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# GEOLOGY AND ORIGIN OF THE PRECAMBRIAN BANDED IRON DEPOSITS AT CLEVELAND GULCH, IRON MOUNTAIN, AND CANON PLAZA, RIO ARRIBA COUNTY, NEW MEXICO

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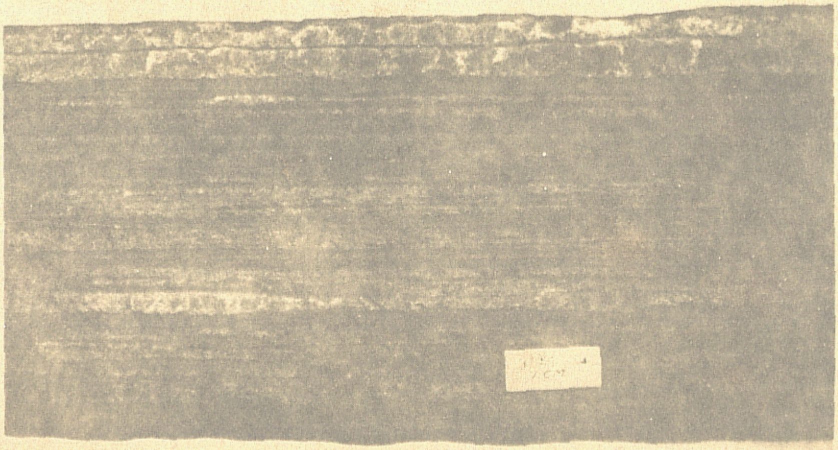
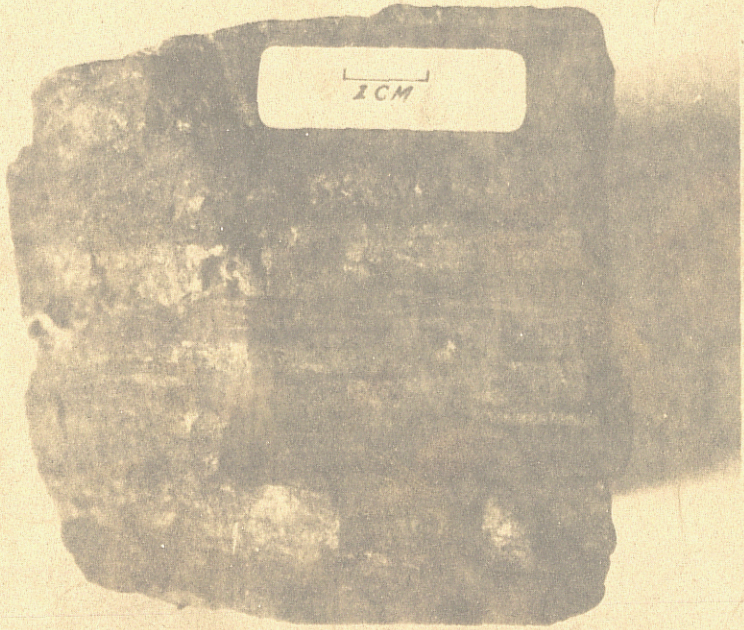


A. Well banded iron formation. Magnetite (martite)-rich bands and quartz-rich bands. Cleveland Gulch.



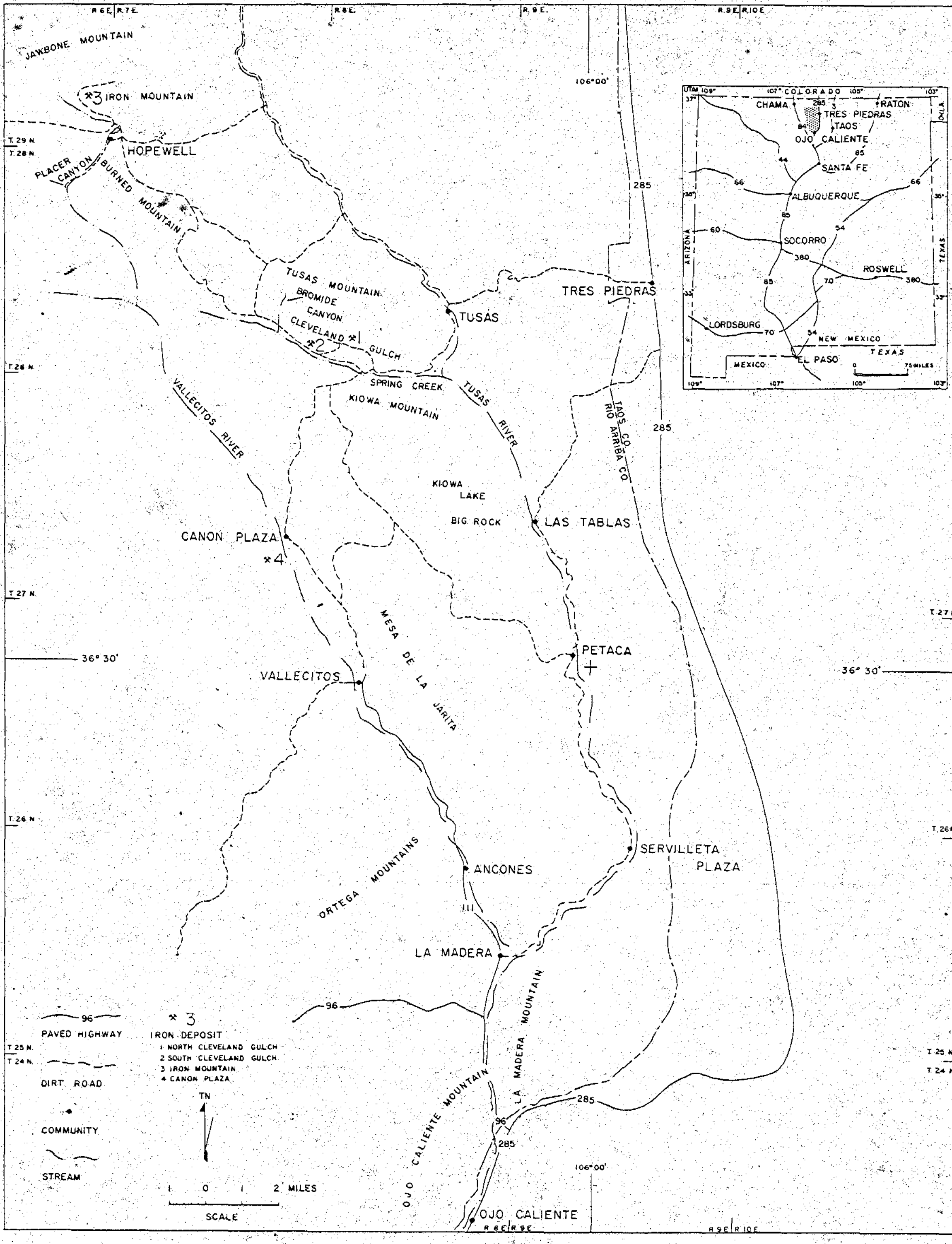

B. Poorly banded iron formation. Magnetite-rich bands and quartz-rich bands. Cleveland Gulch.

# INTRODUCTION

## Location and Accessibility

Banded iron deposits at Cleveland Gulch, Iron Mountain, and Cañon Plaza are 17 to 30 miles south of the Colorado border within the bounds of the Carson National Forest in Rio Arriba County, north-central New Mexico (Fig. 1). Cleveland Gulch is 38 miles northwest of Taos and 70 miles north-northwest of Santa Fe, New Mexico. These deposits are limited to Ranges 6, 7, and 8 East and Townships 27, 28, and 29 North and lie wholely within the Las Tablas Quadrangle, which is bounded by longitude 106° and 106° 15' West and latitude 36° 30' and 36° 45' North.

No all-weather roads extend into the area of the deposits. Cleveland Gulch is on unpaved Route 111 ten miles west-southwest of Tres Piedras, a small community on paved Route 285 (Fig. 1). Iron Mountain is 12 miles northwest of Cleveland Gulch and 2 miles north of Hopewell Lake over fair weather roads (Fig. 2). Cañon Plaza deposit is 8 miles south-southwest of Cleveland Gulch and one mile south of the community of Cañon Plaza which is on unpaved Route 111. Cañon Plaza is 5 miles northwest of the end of pavement on




~~FIGURE 1.~~ INDEX MAP OF CLEVELAND GULCH, IRON MOUNTAIN, AND CANON PLAZA AREA, RIO ARRIBA COUNTY, NEW MEXICO

PLATE I

Route 111 at Vallecitos (Fig. 1).

The New Mexico State Highway Department has started construction of a highway which will connect Taos and Tres Piedras to Route 84 a few miles south of Chama (Fig. 1). When completed, this highway will pass within 2 miles of both Cleveland Gulch and Iron Mountain. In the near future Route 111 may be paved from Vallecitos to Cañon Plaza. Atchison, Topeka, and Santa Fe Railway's nearest tracks are at Lamy which is 10 miles southeast of Santa Fe. Denver and Rio Grande Western Railway is 30 miles north of the area at Antonito, Colorado. Besides the small communities of Cañon Plaza, Vallecitos, and Tres Piedras there are the communities of La Madera, Servilleta Plaza, Petaca, Las Tablas, and Tusas near the area.

## Geography and Vegetation

North-central New Mexico is part of the Southern Rocky Mountains Physiographic Province (Fenneman, 1931, p. 94). Precambrian and Cenozoic rocks of the area are part of the San Juan Mountains, which extend from Ouray, Colorado eastward to the vicinity of Antonito, and then southeastward to the vicinity of Ojo Caliente, New Mexico. From here they swing to the northeast and butt into the Sangre de Cristo Range about 3 miles south of

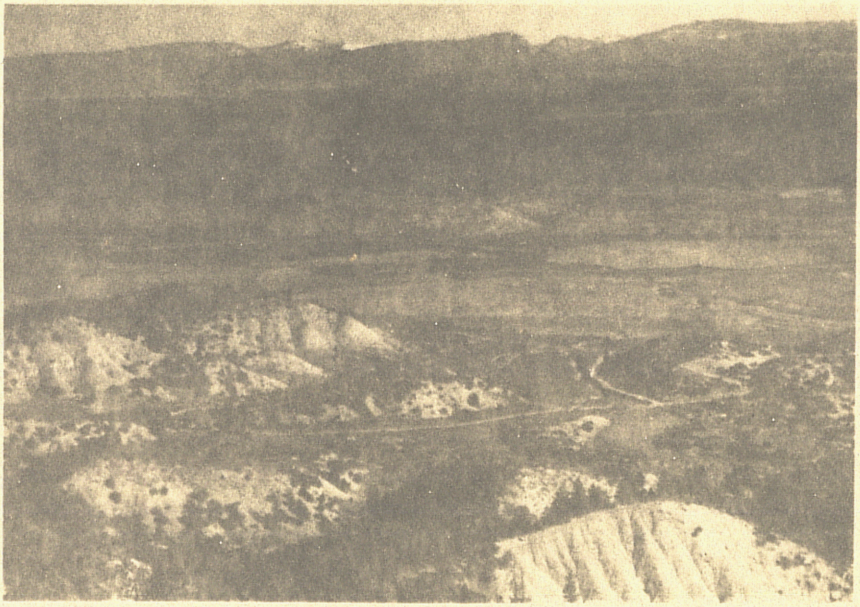
Taos (Fig. 3). Even though many people have emphasized that the name San Juan should be applied in New Mexico (Atwood and Mather, 1932, p. 6; Raisz, 1939), there has been little general use of this name for these mountains in New Mexico. Consequently, the southeastern part of the San Juan Mountains has no accepted general name. Atwood and Mather (1932, p. 8) stated that the local name is Tusas Mountains, but a single peak is also known as Tusas, and this makes the name undesirable for the entire range. Therefore, in this report the name San Juan Mountains is used to refer to the New Mexico extension and the Colorado San Juan Mountains, and appropriate modifiers are used to denote specific areas.

The San Juan Mountains of New Mexico comprise a well defined group of high, not very rugged mountains which are almost surrounded by plateau lands (Plate 2A). Chama River drainage defines the western and southwestern boundary and separates the range from the Jemez Mountains to the southwest. To the east the flat San Luis Valley and its drainage, the Rio Grande, separate most of the New Mexico San Juan Mountains from the Sangre de Cristo Range (Fig. 3). San Luis Valley ends about 10 miles south of Taos where the Rio Grande cuts through the San Juan Mountains separating the Picuris Range, which is the easternmost prong, from the rest of the San Juan

~~PLATE~~ FIGURE 2

A. San Juan Uplift, northern New Mexico in background. Chama Basin in foreground and middleground. Notch in right background was cut by Brazos River through massive Precambrian quartzite. View east.

B. Southern flank of Jawbone Mountain viewed from Iron Mountain. View north.



A

B

Mountains.

Precambrian rocks form windows in the Cenozoic cover in the New Mexico San Juan Mountains. Tusas Mountain, a mile north of Cleveland Gulch, has an elevation of 10,120 feet and is a fairly rugged, heavily forested granitic peak (Fig. 2 and Plate 3A). To the northwest a broad swell with many flat treeless meadows separates Tusas Mountain from Burned Mountain, which is a rounded, heavily forested mountain of 10,189 feet elevation (Plate 3B). Northwest of Burned Mountain are many broad, gently concave meadows separated by heavy forest. The meadow lands become more numerous in the vicinity of Jawbone Mountain (Fig. 1), whose summit at an elevation of 10,605 feet is almost devoid of trees (Plate 2B). Iron Mountain is a small ridge about one mile south of Jawbone Peak. Jawbone Peak is the highest point in the immediate vicinity of the three deposits, and the Vallecitos River at an elevation of around 8,000 feet at Cañon Plaza is the lowest point.

Kiowa Mountain, which is a rugged quartzite peak with an elevation of about 9,600 feet, is south of Tusas Mountain. South of Kiowa Mountain the elevation decreases to around 8,500 feet on a flat plateau known as La Jarita Mesa. La Jarita Mesa is bounded by the Tusas River on the east and the Vallecitos River on

Figure ~~PLATE~~ 3

A. Tusas River Valley. View southeast from Jawbone Mountain. Eastern end of Iron Mountain in foreground. Northwestern flank of Tusas Mountain in right background.

B. View south-southeast from Jawbone Mountain. Burned Mountain in left background. Middle part of Iron Mountain in foreground.

C. Outcrops on Iron Mountain. View north.

A

B

C

the southwest. Several quartzite peaks known as the Ortega Mountains project through the Cenozoic cover south of the Vallecitos River. Cañon Plaza deposit is on one of the quartzite ridges in the Ortega Mountains.

East of Tusas, Burned, and Jawbone mountains is the broad valley of the Tusas River (Fig. 2 and Plate 3A), which has its headwaters a few miles south of the Colorado border. A very deep canyon has been cut by the river east of Cleveland Gulch, and this intersects Spring Creek Canyon along the eastern margin of Kiowa Mountain. Spring Creek has its source on the southwestern flank of Tusas Mountain. Tusas River continues south along the eastern flank of La Jarita Mesa through Petaca to La Madera where it joins the Vallecitos River (Fig. 1). Vallecitos River has its headwaters west of Hopewell and flows through Cañon Plaza and Vallecitos along the southwestern flank of La Jarita Mesa. Vallecitos River (sometimes called the Caliente River south of La Madera) joins the Chama River about 15 miles south of Ojo Caliente, and the Chama joins the Rio Grande a few miles farther southeast. Few other streams are perennial within the area. Stream canyons furnish most of the outcrops, which at best are meager and poorly exposed.

Part of the drainage probably was superimposed onto the Precambrian surface from overlying Cenozoic rocks. However, many of the streams have adjusted themselves to weak zones caused by the great difference in lithologies, the vertical dip of most of the Precambrian rocks, and presence of a prominent northeasterly fault system. Parts of Placer, Spring Creek, and Tusas canyons may have been cut by superimposed streams, since they cut through massive quartzite whereas much more easily erodable rocks were available nearby.

Atwood and Mather (1932, p. 23) believed that the surface produced by the San Juan cycle of erosion reached only late maturity in the New Mexico San Juan Mountains. They attributed this to superior resistance to erosion of the Precambrian terranes in this area. Barker (1958, p. 6) called Kiowa and Tusas mountains monadnocks, but since this area never reached peneplanation, these should not be called monadnocks. Atwood and Mather (1932, p. 25) believed the San Juan erosional cycle occurred in late Pliocene time.

Normally the area receives 15 to 20 inches of precipitation a year (Visher, 1954, p. 197). About half of this comes during the summer months in the form of extremely violent thunderstorms. The remainder comes as occasional spring and fall rains and from heavy winter snows, which normally amount to about 100 inches. The first snowfall is usually around the first of November, but field work often can be carried out until the first of January. Average annual temperature within most of the area is around 45° F (Visher, 1954, p. 23). The normal January minimum is 3° F, and the maximum is 35° F; normal July minimum is usually 35° F, and the maximum is 80° F. A diurnal change in temperature of 40° to 50° F is common throughout most of the year.

Much of the area of study has an elevation between 9,000 and 11,000 feet and is characterized by the Rocky Mountain Subalpine Forest association. The characteristic trees of this association are Engelmann spruce (*Picea engelmanii*), Colorado blue spruce (*Picea pungens*), alpine fir (*Abies lasiocarpa*), and aspen (*Populus tremuloides*) (Plate 4B). These trees are typically in small stands separated by gently undulating grassy meadows. However, the north facing slopes have a very dense forest cover. At the higher elevations in the northwestern part of the area, the dense cover gives way to open grass-covered meadows and small stands of aspen.

Between 8,000 and 8,500 feet trees of the Rocky Mountain Montane Forest are in more or less open stands. This association is characterized by ponderosa pine (*Pinus ponderosa*), Douglas fir (*Pseudotsuga menziesii*), white fir (*Abies concolor*), New Mexico locust (*Robinia neo-mexicana), Gambell oak (scrub) (Quercus gambellii), and Rocky Mountain juniper (Juniperus scopulorum) (Benson, 1957, p. 586). Between 8,500 and 9,000 feet there is a mixing of trees from both associations. In valleys between 8,000 and 8,500 feet grassland predominates except on dry rocky slopes where sagebrush (Artemisia tridentata), mountain mahogany (Cercocarpus montanus), and scrub oak (Quercus gambellii) are common (Dice, 1943, p. 36-42). In the eastern part of the area of study, ponderosa pine forest interdigitates with the piñon (Pinus edulis) - juniper (Juniperus monosperma) association of the San Luis Valley.

At Cleveland Gulch scrub oak was found to be extremely useful in locating the iron deposits. The thickest growths are on slopes with the greatest amount of iron mineralization. The scrub oaks highest on the slope usually define exactly the upper boundary of the iron deposit (Plate 4A). This phenomenon was probably due to soil moisture rather than any dependence of the scrub oak on iron or related elements in the soil.

## Geologic Setting

Cleveland Gulch, Iron Mountain, and Cañon Plaza iron deposits are in the Precambrian rocks of the New Mexico San Juan Uplift. The uplift is flanked on the east by the San Luis Basin, butts into the Sangre de Cristo Uplift on the southeast, plunges into the Rio Grande Trough to the south, and merges into the Chama Basin on the west (Fig. 3).

The San Juan Mountains in this part of New Mexico consist of a sequence of late Precambrian metamorphic and intrusive rocks surrounded by Cenozoic sedimentary and volcanic rocks. In the area under consideration, the oldest unit, the Moppin Formation, is composed of schist, amphibolite, phyllite, and quartzite. Conformable, or nearly so, above this is a thick formation of quartzite with minor schist and amphibolite. This unit is called the Ortega Quartzite. Iron mineralization is present in both the Moppin Formation and the Ortega Quartzite. These formations have been intruded and metamorphosed by the Tusas intrusives ~~Granite~~, which has batholithic dimensions. The formation of the banded iron deposits, base and precious metal vein deposits, and pegmatites apparently accompanied the intrusion.

The rocks have been deformed into large nearly isoclinal folds which plunge steeply toward the west-northwest. The two major folds are the Hopewell Anticline and Kiowa Syncline, which adjoins it on the southwest. Smaller folds modify this structural pattern in some places. Only a few faults are found in the Precambrian, and many of these trend northeast. Schistosity of the rocks parallels the bedding planes in most outcrops.

During late Precambrian time there was extrusion of a thick sequence of volcanic material of felsic to mafic composition. Some of the deposits were intruded by dikes and sills. In local basins and stream beds thin deposits of sand and silt were built up between flows. With little or no erosion, thick deposits of cross-bedded sand followed. Igneous activity continued sporadically during deposition of the sand. These units were indurated, folded, and subjected to low grade regional metamorphism. Some dikes and sills were intruded at this time, and the area was again subjected to regional metamorphism. All the units were intruded by a granitic batholith and subjected to metasomatism over large areas. This large ~~zoned~~ intrusion apparently supplied the materials for most of the mineral deposits of the area, and caused many of the metamorphic effects seen today in the different rock units. Another period of metamorphism, subsequent to the intrusion and metasomatism, was low grade and regional, imparting slight schistosity to the intrusive rock.

In the Paleozoic and Mesozoic eras the area may have received sediments but no trace remains. A few miles to the south and west there is a thick sequence of rocks of Pennsylvanian through Cretaceous age. During the Tertiary the area received fluvial sediments and volcanics, many of which were stripped from the area before late Pliocene time. During the Quaternary more volcanic material and sediments were deposited over parts of the area. At present the Precambrian rocks form a series of windows in the overlying Cenozoic cover.

## Previous Investigations

Evan Just (1937) made a reconnaissance map of the Picuris Uplift and the area from Jawbone Mountain to Ojo Caliente in a study of the pegmatites of the area. Just separated the major rock units with such efficacy that his units still stand as the most workable in most cases.

Atwood and Mather briefly discussed the region under consideration in their physiographic studies of the San Luis Valley (1924) and San Juan Mountains (1932). They were able to recognize the erosion surface which corresponds to the San Juan Peneplain in Colorado on the eastern front of the New Mexico extension of the San Juan Mountains (1932, p. 23).

Cross and Larsen (1935) and Larsen and Cross (1956) studied the Colorado San Juan Mountains, and while these studies were not directly connected with the present area, they do give some idea of the geologic history between the end of Precambrian and the beginning of the Cenozoic. Butler (1946) studied the Tertiary and Cenozoic geology of part of the subject area. Barker (1958) published on the geology of the Las Tablas Quadrangle.

A generalized geologic map of the Rio Chama Country, which included the area under consideration, was published in 1960 by the New Mexico Geological Society (Smith and Muehlberger, 1960). However, it was a compilation from earlier sources and presented no new information. At the same time Muehlberger (1960) published a paper on the Tusas Mountains restating the published work of Just and Barker. In 1965 Bingler published a study of the geology and structural petrology, including a petrologic map, of the La Madera Quadrangle, which adjoins the southern margin of the Las Tablas Quadrangle. Carpenter (1965) did petrologic studies on some of the rock units from Tusas Mountain to Burned Mountain.

Bromide and Hopewell mining districts are between Cleveland Gulch and Iron Mountain. Bromide mining district was discovered in July, 1881 by D. M. Field and J. M. Bonnett (Jones, 1964, p. 75). This camp, about a mile northwest of Cleveland Gulch, produced mostly silver, copper, and a little gold. Little ore has been produced since the turn of the century. Hopewell District was established a few years before Bromide and had both lode and placer mines (Jones, 1964, p. 76). The total value of production was around $200,000. Recently an attempt was made at hydraulic gold mining of the gravel in the lower part of Placer Canyon. However, the operation has apparently closed down before reaching production.

Graton (in Lindgren, Graton, and Gordon, 1910, p. 124-133) studied the Hopewell and Bromide districts and made brief reference to the iron deposits on Iron Mountain. Jahns (1946) in his detailed study of the pegmatites of the Petaca District (Fig. 1) presented information on the general geology of the area. Kelley (1949, p. 230) in a compilation of the iron deposits of New Mexico mentioned the Iron Mountain deposit. Kyanite deposits in the Petaca District were studied by Corey (1960).

The only recent study involving the iron deposits was published by Bertholf in 1960. He studied part of the North Cleveland Gulch deposit (Fig. 1), conducted a magnetometer survey of parts of the deposit, and had beneficiation studies run on a few samples of the magnetite-rich rocks.

## Purpose and Method of Investigation

The presence of a Precambrian banded magnetite-quartz deposit at Iron Mountain has been known for over fifty years (Lindgren, Graton, and Gordon, 1910, p. 128). Despite subsequent publications on the general geology of the area, there has been no study of the geology of this and other surrounding iron deposits. The primary purpose of this investigation is an elucidation of the occurrence, structure, composition, origin, and economic potential of the Precambrian banded iron deposits in the Moppin Formation and Ortega Quartzite at Cleveland Gulch, Iron Mountain, and Cañon Plaza.

A total of eight months was spent in the field. During this time an outcrop geologic map was made of the Iron Mountain deposit and adjacent area at a scale of one inch to fifty feet with the aid of a plane table and alidade (Fig. 4). The Cleveland Gulch deposits were mapped at a scale of one inch to approximately 440 feet on enlarged aerial photographs (Fig. 5). Cañon Plaza deposit was mapped on aerial photographs at a scale of one inch to about 1,320 feet (Fig. 6). Base maps for Cañon Plaza and Cleveland Gulch were prepared from aerial photographs. The area between Cleveland Gulch and Iron Mountain was mapped in the field on aerial photographs at a scale of one inch to about 1,320 feet (Fig. 2). The base maps for Figure 2 were Topographic Sheets of the United States Geological Survey. Heavy vegetation and soil cover imposed limitations on the accuracy of mapping.

Laboratory work consisted of petrographic study of 150 thin sections and 25 polished sections. In addition, approximately 300 hand specimens were studied with a binocular microscope. Semi-quantitative spectrographic analyses for 50 elements were made on 63 samples (Table IV). Specific gravity of 37 samples was determined as a means of ascertaining iron content (Table V). Iron content in four samples was determined by fire assay in order to compare the results of the specific gravity test (Table VI).

## Acknowledgments

Many people have generously given of their time and advice to help with this project. Special thanks are due Charles F. Park, Jr. for his many helpful suggestions throughout the duration of the study. A. J. Thompson, former Director of the New Mexico Bureau of Mines, and Mineral Resources greatly encouraged the project. W. E. Bertholf of the New Mexico Bureau of Mines and Mineral

Resources spent several days in the field with me; his knowledge of the area and iron deposits was freely given then and in subsequent discussions.

T. W. Muraro and S. M. Richards have been especially helpful in reviewing and discussing certain parts of the paper. Evan Just and R. H. Jahns greatly aided with their extensive knowledge of the subject area. Chester R. Longwell has taken considerable time to discuss some aspects of this problem. Carol Ann McLeroy has been a very able field assistant and geologic critic throughout this study.

V. C. Kelley of the University of New Mexico very generously provided several pieces of field equipment during the summer of 1964 and took time from his work to discuss certain facets of the problem. Jack Campbell ably assisted in the field on two different occasions. T. S. Chmura and F. E. Fitzgerald provided valuable technical information for which I am extremely grateful.

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## PRECAMBRIAN ROCKS

### General Statement

Within the Cleveland Gulch, Iron Mountain, and Cañon Plaza areas there are three major Precambrian rock units: the Moppin Formation, the Ortega Quartzite, and the Tusas ~~Granite.~~ Intrusives. The oldest is the Moppin Formation, which apparently is conformably overlain by the Ortega Quartzite. Both of these units were intruded by the Tusas ~~Granite.~~ Intrusives. The iron deposits at Iron Mountain and most of those at Cleveland Gulch are in the Moppin Formation. Part of the southern Cleveland Gulch deposit extends into the Ortega Quartzite (Fig. 2). Iron mineralization at Cañon Plaza is restricted to the Ortega Quartzite (Fig. 6).

Moppin Formation rocks are mainly schist, amphibolite, and phyllite; quartzite is present but is quantitatively unimportant. Ortega Quartzite is cross-bedded, fine to coarse-grained, and locally conglomeratic. Tusas ~~Granite is a~~ Intrusives are partly zoned, ~~intrusive,~~ less silicic in the northwestern part, and has several local apophyses.

The Moppin Formation possibly is correlative with the Irving Greenstone in the Colorado San Juan Mountains; likewise, the Ortega Quartzite might be correlative with the Needle Mountain Group of the same area. This inference is suggested by the following facts: the Irving Greenstone and Needle Mountain Group are the youngest non-intrusive Precambrian rocks in the San Juan Mountains of Colorado (Cross and Larsen, 1932, p. 20); the degree of metamorphism in both areas is virtually the same; and the lithologies are strikingly similar. Larsen and Cross (1956, p. 257) stated that the Uncompaghre Formation (of the Needle Mountain Group) is found as far south as Brazos Canyon, which is about 5 miles northwest of Jawbone Peak. Since the quartzite at Brazos Canyon can be traced almost to Ortega outcrops, it would seem there is a strong possibility that the two formations are equivalent.

## Stratigraphy

### Moppin Formation

#### Definition and distribution

Moppin Formation underlies the area from Cleveland Gulch to Iron Mountain in the Las Tablas Quadrangle (Fig. 2). It is the oldest formation exposed in the Cleveland Gulch - Iron Mountain area. Just (1937, p. 10)

FIGURE
~~PLATE~~ 4

A. North Cleveland Gulch. Scrub oak in the foreground marks the upslope contact of iron formation with enclosing rocks.

B. Cleveland Gulch. Typical vegetation. Metarhyolite outcrop on right.

C. Ortega Quartzite forming the nose of Hopewell Anticline. View northwest.

A

B

C

gave this unit the name Hopewell Series. Since this name was preoccupied, Barker (1958, p. 14) renamed the unit Moppin Metavolcanic Series for exposures in a canyon near "the Moppin Ranch." Unfortunately, he failed to give the location of the ranch on his map, and ownership of the ranch has since changed. After much searching it was found that the canyon he used for the reference locality is Bromide Canyon, ~~and on Figure 5 Moppin Ranch and the canyon have been noted~~. According to the Code of Stratigraphic Nomenclature, Article 13h, the type locality should still be Placer Canyon near the community of Hopewell where Just originally named and defined the formation; the locality Barker used should be termed a reference locality (American Commission on Stratigraphic Nomenclature, 1961, p. 653).

Even though the Code of Stratigraphic Nomenclature stated that a formation should not be named after a ranch, it is judged better to leave the name as it is than to burden the literature with an additional name (American Commission on Stratigraphic Nomenclature, 1961, p. 652). In the future it may be possible to correlate this formation with the Irving Greenstone of the Colorado San Juan Mountains in which case the names "Hopewell Series" and "Moppin Metavolcanic Series" would be predated and therefore superceded by the name "Irving Greenstone."

Because not all the rocks in the Moppin Formation are metavolcanic and the term "Series" is a time-stratigraphic designation and has no proper place in rock-stratigraphic terminology, in the present report the "Moppin Metavolcanic Series" of Barker has been changed to "Moppin Formation" in order to comply with the Code of Stratigraphic Nomenclature.

Barker (1958, p. 11) considered the Moppin Formation to be younger than one quartzite lithosome, but older than the rest of the quartzite. Just (1937, p. 13) believed all the quartzite to be younger than the Moppin Formation. Corey (1960, p. 7) retained Just's nomenclature and proposed geologic sequence. Bingler (1965) did not present a time sequence for the rocks in the La Madera Quadrangle, so we do not know what he considered to be the relative ages of rocks in that area which contains the type locality for the Ortega Quartzite. Barker and Just both agree that the quartzite along the flanks of the anticline from Cleveland Gulch to Iron Mountain is younger than the Moppin rocks. Consequently, since no other quartzite is in contact with the Moppin rocks, the Moppin Formation in the present report is considered to be the oldest unit in the area. Because it has priority (Just, 1937, p. 11), the name Ortega Quartzite is used in this report for the quartzite lithosome.

Moppin Formation crops out in Spring Creek Canyon north to the Tusas ~~Granite~~ rocks and northwest of Burned Mountain, Placer Canyon, and the southeastern flank of Jawbone Mountain (Fig. 2). Part of this formation extends into the Tusas River Valley between Burned Mountain and Tusas Mountain. This outcrop pattern is that of a steeply plunging anticline with Iron Mountain on the northeastern flank and Cleveland Gulch on the southwestern flank. The beds are for the most part either upright and steeply dipping or are vertical. Outcrops of the Moppin Formation are often interrupted by the Tusas intrusions, faults, Tertiary cover, soil, and vegetation.

Muscovite schist, which is interbedded at several horizons in the Moppin Formation, is mainly metamorphosed rhyolite. Just (1937, p. 44) thought these beds unique enough to be considered a separate member and originally named them Vallecitos Rhyolite. He believed they were flows interbedded with the other volcanics of the Moppin Formation. Barker (1958, p. 54) renamed this unit Burned Mountain Metarhyolite and considered it to be intrusive. Present studies support Just's conclusion that these units were deposited as surface flows; they are the most reliable stratigraphic control within the Moppin Formation. Jahns (personal communication) considers the metarhyolites in the Petaca District to be extrusive deposits. ~~These~~

~~metarhyolites have been dated at 1.7 billion years, and the oldest granite in the area, the Tusas Granite, has been dated at 1.4 billion years (Jahns, personal communication~~).

Since these metarhyolites are an integral part of the Moppin Formation, are so interspersed throughout the sequence as to be difficult to map separately, and are also found in the Ortega Quartzite, no special name should be given to them. Both names, "Vallecitos Rhyolite" and "Burned Mountain Metarhyolite," should be abandoned.

## Lithology

Moppin Formation is composed mainly of chlorite, muscovite, and biotite schists, amphibolites with varying amphibole percentages, phyllite, and a minor amount of quartzite. In the field several traverses were made normal to the strike of the beds in order to determine the lithologic units. In typical traverses soil and vegetation covered at least 80% of the area. As a consequence very few of the rocks were available for study except for a small area in the vicinity of the iron deposits north of Cleveland Gulch where the soil had been removed by a bulldozer. After completion of several traverses perpendicular to the strike, an attempt was made to follow the various units along strike. Amphibolite beds were generally found to be impossible to correlate exactly from outcrop to outcrop along strike because successive amphibolites looked so nearly alike and compositional and textural changes along strike were common. Also, there was the possibility that the amphibolite changed in the covered interval along strike to a chlorite schist. Chlorite, biotite, and to some extent muscovite schists presented the same problems.

After much investigation muscovite schists, which were originally rhyolite flows or tuffs, proved to be unique and exceptional in their persistence throughout most of the area of Moppin outcrops. One of these ~~flow~~ units was especially helpful since it was traced from east of Cleveland Gulch northwest along the Moppin-Ortega contact across Burned Mountain, Placer Canyon, around the nose of the anticline to Iron Mountain. Even though in places it is covered for several hundred yards, its distinctive appearance and its position at or near the Ortega Quartzite base permit its recognition almost anywhere (Fig. 5). In the vicinity of Cleveland Gulch another metarhyolite bed, two distinct phyllite beds, and a meta-arkose bed were also used for stratigraphic control.

Cleveland Gulch.--The most varied lithology of the Moppin Formation is in the area around Cleveland Gulch and consequently the petrology and petrography were studied in some detail at this locality. A description of the rocks found in traverses from the base of the Ortega Quartzite south of Cleveland Gulch to the Tusas Contact ~~Granite~~ is presented (Figs. 2 and 5). This composite section is made up of the various lithologies found on five south to north traverses made from about 1,000 feet east of Bromide Canyon to the vicinity of the eastern edge of the map in Figure 5. However, most of the lithologic units are described from samples taken on a traverse starting at the skull-shaped area of iron formation (Fig. 5), and extending almost due north through the widest part of the iron formation north of Cleveland Gulch to the contact of the northernmost intrusive outcrop shown on Figure 5.

The base of the Ortega Quartzite is not well defined in this area. Along the ridge south of Cleveland Gulch is a distinctive phyllite overlain by about 200 feet of meta-arkose beds. This unit is overlain by 200 feet of chlorite schist, succeeded by several hundred feet of feldspathic quartzite beneath pure cross-bedded quartzite which is undoubted and typical Ortega. The top of the chlorite schist was chosen as the top of the Moppin Formation because chlorite schist is more characteristic of Moppin rocks than Ortega rocks. Also, the feldspathic quartzite was traced along the contact for several miles, but the meta-arkose bed pinches out a couple of miles to the northwest. This alternation of meta-arkose beds with chlorite schist and the fact that the Ortega and Moppin formations strike parallel to each other support the conclusion that the two formations are conformable in this locality.

The overall lithologies of the Cleveland Gulch area categorized from youngest to oldest are: (1) chlorite schist; (2) quartz-muscovite schist (meta-arkose); (3) interbeds of phyllite and quartz-muscovite schist (meta-tuffs and metarhyolites); (4) amphibolite, quartz-biotite schist, chlorite schist, some phyllite; (5) banded iron formation, chlorite schist, muscovite schist, amphibolite; (6) amphibolite, biotite schist; (7) muscovite schist (metarhyolite); (8) amphibolite, biotite schist; (9) amphibolite, quartzite. Within each of the categories above, the lithologies are not necessarily listed in chronological order. The lithologic units were grouped in the above manner in order to correspond to the map units on Figure 5. On Figure 5 numbers 1, 4, 6, 8, and 9 are designated ~~p€m~~ M; numbers 3 and 7 are designated ~~p€mt~~ $M_1$; and number 2 is designated ~~p€ma~~ $M_2$. Two units containing iron formation were mapped separately; one is number 5 above, and the other cuts across the meta-arkose, uppermost chlorite schist, and the lower part of the Ortega Formation (Fig. 5).

At Cleveland Gulch the uppermost unit (1) in the Moppin Formation is chlorite schist. It is covered by Tertiary rocks to the southeast, and continues to the northwest ~~for several miles~~ parallel to the contact of the Ortega Quartzite (Fig. 5). In hand specimen it is a fine-grained chlorite-feldspar-quartz schist, with what appear to be relict amphibole phenocrysts.

Below the chlorite schist is about 200 feet of pinkish-gray muscovite schist. This unit (2) is fine to coarse-grained but contains a few metaconglomerate beds. It was an arkose deposit. In thin section the composition was found to be highly variable, but in general the sample contained rounded grains of quartz, potassium feldspar, and plagioclase, and muscovite, chlorite, biotite, and magnetite (Table I 1; Plate 5A). Some of the larger quartz and feldspar grains are slightly crushed (Plate 5B). The muscovite schist pinches out about one and a half miles to the northwest and is covered by Tertiary deposits to the east.

A skull-shaped area of banded iron deposits and amphibolite extends from the upper part of the phyllite-muscovite schist beds (which lie below the meta-arkose), through the meta-arkose, chlorite schist, and into the basal feldspathic quartzite of the Ortega Formation (Fig. 5). The area is about 900 feet in a north-south

TABLE I

| | Cleveland Gulch Unit 2 Muscovite Schist (Meta-arkose) | | Cleveland Gulch (South) Banded Iron Formation | | Cleveland Gulch (South) Iron Deposit Amphibolite | | Cleveland Gulch Unit 3 Phyllite Blue-green | |
|---|---|---|---|---|---|---|---|---|
| | 1 | | 2 | | 3 | | 4 | |
| | % | mm | % | mm | % | mm | % | mm |
| Quartz | 60-80 | 0.1-1.0 | 45-50 | 0.2 | a | | 25-45 | 0.01-0.4 |
| K-Feldspar | 0-5 | | | | | | 20-25 | |
| Oligoclase | 5-30 | | | | 40-55 | 1.0 | | |
| Muscovite | 5-15 | | a | | | | 40-50 | |
| Chlorite | a | | 5 | | 5-10 | | 2-3 | |
| Biotite | a | | | | | | | |
| Hornblende | | | | | 30-40 | 0.2x1.5 | | |
| Epidote | | | a | | a | | | |
| Magnetite | a | | 45-50 | 0.1 | a | | 2-3 | 0.05-0.4 |
| Calcite | | | | | a | | | |
| Other | | | | | | | | |

Explanation: a = accessory amount; 5 = averages 5%; 25-45 = ranges from 25% to 45%; 0.1 = mineral grain averages 0.1 mm; 0.2x1.5 = mineral grain averages 0.2 mm wide and 1.5 mm long; 0.1-1.0 = mineral grain ranges from 0.1 mm to 1.0 mm; 0.1/2.0 = mineral grain has two distinct sizes.

Table I. Mineral composition of rocks from the Moppin Formation, Rio Arriba County, New Mexico.

TABLE I

| | Cleveland Gulch Unit 3 Phyllite Gray-green Crenulated | | Cleveland Gulch Unit 3 Muscovite Schist (Metarhyolite) | | Cleveland Gulch Unit 4 Amphibolite | | Cleveland Gulch Unit 4 Amphibolite | |
|---|---|---|---|---|---|---|---|---|
| | 5 | | 6 | | 7 | | 8 | |
| | % | mm | % | mm | % | mm | % | mm |
| Quartz | 15-25 | 0.08-1.0 | 40-50 | 0.1/2.0 | 3 | | a | |
| K-Feldspar | 20-25 | | 5-10 | 0.5 | | | | |
| Oligoclase | | | | | 40 | 0.3 | 15 | 0.2 |
| Muscovite | 40-45 | | 40-45 | | | | | |
| Chlorite | 2-3 | 0.05 | | | 10 | | 2 | 0.2 |
| Biotite | | | | | | | | |
| Hornblende | | | | | 35 | 0.3x1.0 | 75-80 | 0.5x1.5 |
| Epidote | | | | | 10 | | 2 | 0.2 |
| Magnetite | a | 1.0-2.0 | a | | 2 | | 2 | 0.2 |
| Calcite | | | | | | | 1 | 0.2 |
| Other | a | | | | a | | a | |

TABLE I

| | Cleveland Gulch Unit 4 Phyllite | | Cleveland Gulch Unit 4 Quartz Biotite Schist | | Cleveland Gulch Unit 4 Quartz Biotite Schist | | Cleveland Gulch Unit 5 Quartz Chlorite Schist | |
|---|---|---|---|---|---|---|---|---|
| | 9 | | 10 | | 11 | | 12 | |
| | % | mm | % | mm | % | mm | % | mm |
| Quartz | 50 | 0.07/0.4 | 45 | 0.05/0.1 | 45 | 0.2 | 15 | 0.2 |
| K-Feldspar | 42 | | | | | | | |
| Oligoclase | 3 | | 40 | | 15-25 | | 25 | 0.2 |
| Muscovite | 3 | | 1-2 | | 2 | | | |
| Chlorite | 1 | | 3 | | 4 | | 30 | 0.2 |
| Biotite | 1 | | 5-10 | | 20-25 | | 2 | |
| Hornblende | | | | | | | | |
| Epidote | | | 2 | | 2 | | 2 | |
| Magnetite | | | | | 2 | | 1 | |
| Calcite | | | | | | | 0-25 | 2.0 |
| Other | | | | | | | a | |

TABLE I

| | Cleveland Gulch Unit 5 Biotite Quartz Schist | | Cleveland Gulch Unit 5 Quartzite | | Cleveland Gulch Unit 5 Phyllite Blue-gray | | Cleveland Gulch Unit 6 Amphibolite | |
|---|---|---|---|---|---|---|---|---|
| | 13 | | 14 | | 15 | | 16 | |
| | % | mm | % | mm | % | mm | % | mm |
| Quartz | 30-35 | 0.2 | 98 | 0.1 | 45-50 | 0.08 | | |
| K-Feldspar | | | | | | | | |
| Oligoclase | 30-35 | 0.2 | | | | | 50 | 1.0-2.0 |
| Muscovite | a | | 1 | | 50 | | a | |
| Chlorite | a | | | | | | 0-10 | |
| Biotite | 30-35 | 0.2 | | | | | 0-5 | |
| Hornblende | | | | | | | 20-40 | |
| Epidote | | | | | | | a | |
| Magnetite | a | | | | 2 | 0.2 | 7 | |
| Calcite | | | | | | | a | |
| Other | | | 1 | | | | | |

TABLE I

| | Tusas Mountain-Burned Mountain Amphibolite | | Iron Mountain Chlorite Schist (Metamorphosed Extrusive) | | Iron Mountain Chlorite Schist (Metamorphosed Intrusive) | |
|---|---|---|---|---|---|---|
| | 17 | | 18 | | 19 | |
| | % | mm | % | mm | % | mm |
| Quartz | a | | 30 | 0.1 | 20 | 0.1 |
| K-Feldspar | | | | | | |
| Oligoclase | 45 | 0.08 | 40 | 0.2 | 45 | 3.0-5.0 |
| Muscovite | | | 5 | 0.05 | 5 | 0.05 |
| Chlorite | | | 15-20 | 0.07 | 20 | 0.05 |
| Biotite | 2-3 | 0.05 | | | | |
| Hornblende | 50 | 0.3 | | | | |
| Epidote | a | | 0-10 | | 5-10 | |
| Magnetite | a | | a | | 1 | |
| Calcite | | | 5 | 0.1-5.0 | 2-3 | |
| Other | | | | | | |

direction and a maximum of 650 feet in an east-west direction. The significance of this area is discussed in detail in a later section.

The amphibolite in this outcrop is slightly schistose and contains anhedral to subhedral highly altered oligoclase, subhedral to euhedral hornblende, fine-grained chlorite (usually replacing hornblende), accessory epidote, magnetite, calcite, and quartz (Table I 3).

Associated with the amphibolite is banded iron formation. This rock is composed of equal amounts of quartz and magnetite, about 5% chlorite, and accessory muscovite and epidote (Table I 2). The iron formation in most places is well banded, the dark bands consisting of magnetite and chlorite with a little quartz, the light bands of quartz with a little magnetite and chlorite (Plate 5C). Thickness of the bands ranges from 0.5 mm to 6.0 mm. Bands in some specimens are continuous over considerable distances, but they pinch and swell in other samples and are not persistent. Magnetite is in euhedral crystals averaging 0.1 mm; anhedral, equant quartz grains average 0.2 mm.

Around 800 feet of interbedded muscovite schist and blue and green phyllite underlie the meta-arkose and iron formation (Fig. 5). Three beds constitute the bulk of the rocks exposed in this unit (3): a blue-green phyllite with prominent magnetite and quartz porphyroblasts, a crenulated gray-green phyllite with tourmaline, quartz, and magnetite porphyroblasts, and a muscovite schist. In this unit, as in all the other units examined, other lithologies no doubt are present within this outcrop distance but are covered by soil.

Blue-green phyllite with prominent quartz and magnetite porphyroblasts was found nearest the overlying meta-arkose beds. Fine-grained muscovite, fine-grained quartz, heteroblastic euhedral to anhedral magnetite, fine-grained chlorite, and highly altered feldspar, which is probably potassium feldspar, comprise this rock (Table I 4). Quartz knots are up to 0.4 mm in diameter. Subhedral to anhedral magnetite averaging 0.05 mm is somewhat elongate parallel to the schistosity; this may have been primary magnetite. Euhedral magnetite up to 0.4 mm in diameter is scattered through the rock and cuts across the schistosity.

Between two muscovite schist beds is a gray-green crenulated phyllite. It is approximately 50 feet thick and contains fine-grained quartz, chlorite, fine-grained muscovite, and highly altered fine-grained feldspar (Table I 5). There are porphyroblasts of quartz up to 1.0 mm, euhedral tourmaline 1.0-2.0 mm long, and euhedral magnetite 1.0-2.0 mm in diameter. Many of the tourmaline, magnetite, and quartz porphyroblasts occupy crests of the crenulations.

Muscovite schists are the most distinctive and persistent of the Moppin beds. Along the bottom of Cleveland Gulch (Fig. 5) they are white with prominent white blastoporphyritic quartz. Near the small lake a half mile east of Cleveland Gulch (Fig. 2) this unit in part changes along strike to a deep red, slightly schistose rock with deep red blastoporphyritic quartz. A few hundred feet to the east of the lake the gray-white muscovite schist is again the only rock type. These rocks were originally rhyolitic flows and tuffs, and in places where they are least metamorphosed flow banding is still evident. Red, slightly schistose rock with blastoporphyritic quartz and feldspar is common in the metarhyolites within the Moppin Formation; however, at Cleveland Gulch the most common metarhyolite is white muscovite schist. Along strike to the northwest of Cleveland Gulch and Bromide Canyon this unit becomes more schistose and even the quartz phenocrysts are destroyed. Yet in the vicinity of Burned Mountain, two miles west of Bromide Canyon (Fig. 2), it is again slightly schistose and deep red in color. All intermediates between red, almost non-schistose metarhyolite and fine-grained muscovite schist exist. Loss of the red color in the quartz phenocrysts is the first change to take place in the alteration of the red metarhyolite to white muscovite schist.

At Cleveland Gulch the composition of the muscovite schist is subhedral microcline; heteroblastic anhedral quartz,

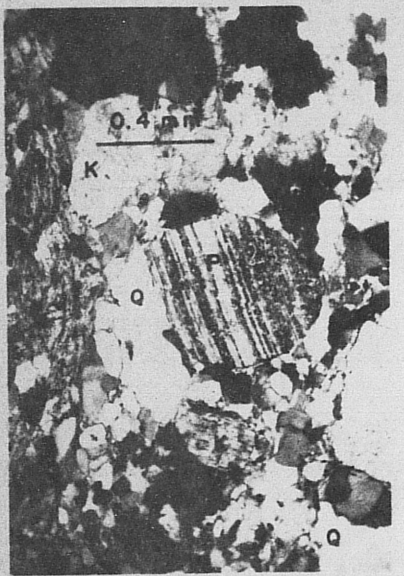
Figure
~~PLATE~~ 5

A. Meta-arkose. Moppin Formation. South Cleveland Gulch. Rounded quartz (Q), plagioclase (P), K-feldspar (K), and muscovite (M). S20 193 41 X-nicols.

B. Meta-arkose (conglomeratic). Moppin Formation. South Cleveland Gulch. Rounded, slightly crushed, quartz pebble (Q), matrix of fine-grained rounded quartz, plagioclase (P), muscovite (M), and euhedral magnetite (Ma). S20 193 42 X-nicols.

C. Banded iron formation. Moppin Formation. South Cleveland Gulch. Quartz-rich band (Q) and magnetite-rich band (Black) with chlorite (C) concentrated in magnetite-rich band. S20 193 61 Plane light.

D. Metarhyolite. Moppin Formation. Cleveland Gulch. Quartz (Q), muscovite (M). S20 193 4 X-nicols.

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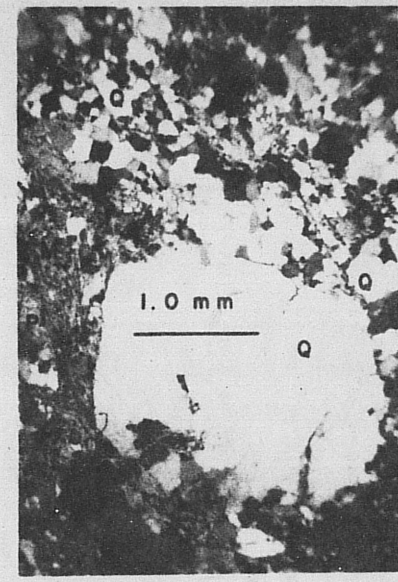


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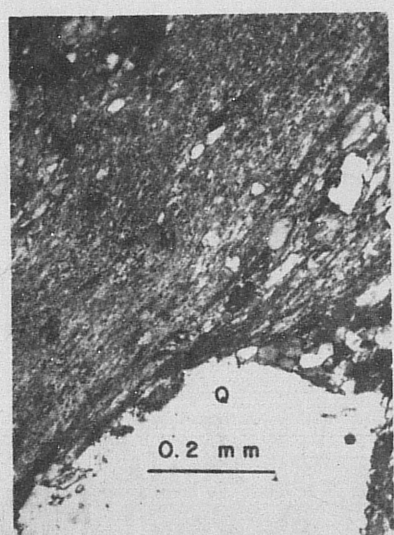



D

0.1 mm in diameter and phenocrysts around 2.0 mm in size; fine-grained muscovite; and a trace of magnetite (Plate 5D) (Table I 6). The rock is slightly banded with muscovite-rich and quartz-rich bands 0.5 mm to 1.0 mm in thickness. Many samples were veined with quartz and a little tourmaline.

A sequence of interbeds of amphibolite, biotite schist, phyllite, and muscovite schist comprises the next unit (4) stratigraphically below the phyllites and muscovite schist. This unit ranges in thickness from around 200 feet to over 1,300 feet. Its thickness has been increased north and east of Cleveland Gulch (Fig. 5) by intrusion of small dikes and sills of granodiorite.

Amphibolites in this unit vary considerably in texture and composition. One sample contained 35% euhedral hornblende, highly altered subhedral oligoclase, chlorite, fine-grained epidote, quartz, euhedral magnetite, and garnet (Table I 7). Banding, which is characteristic of these amphibolites, is formed by alternation of feldspar-rich light layers and amphibole-rich dark layers. Bands average 5 mm and they converge and swell. A couple of hundred feet down section from the aforementioned amphibolite is another outcrop of amphibolite which is greenish-black, only slightly schistose, and not banded. This rock has 75-80% euhedral poikiloblastic hornblende, most of which is randomly oriented or is sub-parallel to the schistosity (Table I 8). Oligoclase is mostly untwinned. The remainder of the rock is made up of chlorite, subhedral to euhedral magnetite, epidote, quartz, garnet, and calcite. Hornblende is apparently being replaced by chlorite. Accessory apatite is found in some specimens.

Other amphibolites in this unit have hornblende percentages intermediate between the two described above. Due to the random orientation of the hornblende crystals and of the partial chlorite pseudomorphs, schistosity is in general poorly developed.

Another lithology in this unit is light tan phyllite. It is composed of heteroblastic, anhedral, equant quartz and highly altered microcrystalline groundmass of probable potassium feldspar and oligoclase. About 3% is 0.1 mm microcline and oligoclase. Fine-grained chlorite, muscovite, stubby biotite, magnetite, and garnet constitute the rest of the rock (Table I 9).

Quartz biotite schists are found in two places in this unit. The upper bed is banded, the lower one is not banded. Quartz accounts for 45% of the rock; in one bed it is heteroblastic with particles 0.05 mm and 0.1 mm in diameter and in the other bed it averages 0.2 mm in diameter. Highly altered subhedral prismatic oligoclase constitutes 40% of the upper bed but only 15% of the lower bed. The remainder of the constituents in the rock are fine-grained chlorite, muscovite, magnetite, epidote, and poikiloblastic biotite, 5-10% in the upper bed and 20-25% in the lower bed (Table I 10 and 11).

Near the bottom of this unit are a few layers of banded iron formation. Most of the specimens have equal amounts of anhedral quartz and euhedral magnetite, both averaging about 0.1 mm. Biotite, chlorite, epidote, and tourmaline are present up to a total of 10%. Continuity of the bands is poor to fair; many pinch out or swell in a distance of a few centimeters. Thickness of the bands ranges between 0.5 mm and 5.0 mm. The lowermost amphibolite bed mentioned above separates this iron formation from the underlying unit which contains most of the banded iron formation north of Cleveland Gulch (Fig. 5).

The unit (5) with the large amount of banded iron formation is highly variable in thickness, ranging from a feather edge to about 450 feet. Other than the iron formation, the unit contains biotite-quartz schist, chlorite schist, phyllite, and amphibolite. These lithologies are not unique to this unit because the boundaries of the unit are based on the presence of banded iron formation and not on the lithology of the contained rocks. Iron formation, which accounts for only a small part of the thickness of the unit, apparently cuts across the strike of the beds (Fig. 5). The rocks within this unit are described in more detail in the section on the iron deposits.

A few of the interbeds with the iron formation north of Cleveland Gulch should be mentioned. An interesting bed of quartz-chlorite-calcite schist is found near the base of the unit. Anhedral, equant quartz; oligoclase, mostly untwinned; chlorite; anhedral calcite in rounded blebs of about 2.0 mm diameter; biotite, which is being replaced by chlorite; and accessory magnetite, garnet, and epidote comprise this rock (Table I 12). Calcite blebs appear to have been amygdules originally.

A well banded greenish-yellow biotite-quartz schist similar to schists in the overlying unit (4) is found in the banded iron formation unit (5). It has equal amounts of anhedral slightly elongate quartz, elongate biotite, and anhedral highly altered oligoclase laths (Table I 13). Accessory subhedral to euhedral magnetite, chlorite, and muscovite are also present. Biotite is responsible for the schistosity and is also present as cross-cutting porphyroblasts. Banding is formed by alternation of biotite-rich and quartz-rich layers. The bands range between 1.0 mm and 5.0 mm in thickness. Continuity of the individual bands is fair to good; the bands usually converge and swell but seldom pinch out in a single hand specimen. Some of the chlorite replaces biotite.

Near the above biotite-quartz schist on the main traverse are a few feet of fairly massive quartzite. This bed is composed of 98% equant quartz grains (Table I 14). The remainder of the rock consists of muscovite and specularite. Muscovite appears to define crude bands in the rock. This bed may be contiguous with a well banded quartz muscovite bed a few hundred yards to the west. There the rock is composed of bands of coarse quartz alternating with bands made up of fine quartz, muscovite, chlorite, biotite, magnetite, and calcite. Both of these beds are relatively thin and discontinuous and are believed to have formed in a small shallow local fresh water basin. These beds are discussed in the section on the origin of the banded iron formation.

A blue-gray phyllite is another distinctive lithology noted in the iron formation unit. It is composed mostly of anhedral equant quartz, fine-grained muscovite, and euhedral magnetite (Table I 15). Amphibolites present within the iron formation unit are much like those in the overlying unit.

Banded iron formation is similar to those mentioned earlier, but different textural and compositional variations are evident. The iron formation rocks are fine-grained and composed of 45-56% magnetite and 45-50% quartz. Besides quartz and magnetite, there are minor amounts of chlorite, apatite, biotite, muscovite, garnet, epidote, and calcite in some specimens. Usually chlorite, biotite, muscovite, and epidote are almost exclusively within the magnetite-rich bands (Plate 6A). When some of these minerals are in the quartz-rich bands, they are in lesser amounts than in the adjacent magnetite-rich bands. Only apatite is more characteristic of the quartz-rich bands. Banding continuity varies from poor to good and thickness of the bands ranges from less than 1.0 mm to over 5.0 mm. Often a single band will pinch and swell within a few centimeters, but in the majority of the cases the bands are fairly continuous over the length of a hand specimen. In some specimens a few bands cut others. In a few specimens secondary euhedral albite cuts and replaces quartz and magnetite bands.

Below the iron formation unit is a sequence of biotite schist and amphibolite ranging from 50 to 600 feet thick. In this unit (6) the amphibolite is similar to those described above. At least one bed is well banded with poikiloblastic hornblende-rich and feldspar-rich bands. Slightly elongate oligoclase forms 50% of the rock, whereas hornblende ranges from 20-45% depending upon the amount of replacement by biotite and chlorite (Table I 16). Euhedral magnetite and accessory epidote, calcite, and muscovite are also present. Some of the coarse-grained amphibolites might have been sills or

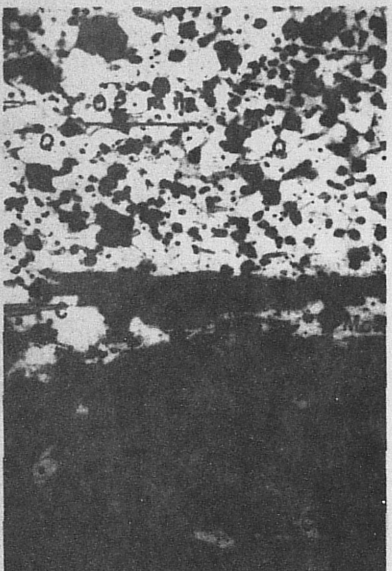
~~PLATE~~ Figure 6

A. Banded iron formation. Moppin Formation. West Cleveland Gulch. Dark band made up of magnetite (Ma) and chlorite (C), light band made up of quartz (Q) and a little magnetite and chlorite. 21 18 9 Plane light.

B. Biotite schist. Moppin Formation. Cleveland Gulch. Plagioclase (P), chlorite (C), and biotite (B). S20 193 26 Plane light.

C. Quartz-muscovite schist (metarhyolite). Moppin Formation. Cleveland Gulch. Blastoporphyritic plagioclase (P), in fine-grained matrix of quartz (Q), muscovite (M), plagioclase, K-feldspar, and magnetite (Ma). S20 193 31 X-nicols.

D. Quartz-muscovite schist (metarhyolite). Moppin Formation. Cleveland Gulch. Blastoporphyritic plagioclase (P) in fine-grained matrix of quartz (Q), muscovite (M), and plagioclase. Plagioclase is being replaced along cleavage planes by an unidentified mineral. S20 193 31 X-nicols.

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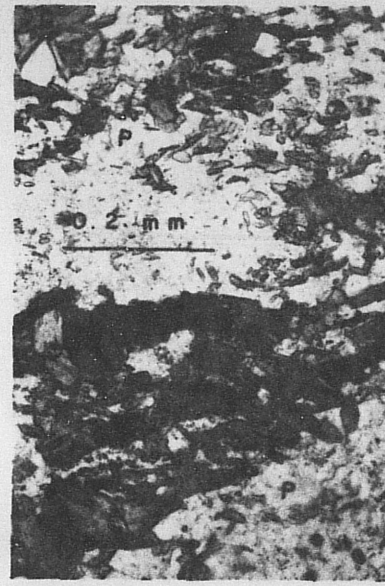


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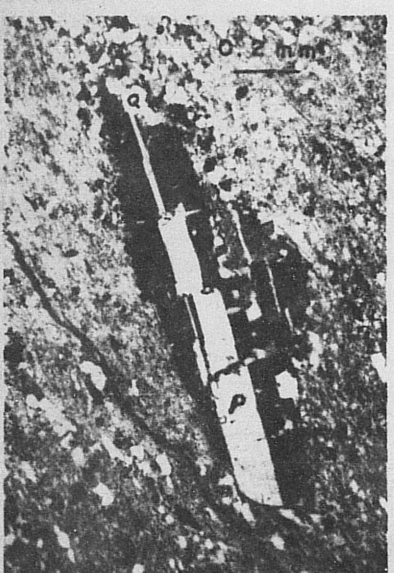



C


D

dikes since they appear to be thin in comparison with adjacent thick-bedded but fine-grained rocks.

Biotite schists are similar in composition and texture to those in the first biotite schist-amphibolite unit (unit 4). They are somewhat coarser, with the biotite and feldspar around 0.3 mm to 0.5 mm. Besides those minerals listed previously, these rocks have accessory pyrite up to 1% and calcite up to 10%. Chlorite replaces some of the biotite (Plate 6B).

Below this second biotite schist-amphibolite unit (unit 6) is a muscovite-quartz schist unit (7), which ranges from a feather edge to 200 feet thick. Feldspar and quartz are heteroblastic, each found as blastoporphyries (Plate 6C) 0.7 mm to 5.0 mm, and in finer grains around 0.08 mm to 0.2 mm. Oligoclase is replaced along cleavages in some specimens (Plate 6D). This unit was originally rhyolite and resembles the metarhyolite described from the bottom of Cleveland Gulch (Fig. 5). In contrast to muscovite-quartz schists already described, this unit does not change to red metarhyolite along strike.

Between this metarhyolite and a large more or less concordant granodiorite body (Fig. 5) is a unit (8) of biotite schist and amphibolite which is about 800 feet thick. Except for the coarser texture, the biotite schist is not different from those previously described. Highly altered plagioclase is anhedral, prismatic, and up to 1.0 mm in size, the quartz ranges to 0.5 mm, and the biotite ranges up to 2.0 mm.

The amphibolites are also much like those previously described but are coarser. Most beds display banding of hornblende-rich and feldspar-rich layers. There is up to 15% quartz present; this may be secondary quartz furnished by the granodiorite intrusion which forms the northern limit of this unit (Fig. 5). The proximity of this fairly large intrusive body may account for the coarser grain of the biotite schist and amphibolite.

Between the granodiorite and the main body of the Tusas intrusions, outcrops are scarce (Fig. 5). The area between the granite and the granodiorite is from 800 to 2,500 feet wide. Amphibolite and quartzite were the only two lithologies found in this unit (9). In hand specimen the quartzite is coarse-grained and somewhat conglomeratic. It contains a little chlorite and up to 20% calcite. Amphibolites are much like the ones described previously, although in general they are not very schistose or banded. In hand specimen most look almost phanerocrystalline with little lineation noticeable. They may have been intrusive rocks, but most likely the texture and composition reflect their position between two large intrusions.

One amphibolite in this area is different enough to deserve separate description. It is about 94% actinolite, 5% chlorite, and 1% euhedral magnetite. Chlorite and probably the magnetite were derived from the alteration of actinolite. Actinolite is euhedral and averages about 3.0 mm long (Plate 7A). The rock may have been a product of metamorphism of a mafic igneous rock, but possibly reflects hydrothermal action on a rock of less mafic composition. Metamorphism associated with the Tusas Granite is discussed later.

These amphibolites and quartzite are the oldest rocks near Cleveland Gulch. They are probably close to the axial plane of the Hopewell Anticline. However, this is not certain because the Tusas batholith has apparently obliterated most of the northern flank and some of the southern flank of the anticline.

All the aforementioned units strike east from Cleveland Gulch and extend for about one and a half miles before they are covered by Cenozoic rocks (Fig. 2). Little lithologic change takes place in this direction. The next good exposure west of the Cleveland Gulch section is at Bromide Canyon, about one mile away. Here the lithologies are of a different metamorphic grade. Also, the rocks of this canyon have been mineralized by hydrothermal fluids, and this possibly accounts for some compositional variations.

Bromide Canyon.--Near the Tusas ~~Granite~~ contact at the upper end of the northeast fork of Bromide Canyon (Fig. 2), there is a coarse-grained amphibolite containing no feldspar and 30-40% epidote. From here, the Tusas-Moppin contact swings north paralleling the northwest fork of the canyon, and the main granite body apparently does not exert much influence on those rocks in the lower canyon which correspond to the section at Cleveland Gulch (Fig. 2).

The rocks in Bromide Canyon from the contact of the granite to the base of the Ortega Formation, a distance of 4,500 feet, are similar to those at Cleveland Gulch. There is much less amphibolite, and the iron formation pinches out before reaching the canyon as does the lower distinctive metarhyolite bed. Most of the rocks are medium-grained chlorite-quartz-biotite-feldspar schists with accessory magnetite, garnet, epidote, and considerable calcite. These are extensions of the chlorite and biotite schists and amphibolites of the Cleveland Gulch section. Muscovite schist with poikiloblastic biotite (Plate 7B) is probably an extension of the metamorphosed silicic tuffs from the Cleveland Gulch area.

An amphibolite bed near the northern contact of the linear intrusive body (Fig. 2) is coarse-grained, and the presence of hornblende reflects the effect of the intrusion. Near the southern contact of the intrusion, chlorite in the schists is pseudomorphic after hornblende. The upper metarhyolite of Cleveland Gulch is represented here by a fine-grained, light gray muscovite schist. Most of the units in Cleveland Gulch between the upper metarhyolite and the base of the Ortega Quartzite pinch out east of Bromide Canyon. In Bromide Canyon the upper metarhyolite is within 100 to 200 feet of the feldspathic quartzite of the Ortega Quartzite.

In summation for Bromide Canyon: the amphibolites are absent except near the intrusions; the small dikes and sills in some units at Cleveland Gulch are not apparent here; the linear intrusion of granodiorite extends into the canyon; and it is possible to trace only one stratigraphic unit in the Moppin Formation from Cleveland Gulch into the canyon--the upper metarhyolite bed. Chlorite, biotite, and feldspar schists characterize the Moppin rocks in Bromide Canyon.

<u>Tusas Mountain-Burned Mountain</u>.--One to two miles northwest of Bromide Canyon, between Tusas Mountain and Burned Mountain, outcrops are scarce; even the

distinctive metarhyolite beds were difficult to find. Vegetation is very dense on the steeper slopes, and the numerous meadows have few outcrops. Due to these difficulties, the geology on Figure 2 for the area between Tusas Mountain-Bromide Canyon and Burned Mountain was mapped on aerial photographs by projections from isolated outcrops to isolated outcrops.

Few different lithologic types were found in the area between Tusas and Burned mountains. Most samples are similar in composition and texture to those at Cleveland Gulch and Bromide Canyon. Near the eastern end of the prong of Moppin Formation extending into the Tusas River Valley (Fig. 2) is a fine-grained black amphibolite. This rock shows no noticeable schistosity in hand specimen and little in thin section. It is composed of euhedral, prismatic hornblende; highly altered, untwinned, anhedral, equant oligoclase; and prismatic biotite which replaces the hornblende (Table I 17; Plate 7C). Accessory quartz, magnetite, and epidote are present. Several other amphibolites crop out along the western margin of the Tusas intrusives ~~Granite~~ (Fig. 2). These are usually coarser grained than the one described above and in general resemble those at Cleveland Gulch.

Interspersed with the amphibolites are chlorite, Figure ~~PLATE~~ 7

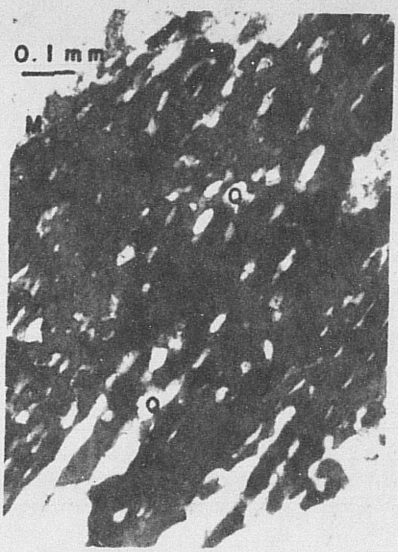
A. Amphibolite. Moppin Formation. North of Cleveland Gulch. All actinolite. 20 194 4 X-nicols.

B. Muscovite-biotite-quartz schist. Moppin Formation. Bromide Canyon. Poikiloblastic biotite (B), quartz (Q), and muscovite (M). 21 19 12 Plane light.

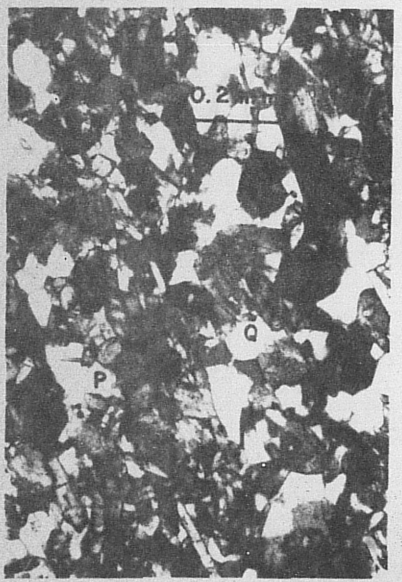
C. Amphibolite. Moppin Formation. West of Tusas Mountain. Hornblende (H), quartz (Q), and plagioclase (P). 21 20 25 X-nicols.

D. Greisen. Moppin Formation-Tusas Intrusives ~~Granite~~ contact. West flank of Tusas Mountain. Euhedral quartz (Qe), fluorite (F), garnet (G), pyrite (Py) in a matrix of quartz and muscovite. 21 20 15 X-nicols.


A




B

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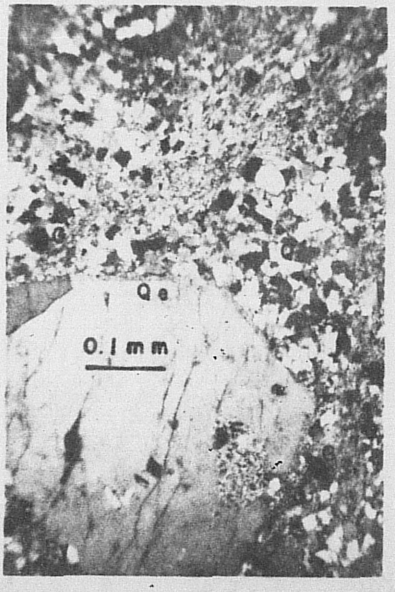



D

muscovite, and biotite schists. At the granite-Moppin Formation contact on the western side of Tusas Mountain is (Fig. 2) an interesting greenish-yellow muscovite schist. Two types of quartz are present; one is anhedral, equant, averages 0.2 mm in diameter, and is believed to be quartz that was present in the original granite. The other is euhedral, about 4.0 mm in size, and is believed to be hydrothermal quartz (Plate 7D). Quartz accounts for 55% of the rock; 3% anhedral microcline, 35% elongate 0.2 mm muscovite, 2% euhedral fluorite, and accessory magnetite, pyrite, and epidote are also present. This appears to be an altered granite, a greisen, with fluorite, garnet, and crystalline quartz added and much of the microcline altered to muscovite. Listed in Table IV are the elements found in a semi-quantitative spectrographic analysis of the rock. This sample was tested for only Be, B, Li, and Sn. Be content was greater than 5 ppm, but less than 20 ppm, B content was around 35 ppm, Li was not detected in the sample, and Sn was between 100 and 200 ppm. It is evident that these elements, expectable in greisens, are here only in very small amounts.

Other rocks between Burned Mountain and Tusas Mountain are similar to those in Bromide Canyon. Most are biotite-chlorite-feldspar-quartz schists (Plate 8A). Usually biotite accounts for 25-35%, quartz 20-30%, heteroblastic feldspar (usually oligoclase) 25-50%, chlorite 5-10%, epidote 2-5%, and magnetite 5-10%. In many specimens the quartz appears as vesicle fillings up to 3.0 mm in size. It is not known if the quartz was the original material of the amygdules or if it replaced some other mineral.

At Burned Mountain, due to faulting and intrusion of ~~the~~ Tusas Granite, very little of the Moppin Formation is preserved. Small outcrops of fine-grained muscovite and chlorite schists were noted. The upper metarhyolite of Cleveland Gulch, however, here reaches a minimum thickness of 200 feet and may attain a maximum of around 400 feet. It is no longer a white muscovite schist as it was at Cleveland Gulch and Bromide Canyon. At Burned Mountain the metarhyolite is red, blastoporphyritic, slightly schistose, and shows What appears to be flow banding in many places.

Placer Canyon.--About two miles west of Burned Mountain in Placer Canyon (Fig. 2) is a thick sequence of metarhyolite. A few beds of chlorite-biotite schists are interbedded with about 3,000 feet of metarhyolite and/or metamorphosed welded rhyolitic tuff. The metarhyolite has been somewhat altered by hydrothermal fluids which were responsible for the gold-quartz mineralization in the canyon.

Metarhyolites in Placer Canyon are composed of 15-30% microcline, 30-35% albite-oligoclase, 30-35% quartz, 5-30% muscovite, with accessory magnetite, pyrite, calcite, chlorite, and epidote. Usually the feldspars are 0.5 mm to 2.0 mm in long dimension; quartz ranges from 0.01 to 1.5 mm in size (Plate 8B and C). The larger grains of feldspar and quartz are probably relict phenocrysts. Enough muscovite is present to make all but the uppermost bed schistose. This uppermost bed corresponds to the uppermost metarhyolite at Burned Mountain, Bromide Canyon, and Cleveland Gulch. To the east toward Burned Mountain and to the west most of these metarhyolite beds pinch out over relatively short distances. This suggests that a volcanic center may have been located in this vicinity during the Precambrian.

Few outcrops of Moppin rocks are found west of Placer Canyon. The top of the Moppin Formation swings around the nose of the Hopewell Anticline about 3 miles northwest of Placer Canyon. The uppermost metarhyolite bed is virtually the only Moppin lithology exposed in this area. On the northern flank of the Hopewell Anticline, a short distance south of Jawbone Peak, Moppin

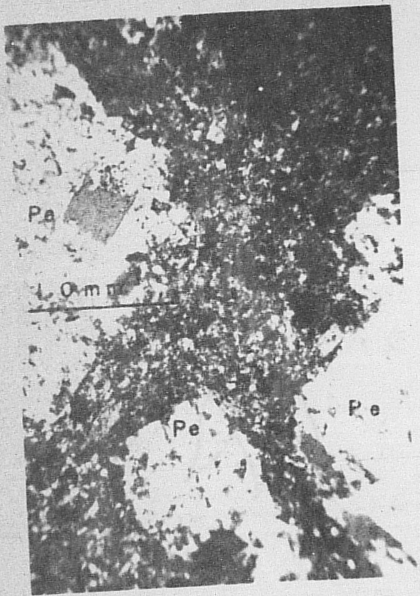
Figure
~~PLATE~~ 8

A. Biotite-quartz-plagioclase schist. Moppin Formation. Between Burned Mountain and Tusas Mountain. Probably a metamorphosed porphyritic dike. Large crystals of plagioclase (Pe) in a matrix of plagioclase, quartz, biotite, magnetite, and epidote. 21 20 8 X-nicols.

B. Quartz-muscovite schist (metarhyolite). Moppin Formation. Placer Canyon. K-feldspar (K) and quartz (Q). 12 109 4 X-nicols.

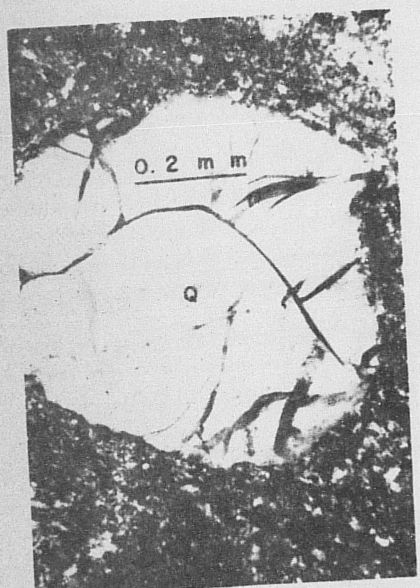
C. Quartz-muscovite schist (metarhyolite). Moppin Formation. Placer Canyon. Large, slightly crushed quartz (Q) grain in a fine-grained matrix of quartz, feldspar, and muscovite. 12 109 5 X-nicols.

D. Chlorite schist (meta-tuff). Moppin Formation. Iron Mountain. Muscovite (M), magnetite (Ma), quartz (Q), chlorite (C), plagioclase (P). 22 203 5 X-nicols.




A


B

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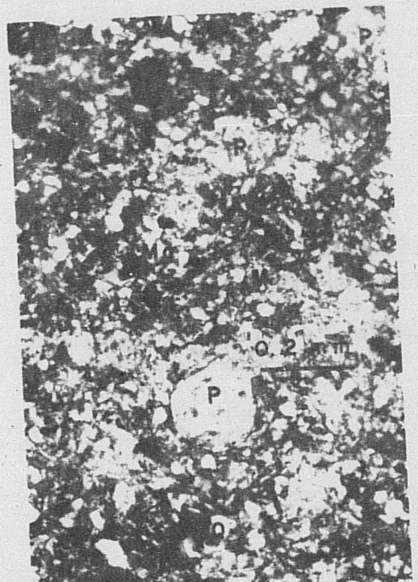



D

Formation schists and metarhyolite again crop out. The metarhyolite is the uppermost bed in the Moppin Formation here and is about 200-300 feet thick. Below the metarhyolite are several hundred feet of schists and phyllites.

<u>Iron Mountain</u>.--Iron Mountain is on the southern slope of Jawbone Mountain. The relative absence of heavy vegetation on the western end of Iron Mountain permitted plane table mapping even though outcrops are not numerous (Fig. 4). The detailed mapping was helpful not only in the elucidation of the origin of the iron formation but also in the determination of the mode of deposition of the Moppin rocks and their subsequent geologic history. Plate 3C shows a typical outcrop of Moppin Formation on Iron Mountain. This exposure is characteristic of Moppin rocks from Cleveland Gulch to Iron Mountain; however, at Iron Mountain outcrops are more numerous than at most other localities.

Figure 4 is an outcrop geologic map of the western half of Iron Mountain. All contacts are visible in the field, except for part of the aplite dike and part of the iron deposit, both of which are indicated by dashed lines on the map. The dashed lines represent covered contacts determined by connecting outcrops which are too small to map at a scale of 1 inch to 50 feet. There are enough of these outcrops to permit good control of the outcrop pattern, but since it was not possible to walk the contact, it has been represented by dashed lines.

Since the rocks on Iron Mountain are representative of the Moppin Formation and are fresh and have more or less continuous outcrops, it is possible to deduce the history of deposition of Moppin rocks in more detail than was possible at other localities. These outcrops show the area was covered by silicic to intermediate tuffs and flows and then intruded by dikes and sills of silicic to intermediate composition. About half of the rocks mapped on Iron Mountain were originally shallow intrusions.

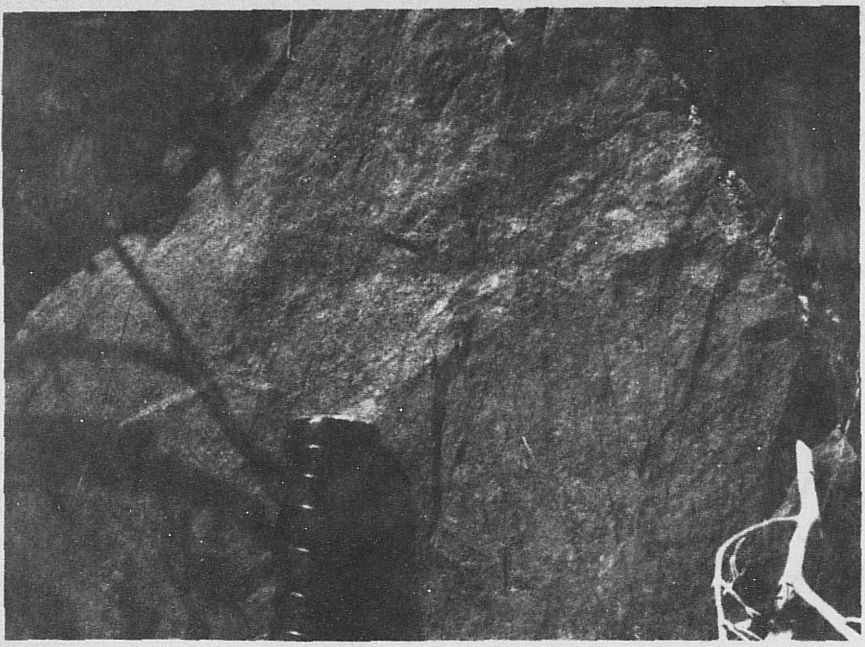
In general the metamorphosed extrusive rocks are medium to dark green, fine-grained, chlorite-quartz-feldspar-epidote-calcite schists (Plate 8D). Almost all outcrops are similar in texture and composition (Plate 9A and B). Most of the grains are less than 0.1 mm in size, but some rocks are more or less porphyritic with a few grains of 0.5 mm to 2.0 mm. The coarser of these may represent flow rocks, but most of the rocks appear to have been tuffs.

Most outcrops have essentially the same mineralogical composition; however, there is considerable range in mineral percentages. Quartz ranges from 5 to 50%,

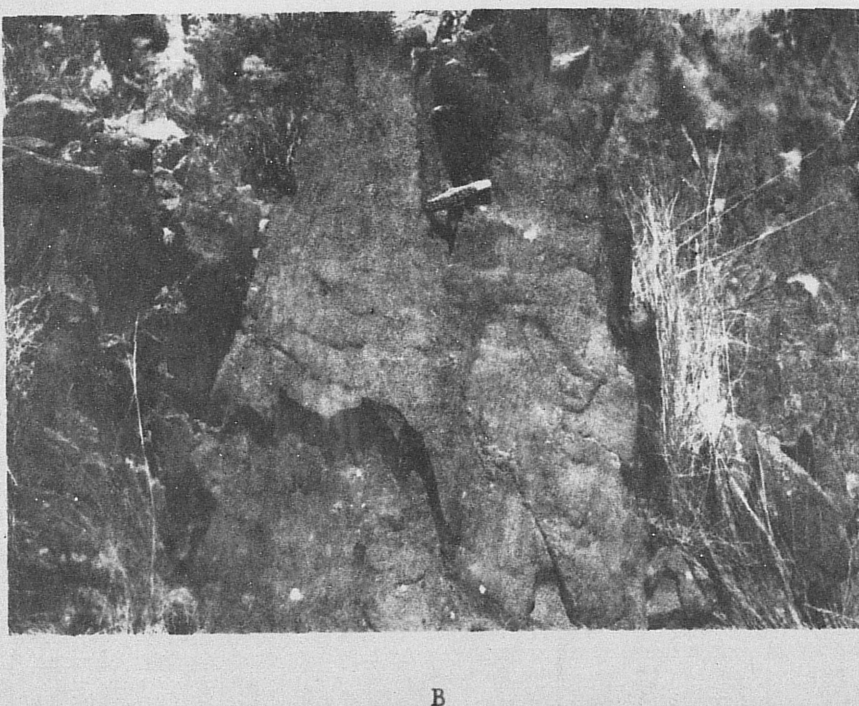
FIGURE
~~PLATE~~ 9

A. Moppin Formation. Iron Mountain. Chlorite schist (meta-tuff). Note fine banding.

B. Moppin Formation. Iron Mountain. Chlorite schist (meta-tuff). Note fine banding.



A



B

Na oligoclase from a trace to 65%, chlorite from 5 to 45%, and magnetite from a trace to 7% (exclusive of the iron formation). Some minerals are common but are not found in every thin section. These are epidote, which ranges up to 40%, calcite from 0 to 35%, and muscovite from 0 to 25%. Other minerals are found in only a few sections. These are biotite (0-3%), hornblende (0-20%), microcline (0-15%), and traces of tourmaline, garnet, and pyrite. Many schist specimens have a crude banding formed by partial segregation of dark and light minerals. The iron formation is composed almost entirely of magnetite and quartz bands. Where these bands are traceable they grade into chlorite-rich and quartz-rich bands.

The beds which enclose the largest body of iron formation on Iron Mountain (Fig. 4) are typical of the metamorphic rocks derived from extrusive rocks. They contain highly altered anhedral oligoclase, anhedral equant quartz, chlorite, muscovite, clusters of equant epidote, equant locally rounded calcite up to 5.0 mm in diameter, and a trace of magnetite (Table I 18; Plate 8D). One thin section of this bed shows large isolated rounded calcite which represents filled vesicles (Plate 10A). It is thought that calcite formed the original amygdules. Many outcrops on Iron Mountain contain rounded isolated grains of calcite or other minerals and these are believed to represent filled vesicles. Some of these may be replaced amygdules.

Metamorphosed intrusive rocks on Iron Mountain are varied in composition and structural form. They range from what appears to be a small "pebble dike" to dikes and sills. However, most are dikes which cut across the strike of the intruded rocks at low angles.

Minerals present in all thin sections of the metamorphosed intrusions studied on Iron Mountain are albite-oligoclase (3-65%), quartz (5-25%), chlorite (10-25%), magnetite from a trace to 3%, and epidote (5-40%). Other minerals are found in a few sections; of these calcite is the most common ranging from 0-20%. Less common are microcline (0-5%) and biotite (0-10%). Biotite becomes more common as the contact with the Tusas Intrusive ~~Granite~~ is approached. This contact is about 300 yards south of the large outcrop of iron formation on Iron Mountain (Fig. 2).

The northernmost metamorphosed intrusive body on Figure 4 contains highly altered subhedral prismatic oligoclase from 3.0 mm to 5.0 mm in length, anhedral equant quartz, chlorite, muscovite in elongate sheaves, euhedral magnetite, epidote replacing plagioclase, and anhedral equant calcite (Table I 19). Muscovite may be an alteration product of potassium feldspar, but no evidence of this was found. The plagioclase grains stand out as large white crystals on the surface of a weathered rock which was probably a hypabyssal porphyritic dike of intermediate composition.

In the field these rocks look like porphyritic dikes cutting the finer grained chlorite schists. In many places fragments of fine-grained chlorite schists are found within the intrusive bodies (Fig. 4). Schistosity of the intrusions is invariably parallel to the walls of the intrusion. The presence of xenoliths and the orientation of the schistosity suggest that the intrusive bodies were emplaced after the initial folding and low grade metamorphism.

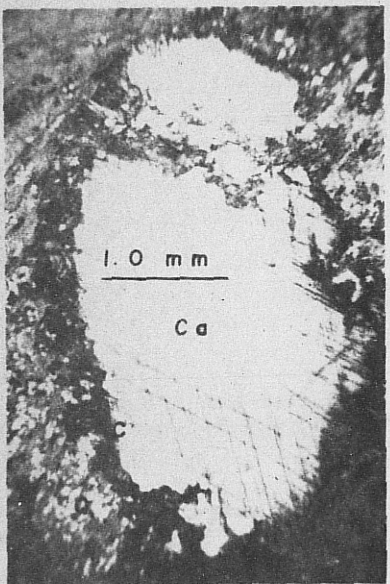
At one small locality near the western end of Iron Mountain is an interesting quartz-pebble-epidote-chlorite schist. In the field it has the appearance of a conglomerate with well-rounded pebbles of quartz, epidote, and feldspar set in a matrix of chlorite and fine-grained quartz, epidote, and feldspar (Plate 10C). However, in thin section the rounded pebbles are found to be mostly lithic fragments and much of the quartz and feldspar has been replaced by epidote (Plate 10B). These rounded to subrounded clasts, which account for 30% of the rock, range from 1.0 mm to greater than 10 mm, but average around 4.0 to 5.0 mm. The matrix

Fig. ~~PLATE~~ 10

A. Chlorite schist with relict amygdule. Moppin Formation. Iron Mountain. Chlorite (C), calcite (Ca), quartz (Q). P22 203 I25 Plane light.

B. Chlorite schist (metamorphosed pebble dike). Moppin Formation. Iron Mountain. Epidote (E) is probably a replacement of a lithic fragment, matrix of fine-grained chlorite, quartz, and plagioclase. P22 203 D83 Plane light.

C. Moppin Formation. Iron Mountain. Chlorite schist (metamorphosed pebble dike). Note the rounded pebbles.




A

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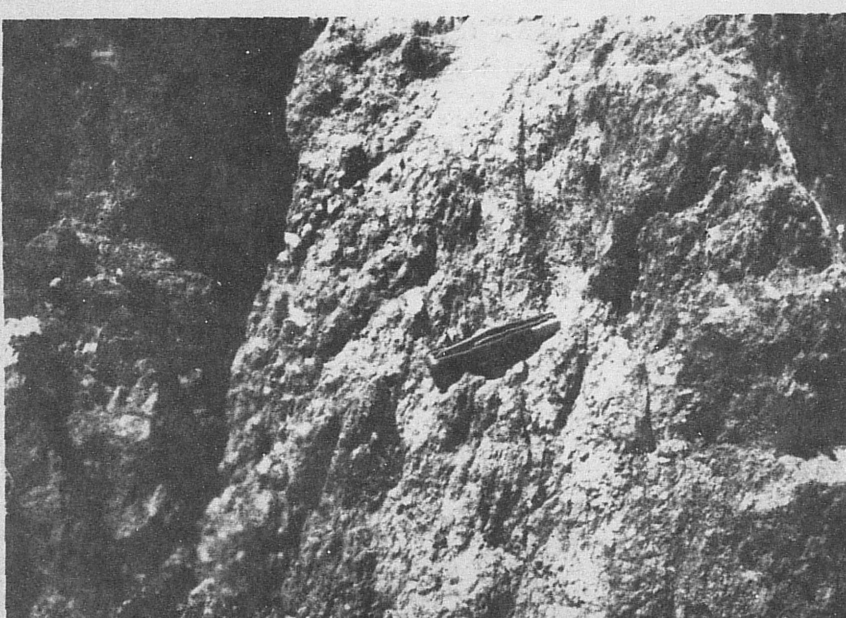


C

looks much like the other metamorphosed intrusive rocks. Since this body appears to be cross-cutting, it is probably a pebble dike (used in the sense of a "pipe"), rather than a conglomerate or agglomerate.

It is often difficult to distinguish between metamorphosed coarse-grained extrusives and intrusives in the field. If it is known that there exist metamorphosed intrusives in addition to the metamorphosed extrusives throughout the Moppin Formation, then, because outcrops are few and poor, it would be doubly difficult to understand the stratigraphy and structure. However, if it is possible to identify these rocks, many strange and seemingly contradictory data can be understood and a more precise picture can be drawn. At Cleveland Gulch metamorphosed intrusions, probably the same age as those on Iron Mountain, plus granitic intrusions which are apophyses of the Tusas ~~Granite~~ Intrusives cause the pronounced thickening in the Moppin rocks (Fig. 2).

Although the present study did not cover La Jarita Mesa in detail, it is suggested that the amphibolites there may be metamorphosed hypabyssal intrusives. If this is so, it would not be possible to utilize them as datum planes in the interpretation of structural relationships as has been attempted (Barker, 1958). Metarhyolites of La Jarita Mesa offer the best stratigraphic control for ascertaining structural relationships of the units there as they do in the Moppin Formation from Cleveland Gulch to Jawbone Mountain.

Thickness

Due to the Tusas intrusion it is difficult to determine the thickness of the Moppin Formation. The fact that the syncline northwest of Jawbone Mountain is slightly asymmetrical with the steeper limb on the north ~~suggests~~ indicates that the Hopewell Anticline ~~may~~ is also ~~be~~ slightly asymmetrical. By projecting the axial trace from the nose of the anticline, which is about 3 miles west of the border of Figure 2, into the area, it is found to pass north of Tusas Mountain. Hopewell Anticline appears to be a simple fold, and there is no reason to believe the axial plane is warped. Measuring the distance from the Ortega Quartzite-Moppin Formation contact at Cleveland Gulch to the projected trace of the axial plane perpendicular to the strike of the Moppin rocks gives a figure of about 12,000 feet. It is difficult to estimate the amount of expansion that has resulted from small intrusions associated with the Tusas ~~Granite~~ Intrusives or how much, if any, flowage there has been along the flanks of the anticline. It is also impossible to determine whether or not there has been omission or repetition of strata due to faults. Since it is impossible to make allowances for these phenomena, a maximum and minimum estimate of thickness will be given.

Just (1937, p. 42) estimated the thickness of the Moppin Formation at about one and one half miles, but recognized that it might be greater. In the Picuris Range, he estimated the rocks to be about three fourths of a mile thick (1937, p. 21). Barker (1958, p. 23) measured the Moppin Formation to be about 3,000 feet on the map, but believed it might be considerably thicker. Between the linear apophysis of the Tusas Intrusives ~~Granite~~ (Fig. 2) and the main body of Tusas Intrusives ~~Granite~~ north of Cleveland Gulch is about 2,500 feet of poorly exposed Moppin rocks. It is difficult to make an estimate of the true thickness of these rocks caught between two intrusions. The minimum thickness of the Moppin Formation would be that from the Ortega Quartzite-Moppin Formation contact to the Tusas-~~Granite~~-Moppin ~~Formation~~ contact, using the lowest possible figure for the thickness of the beds between the linear apophysis and the main intrusion. The maximum thickness of the Moppin Formation would be measured from the Ortega Quartzite-Moppin Formation contact to the projected axial trace north of Tusas Mountain. Therefore, a minimum thickness would be about 3,500 feet, and a maximum thickness would be about 12,000 feet.

## Origin

Just (1937, p. 10) and Barker (1958, p. 23) concluded that the Moppin Formation consists mainly of metamorphosed igneous flows, mostly andesitic or basaltic, but with some interbedded metamorphosed sediments. Barker (1958, p. 23) stated that the schists might represent tuffs. Both recognized and named the extensive beds of distinctive metarhyolite. The present work is in complete agreement with their hypothesis of an igneous origin for Moppin rocks. However, more emphasis should be placed on the presence and extent of the metamorphosed silicic tuffs and flows. There are many of these present, but they are not as distinctive as the red and white muscovite schists which Just named Vallecitos Rhyolite and Barker named Burned Mountain Metarhyolite. Rocks of the Moppin Formation can be categorized into seven broad types: muscovite schist, phyllite, amphibolite, biotite-quartz schist, chlorite-feldspar-calcite schist, chlorite-quartz-feldspar schist, and quartzite.

The amphibolites and chlorite-feldspar-calcite schists are believed to be metamorphosed tuffs and flows of andesitic to basaltic composition. This origin is suggested by the presence of relict ophitic textures, amygdules, relict porphyritic textures, and the chemical composition (Table II). Fine-grained rocks interspersed with coarse-grained rocks of similar composition were probably tuffaceous, and the coarse-grained rocks may have been flows. Absence of much quartz indicates these rocks were relatively mafic.

Muscovite schists and phyllites are metamorphosed rhyolitic tuffs and flows, except for unit 2 at Cleveland Gulch, which is a metamorphosed arkose. The hypothesis of an igneous origin for these units is based on porphyritic textures, flow banding, and relict tuffaceous textures. High silica and potassium content indicates that these units were silicic in original composition. Their extrusive nature is borne out by their concordant relations, flow banding, and very fine-grained matrix~~, and the fact that some have been dated 300 million years older than any of the granitic intrusions in the area~~.

Biotite-quartz schists and chlorite-quartz-feldspar schists were probably intermediate flows and tuffs; however, one biotite-quartz schist in unit 5 of Cleveland Gulch may have been a sedimentary rock since it is closely related to a quartzite. An igneous origin for most of these rocks is based on their relict porphyritic textures, amygdules, and close relationship to other metamorphosed igneous rocks.

TABLE II

| | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 |
|---|---|---|---|---|---|---|---|---|
| $SiO_2$ | 44.44 | 54.57 | 52.53 | 75.47 | 77.25 | 47.4 | 56.63 | 72.80 |
| $TiO_2$ | 1.84 | 1.50 | 1.41 | 0.40 | 0.14 | 2.2 | 0.67 | 0.33 |
| $Al_2O_3$ | 15.04 | 16.27 | 14.22 | 11.42 | 11.56 | 15.6 | 16.85 | 13.49 |
| $Fe_2O_3$ | 3.19 | 3.48 | 4.33 | 2.26 | 0.55 | 3.7 | 3.62 | 1.45 |
| FeO | 9.94 | 6.88 | 7.72 | 0.31 | 0.92 | 9.2 | 3.44 | 0.88 |
| MnO | 0.23 | 0.16 | 0.25 | 0.07 | 0.03 | 0.3 | 0.23 | 0.08 |
| MgO | 7.34 | 2.50 | 5.49 | 0.30 | 0.11 | 8.5 | 4.23 | 0.38 |
| CaO | 8.70 | 6.93 | 9.00 | 0.55 | 0.27 | 10.2 | 7.53 | 1.20 |
| $Na_2O$ | 1.51 | 3.54 | 2.05 | 3.29 | 2.65 | 2.1 | 3.08 | 3.38 |
| $K_2O$ | 0.13 | 2.11 | 0.56 | 4.90 | 5.59 | 0.6 | 2.24 | 4.46 |
| $P_2O_5$ | 0.35 | 0.61 | 0.45 | 0.05 | 0.06 | 0.2 | 0.16 | 0.08 |
| $H_2O+$ | 5.21 | 1.07 | 1.58 | 0.45 | 0.49 | | 0.51 | 1.47 |
| $H_2O-$ | 0.04 | 0.00 | 0.00 | 0.00 | 0.04 | | 0.80 | |
| $CO_2$ | 2.30 | 0.00 | 0.00 | 0.00 | 0.00 | | | |
| TOTALS | 100.26 | 99.62 | 99.59 | 99.47 | 99.66 | 100.0 | 99.99 | 100.00 |

(1) Barker, F. (1958, p. 16) Moppin Formation chlorite schist from Bromide Canyon.
(2) Barker, F. (1958, p. 16) Moppin Formation amphibolite from between Tusas and Burned mountains. Just west of Moppin Formation-Tusas Granite contact.
(3) Barker, F. (1958, p. 16) Moppin Formation amphibolite from Spring Creek Canyon.
(4) Barker, F. (1958, p. 55) Metarhyolite from La Jarita Mesa.
(5) Barker, F. (1958, p. 61) Tusas Granite from Tusas River Canyon S. E. of Cleveland Gulch.
(6) Walker, F. and Poldervaart, A. (1949, p. 649) Average Hebridean olivine basalt.
(7) Clarke, F. W. (1916, p. 455) Augite andesite from Unga Island, Alaska.
(8) Barth, T. F. W. (1952, p. 69) Average effusive rhyolite.

Table II. Chemical analyses of selected rocks from the Cleveland Gulch-Iron Mountain-Cañon Plaza area.

Many of the coarser grained rocks may have been hypabyssal intrusives. Metamorphosed, cross-cutting, porphyritic granodiorite intrusives are common on Iron Mountain (Fig. 4).

Detailed plane table mapping at Iron Mountain revealed another facet of Moppin deposition which was noted by Just (1937, p. 11). This is the presence of large amounts of metamorphosed intrusives of varying compositions which usually cut the strike of the tuffs and flows at low angles or are parallel to the strike. These are very similar in texture and composition to their metamorphosed extrusive equivalents.

Just (1937, p. 11) believed the Moppin rocks were part of a geosynclinal sequence. Barker (1958, p. 14 and 33) stated that the quartzites he believed to underlie and overlie the Moppin Formation were either eugeosynclinal or miogeosynclinal. It is assumed that Barker considered the Moppin rock deposition to be part of the same sequence of events.

The presence of extensive thin beds of metamorphosed rhyolite, some of which display flow banding, interspersed throughout the Moppin Formation and the presence of silicic tuffs indicate that deposition of the Moppin rocks was continental. The uppermost distinctive metarhyolite can be traced for at least 12 to 14 miles along strike; this lateral extent would not be expectable if deposition were subaqueous. Other thin rhyolite flows can be traced for several miles.

Table II gives chemical analyses for a greenschist from near the entrance of Bromide Canyon and an amphibolite from a point west of the granite contact on the western flank of Tusas Mountain (Barker, 1958, p. 16). Columns 6 and 7 are analyses of Hebridean olivine-basalt (Walker and Poldervaart, 1949, p. 649) and augite andesite (Clark, 1916, p. 455) respectively. Composition of the greenschist is near that of many basalts or olivine basalts. The amphibolite is near the andesite composition given here and does not vary significantly from several andesite compositions given by Clark (1916, p. 455). Also included on Table II is an analysis of a metarhyolite from La Jarita Mesa and an analysis of an amphibolite from Spring Creek Canyon located between Cleveland Gulch and Kiowa Mountain (Barker, 1958, p. 16 and 55). The chemical composition of an average effusive rhyolite is included for comparison (Barth, 1952, p. 69); the rhyolite and the metarhyolite are very close in composition, but the metarhyolite has slightly more silica and less Al and Ca.

In summation, the Moppin Formation is a sequence of predominantly volcanic rocks of silicic to mafic composition. A large percentage of the rocks were rhyolitic and dacitic flows and tuffs, which indicate that the Moppin rocks were continental deposits. Local volcanic centers are recognizable. In the vicinity of Placer Canyon a volcanic center apparently extruded mainly rhyolite and some more mafic material. This area might be a section through a metamorphosed composite volcanic cone. A more mafic center may have been present in the vicinity of Cleveland Gulch, but it is not certain how much of the increased thickness of the Moppin Formation there reflects an original accumulation of volcanic material and how much is due to subsequent intrusions.

Local small basins and stream beds were the sites of deposition of sandstone bodies. These deposits are preserved today as quartzite and quartz-muscovite schists which are very limited in areal extent. Intrusive rocks are present and some may have been emplaced contemporaneously with the extrusives. Other intrusives, including the porphyry dikes, probably came after the initial folding and some metamorphism. These intrusives do not include those emplaced much later with the Tusas Intrusives. ~~Granite~~.

## Ortega Quartzite

### Definition and distribution

Ortega Quartzite studied in this work includes the quartzite overlying the Moppin Formation from Cleveland Gulch to Jawbone Mountain (Fig. 2) and the quartzite which encloses the iron deposits at Cañon Plaza (Fig. 6). Just (1937, p. 13) considered all the massive quartzite to be younger than the Moppin rocks and named it Ortega Quartzite for the excellent exposures in the Ortega Mountains (Fig. 1). Barker (1958, p. 10) believed some of the quartzite was older than the Moppin Formation and retained the name Ortega Quartzite for these rocks. The Ortega Quartzite as defined by Barker underlies most of La Jarita Mesa and the Ortega Mountains. Barker considered the quartzite from the Kiowa Mountain-Cleveland Gulch area to Jawbone Mountain to be younger than the Moppin Formation and named it the Kiawa Mountain Formation for the exposures on Kiowa Mountain (1958, p. 24). The accepted spelling for the name of the mountain is Kiowa; therefore, the name Barker gave would have to be corrected. Also this name is so near that of the Kiowa Shale of Kansas that the name "Kiowa Mountain Formation" should be dropped.

La Jarita Mesa is between Cleveland Gulch and

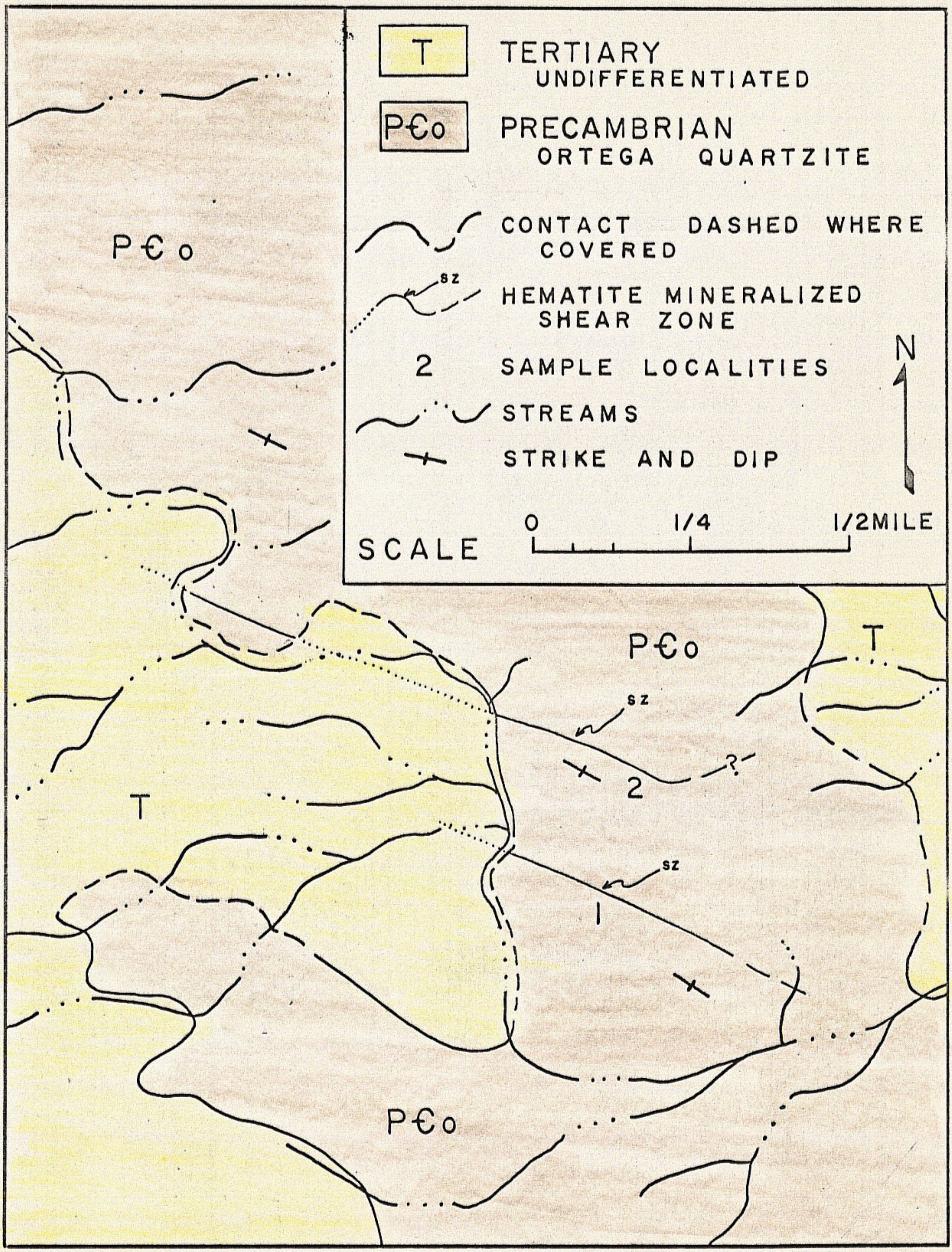



FIGURE 6. GEOLOGIC MAP OF AREA SURROUNDING THE IRON DEPOSITS NEAR CANON PLAZA

the Ortega Mountains. Outcrops on this mesa are not common and many of the rocks there have been extensively metasomatized by fluids from the Tusas ~~Granite~~ Intrusives and the numerous pegmatites. Based on lithology and information on the structure gleaned from the rocks on La Jarita Mesa, Just considered all quartzites to be part of the same formation; Barker considered those to the south of Kiowa Mountain to be older. The present study did not embrace La Jarita Mesa other than in reconnaissance so no new evidence can be brought to bear on this problem. Lithologically the quartzite in the Ortega Mountains and that above the Moppin Formation in the Tusas Mountain area are the same. Consequently, since the problem of relative age is still unresolved and the name Kiowa appears to be preoccupied, the name Ortega Quartzite shall be used as a lithologic term for all the quartzite lithosome. In the following discussion the Ortega Quartzite is understood to be younger than the Moppin rocks at Cleveland Gulch and Iron Mountain, but at Cañon Plaza it may be the same age as that at Cleveland Gulch or it may be older.

Ortega Quartzite is the youngest exposed unit of the Hopewell Anticline and the two flanking synclines. All the Precambrian rocks on La Jarita Mesa are part of the Ortega Quartzite including the metasomatized part known as the Petaca Schist (Just, 1937, p. 13). La Madera Mountain and the Ortega Mountains are also underlain by Ortega Quartzite (Fig. 1). Several knobs of Ortega Quartzite protrude through the Cenozoic cover north and west of the Ortega Mountains. Cañon Plaza iron deposit is in one of these isolated quartzite exposures.

Lithology

Only the Ortega rocks along the contact of the Moppin Formation from Cleveland Gulch to Jawbone Mountain and those in the immediate vicinity of the Cañon Plaza iron deposits were studied petrographically. However, several localities on Jawbone Mountain, Kiowa Mountain, and in the Ortega Mountains were visited in an attempt to understand the regional structure and stratigraphy.

Ortega Quartzite is relatively homogeneous lithologically. In some places it weathers to a tan or flesh color and in other localities it appears white to blue-gray. It is fine to coarse-grained, locally conglomeratic, and usually displays fair to good cross-bedding. On Kiowa Mountain and La Jarita Mesa several thin amphibolite and metarhyolite beds are found in the Ortega Formation.

Along the base of the formation in the vicinity of Cleveland Gulch the quartzite is feldspathic, but this phase pinches out to the northwest. From Burned Mountain to Placer Canyon the basal Ortega is conglomeratic. From Placer Canyon around the nose of the Hopewell Anticline the basal Ortega ranges from medium-grained to conglomeratic. Very few conglomerate lenses are found on Jawbone Mountain. Barker (1958, p. 25) considered the conglomerate distinct enough to have member status in his "Kiawa Mountain Formation." He named this member Jawbone Conglomerate and believed it extended from just northwest of Burned Mountain around the nose of the anticline and underlaid Jawbone Mountain. In the present study the conglomerate was traced from Burned Mountain to about a half mile northwest of Placer Canyon. From this point to Jawbone Mountain, conglomerate is present but it is not typical of the basal Ortega. Less than five percent of the Ortega Quartzite underlying Jawbone Mountain is conglomerate. Consequently, the naming of the Jawbone Conglomerate member seems unwarranted.

At Cleveland Gulch the basal Ortega is a pink-gray, coarse-grained, slightly schistose, feldspathic quartzite. This rock is composed of 85-90% rounded quartz grains averaging 0.2 mm in diameter, 1-2% rounded plagioclase averaging 0.1 mm, 10-15% muscovite (most of which is a product of alteration of potassium feldspar), and trace amounts of chlorite, magnetite, garnet, and epidote.

There is not much variation in composition of the basal Ortega between Cleveland Gulch and the southeastern flank of Burned Mountain (Fig. 2), but the grain size becomes coarser in this direction. In the field the basal Ortega southeast of Burned Mountain in Rock Creek Canyon is pink colored, schistose, and slightly conglomeratic. In thin section 85-90% of the rock is quartz which is well rounded and averages 0.5 mm in diameter. Fine-grained muscovite is present up to 15%, and a highly altered feldspar accounts for about 1% of the rock. Magnetite is in 2 sizes, 0.01 mm and 0.2 mm. The smaller grains are subhedral and may have been detrital, but the larger ones are euhedral and may be secondary.

At Placer Canyon the basal Ortega is vitreous, reddish-gray, and conglomeratic. Some of the pebbles attain a size of 25.0 mm or greater. In thin section fair to well rounded quartz is present in two sizes, around 0.08 mm and 1.5 mm and accounts for 70-75% of the rock. Fine-grained muscovite up to 25%, fine-grained hematite up to 5%, and a trace of magnetite are also present.

At Jawbone Mountain the basal Ortega is not well exposed, but the few outcrops available are vitreous, dense, white, coarse-grained rock composed almost entirely of quartz. Only traces of muscovite are present.

Ortega Quartzite enclosing the iron deposits at Cañon Plaza is dense, vitreous, weathers flesh-color to gray, and is composed of 95% well rounded quartz, ranging in size from 0.1 mm to 1.5 mm. Kyanite and specularite compose the rest of the rock.

The Ortega Quartzite was studied only in the areas near the Moppin Formation contact and surrounding the Cañon Plaza deposit. For the Ortega Quartzite in the La Madera Quadrangle Bingler stated (1965, p. 22):

> In summary, the bulk of the quartzite in the map area is uniform in composition and texture. It has been completely recrystallized to form a xenomorphic aggregate of quartz which in many samples shows indications of later mild cataclasis. The average combined kyanite and specularite content is about 5 per cent with trace amounts of rutile, sphene, and zircon. Subhedral to anhedral kyanite exhibits a high degree of preferred orientation within foliation planes.

This description does not include all phases of the Ortega Quartzite, but it does apply to around 95% of the rocks at the type locality in the Ortega Mountains. This description also is applicable to the rocks from Kiowa Mountain to Jawbone Mountain. In these areas Barker (1958, p. 31) noted that kyanite is absent if muscovite content is more than 10% and muscovite is absent if kyanite content is more than 5%, but in the present study this was not evident. The Petaca Schist phase of the Ortega Quartzite is not exposed in the study area. This quartz-muscovite schist was believed to be metasomatized Ortega Quartzite by Just (1937, p. 43), Barker (1958, p. 34), and Jahns (1946, p. 20). Bingler, however, concluded that in the La Madera Quadrangle the evidence favored a sedimentary source for the muscovite (1965, p. 28).

Thickness

Just (1937, p. 43) estimated the thickness of the Ortega Quartzite in the region from the Ortega Mountains to Jawbone Mountain to be four to five miles. At Picuris, where he had fair cross-bedding control in a transverse canyon, he found the Ortega to be 2 miles or less in thickness (Just, 1937, p. 22). Barker's thickness estimates for the two formations which correspond to the Ortega Formation in this report, are a total of 19,000 to 30,000 feet (Barker, 1958, p. 13 and 32). The difficulty in the determination of the thickness lies in the fact that there are few distinctive beds that can be utilized as markers for determining the structure and the thickness.

Northwest of Placer Canyon near the nose of the Hopewell Anticline about 4,000 feet of Ortega Quartzite is exposed above the Moppin Formation (Plate 4C). Meager cross-bedding indicates that the section is unaffected by folding or faulting and that this is probably the true thickness for the exposure. No evidence was available to ascertain the proximity of the synclinal axis to the outcrops; hence, no top of the Ortega Formation was established. However, where the synclinal axial trace is projected into the area from Barker's Las Tablas map, it falls within 300 to 400 yards of the uppermost outcrop of Ortega rocks. The above figure for the thickness can be classified as little more than an estimate, but this low value gains support from Bingler's work in the La Madera Quadrangle. Considerable evidence for folding and faulting is present in Ortega Quartzite southwest of the Vallecitos River (Fig. 1). Bingler (1965, p. 19-20) concluded that the foliation of the quartzite in La Madera Quadrangle is axial-plane cleavage and not bedding foliation and that repetition of layers by folding and faulting has increased the apparent thickness. He concluded that "in view of the complex folding which these rocks have undergone, the great mass of quartzite now exposed could conceivably have been derived from an original sandstone layer or layers only a few hundred feet thick."

Bingler's belief that all foliation is axial-plane cleavage and that the Ortega might be only a few hundred feet thick appears to be erroneous for the areas studied in this report but his idea that folding and faulting are responsible for the apparent great thickness of the Ortega Quartzite appears to be reasonable. Based on these observations and the studies northwest of Placer Canyon, 5,000 feet is the estimated thickness of the Ortega Quartzite.

Origin

The Ortega Quartzite poses many problems of paleo-environmental significance. Basal feldspathic quartzite apparently conformable on continental volcanic deposits, conglomerate lenses throughout the quartzite, detrital iron and titanium oxides, considerable amounts of aluminum silicate, and interspersed volcanic deposits make interpretation difficult. There is no direct evidence that the sand was continental, although some may have been beach sand. Bingler (1965, p. 21) stated that zircons are rare and usually rounded, but in some specimens euhedral and rounded zircons occur together. Euhedral zircon usually indicates a silicic or intermediate igneous rock provenance, and rounded zircon usually is derived from sedimentary rocks (Krumbein and Sloss, 1963, p. 140). No doubt clastic material from two provenances was supplied to the Ortega Sea.

The basal feldspathic beds represent the initial deposition as the sea encroached on the continent. Where the feldspathic rocks are absent there is usually a pebble conglomerate or coarse sand. As the sea transgressed, finer, more rounded quartz sand was deposited. The cross-bedded nature of the deposits and erratic conglomerate lenses at higher stratigraphic levels suggest deposition fairly close to the shore. Because of the great thickness he believed present and the presence of the Moppin metavolcanics, Barker (1958, p. 33) concluded that deposition could have been either miogeosynclinal or eugeosynclinal. However, he apparently considered the Moppin rocks to be marine volcanics and as such to be part of the geosynclinal marine sequence. Since the Moppin volcanics are evidently continental and unrelated to the marine beds of the Ortega Quartzite, it seems probable that Ortega sedimentation took place in a marginal basin.

## Intrusive Rocks

### Tusas Intrusives ~~Complex~~

#### Definition and distribution

The Tusas intrusive complex is a large body of intrusives of batholithic dimensions. It underlies much of the area from Tusas Mountain to Jawbone Mountain (Fig. 2), and crops out at several places from Tusas Mountain to the Petaca area and at Ojo Caliente (Just, 1937, Pl. III). A few small outcrops are found in the vicinity of Tres Piedras, about 8 miles east of Tusas Mountain. Just (1937, p. 45) believed that the Tusas granite was not very far beneath many of the outcrops of Ortega and Moppin formations in the area. Only those outcrops of Tusas intrusives from Tusas Mountain to Jawbone Mountain were studied; however, most other outcrops were visited.

Just (1937, p. 44-45) defined and named the "Tusas Granite" for the excellent and extensive exposures on Tusas Mountain. He believed the intrusion was zoned and noted the presence of many apophyses. He further pointed out that base and precious metal-bearing veins are found near what he termed the monzonitic phase and pegmatites are found near the granitic phase.

Barker (1958, p. 59) stated ". . . the rock at Tusas Mountain appears to be an atypical very fine-grained porphyritic phase. . . . The granite at Tres Piedras and along the Rio Tusas is here redefined as the Tres Piedras granite, after the excellent exposures in and around that town." According to the Code of Stratigraphic Nomenclature, type localities cannot be changed

(American Commission on Stratigraphic Nomenclature, 1961, Article 13h, p. 653), and established names of rock-stratigraphic units should not be changed (1961, Article 11 and Article 12, p. 652). Barker not only changed the name and type locality for the Tusas Granite, but named it for a locality not included in his Las Tablas Quadrangle. It is difficult to see how the largest outcrop and main body of the Tusas Granite can be considered atypical, while the granite at Tres Piedras, which is the smallest outcrop of granite and is 8 miles away, is considered typical. Consequently, in order to follow the recommendations of the Code of Stratigraphic Nomenclature more closely and note the multiple intrusive nature of the batholith, the name of the large main body of granite from Tusas Mountain to Jawbone Mountain as well as that in the Tusas River Valley from Tusas to Ojo Caliente should be changed ~~back~~ to Tusas Intrusives ~~complex~~. The name Tres Piedras Granite can be retained for those outcrops of granite at Tres Piedras which are separated from Tusas Mountain by 7 or 8 miles of Cenozoic cover.

Barker (1958, p. 56) separated a granodioritic phase from the granitic phase of the Tusas intrusions. He named the unit the Maquinita Granodiorite and concluded that it had been emplaced before the granite. He based this conclusion on the fact that the "granodiorite" displays foliation and lineation (Barker, 1958, p. 59).

He believed the granitic phase is not foliated as much as the "granodiorite." Many of the well-foliated rocks in the area mapped as granodiorite by Barker are at contacts with Moppin rocks, near contacts of dikes within the intrusion, along fault zones, and in apophyses in the Moppin Formation. The part mapped as granite by Barker also displays foliation under these same conditions.

Barker is correct in his conclusion that the composition of the intrusion changes from Tusas Mountain to Jawbone Mountain. However, some of the changes appear to be a zonal effect, and outcrops are too scarce to attempt to map the zones. There are quartz monzonite, monzonite, quartz diorite, granite and granodiorite within the intrusive complex. Considering the relative paucity of data on this complex intrusion, an attempt to delineate the lithologies is premature. Consequently, in the present report the intrusive body is called the Tusas Intrusive ~~complex~~ and is recognized as a partially zoned intrusion with at least part of the complex much older than other parts. The igneous rocks in this report are named according to the classification given by Compton (1962, p. 276).

## Lithology

The southeastern flank of Tusas Mountain is underlain by a rust to tan colored, slightly foliated, medium to coarse-grained granite. Forty-five percent of the rock is quartz in two sizes, 0.5 mm and around 5.0 mm. Subhedral microcline is also present in these two sizes and accounts for 40% of the rock. Subhedral albite is present up to 5%, biotite to 5%, and muscovite to 7%. Trace amounts of magnetite, tourmaline, and epidote are also present.

The linear apophysis of the Tusas intrusion is granodiorite north of Cleveland Gulch (Fig. 2). This rock is pinkish-gray, medium to coarse-grained granodiorite, highly foliated near the contact, but only slightly foliated near the center. In thin section the rock is made up of 25% quartz, 55% highly altered subhedral albite-oligoclase averaging around 0.5 mm, 1% highly altered microcline, 10% muscovite, 7% biotite, 1% magnetite, and traces of epidote and chlorite. Chlorite replaces biotite, and some of the muscovite replaces potassium feldspar (Plate 11A).

This linear apophysis narrows to about 900 feet at Bromide Canyon (Fig. 2) and consequently the center of the intrusion is more foliated than at Cleveland Gulch. At Bromide Canyon the rock is quartz monzonite. Microcline content is around 30% and albite-oligoclase about 45%. No biotite was found in the samples from this area.

Immediately south of the dam at Hopewell Lake (Fig. 2), the Tusas intrusion is granodiorite. Here the rocks are highly sheared near the contact with the Moppin Formation. About 55% of the rock is highly altered subhedral albite-oligoclase of 2 sizes, 0.3 mm and 2.5 mm. Five per cent microcline, 25% quartz, 7% chlorite, 3% muscovite, 5% epidote, and a trace of magnetite compose the rest of the rock.

Northeast of Iron Mountain the intrusion is granite. Immediately south of Iron Mountain it is much like the quartz monzonite at Bromide Canyon, but has about 20% microcline, 3-5% calcite, 35% each of quartz and albite-oligoclase, and 5-7% biotite.

On the northwestern flank of Tusas Mountain the intrusion is light gray fine-grained granite. Here the rock is about 45% microperthitic intergrowths, 6% albite-oligoclase, 45% quartz, 3% muscovite, and traces of biotite, magnetite, and epidote. The feldspars average 1.2 mm, but the quartz averages only 0.3 mm (Plate 11B). Granodiorite, quartz monzonite, and granite west of the prong of Moppin Formation extend into Tusas Valley (Fig. 2).

West of Cleveland Gulch is an amoeboid-shaped area of reddish colored granite (Fig. 5). At first these rocks were thought to be part of the Tertiary igneous rocks which characterize the San Juan uplift. However, upon closer inspection, they showed fair foliation. This rock has 45%

fine-grained microcline, 45% quartz of 2 sizes, 0.1 mm and 5.0 mm, 7% fine-grained muscovite, 1% garnet, and traces of biotite, magnetite, epidote, plagioclase, and zircon (Plate 11C). The rock is fine-grained with large quartz phenocrysts. The apparent cross-cutting amoeboid shape of the outcrop and porphyritic texture indicate that the rock is intrusive. Absence of other Tertiary rocks and the foliation suggest it is Precambrian, and since it is no more or less foliated than the other Tusas intrusives, it was mapped as a phase of the Tusas complex (Fig. 2 and Fig. 5). Table IV lists the trace elements found in this rock.

In summation, Tusas intrusive complex is fine- to coarse-grained and foliated to non-foliated. Its composition varies from granite to granodiorite, although the dominant rock type is granite. The intrusion appears to be zoned, but the zoning was not worked out in detail. Lack of outcrops will probably prevent delimitation of the zones other than on the broadest basis.

Aplite dikes

Aplite dikes were noted in the Tusas granite and in the formations cut by the granite. The only

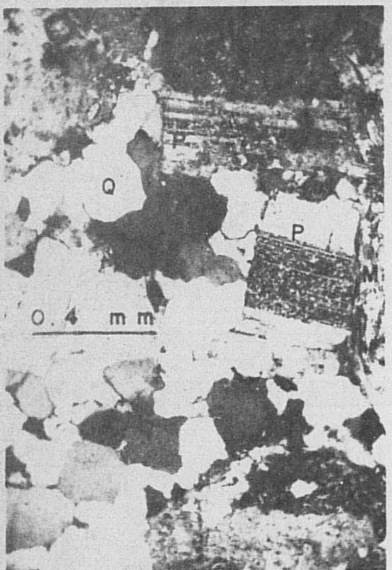
Fig ~~PLATE~~ 11

A. Granodiorite. Tusas intrusion. Linear apophysis north of Cleveland Gulch. Plagioclase (P), muscovite (M), quartz (Q). S20 193 35 X-nicols.

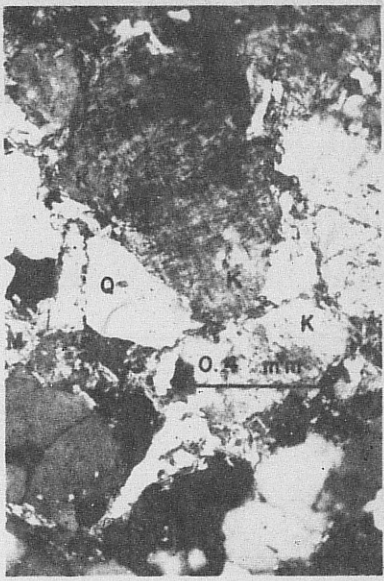
B. Granite. Tusas Intrusives ~~Granite.~~ Northwest slope of Tusas Mountain. K-feldspar (K), quartz (Q), muscovite (M). 21 20 16 X-nicols.

C. Granite ~~Intrusive rhyolite.~~ Tusas Intrusives ~~Granite.~~ West Cleveland Gulch. Quartz (Q) phenocryst in matrix of fine-grained quartz, K-feldspar, and muscovite. 21 18 4 X-nicols.

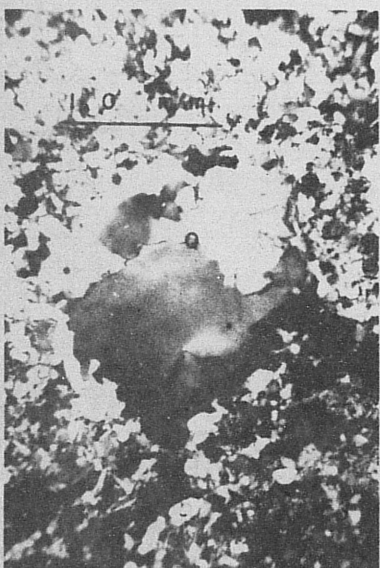
D. Quartz vein, iron formation, and Ortega Quartzite. Cañon Plaza.




A




B




C

D

mapped dike is on the western end of Iron Mountain (Plate 12A). This rock is microcline, quartz, albite, and muscovite. No foliation was visible.

Mafic dikes

Several mafic dikes trend almost north-south in the Tusas rocks north of Hopewell Lake (Fig. 2). These have been metamorphosed to chlorite-plagioclase schist, with the foliation parallel to the sides of the dikes. It is interesting that these dikes are high in magnetite. The magnetite extends into the enclosing Tusas rocks and decreases over a distance of 75 to 100 yards from 5-10% to the normal trace amount of the enclosing intrusion.

Massive quartz veins

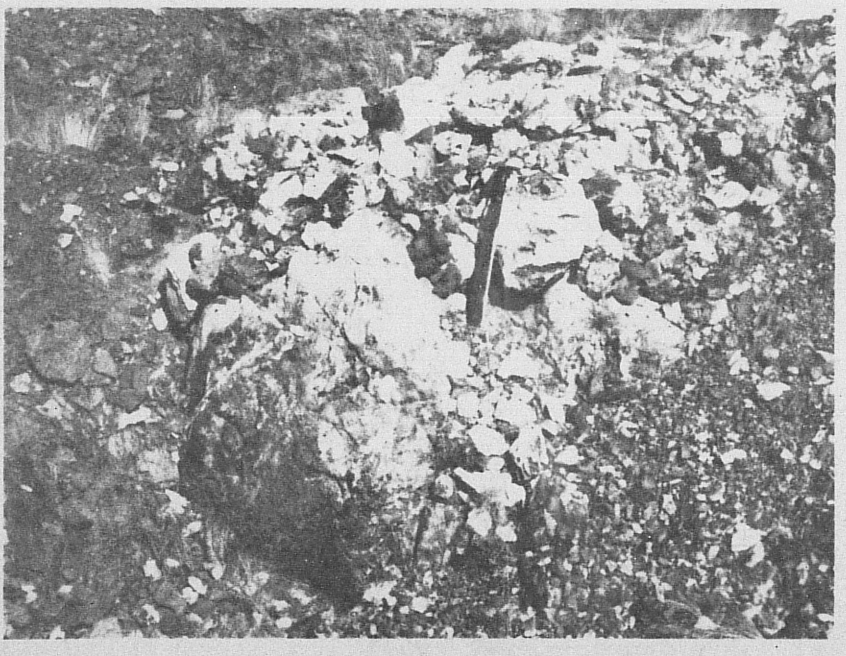
Massive milky quartz veins are the only other igneous-related rock found in the area studied in the present report. These are numerous on Iron Mountain (Plate 12B) and at Cañon Plaza and are in most

Fig. ~~PLATE~~ 12

A. Aplite dike. Iron Mountain. Cutting chlorite schist of Moppin Formation.

B. Massive quartz vein. Iron Mountain. Cutting chlorite schist of Moppin Formation.

A



B

of the Precambrian outcrops. They cut the Moppin Formation, Ortega Quartzite, and Tusas Granite. Where they intersect the magnetite-bearing mafic dikes described above, the dike material is changed to chlorite and euhedral magnetite crystals up to 30.0 mm long. At Cañon Plaza veins occupy cross fractures and bedding planes and grade into the surrounding Ortega Quartzite (Plate 11D). These veins are numerous and widespread; at least some may be metamorphic in origin. At other places they apparently are associated with the Tusas intrusion. Jahns (1946, p. 51) felt that the quartz veins in the Petaca area are genetically related to the pegmatites.

# CENOZOIC ROCKS

## General Statement

Post-Precambrian rocks were studied only in the field mapping. Other than Quaternary Alluvium, the only Cenozoic unit mapped was the Los Piños Formation. Atwood and Mather (1932), Cross and Larsen (1935), Just (1937), Butler (1946), and Barker (1958) have been relied upon for much of the information on the Los Piños Formation.

## Stratigraphy

### Los Piños Formation

This formation was named by Atwood and Mather (1932, p. 93). They attributed the name to Whitman Cross and E. S. Larsen, but since Cross and Larsen's work was not published until 1935, Atwood and Mather are recognized as the first to officially propose the name. The type locality is about 10 miles south of Antonito, Colorado in the Los Piños Creek Canyon near Miguel, New Mexico.

Atwood and Mather (1932, p. 100) considered the Los Piños Formation to have been deposited on the surface produced by the San Juan Peneplain cycle of erosion and to have been deposited before the Hinsdale volcanic rocks. Cross and Larsen (1935, p. 95) included the Los Piños Formation as a member of the Hinsdale Formation. They believed the Los Piños rocks were probably late Pliocene in age.

Butler (1946) was able to establish the stratigraphic relationships in the area of Tusas River. Barker (1958) used Butler's nomenclature for the Los Piños Formation and mapped four members in the Las Tablas Quadrangle.

This formation covers the western flank of the Tusas River Valley, parts of the southern flank of the Hopewell Anticline, part of the area near Hopewell Lake, and part of the area at Cañon Plaza (Figs. 2 and 6). In most places the formation is unconformable on all Precambrian rocks, but in places it is in fault contact with the Precambrian rocks (Fig. 2).

Los Piños Formation is mainly pebble to boulder conglomerate composed predominantly of rhyolitic material. Tuffaceous and sandstone beds are interspersed through the formation. The boulders are up to 5 feet in diameter, but most clasts are 6 inches or less. In the vicinity of Tusas Mountain the formation has a probable maximum thickness of 300 feet, but near Cañon Plaza it is 500 to possibly 1,000 feet thick (Barker, 1958, p. 50).

Barker (1958, p. 50) believed the material of the Los Piños Formation was derived from volcanic centers located on the Taos Plateau or in the San Luis Valley.

## Quaternary alluvium

Alluvium in the area is limited. It forms a thin veneer on bedrock only along the channels of Tusas River, Spring Creek, and Placer Creek. In most places the alluvial material is coarse-grained, but in the wider parts of the valleys it is composed of silt to sand size particles mixed with a few boulders and cobbles from the Los Piños and Precambrian formations. Cobbles and boulders from all the formations transversed by the streams line the edges of the streams in their swifter portions. Gravels along Placer Creek are gold-bearing at places.

# STRUCTURE

## Regional Tectonic Setting

San Juan Uplift is in the Southern Rocky Mountains. In New Mexico the uplift is bounded by the San Luis Basin on the east and butts into the Sangre de Cristo Uplift on the southeast (Fig. 3). The western flank of the uplift plunges into the Chama Basin, and the southern part is probably faulted into the Rio Grande Trough.

Figure 3 is a Tectonic Division Map of the area bounded approximately by parallels 34° 30' and 37° 30' and meridians 105° and 109° 30'; the map covers most of the northwestern one-fourth of New Mexico and adjacent parts of Colorado, Arizona, and Utah. The area west of the Rio Grande Trough, Jemez Caldera, and Chama Basin is in the Colorado Plateau. Most of the remainder of the area is part of the Southern Rocky Mountains.

Tectonic divisions of the Colorado Plateau are characterized by simple structures. The uplifts--Defiance, Zuni, Lucero, Nacimiento--have gentle open anticlines, synclines, and monoclines and many normal faults of small displacement. Thrust faults are present and have large displacements in the Nacimiento and Lucero uplifts where they form the west and east boundaries respectively of the two uplifts. Laccoliths, dikes, sills, and plugs are associated with some of the uplifts. San Juan Basin is structurally simple; it is asymmetrical with the structurally lowest point in the northeast quadrant. The basin is bounded on the north and west by a sinuous monocline, and it is intruded in a few places by volcanic plugs and a dike swarm. The structural simplicity of the basin is modified by a few normal faults and some open folds. Four Corners Platform and Archuleta Arch are broad structural benches with gentle anticlines, synclines, and monoclines; some high-angle and thrust faults; and dikes, sills, plugs, and laccoliths. The Puerco Fault Belt-Mt. Taylor Centers form the only other tectonic division on the Colorado Plateau part of Figure 3. This unit is characterized by scores of high-angle faults and volcanic plugs.

Most of the southeastern Colorado Plateau is underlain mainly by Permian to Tertiary sandstone, shale, and mudstone, although Pennsylvanian limestone is exposed in the Lucero and Nacimiento uplifts. Zuni, Defiance, and Nacimiento uplifts expose Precambrian metamorphic rocks and/or granite. The tectonic divisions of the Colorado Plateau are described in detail in Kelley (1955) and

Kelley and Clinton (1960).

Chama Basin forms a shallow structural sag between the Archuleta Arch on the west and the San Juan Uplift on the north and east. The southern end of the basin merges with the Nacimiento Uplift on the southwest and the Rio Grande Trough on the southeast. The southern end of the basin is cut off by the Jemez volcanic complex. Chama Basin is underlain by Permian to Tertiary sandstone, mudstone, and shale. A gentle arcuate syncline, high-angle faults, and a few intrusions form the structural make-up of the basin.

Cutting ~~Forming~~ the southern end ~~boundary~~ of the Chama Basin is the Jemez Caldera and volcanic field. Volcanics of this area have covered the area bounded by the Nacimiento Uplift on the west, Chama Basin on the north, and the Rio Grande Trough on the east and south. The division is dominated by the Jemez Caldera, which encompasses approximately 250 square miles. Tertiary and Quaternary basalt, rhyolite, tuffs, and sedimentary rocks cut by numerous faults constitute most of the rest of this division.

San Luis Basin is wedged between the San Juan Uplift on the west and south and the Sangre de Cristo Uplift on the east. Probably both of these contacts in New Mexico are fault contacts, but on the western side the evidence for faulting is obscured by late Tertiary and Quaternary deposits except in the vicinity of Ojo Caliente where a large normal fault marks the eastern flank of Ojo Caliente Mountain (Kelley, 1956, p. 110). San Luis Basin is underlain by alluvium, sandstone, conglomerate, and volcanics of Tertiary and Quaternary age. It is a shallow basin, asymmetrical with the steeper limb on the east, with several volcanic centers, and some normal faults.

The Rio Grande Trough is to the South of the San Luis Basin across a narrow arm of the San Juan Uplift. ~~is the Rio Grande Trough.~~ This division trends southwest from the San Juan Uplift on the north to the vicinity of the northern end of the Sandia Uplift where it swings south (Fig. 3). It is a graben, with bounding faults in evidence in most localities except where late Tertiary or Quaternary volcanics or sediments have obscured them. The trough is underlain by Tertiary and Quaternary volcanics, alluvium, sandstone, and conglomerate. Within the graben are several small cinder cones aligned in a north-south direction.

Flanking the Rio Grande Trough on the east from south to north are the Sandia Uplift, Cerrillos Centers, and the southwestern flank of the Sangre de Cristo Uplift (Fig. 3). Sangre de Cristo Uplift trends northward along the flank of the San Luis Basin. It is in fault contact with the San Luis Basin and Rio Grande Trough. On the western flank of this large uplift are Precambrian metamorphic and intrusive rocks, and limestone, sandstone, shale, mudstone, conglomerate, and volcanics of Pennsylvanian to Tertiary age. Besides the bounding normal faults on the west, there are numerous high-angle faults and complex folds within this large unit. Adjoining the Sangre de Cristo on the southwest is the Cerrillos Centers division. This area is dominated by stocks and dikes, but also has some Precambrian rocks, and limestone, mudstone, sandstone, and shale of Pennsylvanian to Tertiary age. High-angle faults are present.

Sandia Uplift merges with the southwestern flank of the Cerrillos division. It is faulted into the Rio Grande graben on the west and plunges gently into the Estancia Sag on the east. The uplift is underlain by Precambrian granite and some gneiss, and limestone, mudstone, sandstone, and shale of late Paleozoic to early Tertiary age. There are normal faults along the western margin and several high-angle faults in the uplift. Small folds are also present. Merging with this unit on the east is the shallow Estancia Sag, which is underlain mostly by Quaternary lake deposits with a few small outcrops of Paleozoic and Mesozoic rocks. The eastern flank of this shallow syncline makes a gentle bench and merges into the low-dipping Pecos Slope. The Pecos Slope is underlain mainly by Pennsylvanian and Permian mudstone, limestone, and shale and some Tertiary and Triassic rocks.

San Juan Uplift has been described in earlier sections of this report. It is a large uplift underlain by Precambrian rocks and a thick section of Paleozoic, Mesozoic, and Cenozoic rocks in Colorado, but only Precambrian and Cenozoic rocks in the area in New Mexico covered by this report. The oldest rocks in depositional contact with Precambrian rocks in New Mexico west of the Rio Grande are on the eastern flank of the Chama Basin where Triassic rocks were mapped in unconformable contact with the Precambrian of the San Juan Uplift (Smith and Muehlberger, 1960). East of the Rio Grande in the Picuris Uplift, Just (1937, Plate II) mapped Pennsylvanian rocks in contact with the Precambrian.

The origin and implications of certain regional tectonic features connected with the San Juan Uplift in New Mexico are relevant here. One puzzling item in the tectonic framework of this area is the northeast-trending part of the San Juan Uplift east of Ojo Caliente--the Picuris Uplift (Fig. 3). The strikingly abrupt change from a general northwest trend to a northeast trend near Ojo Caliente requires discussion. Essentially two possibilities exist: one is that the Picuris Uplift belongs to the Sangre de Cristo Uplift and the other is that it belongs to the San Juan Uplift. The Picuris area has been mapped as a salient of the Sangre de Cristo Uplift by many people including Baltz (1965, p. 2043) and Kelley (1956, p. 110), who considered it also a thrust block. However, Just (1937, p. 20), while not connecting the Picuris Uplift to the San Juan Uplift, stated "the folded series found in both areas would be continuous if the intervening younger rocks were stripped away." He also believed these rocks extend into the Sangre de Cristo Uplift. The similarity of the rocks in the subject area to those in the Picuris Uplift and the proximity of the Precambrian outcrops (less than 10 miles) suggest these two areas are part of the same tectonic division--the San Juan Uplift. The lithologies may possibly extend into the Sangre de Cristo Uplift, but, at present, due to overburden of younger rocks, no unequivocal evidence for this is available. On Figure 3 the tectonic division boundary between the San Juan Uplift and the Sangre de Cristo Uplift was drawn at the Precambrian-Paleozoic contact.

Another alternative exists to explain the peculiar structural setting of the Picuris Uplift if it is not considered part of the Sangre de Cristo Uplift or if it is not the result of the original tectonic trend of the San Juan Uplift. There is a possibility that faulting in the valley between Ojo Caliente and the Picuris Uplift and along the contact of the uplift with the Sangre de Cristo

Uplift may have shifted the structural trend of the Picuris block of the San Juan Uplift from southeast to northeast. Confirmation of either hypothesis must await more detailed work in the San Luis Valley and Sangre de Cristo Uplift.

In conclusion, the similarity of lithologies and proximity lead to the belief that the Picuris Uplift is part of the San Juan tectonic unit; its present structural trend may represent a Precambrian tectonic trend and may continue into the Sangre de Cristo Uplift or it may reflect more recent faulting and rotation to the northeast in the San Luis - Rio Grande downwarp and along the western flank of the Sangre de Cristo Uplift.

There is a suggestion of a relationship between tectonic lineaments and mineralization in the San Juan Mountains of New Mexico. The presence of a southwesterly Precambrian structural trend in the Lake Superior region has long been known (King, 1959, p. 26). King (1959, p. 26) considered that this trend (called continental arch, continental backbone, or transcontinental arch) continues southwest from Minnesota into New Mexico. Absence of lower and middle Paleozoic rocks in central and northern New Mexico but not in southern New Mexico and southwestern Colorado supports this contention. King (1959, p. 26) also stated that "outcrops and drill records indicate similar conditions elsewhere along this feature to the northeast." Similarity of Precambrian age dates along a general northeast trend supports these observations (Gastin, 1960, p. 10).

In New Mexico this lineation is indicated by northeasterly trends of folds, faults, and fractures in Precambrian rocks and by northeasterly trends of series of volcanic plugs, dikes, and faults in post-Precambrian rocks. Apparently, the northwest limit of the virtually complete absence of pre-Carboniferous rocks is the western flank of the Defiance Uplift (Fig. 3) (Lessentine, 1965, p. 2009). Consequently, the northeast trending transcontinental arch is a wide belt in this part of the country.

Kelley (1955, p. 59) found several northwest trending lineaments in the Colorado Plateau and eastern Rockies. The fact that these lineations cross major present day tectonic boundaries and are also often marked by major intrusions and extrusions suggests that they may be basement structural features. In Colorado, the northeast trending Transverse Porphyry Belt also has a general northwest lineation of the intrusive elements (Badgley, 1965, Figs. 9-11, p. 328), which coincide with northwest lineation lines proposed by Kelley (1955, p. 59). San Juan Uplift in New Mexico lies along the Uncompahgre lineament (Kelley, 1955, p. 59). In addition to these northwest trending lineaments and the northeast trending transcontinental arch, there is an east-west girdle of basaltic centers extending from the San Francisco Centers at Flagstaff, Arizona, through the Hopi Centers and the Mt. Taylor Centers in New Mexico.

In the vicinity of Albuquerque, New Mexico three lineaments meet--the northwest belt, the east-west basaltic belt, and the transcontinental arch. Sandia Uplift and the southern end of the Sangre de Cristo Uplift have essentially the same rock-types and both are similarly deformed (Fitzsimmons, 1961; Baltz and Bachmon, 1956). Possibly these may have been continuous in a general north-south line at one time but then offset by right lateral movement along the east-west tectonic zone. The huge Jemez volcanic complex is north of the east-west line connecting the northern end of the Sandia Uplift and the southern tip of the Sangre de Cristo division; this line also coincides with the east-west basaltic belt. Cerrillos intrusive centers are mostly south of this line (Fig. 3). Also, these two igneous complexes lie on what Kelley (1955, p. 59) termed the La Sal Porphyry line. On either side of the east-west line lie the Nacimiento (north) and Lucero (south) uplifts. These uplifts were formed by thrusts, Nacimiento from the east and Lucero from the west. Separating these two uplifts is the Puerco Fault Belt - Mt. Taylor volcanic centers division. Scores of high-angle faults constitute the Puerco Fault Belt (Kelley, 1955, Fig. 2). These may be tension release fractures for the two opposing thrusts and, also, if an east-west shear zone exists here, they might be tension fractures along the shear zone. The numerous volcanic plugs in the Mt. Taylor field are fairly well aligned in a northeast direction, suggesting, again, that the northeast tectonic lineament of the basement played an important part in later geologic events.

In summation for the relationship of local structures to the regional tectonic framework, the Tectonic Division Map (Fig. 3) is crossed by the northeast trending transcontinental arch; the arch was more or less persistent from Precambrian to Pennsylvanian time; prominent northwest lineaments exist in this area; at least one east-west lineament is present; where lineations meet, and other conditions are favorable, there are large igneous intrusions, extrusions, faulting, and/or mineralization. The existence of an east-west shear zone approximately at the 35° 25' parallel along which there may have been right lateral movement is suggested by the similarity in lithology and structure of the Sandia and Sangre de Cristo uplifts; the sharp bending of the Rio Grande graben along this line; the position of the Nacimiento thrust north of the line and the Lucero thrust south of the line; the numerous faults of the Puerco Fault Belt on either side of the zone; the Mt. Taylor volcanic centers which straddle the zone and have a strong northeast alignment. Until much more work is done, the concept of a shear zone near Albuquerque can only be hypothetical.

Even though the interpretation of much of the regional tectonic pattern may be highly conjectural, the facts that major volcanic fields, intrusive complexes, and mineral deposits in many places are aligned in a northeasterly direction and are usually found at intersections of northwest structural lineations with the northeast lineation have great significance in the exploration for mineral deposits. Many of the base and precious metal deposits in the Hopewell and Bromide districts are along northeast trending fault systems. In the Bromide District, the most successful mines were along a major northeast trending canyon and this canyon may have formed as a result of erosion along a fault zone. Outcrops are too poor to determine if there is faulting, and since the beds are nearly vertical, there could have been considerable vertical movement without much apparent displacement. The mineral deposits pinch out away from the canyon. Placer Canyon in the Hopewell District is another northeast trending canyon with base and precious metal mineralization and it possibly formed along a fault zone. There is hydrothermal mineralization along the northeast trending faults in Sheep Gulch and along the eastern flank of Burned Mountain (Fig. 2).

Many of the mineral deposits in the Las Tablas Quadrangle are along prominent northeast or northwest fault-fracture zones. Most of these zones can be located on aerial photographs; this would greatly reduce the cost and increase the effectiveness of future mineral exploration. The iron deposit at Cañon Plaza is along a northwesterly trending fault zone, but iron deposits at Cleveland Gulch and Iron Mountain more or less parallel the strike of the Moppin rocks and no faulting is in evidence.

Figure 7 shows the approximate location of most of the granitic intrusions in the New Mexico San Juan Mountains from Jawbone Mountain to Picuris. The general location of most of the mineral deposits is also included on the figure. Most deposits are clustered around the intrusions and many are along the northeast fault-fracture system, although it is not possible to show this at the scale of the map. The striking correspondence of the intrusions and the mineral deposits to the northeast trending lineament and the proximity of the mineral deposits to the intrusions suggest that the faults and fractures provided ~~lines~~ zones of weakness along which the mineralizing fluids escaped and deposited their metals when the area was intruded by the magma. In conclusion, based on regional and local structures, reasonable targets for mineral exploration in the San Juan Mountains of New Mexico are localities where granite has intruded the metamorphic rocks. Northeast trending shear zones cutting the metamorphic rocks are especially good targets for mineral exploration.

## Local Structures

Cleveland Gulch - Iron Mountain area is dominated by a large nearly isoclinal anticline. The trend of the axial trace is approximately N60°~~-65°~~W west of Cleveland Gulch, but bends sharply east at the Cleveland Gulch area, and is almost east-west from Cleveland Gulch to the point where the structure is covered by Cenozoic deposits. The anticline plunges 35° to 45° to the northwest; however, the northeastern limb flattens somewhat over a short distance into a rather shallow syncline with a low plunge. The axial trace of this northwest trending syncline is north of Jawbone Peak and the syncline was not studied in detail.

Kiowa Syncline (Barker, 1958, p. 65) adjoins the southwestern flank of the Hopewell Anticline. It is a moderately to steeply plunging northwest trending syncline. Only the mutual flank with the Hopewell Anticline was studied in any detail.

Except for a very limited area, the beds in the

Hopewell Anticline are upright. Overturned beds were noted only at one place between Bromide Canyon and Cleveland Gulch (Plate 13A); these are anomalous and may represent local faulting or downslope creep since they are on a very steep slope. No other overturning was noted along the flanks of the Hopewell Anticline.

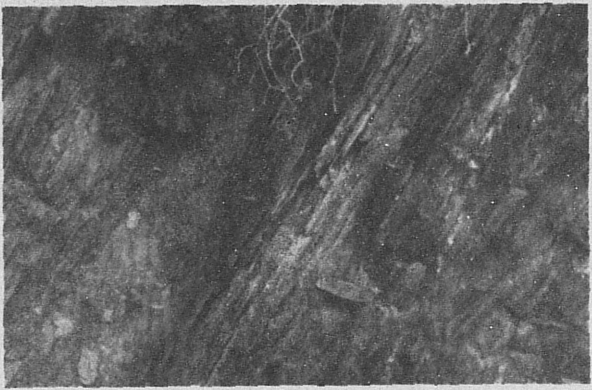
On the Hopewell Anticline schistosity is parallel to the bedding planes. This was ascertained by close comparison of the planes of schistosity to known stratigraphic planes, such as beds of metarhyolite, meta-arkose, or metaconglomerate. Poor exposures on the nose of the anticline prevented determination of the attitude of schistosity there. Schistosity in the Tusas granitic rocks was in general parallel to the contacts of the pluton.

Carpenter, on the basis of petrofabric studies (1965, p. 29), advocated the presence of a major drag fold in the Moppin Formation between Burned Mountain and Tusas Mountain. Stratigraphic information on this area is meager, but the available field data provide no evidence of major drag folding in this locality. No major drag folds were found anywhere in the Moppin Formation. Bingler (1965, p. 24) felt he had evidence from petrofabric data to support the hypothesis that the initial folding in the La Madera Quadrangle to the south was isoclinal and trended northeasterly and this was succeeded by a second episode of

Fig.
~~PLATE~~ 13

A. Prospect pit in iron formation. Moppin Formation. North Cleveland Gulch. Red magnet on magnetite-rich layer. Beds overturned.

B. Banded iron formation. Cañon Plaza. Dark bands are rich in specularite. Light bands are rich in aluminum silicates.



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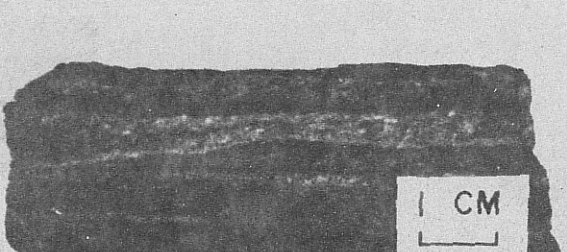


folding which resulted in folds with a northwest trend. However, north of the area in which Bingler worked in the vicinity of Cleveland Gulch and Iron Mountain there was no field evidence that the first fold direction was to the northeast. Based on stratigraphic and structural data in the Cleveland Gulch-Iron Mountain area, it is concluded that the Hopewell Anticline was first folded with a northwesterly trend. Conglomerate particles have been only slightly stretched and in many cases do not show evidence of crushing. Also, the schistosity parallels the bedding planes, making it highly improbable that the Cleveland Gulch-Iron Mountain area was first isoclinally folded in a northeasterly direction and then folded in a northwesterly direction.

A pronounced fissility parallels the schistosity and the bedding in the Moppin rocks. There is a suggestion of a fracture direction perpendicular to the bedding. This may represent planes of stress set up during the initial folding of the Hopewell Anticline. However, the rocks involved in the folding of the anticline were probably too plastic due to the recrystallization to retain tension fractures. After folding and some recrystallization, the rocks of the Hopewell Anticline were intruded by the Tusas Granite. From Cleveland Gulch to Iron Mountain the instrusion is more or less parallel to the strike of the beds; therefore, it may have forced itself up through the axial region of the anticline, assimilated much of the northeastern flank and bulged out the remainder of the Moppin rocks. This bulging out of the limbs of the anticline may have caused the fractures perpendicular to the bedding.

Faulting trends in the same direction as the fractures in most places. Where it was possible to map faults in the field or on aerial photographs, they either more or less parallel the bedding planes in the Moppin and Ortega rocks or trend northeasterly approximately perpendicular to the beds on the southwestern flank of the Hopewell Anticline (Fig. 2). The most prominent transverse faults are ~~in Sheep Gulch and~~ along the eastern flank of Burned Mountain (Fig. 2). Even though these two faults were mapped mostly from study of lineations on aerial photographs, they were noted in a few places in the field. No estimation of the amount of displacement on these two faults can be made, because exposures are so poor, and the beds are nearly vertical in most places. If the movement along the fault was at any time vertical, considerable movement could have taken place without much apparent displacement on the surface. It is believed that the fault flanking the eastern side of Burned Mountain (Fig. 2) had considerable displacement since there is apparently a large offset of beds. The faults ~~in Sheep Gulch and~~ west of Burned Mountain have smaller displacements.

A wedge of Tertiary Los Piños Formation is down-faulted between Precambrian Ortega beds south of Burned Mountain (Fig. 2). The transverse fault immediately west of Burned Mountain, and the longitudinal faults ~~and the fault in Sheep Gulch~~ are possibly post-Los Piños Formation. ~~although not enough outcrops are available at Sheep Gulch to ascertain the relationship of the fault to Los Piños rocks (Fig. 2).~~ The age of the easternmost fault along the ~~eastern~~ flank of Burned Mountain is probably pre-Los Piños Formation, since it apparently does not affect the Los Piños rocks in the Vallecitos River area. Other faults are believed to be present in the Hopewell Anticline, but field evidence is not conclusive. Bromide and Placer canyons may have formed along fault zones, but due to poor exposures it was not possible to determine this positively. Since the apparent displacement would have been small if faults are present in these canyons, no faults are indicated on the map of the area (Fig. 2). Probably the dikes in the Tusas rocks ~~Granite~~ north of Hopewell Lake (Fig. 2) occupy fractures or joints in the granite, but the granite was too poorly exposed to permit determination of the presence of fracture systems in this area.

In summation, for the Hopewell Anticline which has the Cleveland Gulch deposit on the southwestern flank and the Iron Mountain deposit on the northeastern flank, it is a simple more or less symmetrical or isoclinal anticline

(Tertiary cover and Tusas granitic intrusions make it impossible to be more specific); it plunges toward the northwest at about 45°; it is flanked on both sides by simple synclines; it is cut by several northeast and northwest trending faults; schistosity is parallel to bedding planes; besides a fissility parallel to the bedding planes, there is no field evidence for the belief that the northwest trending fold has been superimposed on a previous northeast trending fold system; no major drag folds were found in the field.

Cañon Plaza deposit is in a northwest trending shear zone and is the smallest of the iron deposits. A few hundred feet southwest of the mineralized shear zone is another shear zone which contains less iron, but iron content is higher than in the country rock (Fig. 6, Localities 1 and 2).

Based on observations on conglomerate beds, the strike of the enclosing Ortega Quartzite is N 67° W; the shear zones trend about N 78° W (Fig. 6). Schistosity in the quartzite here is parallel to the bedding planes. Referring to the Ortega Quartzite in the La Madera Quadrangle, Bingler (1965, p. 19) stated "the foliation is an axial-plane cleavage and not bedding foliation," but the Ortega Quartzite near Cañon Plaza apparently has foliation parallel to the bedding planes. Foliation in the Ortega Quartzite is parallel to conglomerate beds and bedding planes as outlined by heavy minerals and cross-beds. Along the southwestern flank of the Hopewell Anticline the foliation in the Ortega Quartzite is parallel to the bedding planes. Bingler undoubtedly is correct in his analyses in areas where there is tight folding; this tight folding could result in axial-plane cleavage, and the foliation would also be more or less parallel to the bedding.

In summation for the Cañon Plaza area, the beds of Ortega Quartzite strike approximately N 67° W; they are cut by northwest trending shear zones along which there is mineralization; foliation is parallel to the bedding planes.

The prominent northwesterly fold system and the northeasterly fault system parallel the tectonic lineation of the Uncompahgre line and the transcontinental arch respectively. The tectonic outline of the New Mexico San Juan Mountains (Fig. 3) does not exactly parallel the Uncompahgre lineament, but the major folds and some faults are parallel to the lineament. It is believed that the transverse faults which cut the major folds of the area reflect the northeastern transcontinental arch lineament. This northeastern lineation is present in the Picuris Uplift, Sangre de Cristo Uplift, and possibly in the Sandia Uplift. This fact would support Bingler's contention that there was a period of tectonism which resulted in northeast trending structures in northern New Mexico (Bingler, 1965, p. 124). However, as stated above, there is no field evidence in the vicinity of the Cleveland Gulch-Iron Mountain iron deposits which would indicate that the northeast trending structures were formed earlier than the northwest trending structures.

# IRON DEPOSITS

## General Statement

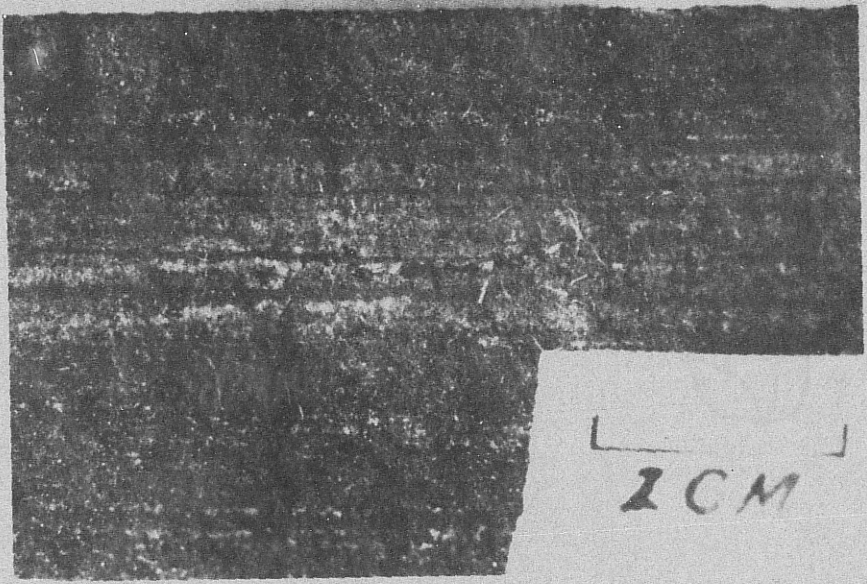
Most samples of the iron deposits at Cleveland Gulch, Iron Mountain, and Cañon Plaza are banded. Banding varies in the degree of continuity as well as in the degree of concordance. Plates 1A and 14A are examples of bands which are concordant and continuous over the length of the hand specimen. In prospect pits it is possible to trace individual bands of this type for distances of a couple of yards. However, because these exposures are the maximum continuous specimens, it is not possible to determine whether banding continuity exceeds two yards. Plates 2B and 14B show specimens in which the bands are not concordant and are discontinuous. All variations between these two extremes are found in the deposits. Plate 13B is an example of banded specularite-quartz-kyanite-sillimanite-andalusite rock from Cañon Plaza. Banding is poor but still discernible in this specimen. Most of the specimens from Iron Mountain and Cleveland Gulch resemble the specimen in Plate 14A.

Throughout this report the term "banded iron formation" has been used interchangeably with "banded iron deposits" or "iron deposits." The stratigraphic term

Fig ~~PLATE~~ 14

A. Well banded iron formation. South Cleveland Gulch. S20 193 61.

B. Poorly banded iron formation. Iron Mountain. P22 203 F20.

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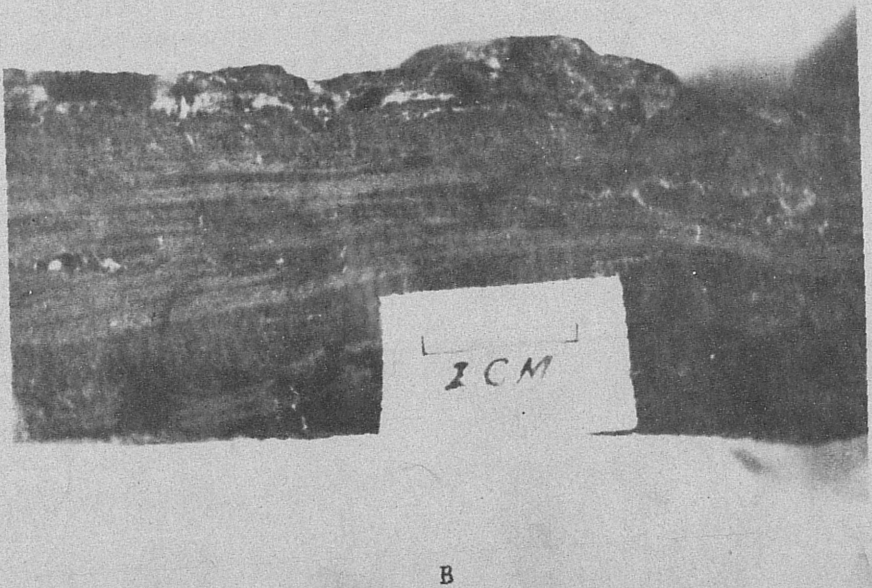



B

"formation" is so deeply entrenched in the literature and usage that any attempt to abandon the term probably would be unsuccessful. The most unfortunate aspect of the usage is that there is no generally accepted definition for "banded iron formation."

The origin of the term "iron formation" is obscure. Winchell and Winchell used the term in 1891 (p. 234) but did not define it. As far as has been ascertained, the term probably was derived from a stratigraphic diagram by Foster and Whitney (1851, p. 2). Their "Classification of the Rocks" was broken down into "Formations" and this was subdivided into "Aqueous," "Metamorphic," and "Igneous." The "Igneous" was further divided into "Plutonic Rocks" and "Trappean or Volcanic Rocks." One unit under the "Trappean and Volcanic Rocks" was "Masses of Specular and Magnetic Oxide of Iron." One can see that "masses of specular and magnetic oxide of iron of the trappean or volcanic rocks of the igneous formation" might readily have been shortened to "iron formation," which in those days of fairly uncomplicated terminology would have meant "iron masses in volcanic rocks." Foster and Whitney (1851, p. 2) define formation as an explanation of origin of the rocks. If this is the origin of the term "iron formation," it is interesting that the first indirect usage indicated the iron was volcanic or trappean in origin.

James (1954, p. 239) pointed out that rocks called "iron formation" have many different mineralogical variations and not even the commonly attributed "chert" is universally present. He considered that the primary feature in the definition of iron formation was "iron" and that the iron was sedimentary and most often was laminated. Consequently, he defined iron formation as "a chemical sediment, typically thin-bedded or laminated, containing 15 percent or more iron of sedimentary origin, commonly but not necessarily containing layers of chert." This definition was suitable for his classic work on the sedimentary iron deposits around Lake Superior.

It would be desirable to abandon the term "iron formation," but this seems impossible and, besides, a general term for banded iron deposits is needed. The following definition will be used in this report and will follow the guidelines set down by James (1954, p. 239), but without the stipulation of a sedimentary origin. Iron formation is an iron-rich rock mostly thin-bedded or laminated, containing 15% or more iron of sedimentary or epigenetic origin, and typically containing interlayers of quartz. This definition excludes basalts and other mafic rocks which have a high iron content, yet includes the rocks covered in James' work and those banded quartz and iron oxide rocks which may have originated by replacement processes.

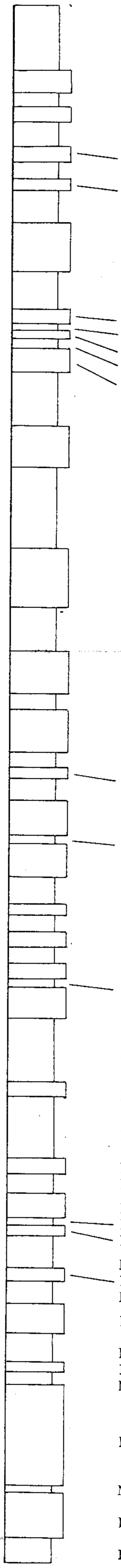
Use of the word "band" has come under discussion recently. Calkins (1941, p. 345) and Gundersen (1960, p. 565) both stated that "band" refers to two dimensions. Neither the "American Geological Institute Glossary of Geology" nor "Webster's Unabridged Dictionary" completely supports the 2 dimensional idea, although most people would agree that band does refer to what one sees on a flat surface. Nonetheless, there is little likelihood that anyone will misunderstand what is meant by "banded iron formation."

## North Cleveland Gulch Deposit

North Cleveland Gulch deposit is a series of lenses of banded iron formation that more or less parallel the enclosing Moppin rocks from the Cenozoic cover on the east to a short distance east of Bromide Canyon (Fig. 5). Over this 2 mile distance the thickness of iron formation varies from 0 to about 15 feet, although the outcrop width for this 15 feet of iron formation and the intervening beds is up to 400 feet. Most of the iron formation and intervening beds are nearly vertical. Thickness of the iron formation probably averages about 6 to 7 feet with 32% iron over an outcrop width of 100 to 150 feet. This estimate is based on measurements in three bulldozer pits, numerous isolated outcrops, and several prospect pits along the 2 mile length of the deposit. There is a possibility that more iron formation exists in the area outlined north of Cleveland Gulch on Figure 5. The above figures are to be considered as reliable only where enough of the deposits were exposed in prospect pits, outcrops, or the bulldozer slices to distinguish the different rock types and to measure the iron formation. Other parts of the Moppin Formation north of Cleveland Gulch have iron formation present but in smaller amounts.

The best exposure of the North Cleveland Gulch deposit was in a bulldozer cut in the widest part of the iron deposit mapped on Figure 5 (Plate 16A). Table III lists the lithologies found in a continuous section of about 13 1/2 feet through the most iron-rich part of the outcrop. The intervening schists may be meta-pelites or meta-tuffs. This sequence listed on Table III is somewhat atypical in that most of the rest of the iron-rich unit (unit 5 in the petrographic descriptions) have more chlorite-biotite schists and amphibolite beds intervening but are not as well exposed as the sequence listed on the table. Petrographic descriptions for the intervening schists, phyllite, quartzite, and amphibolite have been presented in the section on Moppin stratigraphy.

Most samples from the iron deposits north of Cleveland Gulch are banded (Plate 15A). Banding is formed by




Total: 161.4 inches

Scale: 0.1":1.0"

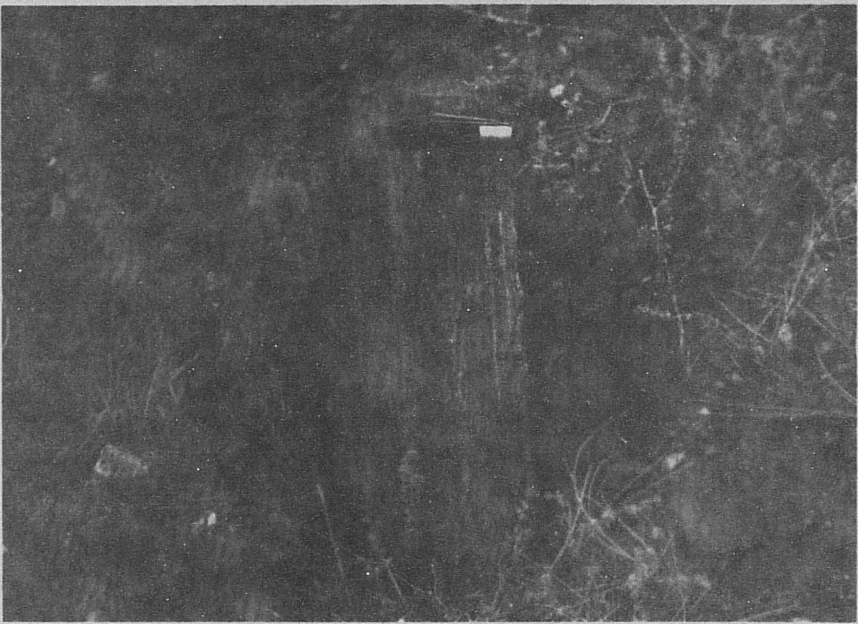
(164)* Samples having spectrographic analyses. Number is the last 2 or 3 numbers of 20193 series north of Cleveland Gulch (See Table I).

Table III. Section of the part of Moppin Formation with the highest iron formation content north of Cleveland Gulch.

~~PLATE~~ Fig. 15

A. Banded iron formation. North Cleveland Gulch.

B. Altered iron formation. Vuggy with earthy hematite. North Cleveland Gulch.



A

B

the alternation of magnetite-rich layers and quartz-rich layers; individual layers range from 0.1 mm to over 10.0 mm in thickness, but typical hand specimens do not show this range. Most bands are in the 1.0 mm to 5.0 mm range. Banding continuity and concordance range from good to poor (Plate 1A and B), but most samples display good banding. However, some bands pinch and swell, others join, and some cut across other bands. Near one contact with the enclosing rocks banding is poor or absent, iron is mostly in the form of earthy hematite, and the rock is vuggy (Plate 15B). This vuggy rock with the earthy hematite may have resulted from meteoric or hydrothermal action along the contact of the iron formation and the enclosing rock. At any rate this phenomenon is rare.

Magnetite, partly weathered to martite, is in general euhedral, ranges from 0.05 mm to 0.5 mm, and accounts for 25-50% of the rock. In some specimens, magnetite displays two distinct size ranges, one of them usually quite small. Quartz forms an equigranular mosaic ~~of grains~~ ranging from 0.07 mm to 0.4 mm in diameter and accounts for 30-70% of the rock. Other minerals in the iron formation are biotite, 0-20%; calcite, 0-25%; chlorite, 0-10%; feldspar, 0-15%; muscovite, 0-7%; epidote, 0-2%; tourmaline, 0-3%; apatite 0-3%; rutile a trace; and garnet a trace. In the order of decreasing frequency, chlorite, muscovite, calcite, and biotite are the most common. Tourmaline and feldspar cross-cut mineral grains in a few specimens.

Quartz-rich bands usually contain a few percent magnetite, some of any other minerals present in the total rock specimen, and almost all of the apatite. Magnetite-rich bands usually contain a small percentage of quartz and almost all the other minerals--especially chlorite, muscovite, biotite, and calcite if they are present in the sample. The affinity of the magnetite for micas and chlorite is so pronounced that a genetic significance might be attributed to this relationship. This is discussed in the section on the origin of the iron formation.

Near the eastern end of the iron formation outcrops (Fig. 5), the specimens are moderately to well-banded. Individual layers are 0.1 mm to 2.0 mm thick, and some pinch and swell. Quartz forms an equigranular mosaic, with individual grains averaging 0.1 mm, and accounts for about 50% of the rock. Euhedral magnetite, averaging 0.08 mm, makes up 40% of the rock. Chlorite and calcite each account for 5%, and, although chlorite and calcite are present in both light and dark bands, they are most common in the magnetite-rich bands.

About a thousand feet west of the above outcrop, in a large bulldozer cut, one typical sample had 45% equant quartz grains averaging 0.2 mm, a trace of muscovite and garnet, 40% euhedral magnetite averaging 0.1 mm, and about 15% cross-cutting anhedral albite. Banding was fair to good and ranged between 3.0 mm and 10.0 mm in thickness. The albite is somewhat anomalous, because it was found only in two specimens in the North Cleveland Gulch area. A quartz vein cuts the iron formation which contains the albite and it is believed that this vein contributed the albite. Most other samples in the immediate area have small amounts of muscovite and epidote but no feldspar. Some samples have a few plates of specularite.

Another thousand feet west is an outcrop of a poorly banded part of the iron formation (Plate 1B). In this locality banding is formed by layers of magnetite with chlorite and quartz and layers of quartz with magnetite (Plate 16B). Banding continuity and concordance is mostly poor, but a few good bands are present; band thickness varies from 0.5 mm to over 10.0 mm. Equant quartz averaging 0.3 mm in diameter accounts for 50-70% of the rock; euhedral magnetite averaging 0.07 mm, 30-45%; chlorite, 2-7%; and apatite, 0-3%. Chlorite is mostly in magnetite bands, and apatite is virtually restricted to the quartz bands.

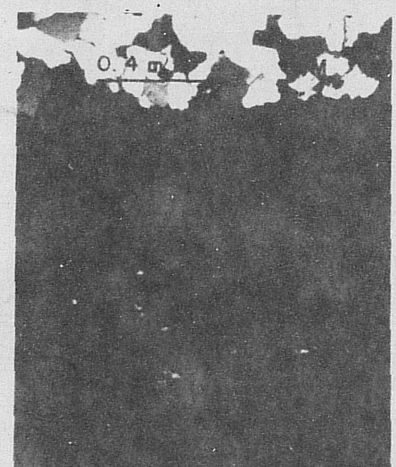
West of the irregular intrusive ~~rhyolite~~ (Fig. 5), the iron formation displays fair to good banding; in one specimen some bands cut others. Individual bands are from 0.1 mm

~~PLATE~~ Fig. 16

A. Bulldozer pit in iron formation. Moppin Formation. North Cleveland Gulch. Darker units are iron formation. Lighter areas are mostly schist. Continuous section on Table III described from this exposure.

B. Banded iron formation. Moppin Formation. North Cleveland Gulch. Quartz (Q), magnetite (M), chlorite (C). S20 193 44 Plane light.

A




B

to about 3.0 mm thick. Quartz averaging 0.1 mm in diameter accounts for 40-50% of the rock; euhedral magnetite averaging 0.07 mm, 40-50%; chlorite, 3-7%; muscovite, 0-3%; and a trace of apatite compose the remainder of the rock. In these specimens the chlorite and muscovite are consistently more common in the magnetite-rich bands (Plate 6A). One sample from this area has crushed quartz grains and microfaulted bands. This deformation may have been the result of the emplacement of the adjacent intrusive ~~rhyolite~~ (Fig. 5). ~~since it is adjacent to this section of iron formation.~~

None of the hand specimens of the iron formation from different localities in the North Cleveland Gulch deposit look alike. Grain size of the magnetite varies, and layer thickness is variable as well as the range of layer thickness in a single specimen. The degree of weathering of the magnetite is also highly variable. Kind and amount of accessory (non-magnetite, non-silica) minerals are also highly variable. Consequently, it is not possible to identify an individual lens from one outcrop to the next where covered areas intervene. The two things that samples from various outcrops have in common is banding formed by alternation of magnetite-rich layers with quartz-rich layers and the concentration of chlorite and the micas in the magnetite-rich bands.

For the most part, weathered schists and phyllites are associated with the iron deposits. Due to weathering characteristics, iron formation commonly is the only constituent or at least the most prominent component of most outcrops. This condition made it difficult to determine degree of concordance or discordance of the iron formation to surrounding rocks. However, at one locality, the iron formation had an outcrop length of several hundred feet and in this distance the strike changed from about N 85° E to S 80° E. A fine-grained amphibolite bed which was about 100 feet below maintained an east-west strike over this same distance. This example shows some discordance between the iron formation and the enclosing silicate rocks. Over a distance of about 10 feet in the bulldozer slices and in the prospect pits the iron formation was conformable. Outcrops were too scarce to determine the structural relationships in any more detail. The pronounced weaving shown by the limits of iron mineralization on Figures 2 and 5 could reflect lensing of the iron formation and some of the enclosing rocks as much as any discordance.

## South Cleveland Gulch Deposit

Across Cleveland Gulch to the south of the large iron deposit described above is a smaller body of iron formation (Fig. 5). This skull-shaped area with iron formation is confined mainly to the meta-arkose unit in the upper part of the Moppin Formation, but it also extends into the basal Ortega Quartzite. This area is about 900 feet in its longest north-south dimension and about 650 feet in its widest east-west dimension. Presence of a deposit south of Cleveland Gulch was first recognized when several pieces of banded iron formation float were found. However, the exact locality of the strata from which the float pieces originated was difficult to find due to soil and vegetation cover. A curious skull-shaped dark area was noted on the aerial photographs, and upon inspection the area of iron formation was located in the field.

Soil cover on the meta-arkose unit is a light shade all along the strike and only meta-arkose float or outcrops are found. The soil of the mineralized area, which cuts the meta-arkose, is dark gray-black and here banded iron formation float is abundant. This change is abrupt and can be traced in the field. No outcrops of iron formation were found, although pieces weighing up to about a half ton are present. Absence of iron formation outcrops is probably due to the fact that the deposit is surrounded by meta-arkose and quartzite, both of which are relatively more resistant in the climate of the area. The North Cleveland Gulch iron deposits, on the other hand, are surrounded by schists and amphibolite which are less resistant than the iron formation, and hence, the iron formation crops out.

The possibility that the skull-shaped area represents an ancient float or stream deposit instead of an *in situ* deposit has been discarded. This is based on geometrical relationships, weathering characteristics, and trace element dissimilarities. The limited areal extent of the skull-shaped area, the total absence of iron formation float in adjacent areas, and the fact that the skull-shaped area narrows at its southern (downslope) end away from the only possible allochthonous source for the iron formation float preclude a float origin from the North Cleveland Gulch deposit. In addition, upon exposure to the atmosphere, the magnetite in the iron formation usually weathers rapidly to martite. Almost all the float from the North Cleveland Gulch deposit in the present stream channel at the bottom of Cleveland Gulch has magnetite grains which are 75% or greater martite. If the magnetite of the South Cleveland Gulch area had been derived from the North Cleveland Gulch deposit, one would expect the magnetite to contain a higher percentage of martite than the float in the bottom of Cleveland Gulch and a higher percentage of martite than the North Cleveland Gulch deposit because the time and distance of travel involved would have been longer. This, however, is not the case. South Cleveland Gulch deposit has magnetite grains which have less martite than most of the magnetite north of Cleveland Gulch. Therefore, for these reasons and trace element differences which will be discussed later, a North Cleveland Gulch origin for these deposits has been discounted.

Few float rocks that resemble meta-arkose or quartzite are found in the iron deposit area. The only other common rock type in the skull-shaped area is an amphibolite which was described in the section on the Moppin Formation. In appearance and mineral composition, the amphibolite is similar to those north of Cleveland Gulch.

Banded iron samples from the skull-shaped area usually have about 45% quartz, 45% euhedral magnetite, 8% chlorite, and accessory muscovite and epidote. Chlorite is virtually restricted to the magnetite bands (Plate 18A). Banding continuity and concordance are usually good (Plate 14A). Because no outcrops are available, it is not possible to determine any variation from the outside of the deposit to the center or upon crossing the contact of the meta-arkose and chlorite schist and the contact between the schist and the feldspathic quartzite of the basal Ortega Quartzite.

## Iron Mountain Deposit

Iron deposits and the enclosing rocks on the western flank of Iron Mountain were mapped with the aid of plane table and alidade (Fig. 4). Moppin Formation here is mainly chlorite schists which were tuffs intruded by dikes and sills of silicic to intermediate composition. Banded iron formation is restricted to the meta-tuffs. On Figure 4 the prospect pits are in the thickest part of the iron formation on Iron Mountain. The body is about 10 feet thick (Plate 17A) and is believed to extend northeast and southwest of the outcrop for several hundred feet. Few silicate beds are interbedded in this 10 feet of iron formation. Iron deposits on Iron Mountain contain from 22% to 37% iron and average 28%.

Other areas of iron mineralization on Iron Mountain are smaller in thickness and areal extent. The largest body southeast of the main body on Figure 4 is up to 15 feet thick, but only about 1 or 2 feet of this thickness is banded iron formation. The main body of iron mineralization cuts across the strike of the meta-tuff bed (Fig. 4). Strike of the meta-tuffs was determined

~~PLATE~~ Fig. 17

A. Prospect pit in main banded iron formation deposit on Iron Mountain.

B. Cross-cutting small body of banded iron formation on Iron Mountain. Black line approximately along the iron formation. Camera facing parallel to the schistosity of the enclosing chlorite schist.

A

B

from the strike of the schistosity which is parallel to the strike of the overlying metarhyolite bed about 100 yards north of Iron Mountain. The strike of the Ortega Quartzite, which is stratigraphically above the Moppin Formation and everywhere strikes parallel to the underlying Moppin rocks, was used to support the strike of the metarhyolite. At no locality on the southwestern flank of the Hopewell Anticline was the schistosity other than parallel to the bedding. East of Iron Mountain and lower in the stratigraphic section are some thin metarhyolite and quartzite beds which strike parallel to the schistosity. These data support the contention that the schistosity and the strike of the bedding are parallel. Consequently, the main banded iron deposit is a cross-cutting vein-like body.

Many of the smaller iron formation bodies cut across the strike of the enclosing beds (Plate 17B). In Plate 17B the camera faces along strike of the chlorite schist and the various markers are at bend points of the strike of a small banded iron body. Where it is possible to trace one of these small bodies in the field, it grades into banded chlorite-quartz schist (Plate 19A).

Banding is poor (Plate 14B) to good, but poor banding is rare. Bands range from 0.1 mm to 5.0 mm in thickness. In the poorly banded specimens the layers pinch out over short distances, cut across other layers, and are very irregular under the microscope (Plate 18B). Banding is formed by an alternation of quartz-rich and magnetite-rich layers. In whole rock samples, quartz ranges from 40-50%, magnetite from 30-50%, chlorite from 0-25%, muscovite from 0-7%, and epidote is present in traces.

Chlorite is in both light and dark bands, but it is ~~generally~~ concentrated in the dark layers. Muscovite is in fine-grained flakes parallel to the schistosity and in euhedral books at various angles up to 90° to the planes of schistosity. These books of muscovite show interesting textural relationships with the quartz and magnetite bands. Quartz bands pass through muscovite books. Magnetite bands persist through the muscovite books with the only effect on the magnetite being a reduction in grain size (Plate 18C). In places only "ghosts" of muscovite books remain (Plate 18D); these "ghosts" are outlined by magnetite grains which are reduced in size. The muscovite "ghosts" and partly altered books of muscovite may be together in the same thin section. ~~Apparently~~ the muscovite books originated after the development of schistosity since they are cross-cutting. ~~Possibly~~ the magnetite was probably introduced after the muscovite books as is strongly suggested by the muscovite

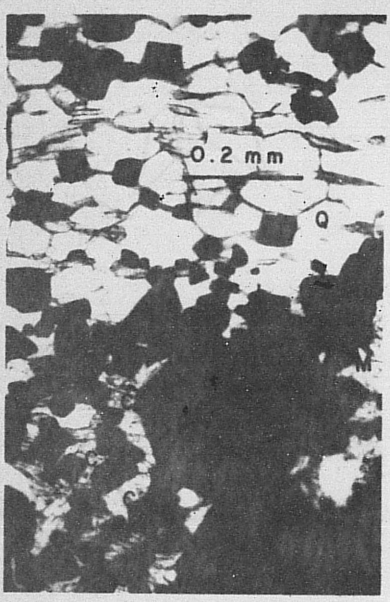
Fig. ~~PLATE~~ 18

A. Banded iron formation. Moppin Formation. South Cleveland Gulch. Magnetite (M), chlorite (C), quartz (Q). Chlorite concentrated in magnetite-rich bands. S20 193 61 Plane light.

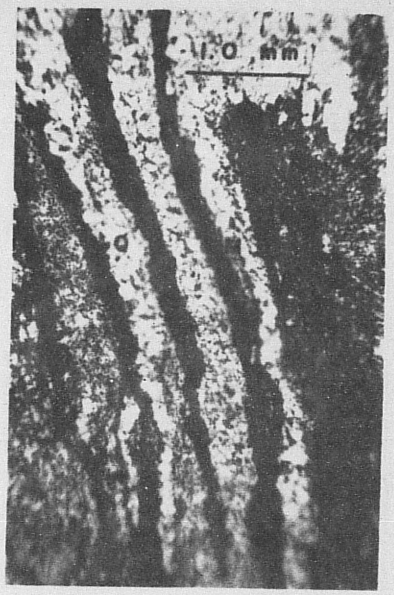
B. Poorly banded iron formation. Moppin Formation. Iron Mountain. Quartz (Q), magnetite (M). Note irregular banding. P22 203 F20 Plane light.

C. Banded iron formation. Moppin Formation. Iron Mountain. Cross-cutting muscovite book (M) partially replaced by magnetite (M) and quartz (Q). 22 203 12 Plane light.

D. Banded iron formation. Moppin Formation. Iron Mountain. Cross-cutting muscovite book completely replaced by magnetite and quartz bands. 22 203 15 Plane light.




A




B


C

D

"ghosts" and their relationship to the magnetite.

Small iron bodies were found on the eastern flank of Iron Mountain (not mapped by plane table) and in the Moppin rocks east of Iron Mountain. None of these exposures is large, and from the paucity of float it is believed that the deposits are thin and limited in extent.

## Cañon Plaza Deposit

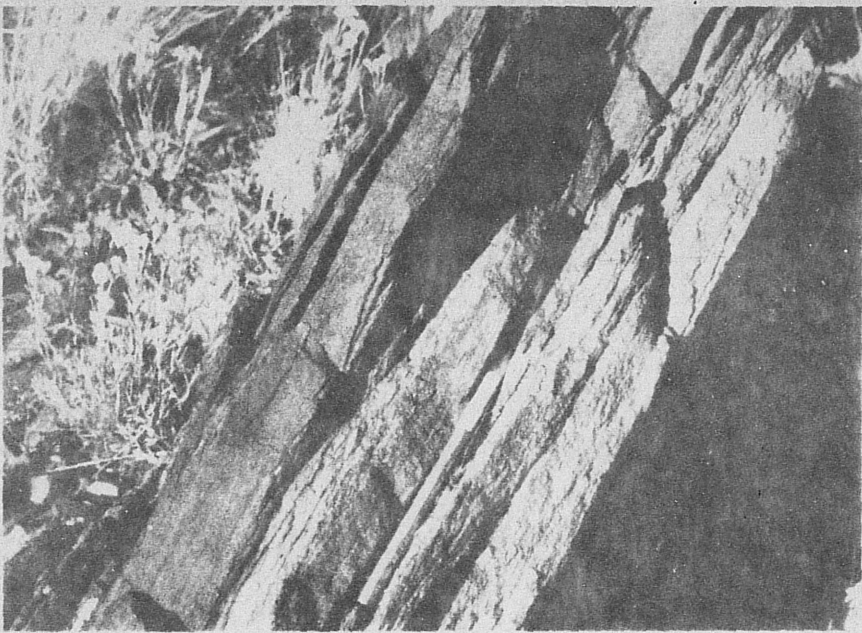
Cañon Plaza deposit is limited in extent; it is also strikingly different in character from the three previously described deposits. Outcrops are scarce, and throughout most of its length the deposit is only about 1 to 2 feet wide. At only one locality was it up to 10 feet wide (Plate 19B). The deposit is in a shear zone which in many places is nearly parallel to the strike of the enclosing Ortega Quartzite. At other localities the shear zone cuts across the strike of the Ortega beds. In Plate 19B the black pencil is laying along the strike of the Ortega beds and the schistosity of the iron deposit is nearly vertical. The strike of the iron deposit is perpendicular to the plane of the photograph. Consequently, the cross-cutting nature of the iron deposit is apparent.

About 300 yards south of this mineralized shear zone is another zone which is like the one described

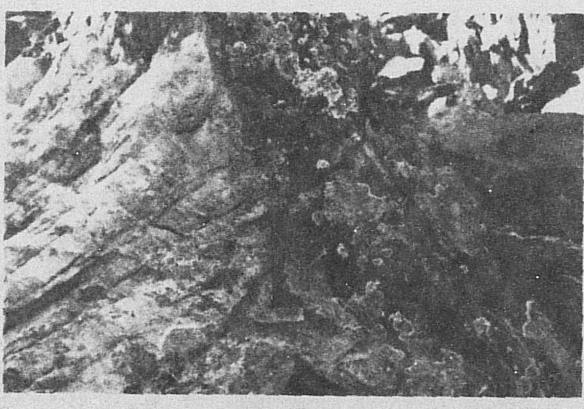
~~PLATE~~ Fig. 19

A. Iron Mountain. Gradation of small lens of banded iron formation (to immediate left of pencil) into chlorite schist (at top of photograph above point of pencil).

B. Cross-cutting iron formation at Cañon Plaza. Strike of enclosing Ortega Quartzite is parallel to pencil. Strike of iron formation normal to plane of the photograph.



A



B

above in every respect except for absence of much specularite and metasomatic minerals. Both of these zones were formed by northwest trending faults, but only the northern zone was mineralized.

Banding is poor to fair (Plate 13B) with component layers usually pinching and swelling over short distances. Banding is formed by alternation of specularite-rich layers and quartz-aluminum silicate-rich layers. In many places the banding is highly contorted indicating post-band deformation. Also, the specularite causes pronounced schistosity in the deposit. Iron content ranges from 17% to 25% (Table V) and averages 22%. Projecting this deposit to a depth of 200 feet would yield only about 80,000 tons of "ore."

Cañon Plaza has a unique mineral assemblage for the iron deposits of the area. Instead of magnetite, specularite is the iron oxide (Plate 20A). In addition to quartz, the rock contains andalusite, kyanite, sillimanite, vesuvianite, gahnite, tourmaline, rutile, garnet, muscovite, and apatite. Besides the unique mineral assemblage, several interesting textural relationships are present. One of the most interesting features is the presence of andalusite, kyanite, and sillimanite in a single specimen. In some hand specimens, andalusite up to 7 mm and kyanite up to 5 cm long are present. Kyanite usually changes from blue to pink along a single grain. Sillimanite forms feathery masses in quartz grains and was observed only in thin section. Presence of the three polymorphs of aluminum silicate in a single specimen is rare. Hietanen (1956) made a complete study of an assemblage of the three aluminum silicates in the Belt Series of Idaho and attributed their presence to fluctuation of temperature and pressure around a field where all modifications may exist in equilibrium, and this occurred during complex regional and thermal metamorphism (1956, p. 27). Probably disequilibrium could account for their presence together as well. At Cañon Plaza, no intergranular sillimanite was found. This suggests that equilibrium may not have been attained. If this assemblage does represent an equilibrium assemblage, it would have formed near the triple point for the three polymorphs on a pressure-temperature diagram and would indicate a temperature of about 300° C at a pressure of about 8 kilobars (Morey, 1964, p. 25). However, these phase boundaries have not been established experimentally and should not be accepted as definite until additional laboratory work has been completed. It is highly improbable that the assemblage at Cañon Plaza was in equilibrium, and as a consequence, phase boundaries would be meaningless for this area.

Samples of Ortega Quartzite in the vicinity of the Cañon Plaza deposit contain kyanite and sillimanite grains. Possibly much of the kyanite and sillimanite in the shear zone was inherited from the quartzite. However, there is apparently more kyanite than is present in the quartzite. This excess kyanite probably was introduced by aluminous-rich mineralizing fluids that invaded the shear zone. Even though high pressure during metamorphism favors the formation of kyanite over the other polymorphs, thus suggesting that kyanite is formed only under this condition, kyanite is found in pegmatites where no stress seems to have been present (Barth, 1952, p. 257) and also in gold-quartz veins (C. F. Park, personal communication). Corey (1960, p. 53) believed that kyanite deposits on La Jarita Mesa were formed in part by "injection of siliceous hydrothermal solutions containing assimilated aluminous material." Consequently, the andalusite and at least part of the kyanite are believed to have been emplaced by the mineralizing fluids which entered the shear zone.

Metasomatism is supported by the presence of gahnite. Deer, Howie, and Zussman (1962, p. 67) stated "The zinc spinel, gahnite, occurs chiefly in granitic pegmatites. . . ., but is also found in contact altered limestones and in metasomatic replacement veins and ore bodies." Presence of gahnite, presence of a shear zone, irregular pinching and swelling of the deposit, the cross-cutting nature of the zone, and the apparent absence in the unaffected Ortega rocks of the elements necessary to make up the unique mineral assemblage indicate that the shear zone has received some form of epigenetic mineralization.

Near the contact of the deposits with the Ortega Quartzite, the rocks are composed of about equal amounts of quartz, andalusite, and specularite with small amounts of muscovite, kyanite, rutile, sillimanite, gahnite, vesuvianite, garnet, tourmaline, and apatite in the order of decreasing abundance. Near the center of the deposit, quartz accounts for the largest amount of the non-opaque minerals; specularite is present but accounts for only about a third of the rock. Kyanite is the next most common mineral, and there are small amounts of andalusite, muscovite, and vesuvianite.

The aluminum silicates, in general, have their long axes aligned parallel to the schistosity of the specularite. Near the contact of the deposit, the layers are highly contorted indicating that there was deformation during the late stage of band formation or after the formation of the mineral suite. However, near the center of the deposit the layers show very little effect of late deformation. Quartz recrystallized to large optically continuous grains with muscovite or sillimanite grains usually within the quartz. There is marked contrast between these large clear quartz grains and the fine-grained mosaics of quartz which typify the iron formation at Cleveland Gulch and Iron Mountain. Near the contact, but away from the highly contorted part, the specularite bands pass through some of the large quartz grains and other minerals without apparent deflection or other effect. Nearer the middle of the deposit the components of the bands become more segregated, and the large optically continuous quartz grains appear to be partly changed to a fine-grained equidimensional mosaic resembling the quartz mosaics of the Cleveland Gulch-Iron Mountain iron formation.

## Spectrographic Analyses

Table IV is a compilation of semi-quantitative spectrochemical analyses of 63 whole-rock samples from the iron deposits and enclosing rocks at Cleveland Gulch, Iron Mountain, and Cañon Plaza. Sample 212015 was checked only for Be, B, Li, and Sn. Each of the other samples was checked for 50 elements, but only 26 of the elements are present in one or more of the samples. Amounts of the elements were determined in the following

BANDED IRON FORMATION
NORTH CLEVELAND GULCH

2019327
2019359
2019366
2019377
2019380
2019388
2019397
20193103
20193107
20193116
20193133
20193142
20193164
20193167
20193185
20193195
S2019320
S2019337A
S2019337B
S2019344A
S2019344B
S2019347
S2019348
S2019349
21185
21186
21189
211810

BANDED IRON FORMATION
SOUTH CLEVELAND GULCH

S2019350
S2019351
S2019352
S2019353
S2019361

BANDED IRON FORMATION
IRON MOUNTAIN

22203C87
22203I18
22203I19
22203L86
222030177

BANDED IRON FORMATION
CANON PLAZA

2110M
2110M1
2110M2
2110C3
2110C6
2110C7

METAMORPHIC ROCKS
NORTH CLEVELAND GULCH

| | |
|---|---|
| 2019330 | Amphibolite |
| 2019351 | Chlorite Schist |
| 2019366A | Biotite Schist |
| 2019373 | Quartzite |
| 2019382 | Biotite Schist |
| 2019385 | Biotite Schist |
| 20193113 | Quartzite |
| 20193188 | Muscovite Schist |
| 20193198 | Phyllite |
| S2019344C | Banded Quartz-Muscovite Schist |

METAMORPHIC ROCKS
SOUTH CLEVELAND GULCH

| | |
|---|---|
| S2019341 | Meta-arkose |
| S2019342 | Meta-arkose |
| S2019345 | Amphibolite |
| S2019346 | Meta-arkose |
| S2019354 | Amphibolite |

METAMORPHIC ROCKS
CANON PLAZA

| | |
|---|---|
| 2110SZ | Sheared Quartzite |
| 2110CYR | Quartzite |

INTRUSIVE ~~RHYOLITE~~
NEAR CLEVELAND GULCH

21184
212015 Greisen - Tusas Mt.

| SYMBOLS | PERCENTAGE RANGE |
|---|---|
| +++++ | 1% or more |
| ++++ | 0.1% - 1.0% |
| +++ | 0.01% - 0.1% |
| ++ | 0.001% - 0.01% |
| + | Less than 0.001% |
| D | Detected |

Besides the elements listed, each sample was checked for: Li, Rb, Cs, Sr, Ba, Cr, Au, Cd, Hg, Bi, As, Te, Hf, Nb, Ta, Re, Pt, Pd, Sc, Y, R.E., Th, U, In, Tl, Zr. Except for Cr and Zr, none of these were present.

Table IV. Spectrographic analyses of 63 samples from Cleveland Gulch, Iron Mountain, and Cañon Plaza, Rio Arriba County, New Mexico.

TABLE IV

| | 2019327 | 2019359 | 2019366 | 2019377 | 2019380 | 2019388 | 2019397 | 20193103 |
|---|---|---|---|---|---|---|---|---|
| Be | | | | | | | | |
| B | | | | | | | | |
| Na | +++ | +++++ | +++++ | +++++ | +++++ | ++++ | ++++ | +++++ |
| Mg | ++++ | ++++ | ++++ | +++++ | +++++ | +++ | ++++ | +++ |
| Al | +++++ | +++++ | +++++ | +++++ | +++++ | ++++ | ++++ | ++++ |
| Si | +++++ | +++++ | +++++ | +++++ | +++++ | +++++ | +++++ | +++++ |
| P | | | | | | | | |
| K | +++ | ++ | + | D | D | ++ | ++ | |
| Ca | D | | | D | D | | | |
| Ti | | + | + | + | + | ++ | ++ | |
| V | +++ | +++ | +++ | ++ | ++ | D | D | |
| Mn | ++ | + | ++ | + | + | ++ | +++ | + |
| Fe | +++++ | +++++ | +++++ | +++++ | +++++ | +++++ | +++++ | +++++ |
| Co | | | ++ | | | | | |
| Ni | + | ++ | ++ | D | D | +++ | +++ | |
| Cu | ++++ | ++++ | ++++ | ++++ | ++++ | +++++ | +++++ | ++++ |
| Zn | | D | D | D | D | | D | |
| Ga | | + | + | + | + | | | |
| Ge | | | | | | | | |
| Mo | + | +++ | ++ | | + | +++ | +++ | |
| Ag | D | ++ | + | D | + | ++ | +++ | D |
| Sn | | | | ++ | | | ++ | |
| Sb | | | | | | | | |
| Pb | D | D | D | ++ | ++ | D | +++ | |

| | 20193107 | 20193116 | 20193133 | 20193142 | 20193164 | 20193167 | 20193185 |
|---|---|---|---|---|---|---|---|
| Be | | | | | | | |
| B | | | | | | | |
| Na | +++ | +++ | ++++ | ++++ | +++ | +++ | ++++ |
| Mg | +++ | +++ | +++ | +++ | ++++ | +++ | +++ |
| Al | ++++ | +++ | +++++ | +++++ | ++++ | ++++ | ++++ |
| Si | +++++ | +++++ | +++++ | +++++ | +++++ | +++++ | +++++ |
| P | | | | | | | |
| K | | | | | | | ++ |
| Ca | | | | | | | |
| Ti | | | + | D | D | | |
| V | | | +++ | ++ | ++ | ++ | |
| Mn | ++ | ++ | ++ | ++ | ++++ | ++++ | ++ |
| Fe | +++++ | +++++ | +++++ | +++++ | +++++ | +++++ | +++++ |
| Co | | | | | | | |
| Ni | +++ | ++ | +++ | + | +++ | +++ | D |
| Cu | +++++ | ++++ | ++++ | ++++ | ++++ | ++++ | ++++ |
| Zn | | | | | D | | |
| Ga | | | D | D | +++ | ++ | |
| Ge | | | | | | | |
| Mo | D | + | ++ | D | | | |
| Ag | + | D | D | D | ++ | ++ | + |
| Sn | | | | | | | |
| Sb | | | | | | | |
| Pb | | | D | D | +++ | ++ | + |

| | 20193195 | S2019320 | S2019337A | S2019337B | S2019344A | S2019344B | S2019347 |
|---|---|---|---|---|---|---|---|
| Be | | | | | | | |
| B | | | | | | | |
| Na | | ++ | | | +++++ | | +++++ |
| Mg | +++ | +++ | +++ | ++++ | +++ | ++++ | +++ |
| Al | ++++ | ++++ | +++ | ++++ | ++++ | +++ | ++++ |
| Si | +++++ | +++++ | +++++ | +++++ | +++++ | +++++ | +++++ |
| P | | | | | | | |
| K | ++ | | | | | | |
| Ca | | | | | | | |
| Ti | D | + | | | + | | + |
| V | ++ | + | | + | ++++ | | ++++ |
| Mn | ++++ | +++ | +++ | +++ | ++ | D | ++ |
| Fe | +++++ | +++++ | +++++ | +++++ | +++++ | +++++ | +++++ |
| Co | | | | | | | |
| Ni | +++ | | | ++ | | | |
| Cu | +++++ | | ++++ | ++++ | ++++ | +++ | ++++ |
| Zn | | | | | | | |
| Ga | | | | | D | | D |
| Ge | | | D | | D | | D |
| Mo | D | | | | | | |
| Ag | | | | ++ | | | |
| Sn | | | | | | | |
| Sb | | | | | | | |
| Pb | | | | | ++ | | ++ |

| | S2019348 | S2019349 | 21185 | 21186 | 21189 | 211810 | S2019350 | S2019351 |
|---|---|---|---|---|---|---|---|---|
| Be | | | | | | | | |
| B | | | D | D | | | | |
| Na | +++++ | ++ | ++++ | ++++ | +++ | +++ | | +++ |
| Mg | +++ | +++ | +++++ | +++++ | ++++ | ++++ | +++ | ++++ |
| Al | ++++ | +++ | ++++ | ++++ | +++++ | +++++ | +++ | ++++ |
| Si | +++++ | +++++ | +++++ | +++++ | +++++ | +++++ | +++++ | +++++ |
| P | | | | + | | | | |
| K | | | | | +++ | +++ | | |
| Ca | | | | | | | | + |
| Ti | | | + | + | | | | + |
| V | | + | +++ | +++ | + | ++ | | + |
| Mn | +++ | +++ | +++ | +++ | ++ | +++ | ++ | +++ |
| Fe | +++++ | +++++ | +++++ | +++++ | +++++ | +++++ | +++++ | +++++ |
| Co | | | | | | | | |
| Ni | | | + | + | +++ | ++++ | | |
| Cu | ++++ | | | ++++ | ++++ | +++++ | +++++ | |
| Zn | | | | | D | D | | |
| Ga | D | | + | + | + | ++ | ++ | |
| Ge | D | | | | | | | |
| Mo | | | | | + | +++ | | |
| Ag | | | | | +++ | +++ | | |
| Sn | | | | | | | | |
| Sb | | | | | | | | |
| Pb | | | | + | D | + | | |

| | S2019352 | S2019353 | S2019361 | 22203C87 | 22203I18 | 22203I19 | 22203L86 |
|---|---|---|---|---|---|---|---|
| Be | | | | | | | |
| B | D | | | | | | + |
| Na | | | +++ | +++++ | | +++++ | +++ |
| Mg | ++++ | ++++ | ++++ | ++++ | ++++ | +++++ | ++++ |
| Al | +++ | +++++ | ++++ | +++++ | +++++ | ++++ | +++ |
| Si | +++++ | +++++ | +++++ | +++++ | +++++ | +++++ | +++++ |
| P | | | | | + | | |
| K | | | ++ | | | | |
| Ca | | | | | | + | |
| Ti | | | | +++ | | | |
| V | + | | + | +++++ | | + | +++ |
| Mn | +++ | +++ | +++ | ++ | ++ | + | +++ |
| Fe | +++++ | +++++ | +++++ | +++++ | +++++ | +++++ | +++++ |
| Co | | | | ++ | | | |
| Ni | + | | +++ | + | | | + |
| Cu | | ++++ | ++++ | +++++ | +++++ | | |
| Zn | | | | | | | |
| Ga | | | ++ | +++ | D | + | + |
| Ge | | D | | | D | | |
| Mo | | | ++ | | | | |
| Ag | | | +++ | +++ | | | |
| Sn | | | | | | | |
| Sb | | | | | | | |
| Pb | | | | D | ++ | | |

| | 222030177 | 2110M | 2110M1 | 2110M2 | 2110C3 | 2110C6 | 2110C7 | 21184 |
|---|---|---|---|---|---|---|---|---|
| Be | | +++ | | | | ++++ | | |
| B | | | D | +++++ | +++++ | +++++ | | |
| Na | | ++++ | ++++ | +++ | ++++ | +++++ | | +++++ |
| Mg | +++ | +++ | +++++ | ++++ | +++ | +++ | +++ | +++ |
| Al | +++ | +++ | +++++ | +++++ | +++++ | ++++ | ++++ | +++++ |
| Si | +++++ | +++++ | +++++ | +++++ | +++++ | +++++ | +++++ | +++++ |
| P | | +++ | +++ | ++ | +++ | +++ | | |
| K | | +++ | ++++ | +++ | ++ | ++++ | ++++ | D |
| Ca | | + | +++ | | ++ | ++ | | |
| Ti | | ++++ | +++++ | ++++ | +++++ | ++++ | +++++ | + |
| V | | ++ | +++ | ++++ | +++ | ++++ | + | |
| Mn | ++ | ++ | ++++ | | ++ | ++++ | | +++ |
| Fe | +++++ | +++++ | +++++ | +++++ | +++++ | +++++ | +++++ | ++++ |
| Co | | | +++ | ++ | | | | |
| Ni | | +++ | +++++ | ++++ | +++ | ++++ | ++++ | + |
| Cu | ++++ | ++++ | ++++ | ++++ | ++++ | ++++ | ++++ | +++ |
| Zn | | +++ | ++++ | ++++ | ++ | +++ | ++ | |
| Ga | D | +++ | +++ | + | +++ | ++++ | + | + |
| Ge | | | | | | | | |
| Mo | | +++ | | | | ++++ | ++ | |
| Ag | | ++++ | | | | ++++ | + | + |
| Sn | | D | | | + | D | | |
| Sb | | + | + | D | + | + | | |
| Pb | | +++ | ++ | | +++ | ++++ | | D |

(Zr— D)

| | 2019330 | 2019351 | 2019366A | 2019373 | 2019382 | 2019385 | 20193113 | 20193188 |
|---|---|---|---|---|---|---|---|---|
| Be | | | | | | | | |
| B | | | | | | | | |
| Na | +++++ | | +++++ | | +++++ | +++++ | | ++++ |
| Mg | ++++ | +++++ | ++++ | ++++ | +++++ | ++++ | +++ | +++ |
| Al | +++++ | +++++ | +++++ | ++++ | +++++ | +++++ | +++ | +++ |
| Si | +++++ | +++++ | +++++ | +++++ | +++++ | +++++ | +++++ | +++++ |
| P | | | | | | | | |
| K | ++ | ++++ | +++ | D | D | +++ | | + |
| Ca | D | +++ | ++++ | | | | | |
| Ti | ++++ | | + | | + | +++ | | |
| V | +++++ | +++ | +++ | | ++ | +++ | | |
| Mn | +++ | ++++ | ++ | +++ | ++ | | +++ | ++ |
| Fe | +++++ | +++++ | +++++ | +++++ | +++++ | +++++ | +++++ | +++++ |
| Co | ++ | +++ | ++ | | + | | | |
| Ni | +++ | +++++ | ++ | + | ++ | ++ | + | +++ |
| Cu | ++++ | +++++ | ++++ | ++++ | ++++ | ++++ | ++++ | ++++ |
| Zn | D | ++ | D | | D | | | |
| Ga | ++ | + | ++ | | + | | | |
| Ge | | | | | | | | |
| Mo | ++ | + | ++ | D | D | | | ++ |
| Ag | ++ | D | + | D | D | ++ | D | +++ |
| Sn | | | | | + | | | |
| Sb | | | | | | | | |
| Pb | D | D | D | | ++ | D | | + |

| | 20193198 | S2019344C | S2019341 | S2019342 | S2019345 | S2019346 | S2019354 |
|---|---|---|---|---|---|---|---|
| Be | | | | | | | |
| B | | | | | | | |
| Na | ++++ | +++++ | +++++ | ++++ | +++++ | +++ | +++++ |
| Mg | ++++ | ++ | ++ | ++++ | ++ | ++ | +++++ |
| Al | +++++ | ++++ | +++++ | ++++ | +++++ | +++ | ++++ |
| Si | +++++ | +++++ | +++++ | +++++ | +++++ | +++++ | +++++ |
| P | | | | | | | |
| K | ++++ | ++ | +++ | ++ | ++ | ++ | ++++ |
| Ca | | | | | | | + |
| Ti | +++ | +++ | +++++ | ++++ | ++++ | ++++ | ++++ |
| V | +++++ | ++ | ++++ | +++++ | +++++ | ++++ | +++++ |
| Mn | ++ | +++ | +++ | +++ | ++++ | ++++ | ++++ |
| Fe | +++++ | +++++ | +++++ | +++++ | +++++ | +++++ | +++++ |
| Co | + | | | | D | | +++ |
| Ni | +++ | +++ | +++ | +++ | ++++ | ++++ | +++++ |
| Cu | ++++ | ++++ | ++++ | ++++ | +++++ | ++++ | ++++ |
| Zn | | | | | | | |
| Ga | ++ | + | + | | ++++ | D | +++ |
| Ge | | | | | | | |
| Mo | | + | + | | ++ | + | |
| Ag | ++ | ++ | | | | + | |
| Sn | | | | | | | |
| Sb | | | | | | | |
| Pb | | | | | | D | |
| | | | | (Cr - D) | | | (Cr - ++) |

| | 2110SZ | 2110CYR |
|---|---|---|
| Be | | |
| B | ++ | |
| Na | | |
| Mg | +++ | +++ |
| Al | +++ | ++++ |
| Si | +++++ | +++++ |
| P | | |
| K | +++ | + |
| Ca | | |
| Ti | ++++ | +++ |
| V | +++ | |
| Mn | | |
| Fe | +++++ | +++++ |
| Co | | |
| Ni | +++ | +++ |
| Cu | ++++ | ++++ |
| Zn | | |
| Ga | + | |
| Ge | | |
| Mo | | |
| Ag | +++ | ++ |
| Sn | | |
| Sb | | |
| Pb | D | |

| | 212015 |
|---|---|
| Be | Greater than 5 ppm, less than 20 ppm |
| B | 35 ppm |
| Li | |
| Sn | 100 to 200 ppm |

ranges:

| Symbols | Percentage Range |
|---|---|
| +++++ | 1% or more (Large Amount) |
| ++++ | 0.1% - 1.0% (Moderate Amount) |
| +++ | 0.01% - 0.1% (Small Amount) |
| ++ | 0.001% - 0.01% (Trace) |
| + | Less than 0.001% (Faint Trace) |
| D | Detected |

Percentage ranges such as these have the intrinsic disadvantage of being difficult to represent in tables and on graphs. However, values of 0 - 5 (0 - +++++) have been arbitrarily assigned to the percentage ranges on Table IV. A concomitant difficulty in this arrangement is the inherent ambiguity of meaning of an assigned value. For example, if a value of 2 (++) is given to two elements, one element might actually be present in the amount of 0.01% and the other in the amount of 0.001%. In using the assigned values in the comparison of elements, a somewhat distorted picture would emerge. In the present study, however, emphasis is placed on the elements in each sample and their relative amounts rather than actual amounts of each element.

The following descriptive material points out the element variations in individual iron deposits, suggests the minerals which might contain the elements in their structure, compares the content of the various iron deposits, and indicates differences between the iron deposits and surrounding rocks. A brief comparison of Mo, Ag, Pb, Ga, Cu, and ferride content (Ti + V + Mn + Co + Ni) of the deposits to their content in country rocks is included because these elements are present in most analyses and do not form major components in the minerals present.

Twenty eight samples of banded iron formation from North Cleveland Gulch were analyzed. Samples S2019337A and B are from the eastern end of the deposit; 21185-6 and 21189-10 are from the western end; the rest of the samples are from the middle of the deposit near the large bulldozer cut and about 1000 feet west of the slice. No significant changes are observable along the strike of the deposit.

North of Cleveland Gulch, boron is found in only two samples from near the contact with the irregular intrusive body ~~rhyolite~~ (Fig. 5). Less than 0.001% was detected and it probably is from tourmaline. Sodium, aluminum, potassium, and calcium are possibly concentrated in feldspar grains, although calcite, muscovite, and biotite are also present and may account for most of these elements. Magnesium is present in some samples in moderate amounts, probably in chlorite and biotite. Phosphorous is noted in only one sample and is from apatite. Titanium is present in small amounts in 15 of the 28 samples. The Ti may have replaced ferric iron in magnetite, but it is probably for the most part in the form of rutile. Vanadium, which is present in 21 of the 28 samples, averages 0.001% - 0.01%. This element may be substituting for ferric iron in magnetite or possibly for aluminum. Chromium was not detected in the 28 samples. Manganese is found in small to moderate amounts in all 28 samples and probably substitutes for ferric iron in magnetite. Cobalt is found in only one sample and probably substitutes for iron in magnetite. Nickel is common and is found in 20 of the 28 samples, mostly in small amounts. The Ni ion could substitute for ferrous iron in magnetite, but it may be in another mineral structure. Copper is common in moderate to large amounts and is found in 25 of the 28 analyses. Although not observed in either thin or polished sections, copper is believed to be in the form of native copper or cuprite. Zinc was detected in very faint traces in 8 samples and possibly substitutes for ferrous iron. Gallium is present in 15 samples, mostly in trace amounts, and probably ~~it~~ substitutes for aluminum.

Very small amounts of Ge were detected in 4 samples and these may have been in the rutile. Molybdenum is found in 13 samples, mostly in trace amounts. It is uncertain where the Mo is located in the mineral suite; it may form its own mineral but none was noted in thin or polished sections. Silver is present in 18 samples, mostly in trace amounts, and apparently is associated with Mo or with the copper. There is a trace amount of Sn in 2 samples, and it may be in chlorite or biotite. Lead is in 18 of the samples and possibly substitutes for sodium or calcium.

At Iron Mountain and South Cleveland Gulch the elements and minerals are similar to those north of Cleveland Gulch. At Cañon Plaza a few additional elements are present, quantities are much greater than at the other deposits, and the mineralogy is different. The chief difference in the mineralogy is the presence of aluminum silicates and of specularite instead of magnetite. In 3 of 7 samples analyzed from Cañon Plaza, Be is present. Beryl may have been present, but it was not noted in the thin sections from this deposit; vesuvianite was possibly present and the Be could be in this mineral. The high percentage of B in most specimens from Cañon Plaza probably is from tourmaline. The mineral distribution of the other elements is probably similar to that at Cleveland Gulch and Iron Mountain.

In thin section the iron formation at Cañon Plaza changes somewhat from the edge of the deposit to the center, but the trace element content at the contact and in the middle is similar. The middle part of the deposit has, in general, slightly more Na, Mg, P, V, Ni, Zn, and Pb. It also has some Co which is lacking in the contact part. The middle section has less Be, B, Ti, Ga, Mo, and Ag. Tin, which was not detected in the middle, is present at the contact. In general these differences are minute and probably result from the change in the relative amounts and kinds of minerals.

Trace element contents of North and South Cleveland Gulch deposits are of special interest because it is important to establish the relationship of these two deposits. South Cleveland Gulch has much less Na and V; less K, Ni, Cu, Mo, Ag; more Mn; and no Pb. The most significant differences are the low V and absence of Pb south of Cleveland Gulch. These differences possibly reflect different environments of emplacement of the iron. Iron Mountain deposit differs from North Cleveland Gulch deposit in that it has no K and Mo, less Ni and Cu, more Ga, and Co is present.

Mason (1958, p. 44) listed average amounts of elements in crustal rocks. A comparison of the 26 elements present in the iron deposits with his list shows that all the iron deposits are low in Na, Ca, Mg, K, and Al; this condition is expectable considering the mineralogy of the deposits. Of more interest are the ferrides. The amounts of V and Mn in the iron deposits are of the same order of magnitude as the average for the crust; Ti, Cr, Co, and Ni tend to be lower than the average for the lithosphere. Gallium and Pb are around the average for the lithosphere, but Mo and Ag show enrichment in most samples. Copper is highly enriched, averaging over 0.1%. Turekian and Wedepohl (1961, Table 2) listed the distribution of the elements in the crust by rock types. The highest copper content in the crustal rocks they list is 250 ppm in deep sea clays, much below the amount of copper in these samples.

North of Cleveland Gulch the iron formation, in general, has less K, Ca, Ti, V, Ni, and Ga than the enclosing amphibolite, biotite and chlorite schists, and phyllite. The iron deposit has more Mn than the biotite and chlorite schists and phyllite, and somewhat more Pb than the three enclosing rock types. These differences are attributed to the difference in mineralogy of the iron formation and the enclosing rocks, although the Mn and possibly the Pb in the iron formation may have come with the iron. It is interesting that the copper content in the iron formation and the enclosing metamorphosed volcanic rocks is nearly the same. On the Olympic Peninsula, Washington, and other localities, basalts and tuffs have high native copper contents (C. F. Park, Jr., personal communication). Copper in the samples in the subject area is thought to have been present as native copper in the volcanics before the iron mineralization.

The Cañon Plaza iron formation has Be, B, Na, P, Ca, V, Mn, Co, Zn, Ga, Mo, Sb, and Pb, all of which the enclosing Ortega Quartzite lacks. In general, there is more Mg, Al, K, Ti, and Ni in the iron formation than in the enclosing rocks. For some reason, in most of the samples studied, silver is slightly more abundant in the quartzite than in the iron formation.

A comparison of Mo, Ag, Pb, Ga, Cu, and ferride content (with the exception of Fe) of samples from the four iron localities and country rocks from North Cleveland Gulch, South Cleveland Gulch, and Cañon Plaza is included because these particular elements do not commonly form major structural parts in the minerals present. The iron deposit at North Cleveland Gulch has about the same amount of Mo, Ag, Pb, and Ga as the schists; the schists have more copper and ferrides. The amphibolite has more Mo, Ag, Ga, Cu, and ferrides, but contains no lead. This trace element difference between the iron deposits and the country rocks of North Cleveland Gulch is explained by the fact that more minerals which can accommodate these elements in their structure are present in the country rocks.

Meta-arkose south of Cleveland Gulch, which encloses ~~most~~ much of the iron deposit there, has about the same amount of Mo, Pb, and Ag as the iron formation, but it has higher amounts of Ga, Cu, and the ferrides. The amphibolite south of Cleveland Gulch has less silver, more Mo, much more Ga, Cu, and the ferrides than the iron formation. The relationship between the iron formation at Cañon Plaza and the enclosing quartzite has been mentioned before. The deposit has more Mo, Pb, Ga, and the ferrides, and the quartzite has more Ag and Cu.

In summation, only 26 elements are recorded in one or more samples of the iron formation and enclosing rocks at Cleveland Gulch, Iron Mountain, and Cañon Plaza. Of these 26 elements only Na, Mg, Al, Si, K, Ca, Ti, V, Mn, Fe, Ni, Cu, Ga, Mo, Ag, and Pb are present in most of the samples; in addition to the high concentration of iron, Cu, Mo, Ag, and possibly Pb are concentrated in the area (are above the average for the crust); five of the other elements are significantly lower than the average for the lithosphere; the most prominent differences at North and South Cleveland Gulch are the low vanadium and absence of lead south of the gulch; elements in the iron formation from Iron Mountain are similar to those at Cleveland Gulch; a comparison of trace elements from

Cleveland Gulch and Iron Mountain with those of Cañon Plaza would not be meaningful because the mineralogy and geologic setting are unique at Cañon Plaza; there is little difference in element content and quantity between the iron formation and enclosing rocks at Cleveland Gulch, and the variations that are present could be due to the quantity and types of minerals in each rock type. It is believed that variations in quantity and content of most elements in the analyses are due to mineralogical variations in the rocks analyzed. At Cleveland Gulch and Iron Mountain most of the elements in the iron formation could have been in the host rock before iron mineralization, but at Cañon Plaza many of the elements must have been deposited by the mineralizing fluid.

Except for the conclusions that most of the trace elements at Cleveland Gulch and Iron Mountain were probably present before iron mineralization and many of those at Cañon Plaza probably were not present before mineralization, no genetic significance has been attached to the trace elements and their distribution in this report. Control on the sampling, sample preparation, and determination of amounts were not designed to allow a rigorous analysis of trace element distribution. These analyses were completed and are included here (Table IV) with the purpose of presenting the kinds and relative amounts of the trace elements in a deposit as a whole, and as compared with other deposits and country rocks. Also, the trace element analyses are presented with the hope that they will be useful when comparing other banded iron deposits with those in the present study.

## ORIGIN OF THE BANDED IRON FORMATION

### General Statement

Since the early part of the last century banded iron formations have been studied by geologists, and a plethora of theories of origin has emerged. Foster and Whitney (1851, p. 2) considered the Lake Superior iron deposits to be volcanic in origin. However, the banding of the Lake Superior iron deposits led most other geologists to consider the deposits as sedimentary in origin. Although there are apparently some replacement ores, the bulk of the Lake Superior deposits are thought to be sedimentary in origin. These deposits, along with other banded Precambrian iron deposits throughout the world, have been the subject of considerable controversy for scores of years.

After reviewing 18 different hypotheses on the origin of the Lake Superior deposits, Winchell and Winchell (1891, p. 255) observed as early as 1891 that:

> It is also evident that no thoughtful person can ever again attempt to explain all deposits of iron ore on any one theory, because iron is a metal of such wide distribution and ready chemical affinity, and of such varied forms of combination that it may be acted upon by every agent of solution or decay as well as of precipitation or mechanical deposition.

A brief review of the various major theories of the origin of banded iron deposits and a brief discussion of their limitations follow.

All but one of the major theories assume that the Precambrian banded iron deposits are sedimentary. Many of the theories consider that the time of formation of these deposits involved a unique set of conditions in the history of the earth which has never been repeated or may never be repeated. This stipulation imposes serious limits on one of the most basic geologic principles--uniformitarianism. One of the main difficulties in the formulation of theories of origin is the explanation of the manner in which banding occurred. Certainly this banding resembles a sedimentary feature. Consequently, sedimentary processes are the subject most workers have pursued in their endeavors to unravel the origin of Precambrian banded iron formations.

Any comprehensive theory of origin must answer some basic questions about banded iron deposits. Where could large quantities of iron have originated? Where did the silica originate? What was responsible for bringing these two phases together and how did the banding form? Why are most banded iron deposits restricted to Precambrian rocks? Why are most of these deposits associated with a volcanic sequence to some degree? Why are most Paleozoic and later sedimentary iron deposits oolitic, deposited near shore, and unbanded? Why is the mineralogy of most deposits so simple--mostly iron oxide, (silicate, carbonate, or sulfide), and quartz? Most explanations of origin do not answer all of these questions satisfactorily. In fact, as Winchell and Winchell indicated in 1891, it is really not possible to explain all banded iron deposits by a single theory. The presence of large amounts of iron oxide and quartz--usually banded--in widely separated deposits and diverse geologic settings does not necessitate a unique origin.

Van Hise and Leith (1911, p. 516) are generally considered as the fathers of the theory that the presence of the iron is a result of direct contribution of magmatic waters from basic igneous rocks and also of direct reaction of waters upon hot lavas. This would mean that the basic igneous rocks so commonly associated with banded iron deposits had a direct influence on the formation of the deposits. A careful reading of their report shows that they considered that direct addition of iron-rich magmatic waters from basic igneous rocks accounts for only a small part of the iron; most came from normal weathering and sedimentary processes.

Gruner (1922, p. 459) suggested that most of the iron was derived from weathering of greenstone and basalt and was transported (stabilized by organic compounds) to places of shallow clear water (oceanic or fresh). Climate was tropical or subtropical. Precipitation was caused by algae and bacteria, and much oolitic material was formed. Part of the silica was derived from weathering and colloidal precipitation, and part came by direct contribution to the sea by magmatic springs or hot submarine lava flows.

Moore and Maynard (1929, p. 276-277) concluded that "carbonated water is able to dissolve sufficient iron and silica from a basic terraine to form a large sedimentary iron deposit" and "that the iron going to make up some of the large sedimentary iron formations was transported principally as a ferric oxide hydrosol, stabilized by organic matter, and . . . that the greater portion of the silica was transported as a silica hydrosol." Banding resulted from a differential rate of precipitation helped by seasonal variation in amount of hydrosols (Moore and Maynard, 1929, p. 524). They believed that these deposits were the result of a unique set of conditions existing only in the Precambrian.

Woolnough (1941, p. 465) felt that Moore and Maynard's experimental work offered ample support for his hypothesis of epicontinental formation of banded iron deposits "from cold natural solutions in isolated closed basins on a land surface that had been reduced to the last limit of peneplanation." According to Woolnough, the closed basins would offer a quiet place for banding and influx of seasonal rainfall with a lag in precipitation of iron (caused by the absence of electrolytes) would account for alternation of bands. He believed in a single origin for all banded iron formations (Woolnough, 1941, p. 469). It would seem that the large "fresh water" lakes would actually contain very high amounts of electrolytes if rainfall was seasonal and the area was reduced to the ultimate stages of peneplanation; this concentration of electrolytes would probably promote rather than inhibit immediate precipitation of iron. Woolnough (1941, p. 482) also suggested that dilution of iron-rich water issuing into the sea would disperse rather than concentrate the iron. This does not account for such obviously marine iron deposits as the Clinton Formation. This objection to deposition in the open sea does not hold up. He also considered that the Precambrian period of peneplanation was unique in the history of the earth.

Sakamoto (1950, p. 463-464) concluded that the iron in iron formation was derived from a land surface undergoing mature weathering in a monsoon-type climate. Iron and silica were periodically carried to a wide shallow basin separated from the sea by a narrow barrier.

Iron was delivered in acid surface waters in the wet season and remained in solution until the dry season when the water became alkaline by seepage of alkaline, silica-rich ground-water. The silica remained in solution until the rainy season when the water became acid. In the formulation of this theory, Sakamoto presupposed that the Precambrian was a unique period in earth history.

Hough (1958, p. 416, 428) objected to Sakamoto's lake theory; he postulated deposition of iron and silica in fresh water monomictic lakes during a period of mature weathering in a subtropical climate. Banding came about by oxidation and precipitation of iron in the epilimnion in the summer, then reduction and dissolution on reaching the hypolimnion; in winter when the lake was not stratified, the iron would be oxidized and deposited. Silica would possibly be deposited more or less constantly throughout the year. To explain the absence of post-Precambrian banded iron deposits, Hough (1958, p. 429) also invoked a unique non-recurrent geologic situation. The chemistry of iron and silica under the conditions postulated by Hough is questionable as is the reason that no deposits such as these have been found in extant lakes.

Alexandrov (1955, p. 463) believed that seasonal changes of temperature, amount of rainfall, and consequently alternately higher and lower pH of the leaching solution (caused by change in "humic" acids due to changes in temperature and amount of water) caused the soil to yield solutions carrying silica in the warm season and chiefly iron oxide during the cold season. Alexandrov assumed that conditions favorable for this process were present only in the Precambrian, that large amounts of organic matter were present on the land surface, and that the area was near the ultimate base level. Some of his laboratory results do not agree with later chemical data or natural conditions. He stated (1955, p. 461) that both ferric oxide and silica decrease with increased pH; this does not hold true for silica (Mason, 1952, p. 160). In Alexandrov's experiments at constant temperature and a pH of 6.1 ("humic" acid), 8.8 ppm of ferric oxide and 4.4 ppm of silica were leached. Ruckmick (1963, p. 234) measured the pH of a spring issuing from the Cerro Bolivar, Venezuela iron deposit. The water had a pH of 6.1 and contained 0.05 ppm of ferric oxide and 10.5 ppm of silica. From this it would seem that the results Alexandrov obtained should be applied to Precambrian deposits with care since they do not seem to stand up for present-day environments which are similar to those postulated for the Precambrian.

Each of the aforementioned theories of origin for banded iron formation has some of the following weaknesses: a unique set of conditions, not repeated in later geologic history, is invoked; the theories assume a great abundance of land vegetation, which has not been proven to have been present in the Precambrian; the theories fail to provide any good mechanism for the separation of the quartz from the iron and for the formation of the bands; some of the observations and calculations related to the dissolution and transportation of silica and iron are not supported by later controlled experiments; the results of experiments are not corroborated in the field.

The most definitive study of the origin of banded iron formations in the Lake Superior region is by James (1954). He proposed (1954, p. 242) that the iron deposits around Lake Superior were deposited in a restricted deep basin and that precipitation of the oxides, carbonates, and sulfides of iron depended on the Eh and pH conditions in the basin. James (1954, p. 242) also pointed out that a major river with 5 ppm ferric oxide could easily have supplied all the iron in the Lake Superior region in a relatively short time. The only drawback to this work is James' failure to provide in his model for the banding of the iron formation and the dismissal of any relationship of the iron deposits to the volcanic sequence in which many of the deposits are located. Still, James' hypothesis is the most plausible theory of origin for the Lake Superior sedimentary iron formation.

Dunn (1935, 1941) proposed a non-sedimentary origin for the banded iron ores in Singhbhum, India. Despite the objection of Spencer and Percival (1952) to Dunn's theory, it is difficult to disregard a quarter of a century of field observations by Dunn. Dunn stated (1935, p. 653) that the banded iron deposits are mainly in a sequence of meta-tuffs and flows and that these were silicified during thermal activity soon after deposition. This silicification proceeded along bedding planes causing a banded rock. The ferro-magnesian minerals originally present in the tuffs were then oxidized to iron oxides. It is not clear what processes were involved in the change from a ferromagnesian silicate to iron oxide, and herein lies the main objection to Dunn's thesis.

As yet apparently no one has advocated a hydrothermal replacement origin for the iron in banded iron formation. For deposits of large areal extent, such as Lake Superior, this hypothesis would no doubt be untenable. However, numerous iron deposits of contact-replacement origin display excellent banding in certain environments. The bands are usually magnetite-quartz or magnetite-carbonate. Callahan and Newhouse (1929, p. 407) presented a photograph of banded magnetite rock from the replacement zone at Cornwall, Pennsylvania. Lamey (1945, 1961) found some well-banded magnetite-carbonate rocks at Iron Mountain, California and Silver Lake, California. He suggested (1961, p. 675) that at Silver Lake the magnetite has replaced the more magnesium-rich layers in the carbonate sequence. At Iron Mountain, Utah, some of the limestone surrounding the iron-bearing Three-Peaks Intrusion has laminations of carbonate-rich layers and quartz-rich layers. Where these laminations were invaded by the iron-rich fluids, only the carbonate layers are replaced (Plate 20B).

From these examples, it is obvious that a mechanism exists for formation of banded iron deposits by replacement processes. A replacement origin should be considered for banded iron deposits of limited areal extent and limited tonnage associated with plutons and located in rocks with some sort of previous banding. This mode of origin is here proposed for the Cañon Plaza, Iron Mountain, and Cleveland Gulch banded iron formation.

## Cleveland Gulch, Iron Mountain, Cañon Plaza

Some of the reasons for advocating a replacement

PLATE 20

A. Iron formation. Ortega Quartzite. Cañon Plaza. Quartz (Q), andalusite (A), specularite (S), Kyanite (K). 21 10 X-nicols.

B. Banded magnetite-quartz from replacement deposit at Iron Mountain, Utah.


A

origin for the banded iron deposits of Cleveland Gulch, Iron Mountain, and Cañon Plaza have been briefly discussed in the descriptions of the iron deposits. One significant thing about these deposits is their location in what is apparently a continental volcanic and hypabyssal sequence with few sedimentary interbeds. The continental nature of the sequence is indicated by the numerous beds of metamorphosed silicic tuffs, welded tuffs, and flows, all of which are thin but widespread. The presence of a high percentage of igneous rocks and the continental character of the sequence contrast markedly with the requirement of the more plausible banded iron formation theories that the deposits are in either marine or fresh water sedimentary rocks.

The relatively small size and limited areal extent of the iron deposits at Cleveland Gulch, Iron Mountain, and Cañon Plaza also suggest they differ from the large banded iron formation deposits discussed earlier. North Cleveland Gulch deposit is the largest of the four deposits. Table V shows that the average iron content of the iron formation is about 32%. Iron formation with this average percentage is distributed through an outcrop width of a few feet to around 400 feet, with lenses of iron formation totaling up to a maximum thickness of 15 feet. Probably 6 to 7 feet of banded iron formation

| SAMPLE NUMBER | IRON CONTENT (%)* |
|---|---|
| **North Cleveland Gulch Deposit** | |
| 2019326 | 45.0 |
| 2019359 | 34.5 |
| 2019362 | 31.0 |
| 2019368 | 29.0 |
| 2019380 | 31.0 |
| 2019388 | 32.0 |
| 2019396 | 44.5 |
| 20193103 | 35.0 |
| 20193116 | 26.5 |
| 20193119 | 26.5 |
| 20193128 | 28.0 |
| 20193133 | 31.5 |
| 20193142 | 32.0 |
| 20193157 | 29.0 |
| 20193164 | 31.0 |
| 20193167 | 33.5 |
| 20193182 | 32.0 |
| 20193185 | 35.0 |
| 20193194 | 32.0 |
| 20193195 | 30.5 |
| S2019337 | 34.5 |
| S2019344A | 37.0 |
| S2019344B | 40.0 |
| S2019348 | 34.5 |
| 21185 | 32.0 |
| 211810 | 31.0 |
| **South Cleveland Gulch Deposit** | |
| S2019353 | 39.0 |
| S2019361 | 40.5 |
| **Iron Mountain Deposit** | |
| 2220318 | 28.5 |
| 2220322 | 24.0 |
| 2220324 | 29.0 |
| 2220330 | 29.5 |
| 2220331 | 37.5 |
| 2220338 | 22.0 |
| **Cañon Plaza Deposit** | |
| 2110M | 25.5 |
| 2110M2 | 22.5 |
| 2110M5 | 17.0 |

*Determined by specific gravity

Table V.. Iron content of banded iron formation, Rio Arriba County, New Mexico

| SAMPLE NUMBER | IRON CONTENT (%) |
|---|---|
| North Cleveland Gulch Deposit | |
| 2019359 | 36.12 |
| South Cleveland Gulch Deposit | |
| S2019361 | 40.37 |
| Iron Mountain Deposit | |
| 2220324 | 34.04 |
| Cañon Plaza Deposit | |
| 2110M5 | 17.12 |

Table VI. Iron content of banded iron formation, Rio Arriba County, New Mexico. (Determined by Metallurgical Laboratories, Inc., San Francisco, California).

distributed through an outcrop width of 100 to 150 feet would be average. If this much iron formation is present to a depth of 200 feet, there would be about one and a half million tons of iron formation. If the South Cleveland Gulch iron-rich area, which contains an average of 38% iron (Table V), continues to a depth of 200 feet, and an estimated 1/4 of this volume is iron formation (estimation based on the amount of iron formation float), this would yield a little less than a million tons of iron formation.

The main Iron Mountain iron deposit has an average of 28% iron (Table V) and averages 10 feet thick through an outcrop distance of about 200 feet. This would give about forty thousand tons of "ore" if the deposit were present to a depth of 200 feet. In addition, this deposit is known to extend for at least several hundred feet in both directions from the outcrop. From float information, a rough estimate of a minimum of 1,000 feet is given for the strike distance of this deposit. Consequently, the total tonnage might be on the order of two hundred thousand tons. Cañon Plaza deposit averages 22% iron and has a maximum of eighty thousand tons of iron formation if projected to a depth of 200 feet. All of these tonnage figures represent the minimum amount of iron formation. There could be considerably more iron formation in covered areas, but there is probably not less than the indicated amounts if the deposits extend to a depth of 200 feet. The only exception might be South Cleveland Gulch where the cross sectional area of the deposit was estimated on the basis of float and not on outcrops. Larsen and Cross (1956, p. 23) did not report any iron formation in the Irving Greenstone in the Colorado San Juan Mountains. Neither Just (1937) nor Montgomery (1953) reported any iron formation in the Picuris area. Obviously, the deposits at Cleveland Gulch - Iron Mountain - Cañon Plaza are very small in areal extent and tonnage when compared with the hundreds of millions of tons of ore common to the banded iron formations alluded to by authors of theories on iron formation origin outlined above.

Another prominent characteristic of the subject deposits is their closeness to a major intrusion. Figure 7 is a geologic map of the area from the vicinity of Hopewell to the Picuris Uplift. This map has been modified from Just's map (1937, Plate I) with the addition of the major fold axes, the general outline of the Precambrian granite outcrops, and the locations of the major mineral deposits or mining areas. This map demonstrates that most of the ore deposits and all the banded iron deposits except the Cañon Plaza deposit are clustered around the granitic intrusion. Even though the Cañon Plaza deposit is not near granite outcrops, it bears evidence of metasomatism. This association of plutons with banded iron deposits, though considered significant, is not unique to these deposits. Plutons intrude many of the large banded iron formation areas throughout the world.

The primary field evidence on which an epigenetic origin for the banded iron formation is based is the cross-cutting nature of the banded iron deposits. The discordance of the main deposit on Iron Mountain is obvious on the plane table map (Fig. 4). Banded iron formation here cuts across the strike of a metamorphosed tuff unit. On Iron Mountain many smaller bodies are also transverse to the strike of the enclosing rocks (Fig. 4; Plate 17B). South Cleveland Gulch deposit forms a skull-shaped area which transects the strike of a meta-arkose, chlorite schist, and metamorphosed feldspathic sandstone. Only in one locality is the North Cleveland Gulch deposit exposed for enough distance to determine whether it cuts across the enclosing rocks; at this locality it is probably very slightly cross-cutting. Cañon Plaza deposit obviously cuts across the strike of the enclosing Ortega Quartzite (Plate 19B). For three of the four deposits, discordance of the iron formation to the enclosing rocks is well established and is incontrovertible; the fourth, at North Cleveland Gulch, is also possibly cross-cutting. Cross-cutting banded iron formation is not considered characteristic of large sedimentary iron formation deposits.

None of the iron-rich metamorphic minerals characteristic of the Lake Superior region was found in the iron deposits of this area. Yoder (1957, p. 233) concluded that sedimentary iron deposits of greenalite react with quartz to yield minnesotaite when subjected to increasing temperature during metamorphism. With further increase in temperature, grunerite appears, and at the highest temperatures, the assemblage consists of fayalite and quartz. If the sediments were rich in chamosite, stilpnomelane is the first mineral to appear with increasing temperature, and chloritoid is the next to appear. Biotite and garnet form at the expense of chloritoid with further increase of temperature. No minnesotaite, grunerite, fayalite, stilpnomelane, or chloritoid was noted in any thin section, and biotite and garnet, if present, were in very small or trace amounts. This suggests that greenalite or chamosite-rich sediments were not the primary iron deposits.

The presence of the iron deposits in a metamorphosed continental volcanic sequence, the small size of the deposits, their limited areal extent, their closeness to a major granitic pluton, and the cross-cutting nature of the deposits indicate that they are different from the major banded iron deposits and probably did not form in the same manner.

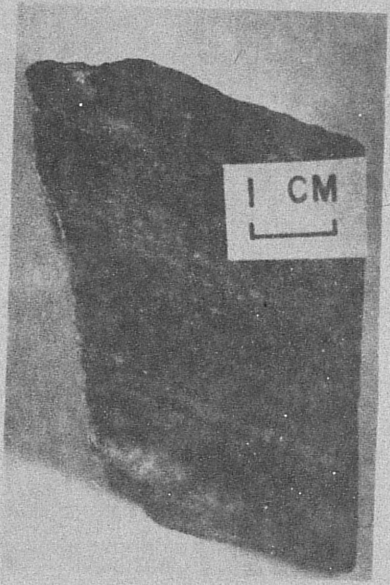
Evidence of the formation of banded magnetite-quartz and magnetite-carbonate deposits associated with massive replacement deposits has been outlined previously. The manner in which replacement in the subject area took place and the physical and chemical conditions during deposition are somewhat obscure. The prime requisite for the formation of banding is lamination of some type in the pre-existing rocks. Since the Moppin Formation and Ortega Quartzite were folded and metamorphosed before the intrusion of the Tusas ~~Granite,~~ ~~rocks~~ intrusives, rocks with laminae of varying composition were available at the time of the intrusion. An examination of relations at each of the four deposits gives an indication of the original composition of the bands. At Iron Mountain, where it is possible to trace the smaller deposits of banded iron formation into the chlorite schist, the iron bands grade into chlorite bands and the quartz bands continue into quartz bands. At South Cleveland Gulch the banding in the meta-arkose is formed by muscovite-rich layers and quartz (with some feldspar)-rich layers. At Cañon Plaza there is evidence that banding is due to shearing and segregation coupled with replacement or rearrangement of the iron oxide.

North of Cleveland Gulch numerous banded rocks were available for selective replacement, and possibly different types were involved in the formation of the iron deposits. About 1,000 feet west of the widest part of the iron deposit (Fig. 5), the iron formation has poor to good banding (Plate IB). Quartz bands in this deposit are unique in that they are coarse-grained and fairly thick. Associated with the banded iron formation is a banded quartz-mica, chlorite rock (Plate 21A). Outcrops were not numerous enough to permit the determination of the geometric relationship of these two rock types. In hand specimen the banded quartz-mica, chlorite rock was identical in appearance to the iron formation rock except for the exchange of mica for magnetite; the same distinctive coarse-grained quartz bands were present in both. In thin section the similarity is even more striking. In the quartz-mica, chlorite rock, coarse-grained quartz bands alternate with bands composed of fine-grained quartz, biotite, muscovite, chlorite, and microcline (Plate 21B). The iron formation in thin section shows coarse-grained quartz bands alternating with bands composed of magnetite, biotite, chlorite, and fine-grained

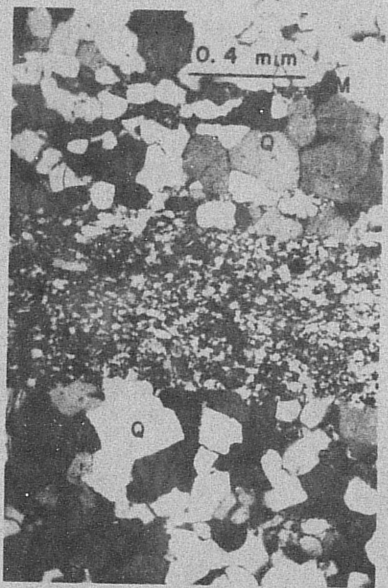
PLATE 21

A. Banded quartz-mica, chlorite rock. Moppin Formation. North Cleveland Gulch. This rock may have been the host rock for much of the iron mineralization in this area.

B. Banded quartz-mica, chlorite rock. Moppin Formation. North Cleveland Gulch. Quartz (Q), magnetite (M), with fine-grained quartz, biotite, and chlorite. S20 193 44C X-nicols.




A




B

quartz (Plate 16B). Quartz bands in the iron formation have some apatite. The similarity of these two rocks is so striking in hand specimen and thin section that, for this particular area of the North Cleveland Gulch deposit, the banded quartz-mica, chlorite rock is considered to have been the host for the replacement of the fine-grained mica, chlorite-rich bands by iron oxide. Because it is impossible to know the exact mineral composition of the particular mica, chlorite bands which were replaced by the magnetite, it is possible to say only that the magnetite probably replaced fine-grained bands made up of chlorite, biotite, muscovite, and feldspar and that most minerals were either converted to chlorite or biotite or broken down and transferred by the mineralizing fluids.

Some indication of what was lost and gained in this process can be seen on Table IV. Samples S2019344A and B are the banded iron formation, and S2019344C is the banded quartz-mica, chlorite rock from this locality. These analyses show that Na, K, Ti, Mn, Ni, Ga, Mo, and Ag are higher in the quartz-mica, chlorite rock and Mg and Pb are higher in the banded iron formation. It is believed that most of the elements which are present in smaller quantities in the iron formation than in the quartz-mica, chlorite rock were removed during the replacement process. Sodium, potassium, titanium, nickel, and gallium could easily have been removed with the break down of muscovite, biotite, and feldspar. Lead and magnesium may have been introduced with the magnetite. It is uncertain why the host rock should have more Mn, Mo, and Ag than the iron-rich rock. The Mo and Ag may have been associated with the copper in the quartz-mica, chlorite rock, but it is uncertain why Mo and Ag should be removed and not the copper during invasion by iron-rich fluids. Why Mn content should be reduced with the invasion of the iron is also problematical. These problems might be solved if more samples were analyzed for their trace element content.

Holser and Schneer (1961, p. 382) pointed out that experimental work shows "that geologically significant concentrations of iron can be mobilized at temperatures and pressures similar to those at which hydrothermal deposits were formed and in solutions two orders of magnitude more dilute in HCl than these natural fluids." They pointed out (1961, p. 382) that the concept of an acid solution is difficult to reconcile with the arguments (Fyfe, Turner, and Verhoogen, 1958, p. 144) that the last fluids of many magmas are alkaline. At any rate, their experiments showed that a weakly acid solution would be an effective transportation agent and deposition could be accomplished by neutralizing the solution. Krauskopf (1964) has suggested that, for the ferrides, volatilities might play an important role in separating and transporting these chalcophile elements from others in a cooling magma. Whatever the means of transportation and deposition, the total sulfur content of the fluid was probably low as indicated by the absence of pyrite, and the partial pressure of oxygen was low enough for the formation of magnetite while excluding hematite, but high enough to prevent the ferrous silicates from forming. These limitations are obvious from the mineralogy of the deposits.

Winkler (1965, p. 62) gave the temperature of granitic intrusions as 700° to 800° C. He further stated that the temperature of the country rocks at 0.2 of the thickness of the intrusion away from the contact would be around 500° C. Therefore, it is possible that the country rocks at Cleveland Gulch and Iron Mountain, which are within the 0.2 limit, were at a fairly high temperature when the mineralizing fluids penetrated them. If the fluid was somewhere between 500° and 600° C, little alteration of the enclosing rocks would result because they would have been in approximate thermal equilibrium with the fluid. This might account for the lack of alteration of the enclosing rocks; however, alteration around a replacement body involves so many variables that other reasons for lack of alteration effects might also be sound.

Besides the granitic pluton, there is possibly one other source for a small part of the iron. Barker (1958, p. 89) noted that an amphibolite layer in the Ortega Quartzite (his "Kiawa Mountain Formation") showed progressive metasomatic alteration toward the Petaca pegmatite deposits. His sequence of alteration was (1958, p. 90)

Hornblende-plagioclase amphibolite (unaltered)

Hornblende-plagioclase amphibolite with knots of chlorite

Biotite-epidote-quartz-plagioclase schist

Muscovite-biotite-garnet skarn rock

Chemical analyses of these rocks showed a gain in potash, alumina, water, and phosphorous pentoxide but a loss of silica, lime, magnesia, ferric oxide, and soda. Albeit, these figures might not represent net losses and gains, but from the modal analyses (Barker, 1958, p. 91) it would seem that these losses and gains are probable. Barker noted a loss of 1.0% ferric oxide during the change from an unaltered amphibolite to a muscovite-biotite-garnet rock. No rocks in the Cleveland Gulch area with exactly this mineral assemblage were noted, but metasomatism of some of the rocks in the Moppin Formation is suspected. Close to the granite contact are numerous obvious metasomatic effects as have been outlined in the description of the lithology of the Moppin rocks. However, away from the contact the only rock types which show alteration strongly suggesting metasomatism are the metarhyolites. Along strike these rocks change from a dark red slightly schistose rock with red quartz phenocrysts to a white schistose rock with white quartz eyes. This change is often abrupt and persists for only a short distance. Also, they have no apparent relationship to the regional metamorphic sequence. Regional metamorphic processes would tend toward equilibrium in the rocks, and there would be approximate uniformity in textural and mineralogical make-up for a particular rock type in a particular metamorphic zone. Metasomatism on the contrary may be selective and non-uniform in its results. Since these abrupt changes in the texture and mineralogy of the metarhyolites are more or less random, it is possible that they were caused by metasomatizing fluids from the granite. Just (1937, p. 43) suggested that the Petaca muscovite schist, which covers a large part of La Jarita Mesa, was formed by metasomatism of the Ortega Quartzite by fluids from the Tusas ~~Granite.~~ Intrusives. This would require an alkali-rich fluid, which, if in contact with the metarhyolite, would probably be in equilibrium with the minerals in the metarhyolite. This could result in a textural change without any mineralogical changes.

The presence of metasomatic activity as indicated by the metarhyolites suggests that reactions such as Barker proposed for metasomatic alteration of the amphibolite on La Jarita Mesa may have taken place in the vicinity of the iron deposits at Cleveland Gulch. Soil-covered areas in these localities are usually underlain by schistose rocks which could have been altered from amphibolite. If there was a net loss of one or more percent of iron during alteration of the amphibolites, it would not require alteration of a very large volume of rock to furnish significant amounts of iron. If iron were liberated in this manner, it probably would not migrate far, and many of the small isolated bodies of iron formation might conceivably have been formed in this manner. This is a possible mode of origin of the small isolated iron formation bodies because there is evidence of widespread alkali metasomatism, and Barker (1958, p. 92) found that there was a loss of iron during metasomatism of amphibolite on La Jarita Mesa. A hydrothermal replacement origin by iron-rich fluids from the Tusas Intrusives ~~Granite~~ is advocated for the large iron formation bodies, and may account as well for the small isolated bodies. Carefully selected chemical analyses of some of the putative metasomatically altered rocks at Cleveland Gulch will have to be conducted before this mechanism can be proven to have had any influence in this area.

Cañon Plaza is different from the other iron deposits; it occupies a shear zone and has an entirely dissimilar mineral suite. Its mineral assemblage suggests metasomatism as discussed under the description of the deposit. Quantities and kinds of trace elements could not have originated in the enclosing rock. Banding probably resulted from movement along the shear zone with crystal rearrangement.

In summation, various lines of evidence indicate that the origin of banded iron deposits at Cleveland Gulch, Iron Mountain, and Cañon Plaza is different from those proposed for major banded iron deposits of the world. The evidence, from both field and laboratory, indicates that the deposits were formed by hydrothermal replacement processes. Evidence includes 1) the cross-cutting relationships of the banded iron formation bodies to the enclosing rocks; 2) the presence of the banded iron formation bodies in a continental volcanic sequence; 3) the relatively small tonnage and limited areal extent of the iron formation; 4) the proximity of the banded iron formation bodies to a large granitic intrusion;

5) the close relationship of some of the iron bodies to faults and fractures; 6) the association at North Cleveland Gulch of magnetite-quartz banded iron formation with mica, chlorite-quartz banded rock, and the virtual compositional and textural equivalence of the two rock types except for the exchange of chlorite and mica for magnetite; 7) the continuation of magnetite-rich layers of the banded iron formation into chlorite-rich layers of chlorite schist, and the continuation of quartz-rich layers of the banded iron formation into quartz-rich layers of chlorite schist on Iron Mountain; 8) in the iron formation at Iron Mountain, the continuation of magnetite bands through cross-cutting muscovite books with a reduction in grain size of the magnetite in the muscovite and the presence of completely replaced muscovite books ("ghosts"--whose presence is known by the reduced size of the magnetite) suggesting that the cross-cutting muscovite came after metamorphism and that the magnetite came after the muscovite; 9) the presence of a metasomatic mineral assemblage at Cañon Plaza; 10) absence in the enclosing rocks of trace elements present in significant quantities in the iron deposit at Cañon Plaza; 11) the absence of iron-rich metamorphic minerals associated with the metamorphism of iron-rich sediments. While it is conceivable that some of the evidence con-

sidered singly might not be diagnostic of a replacement origin, it is concluded that the totality of evidence renders a replacement origin ~~tenable~~ the most probable.

Banding of the iron formation is believed to have originated by selective replacement of mica, chlorite-rich layers in banded quartz-mica, chlorite rocks and of mica, chlorite-rich layers in chlorite schist, except at Cañon Plaza where banding probably was the result of segregation due to metasomatism and movement along a fault zone. For the most part, the iron was derived from the granite; however, a small amount may have been derived by alteration of amphibolite, but no good evidence for this exists in the area of the iron deposits. The host rocks were probably near thermal equilibrium with the mineralizing fluids; this would account for the absence of alteration effects in the country rocks associated with the banded iron deposits.

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DEPARTMENT OF GEOLOGICAL SCIENCES



Feb. 23, 1970

VIA AIR MAIL



Mr. D. H. Baker, Jr., Director
New Mexico Bureau of Mines and
Mineral Resources
Socorro, New Mexico 87801

Dear Mr. Baker:

Thank you for your letter of January 23, 1970. It looks as though my manuscript was caught up in the changes which have been taking place at the New Mexico Bureau of Mines and Mineral Resources.

In 1963 while I was at Stanford University working toward a doctorate degree, my advisor, Charles F. Park, phoned Mr. Thompson and asked if the Bureau would have funds for a study of the iron deposits in the Carson National Forest. The Bureau made available $1000 for this study with the understanding that in exchange the Bureau would receive a copy of the manuscript for publication. Dr. Bertholf assisted with several visits to the field area while I was working there. During the field seasons of 1963 and 1964 there was considerable interest shown by several groups in the mineral deposits of that part of New Mexico--iron, copper, and gold. Due to this interest, I was encouraged to increase the scope of the study to include the gold and copper areas at Bromide and Placer canyons.

The dissertation was submitted to Stanford University in the summer of 1966. Shortly thereafter, I mailed a copy of the manuscript to the New Mexico Bureau of Mines and Mineral Resources. It was returned in the spring of 1967 with several editorial comments. In the meantime, I had conducted some additional spectrographic analyses to substantiate the presence of high copper contents in some of the samples (some were greater than 1% copper) and also to establish whether or not there was beryllium in the greissen deposits around the Tusas intrusives.

The results of these additional analyses were suggestive that the deposits might be of significant economic importance and I felt they should be checked out in the field again before the paper was published. I requested Dr. Bertholf to allow me to check out the copper and beryllium-rich areas again in the field and perhaps do some additional mapping before returning the manuscript. The Bureau of Mines made available $400 additional for this field work. This money was matched by funds provided by Lehigh's Institute of Research and the Department of Geological Sciences. These funds were sufficient for me to return to New Mexico to conduct several weeks additional field work in the summer of 1967.

This additional field work made possible the upgrading of some of the maps, establishing that the high copper contents are irregularly distributed and associated with post-iron ore quartz stringers, and the beryllium is very irregularly distributed in the greissen. Dr. Bertholf indicated that Mr. Thompson was including money for publication of the manuscript in the 1968-1969 budget. In the spring of 1969, I sent the revised manuscript to Dr. Bertholf with the understanding that he would pass it on to the editor for the Bureau. He wrote back shortly thereafter and stated that he had turned the manuscript over to Dr. J. R. Renault, secretary to the editorial committee for the Bureau. He also stated that he was leaving the Bureau.

Since I had not heard anything regarding the manuscript, I phoned Dr. Renault during the fall, 1969 and inquired about the status of the manuscript. He located it in his file and said that nothing had been done and that he would get it finished soon. After this I heard nothing until your letter of January 23, 1970.

I would like for the Bureau to publish the manuscript since it does have immediate value for Rio Arriba County geology. This is especially important since there is renewed activity in regard to the gold deposits in and around Placer Canyon. The Bureau has expended considerable support in the form of financial aid and time for Dr. Bertholf on this work and I have always received every indication that it was the Bureau's intention to publish the final results. If this were not the case, I certainly would not have spent the time, energy, and additional finances to return to New Mexico in 1967 and revise all the maps and the text for the Bureau to publish as a completed unit.

I fully realize that there have been very significant changes within the administration at New Mexico Bureau of Mines, and the Bureau, like here and all other schools, is certainly faced with a budget squeeze. However, I would like for you to consider this work for publication; the size of the article can be reduced by tabulation of the raw data and by removing certain parts not directly related to the Rio Arriba County area. This could reduce the size of the manuscript by 25 to 50%. The information contained in the manuscript should certainly be made available to the public under any circumstances because of the mineral exploration activity in that part of New Mexico. None-the-less, I have spent so much time, money, and energy on this work that I feel I should publish at least those parts which are new and unique prior to public promulgation by open filing. If the Bureau would be willing to publish these parts of the publication as

Circulars, which certainly would be only a fraction of the cost of a Bulletin, then I would be very happy to have the entire work placed on open file report at a subsequent date.

Sincerely,

Donald F. McLeroy

Donald F. McLeroy
Assistant Professor of Geology

Bingham 1968 Rio Arriba Co. bulletin describes same area & has geological maps; does McLeroy's differ enough?

January 23, 1970

Mr. Donald F. McLeroy
Dept. of Geological Sciences
Lehigh University
Bethlehem, Pennsylvania

Dear Dr. McLeroy:

Mr. Nicholson, the Bureau's new editor, came across your manuscript entitled "Geology and Origin of the Precambrian Banded Iron Deposits at Cleveland Gulch, Iron Mountain, and Canon Plaza, Rio Arriba County, New Mexico". I have no idea how long it has been here. I understand that Dr. Bertholf, who left the staff before I arrived, had received it from you.

In reviewing this manuscript, we found it to be of interest but quite long and detailed. Our limited printing budget will not permit us to publish it. However, if you do not object, we would like to place it on open file for use by students and the public. If you object to this, please advise me and the manuscript will be returned.

Thank you for your patience.

Sincerely,

Don H. Baker, Jr.
Director

DHBjr:jd

March 11, 1970

Prof. Donald F. McLeroy
Assistant Professor of Geology
Dept. of Geological Sciences
Lehigh University
Bethlehem, Pennsylvania 18015

Dear Prof. McLeroy:

Thank you for your letter dated February 23, 1970. Inasmuch as there is a large quantity of mineral exploration in Rio Arriba County, it may be that some of your material could be helpful to the companies and individuals that are exploring the Bromide and Hopewell districts. Both of these areas, their mineral deposits, and the possibilities for future ore have been described in some detail by Dr. Edward Bingler in our Bulletin 91, the Geology and Mineral Resources of Rio Arriba County. If you could pull out the parts of your manuscript that you believe are (a) new, (2) unique, and (3) helpful to mineral exploration, we could probably publish this material as a Circular. The NMBM&MR's main purpose is to encourage mineral exploration with publication of technical reports that give the correct scientific facts. Even facts, such as yours, that both the high copper and the beryllium content are irregularly distributed (which does not encourage exploration) can be useful against their background of association with quartz stringers and the greissen. Suggest you compare your work with Bingler's published bulletin, and send us the material you believe pertinent for a Circular. This could be a xeroxed copy of your original manuscript with the excess material cut out or crossed out.

Your cooperation is greatly appreciated.

Sincerely,

Don H. Baker, Jr.
Director

DHBjr:jd

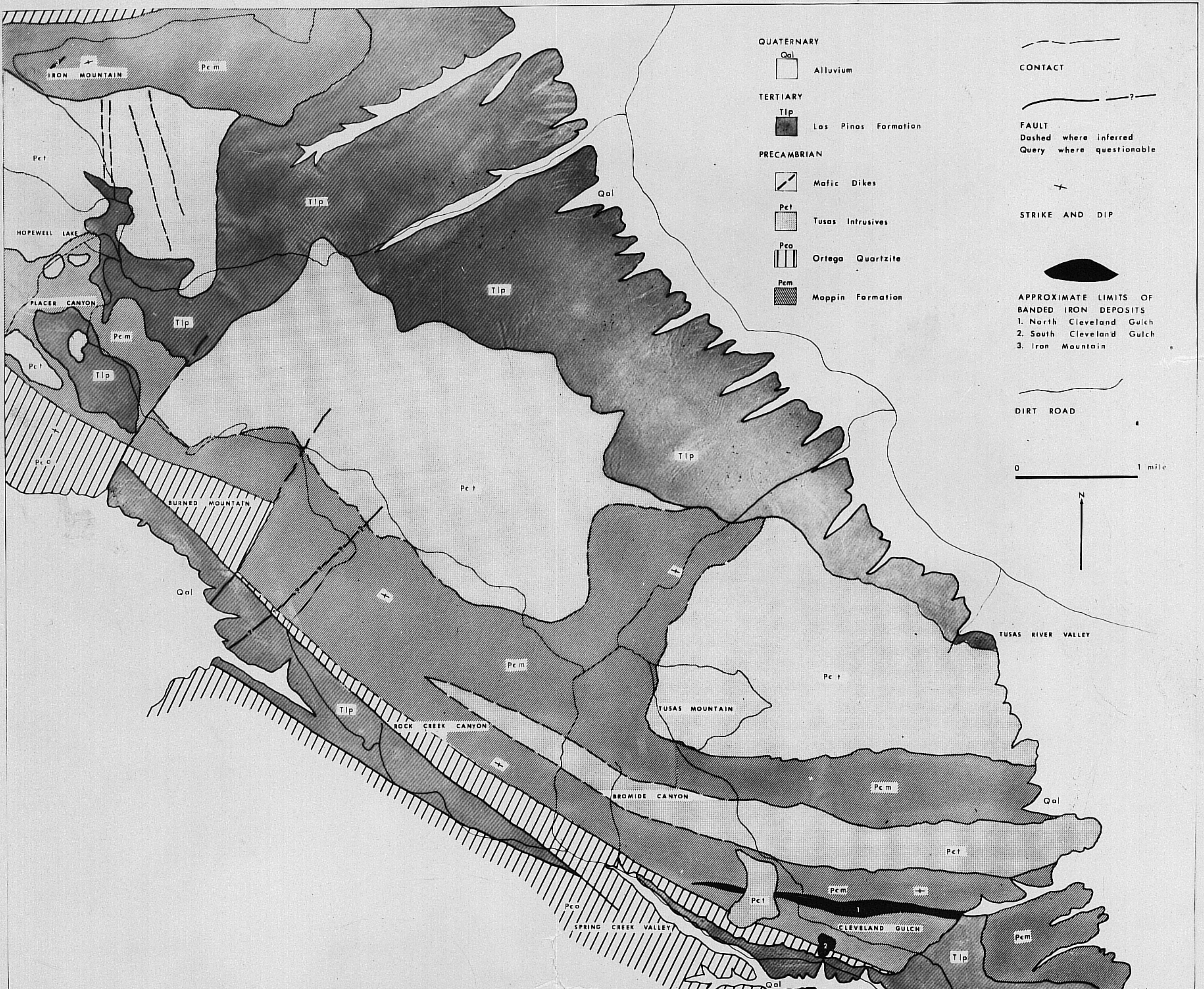
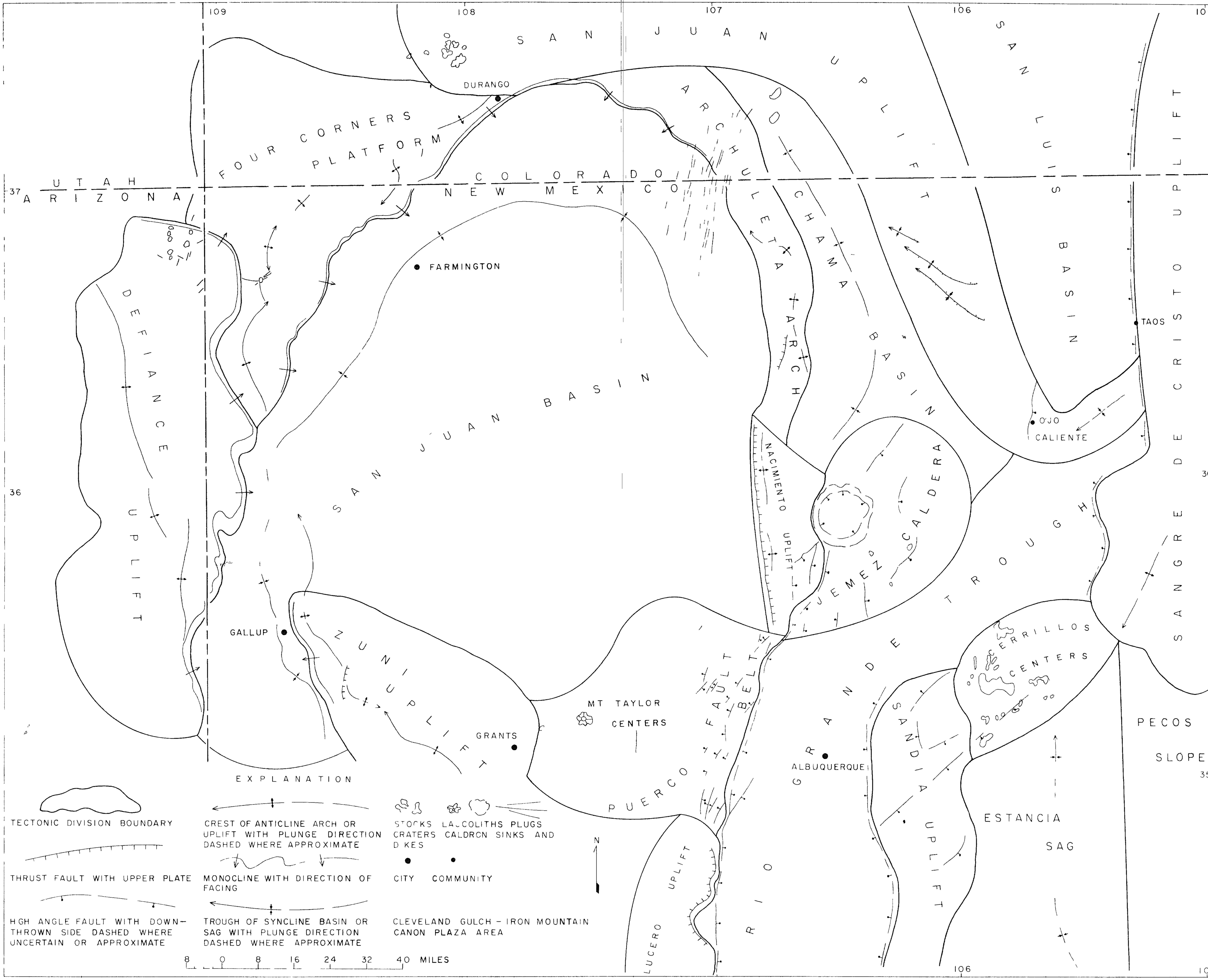
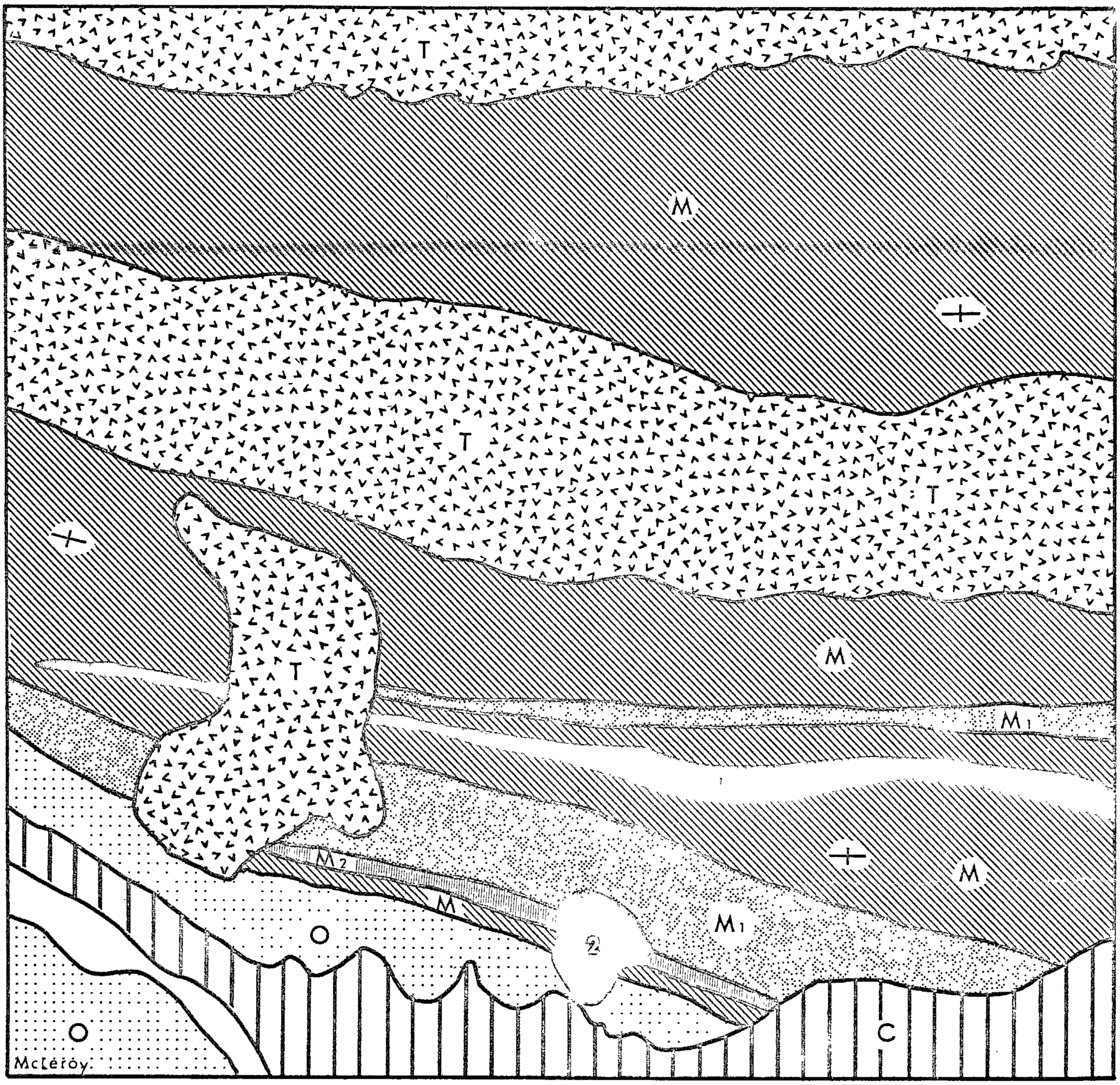



FIGURE 2. GEOLOGIC MAP-CLEVELAND GULCH TO IRON MOUNTAIN




~~FIGURE~~ PLATE 3 TECTONIC DIVISION MAP OF THE REGION AROUND CLEVELAND GULCH-IRON MOUNTAIN-CANON PLAZA (COMPILED FROM KELLEY 1955 1961 DARTON 1928 AAPG TECTONIC MAP 1944 NEW MEXICO GEOLOGICAL SOCIETY GUIDEBOOKS 1960 1961 JUST 1937)


FIGURE 4. GEOLOGIC OUTCROP MAP OF THE WESTERN HALF OF IRON MOUNTAIN, RIO ARRIBA COUNTY, NEW MEXICO




N

0 ——— 1000 FEET

CENOZOIC

ALLUVIUM

UNDIFFERENTIATED

PRECAMBRIAN

TUSAS INTRUSIVES

ORTEGA QUARTZITE

MOPPIN FORMATION

M METAMORPHOSED INTERMEDIATE TO MAFIC FLOWS AND TUFFS

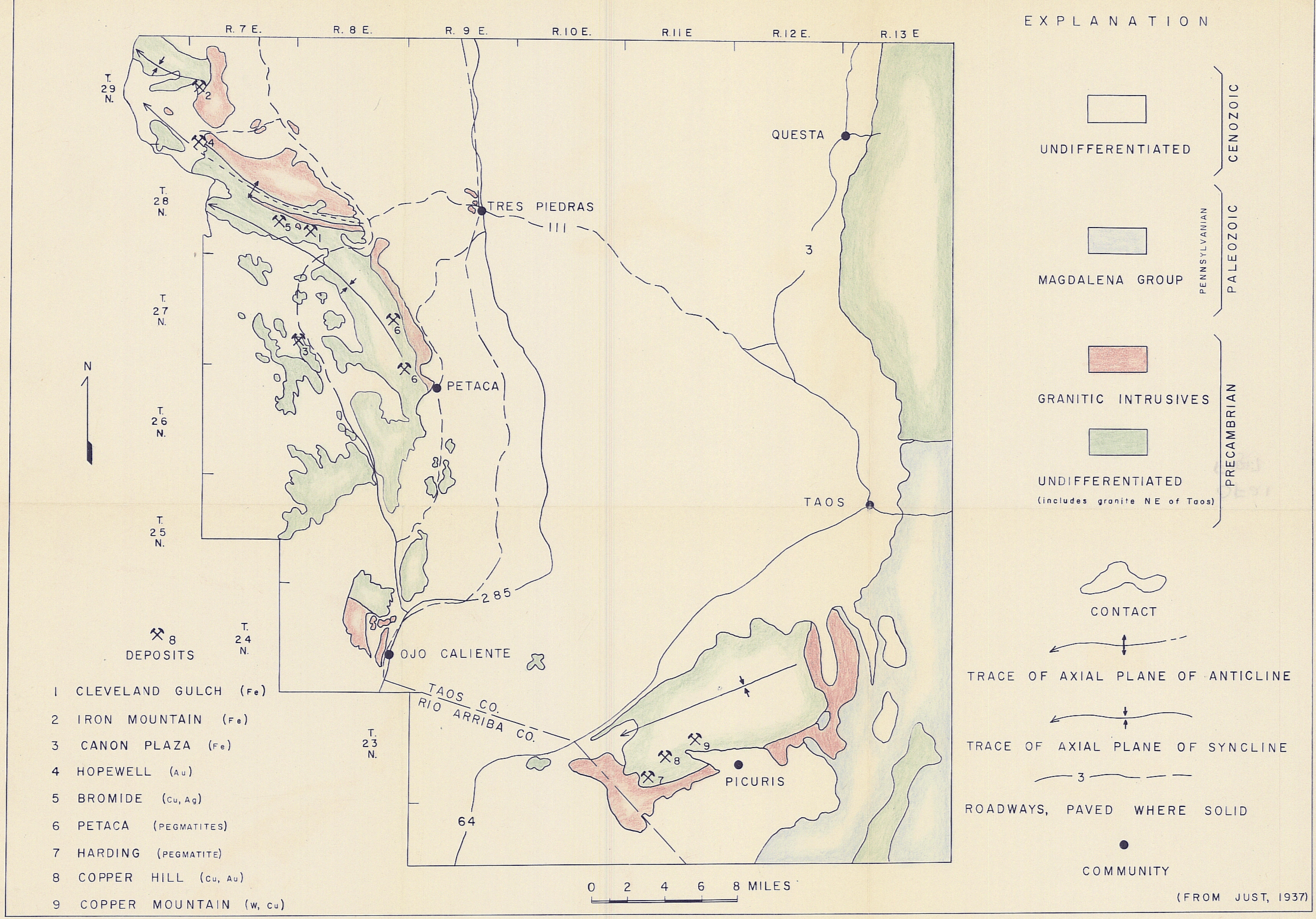
$M_1$ METAMORPHOSED RHYOLITIC TUFFS AND FLOWS

$M_2$ METAMORPHOSED ARKOSE

LIMITS OF ROCKS CONTAINING BANDED IRON DEPOSITS

1. NORTH CLEVELAND GULCH

2. SOUTH CLEVELAND GULCH




~~FIGURE~~ PLATE 7. OCCURRENCE OF MINERAL DEPOSITS AND GRANITIC INTRUSIONS IN EAST-CENTRAL RIO ARRIBA AND WEST-CENTRAL TAOS COUNTIES, NEW MEXICO