brecciation, and neomorphism; anadiagenetic, which includes tectonic fracturing and brecciation, stylolitic solution, emplacement of ore minerals, white gangue dolomite; and epigenetic, which includes solution and oxidation with subsequent precipitation of oxide-enriched ore minerals.

5. The host dolostone which comprises the ore horizons of the Shady Formation were deposited in a hypersaline environment.

6. The microbrecciated white gangue dolomite (met) and solution margins between sphalerite and the gangue (met) indicate that fluids capable of dolomite dissolution were active during and after the main period of mineralization.

7. The absence of appreciable CL zoning in the white gangue dolomite makes CL work an uncertain exploration tool in the district.

8. The microbrecciation and crystal habit of the white gangue dolomite suggest that the crystals grew in an open-space.

9. The primary ore minerals appear to have been emplaced during one main period of mineralization.

10. Open-space filling is the most significant type of sulfide mineralization; sulfide mineralization by replacement of preexisting carbonate minerals is quantitatively minor.

11. Some of the sulfide ores were apprently emplaced after or during orogenic movements.

12. In the samples examined by the author, the microspar appears to be calcitized dolostone (dedolomite) rather than a primary limestone.

13. The invertebrate fossil Salterella conulata appears to be a possible guide fossil to the Lower Cambrian as first suggested by Yochelson (1971).

II. RECOMMENDATIONS

This study has indicated that the ore horizons of the Shady Formation are related to permeability zones caused by solution collapse brecciation possibly related to sulfate dissolution. Another possibility exists that some of the brecciation in highly faulted areas may be tectonic in origin. The direct relationship of the white gangue dolomite and primary ore minerals is well established. The potential of petrologic studies as a possible exploration guide warrants further research in mineralized areas of the Shady Formation. This research should include:

1. Petrologic study of white gangue dolomite areas away from mineralized zones to determine if the met in the nonmineralized zones is baroque. If substantial differences such as an absence of baroque dolomites are observed, a simple exploration guide would be gained.

2. Petrologic study areas of the Shady Formation which are outside of the Austinville-Ivanhoe district and are similarly mineralized (i.e., Embreeville, Tennessee).

3. Petrologic study of core samples to determine environments of deposition of the major ore horizons in the Shady Formation. This information could then be applied to the "facies" change to better understand its significance.

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APPENDIX

SAMPLE LOCATIONS AND SAMPLING PROCEDURES

One hundred and eighty-one hand specimens were collected from 10 ore bodies within the Austinville-Ivanhoe mining interval (Figure 2, p. 4). The spatial distribution of the various ore bodies is best seen in reference to Figure 30. The sampling localities and the sampling patterns used were those suggested to the author by the Geology Staff at Austinville (New Jersey Zinc Company). All stopes and drifts were sampled on 10 foot intervals with the exception of samples provided by Michael L. Ebers (Mine Geologist, Austinville). The following is a list of the 181 samples collected and their locations. Each sample location in the stopes was marked with red paint.

The samples collected in the different stopes were of two types: (1) representative ore samples containing sphalerite, pyrite, galena (when present), met (at least 25 percent) and ribbony dolostone; (2) host rocks which did not contain any sulfides but contained at least 50 percent met. Whenever possible, the specimens were at least 2" × 2" × 2", so that polished slabs and thin sections could be made.

The New Jersey Zinc Company provided the author with a 30 scale copy (1 inch - 30 feet) of the 11-21-90 crosscut, 11-03-42 stope, and the 8th level drift Austinville, showing the sample localities. Samples provided by Michael L. Ebers are located by their gross relationships to specific ore horizons.

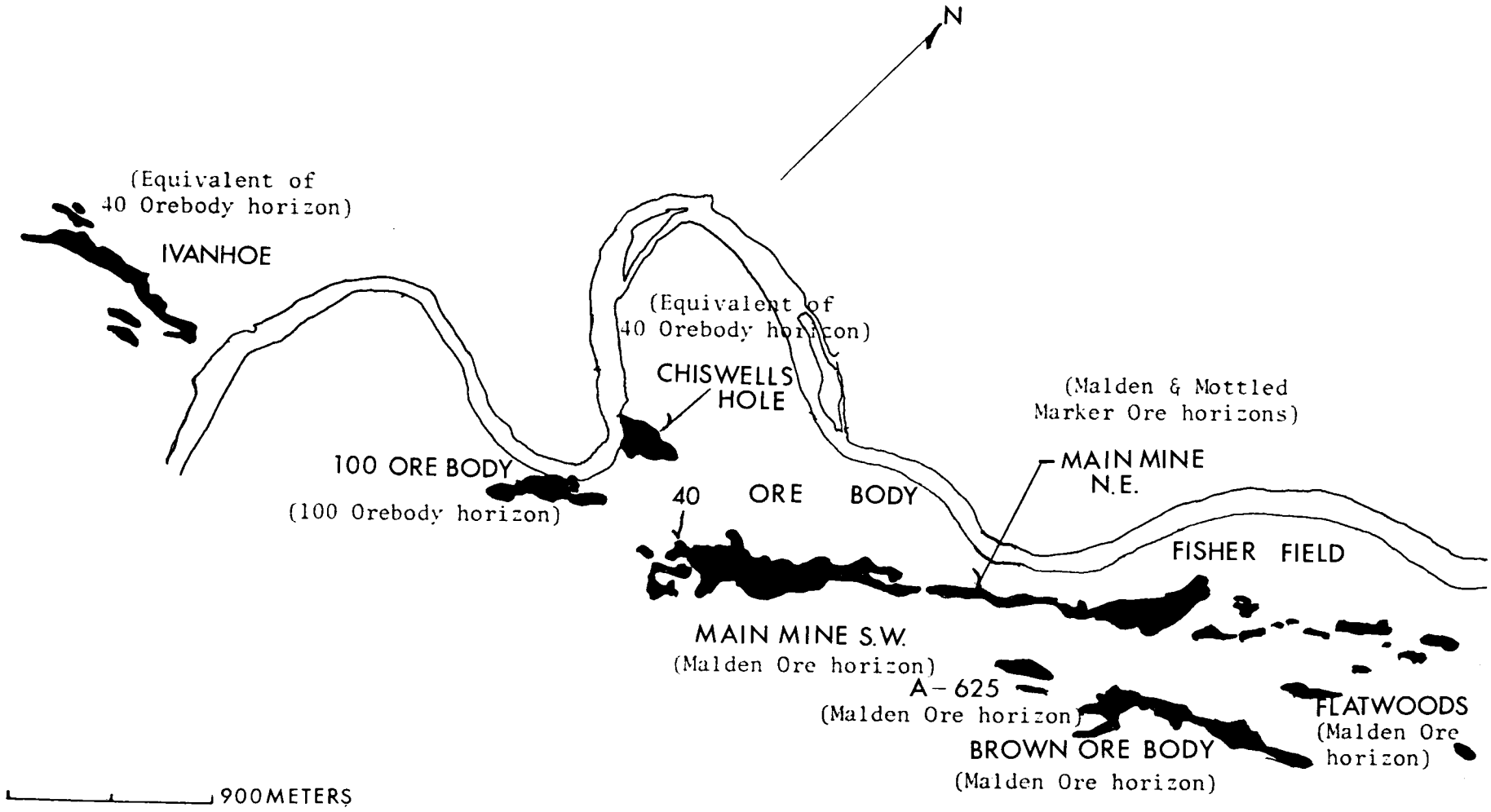

Figure 30. Plan view of Austinville-Ivanhoe Orebodies, after Brown and Weinberg (1968).

LOCATION OF SPECIMENS COLLECTED BY THE AUTHOR

South-limb Facies (Austinville)

Sample Number	Stratigraphic Position
1 through 11	The first 11 samples were collected in 11-03-42 stope (Figure 31). Sample numbers 1 and 2 were collected approximately 6 feet up the pillar (235' NW of base line). Samples 3 thru 11 were collected along the upper wall of 11-03-42 stope from spad 44 #4 to spad 42 #3. The sample interval begins at the top of the Lower Ribbon Sand Grain (L.R.S.g.) with sample number 1 and terminates 80' true thickness above the L.R.S.g. in the 40 orebody (Figure 2, p. 4).
12 through 26	This group of samples was collected at 10' invervals along 11-21-90 crosscut on 1100 level in the 100 orebody Figure 32). The sampling intervals were measured from a spad in 11-21-100 drift. The sample numbers were marked on the rib of the drift with red paint where they were collected. The sample interval begins 145' below the L.R.S.g. with sample 12, and terminates 200' below the L.R.S.g. with sample 26 (Figure 2, p. 4).
27 through 32	These samples were collected from muck at the base of pillar 76 #2 in stope 11-19-74. The hanging wall of the stope is 10' below the L.R.S.g. (Figure 2, p. 4).
H-1 through H-15	These samples were collected 500' S 45° W from the 100 orebody (Figure 30). The samples were collected at 10' intervals along the crosscut on the 1100 level. The sampling intervals were measured from a spad in the drift. The sample numbers were marked on the rib of the crosscut with red paint where they were collected. The sample interval begins 190' below the L.R.S.g. and terminates 100' below the L.R.S.g. (Figure 2, page 4).

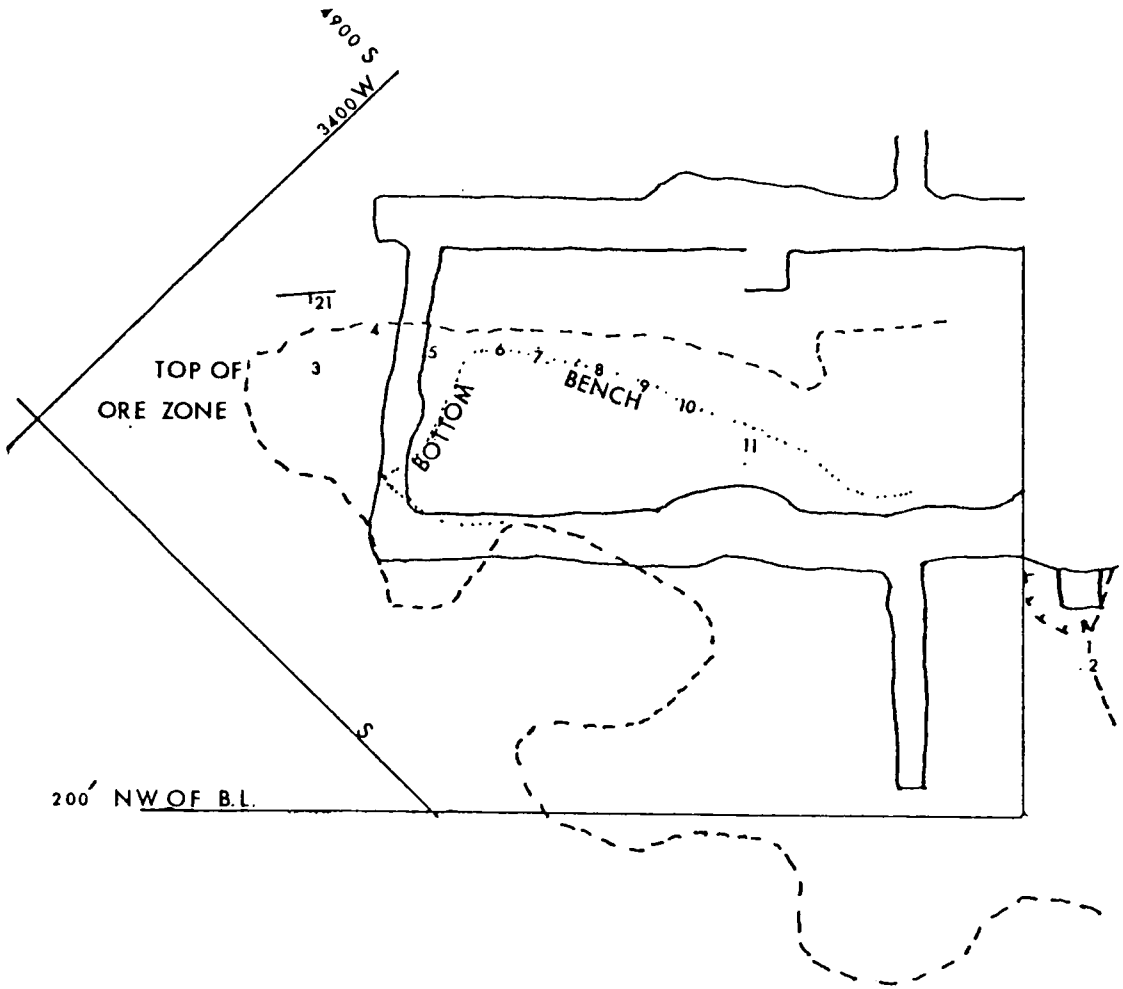

Figure 31. Sample location map of 11-30-42 stope, modified mine map supplied by New Jersey Zinc Company.

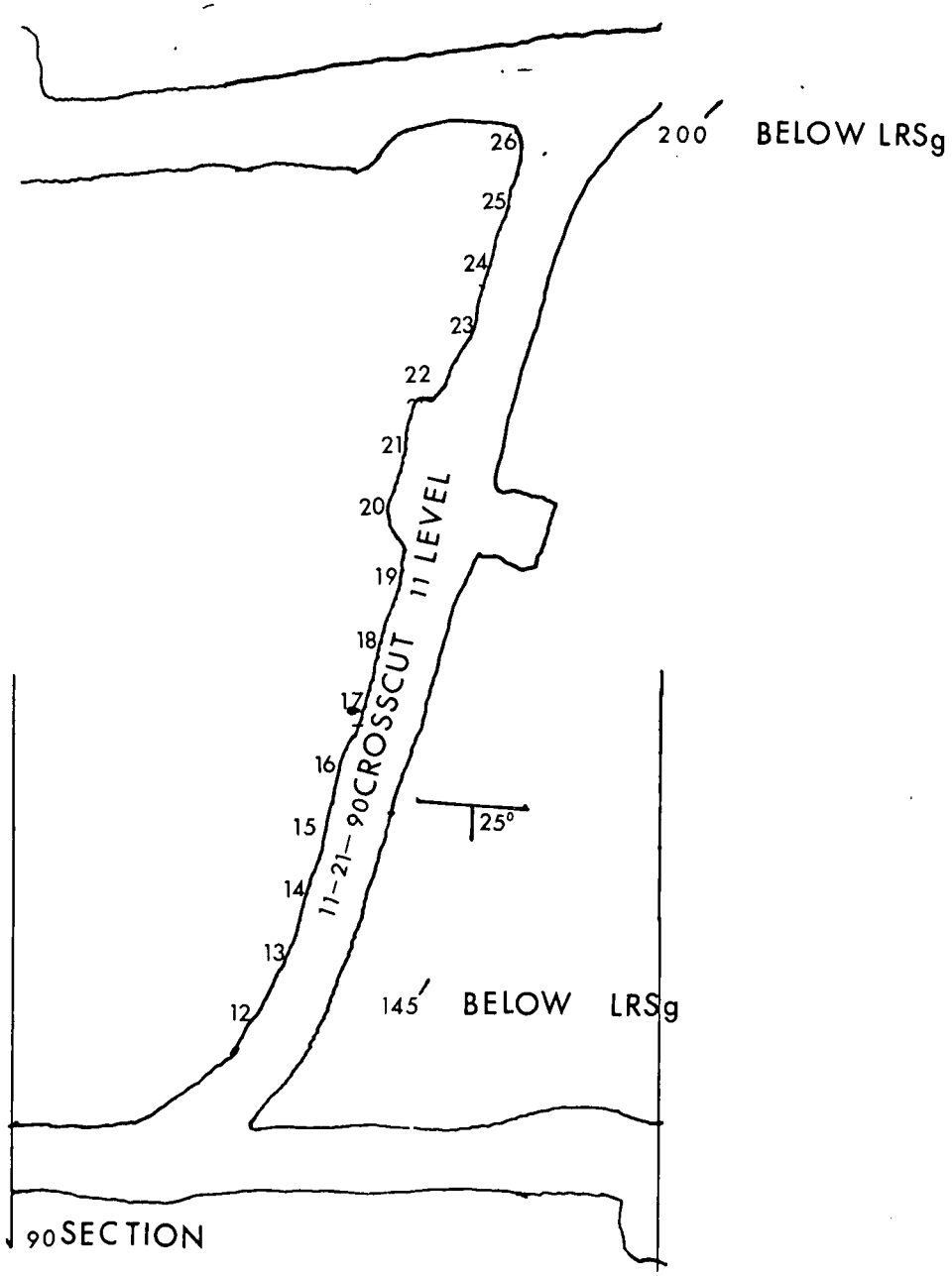

Figure 32. Location map of 11-21-90 crosscut on 1100 level in the 100 Orebody. Map is a modified version of one supplied by New Jersey Zinc Company.

North-limb Facies (Ivanhoe)

The following samples are rough stratigraphic equivalents to the 40 Ore Horizon south-limb facies (Figure 2, p. 4).

Sample Number	
33 through 36	These samples were collected from the face of 11-0-07 stope 100' SE of the base line near pillar 09 #1.
37 through 38	Both of these samples were collected in 11-01-00 stope. These samples are representative of Ivanhoe Ribbon ore.
39 through 40	These samples were collected from the muck in the SW side of 11-05-16 stope.
41 through 42	These samples were collected from 11-12-41 stope.
43 through 53	These samples were collected at 10' intervals from section 29+ 90' to section 31+ 80' along SE side of the 8th level drift at Ivanhoe. The sample intervals were measured from a tagged spad (NE-URBA) and the sample numbers marked on the SE side of the drift where the samples were collected (Figure 33).
54 through 57	These samples were collected from the muck in Chiswells Hole near pillar 62 #1 and #2 (Figure 28, p. 59).
8-1 through 8-32	These samples are stratigraphic equivalents to the samples collected in the 9th level drift (Figure 33). The samples were collected 800' NE of the Ivanhoe shaft on 1100 level. The sample intervals were measured from tagged spad every 10 feet and the sample numbers marked on the SE side of the drift where the samples were collected.

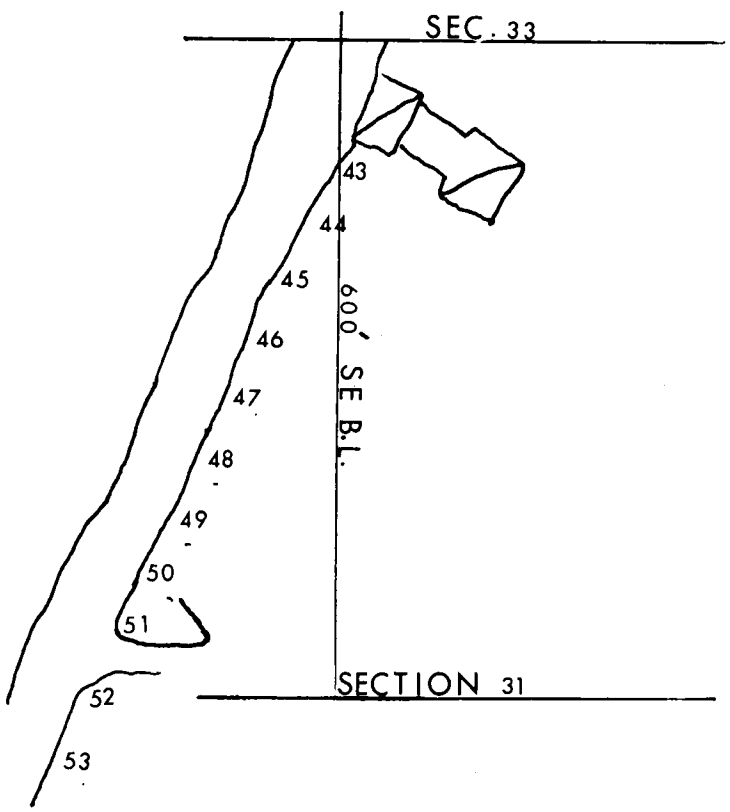

Figure 33. Location map of 8th level drift Ivanhoe, modified from map supplied by New Jersey Zinc Company.

TABLE II

LOCATION OF SPECIMENS PROVIDED BY MICHAEL L. EBERS,
MINE GEOLOGIST (AUSTINVILLE, VIRGINIA)

Sample Number	Orebody (Fig. 29, p. 60)	# of Specimens	Remarks (Fig. 2, p. 4)
stope 4-03-46	open pit	4	ore samples, Malden Ore Horizon
stope 5-02-09	Main Mine N.E.	3	ore samples, Malden Ore Horizon
stope 5-02-16	Main Mine S.W.	3	ore samples, Malden Ore Horizon
stope 5-03-30	Main Mine S.W.	2	ore samples, Malden Ore Horizon
stope 5-04-20	Main Mine S.W.	3	ore samples, Malden Ore Horizon
stope 5-34-59	A-625 Orebody	1	ore sample, Malden Ore Horizon
stope 5-34-61	A-625 Orebody	2	ore samples, Malden Ore Horizon
stope 6-11-06	not available	1	Ivanhoe, Malden Ore Horizon
stope 7-04-05	not available	1	ore sample, Malden Ore Horizon
stope 7-04-18	Main Mine S.W.	4	ore sample, Malden Ore Horizon
stope 7-10-69	Main Mine N.E.	1	ore sample, Malden Ore Horizon
stope 7-12-77	Main Mine N.E.	2	ore samples, Malden Ore Horizon
stope 7-14-89	Main Mine N.E.	2	ore samples, Malden Ore Horizon
stope 7-14-101	Main Mine N.E.	1	ore sample, Malden Ore Horizon
stope 7-22-195	Flatwoods	2	ore samples, Malden Ore Horizon
stope 7-40-65	A-625 Orebody	8	ore samples, Malden Ore Horizon
stope 7-44-75	Brown Orebody	3	ore samples, Malden Ore Horizon
stope 11-01-09	not available	1	Ivanhoe Ribbon, 40 Orebody Horizon
stope 11-03-42	40 Orebody	1	ore sample, 40 Ore Horizon
stope 11-10-09	unknown	1	location unavailable
stope 11-10-17	unknown	1	location unavailable
stope 11-12-41	Ivanhoe	3	ore samples, Ivanhoe Ribbon, 40 Orebody Horizon
stope 11-19-74	100 Orebody	1	ore specimen, 100 Ore Horizon
stope 11-19-78	100 Orebody	1	ore specimen, 100 Ore Horizon

TABLE II (Continued)

Sample Number	Orebody (Fig. 29, p. 60)	# of Specimens	Remarks (Fig. 2, p. 4)
stope 11-19-80	100 Orebody	2	ore samples, 100 Ore Horizon
stope 11-21-80	100 Orebody	2	ore samples, 100 Ore Horizon
stope 11-21-84	100 Orebody	1	ore sample, 100 Ore Horizon
stope 11-23-106	100 Orebody	5	ore samples, 100 Ore Horizon
stope 11-25-80	100 Orebody	1	ore sample, 100 Ore Horizon
stope 11-40-89	Brown Orebody	4	ore samples, Malden Ore Horizon
stope 11-48-89	----------	1	ore sample, Malden Ore Horizon
stope 11-48-91	Brown Orebody	4	ore samples, Malden Ore Horizon
stope 11-48-113	Brown Orebody	1	ore sample, Malden Ore Horizon
91 × cut	Main Mine N.E.	1	ore sample, Mottled Marker
Chiswells Hole	Chiswells Hole	1	ore sample, 40 Orebody Equivalent
100 O.B.	100 Orebody	1	
1940	Ivanhoe 1100 level	1	ore sample, 40 Orebody Horizon
V-1946	Fisher Field	1	ore sample, Malden Ore Horizon
-----	unknown	5	5 ore samples, jasper replacement common

VITA

James B. Hersch was born in Bethlehem, Pennsylvania, on April 15, 1952. He graduated from Freedom High School, Bethlehem, Pennsylvania, in June of 1970. In September 1970 he entered Appalachian State University and received a Bachelor of Arts degree in 1975. In September 1975 he entered the Graduate School at the University of Tennessee, Knoxville, Tennessee, to begin work on a master's degree with a major in geology. Upon completion of his master's degree, he took a position with Exxon Company U.S.A. as an exploration geologist.