

Precambrian Geology
of the Tusas Mountain Area
Rio Arriba County, New Mexico

by

Susan C. Kent

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ABSTRACT

Detailed mapping in the Tusas Mountain area has revealed the occurrence of a Proterozoic metavolcanic-metasedimentary supracrustal terrane which has been complexly deformed and intruded by early- to syn-tectonic granodiorites and late- to post-tectonic granites. The supracrustal sequence contains a mafic volcanic-volcaniclastic rock sequence, felsic schists, and arkosic sedimentary rocks. The mafic volcanic-volcaniclastic series consists in large part of chlorite-hornblende phyllites, schists, and gneisses with minor horizons of hornblendites, metapelites, a mafic pyroclastic breccia, meta-argillites, and two types of iron formation (hematite-quartz and magnetite-quartz ironstones). The mafic sequence is unconformably overlain by two felsic schists, one interpreted as a metarhyolite and the other as a distal, reworked facies of the metarhyolite. Overlying the felsic schists are arkosic sedimentary rocks and juxtaposed by faulting are subarkoses, pelitic schists, and a thick sequence of quartzites. The relative age relationships between the faulted sections cannot be determined within this study area.

The supracrustal succession has been deformed at least once by isoclinal folding and evidence for complex folding is discussed. Metamorphism in the Tusas Mountain area increases from northwest to southeast. Metamorphic

assemblages in the northwest are typically chlorite + albite (or oligoclase); in the southeast the mafic assemblages contain hornblende + oligoclase and the pelitic schists contain garnet + kyanite + muscovite + staurolite.

Major element analyses reveal that the volcanic rocks in the Tusas Mountains are bimodal, with calc-alkaline or tholeiitic basalts and tholeiitic rhyolites. Two granodiorites which can be distinguished by field criteria are also chemically distinct.

Evidence for two tectonic settings is proposed: 1) an extensional environment to yield a supracrustal succession of bimodal volcanic rocks and arkosic sedimentary rocks, and 2) a compressional environment to produce the events following the deposition of the supracrustal rocks (eg., the deformation and metamorphism).

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INTRODUCTION

General Statement

The Precambrian exposures in the Tusas Mountains are the most northerly occurrence of Precambrian rocks in New Mexico west of the Rio Grande. Major rock units in the Tusas Precambrian terrane are felsic intrusives (including trondhjemite, granodiorite, quartz monzonite, and granite compositions), pelitic schists, quartzites, felsic schists, and a stratified sequence of mafic volcanic and volcanoclastic rocks (the Moppin series). Conclusions of previous studies as to stratigraphic sequence, nature of contacts, and amount of deformation in this area have been based only on reconnaissance-type mapping. This project, however, has involved mapping portions of the Moppin series, felsic schists, and arkosic sedimentary rocks in detail and uses supportive petrographic and major element geochemistry to: 1) tentatively define the Precambrian stratigraphy of the Tusas Mountain area, 2) identify two phases of syntectonic granodiorite intrusions and one post-tectonic granite intrusion, 3) identify two types of iron formation, 4) interpret the nature and grade of metamorphism, 5) examine structural fabrics within the area, and 6) evaluate this terrane with respect to other Proterozoic sequences in the southwestern United States.

Location and Access

The study area includes approximately twelve square miles in east-central Rio Arriba county, northern New Mexico (figure 1). Four 7 1/2-minute quadrangles (Burned Mountain, Canon Plaza, Las Tablas, and Mule Canyon) include portions of the study area (figure 2). The eastern boundary is slightly west of the Rio Tusas and extends westward to the ridge between Rock Creek and Sheep Gulch; it is bounded on the north by Tusas Mountain and on the south by Quartzite Peak and the northern flanks of Kiowa Mountain.

New Mexico Highway 64 between Tres Piedras and Tierra Amarilla skirts the eastern edge of the map area and direct access into the Tusas Mountain area is provided by state highway 111 (the Burned Mountain road) and by a very good logging road along Cunningham Gulch Ridge. Forest roads along Cow Creek and American Creek permit access by car to the eastern edge of the study area; the Cleveland Gulch road provides access to the middle part of the area by truck or four-wheel-drive vehicles. New or old logging roads are ubiquitous.

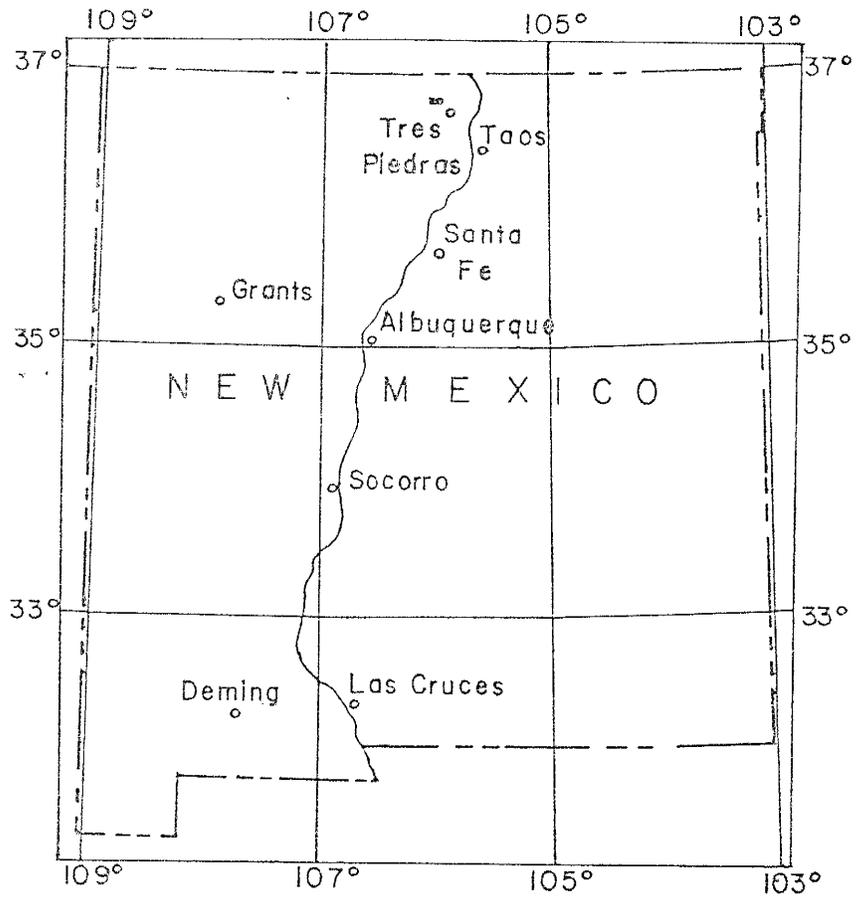
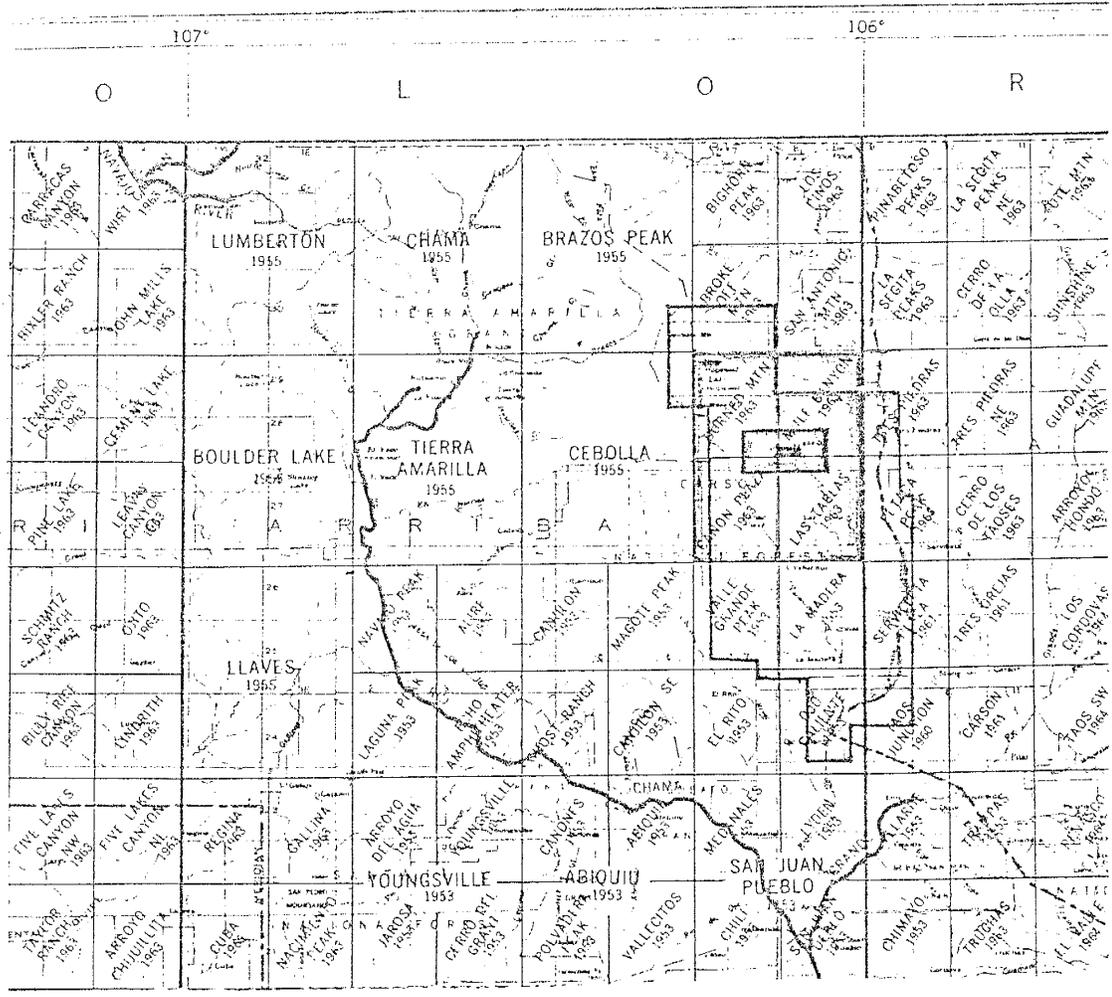


Figure 1. Index map of New Mexico.  is location of this study.



- Just, 1937
- Barker, 1958
- McLeroy, 1970
- This study

Figure 2. Location map of previous investigations which included part or all of this study area.

Previous Investigations

The study area of this report has been mapped previously four times (figure 2). Just (1937) named the major Precambrian units in an early investigation of the Tusas Mountains. He interpreted a layered sequence of metasedimentary rocks (the Hopewell series) to be the oldest rocks; interlayered mafic and felsic flows were called the Picuris basalts and the Vallecitos rhyolites, respectively. The youngest Precambrian unit in Just's stratigraphic succession was a thick sequence of quartzites which he named the Ortega quartzite. Barker (1954, 1958) remapped a portion of Just's area (the Las Tablas 15-minute quadrangle). He combined the Hopewell series and the Picuris basalts into the Moppin metavolcanic series, renamed the Vallecitos rhyolite the Burned Mountain metarhyolite, and subdivided the Ortega quartzite of Just into two distinct units. Barker restricted the "Ortega quartzite" proper to quartzites he interpreted as older than the Moppin, while quartzites interpreted as younger than the Moppin metavolcanic series were renamed the Kiawa Mountain formation. (This report uses the spelling on USGS quadrangle maps: Kiowa Mountain.) Barker divided Just's Tusas granite into two different units, the Maquinita granodiorite and the Tres Piedras granite. He also noted that a fine-grained granite underlying Tusas Mountain was an atypical porphyritic phase of the Tres Piedras granite. Carpenter (1968) studied the metamorphic mineral assemblages

in a small area west of Tusas Mountain and suggested that the Moppin had undergone two periods of deformation (rather than a single period of isoclinal folding as proposed by Barker). Carpenter presented evidence for a period of strike-slip shearing which followed isoclinal folding. McLeroy (1970) studied the Precambrian iron deposits to the southeast of Tusas Mountain and on Iron Mountain (northwest of this study). He subdivided the Moppin metavolcanic series into metamorphosed rhyolites, mafics, and arkoses, and proposed a hydrothermal origin for the iron deposits.

Parts of Just's and Barker's map areas which do not overlap this study area have been remapped in greater detail. Lindholm (1964) studied the structural petrology of the Ortega quartzite in the Ortega Mountains. Bingler (1965) mapped the LaMadera 7 1/2 minute quadrangle, Hutchinson (1968) the Burned Mountain area, and Ritchie (1969) a portion of the new Las Tablas 7 1/2 minute quadrangle. Shoemaker (1948) and Treiman (1977) mapped identical areas of the Ojo Caliente 7 1/2-minute quadrangle. These various investigators differ in detail from Barker and either partly or completely reinterpret his petrographic, stratigraphic, and structural conclusions. Whereas Hutchinson essentially retained Barker's nomenclature, Bingler, Ritchie, and Treiman referred to units by their mineralogical compositions (see Table 1 for an approximate correlation of these different units). Gresens and Stensrud (1974) and Gresens (1976) interpreted

Just 1937	Shoemaker 1948	Barker 1952, 1958	Hutchinson 1968	Ritchie 1969	Bingler 1965, 1974	Treiman 1978	Present Study
Hopewell Series	Hopewell series -knoten schist -garnitiferous -q-f-m schist -q-m schist	Moppin Metavolcanic series	Moppin Metavolcanic series -chl schist -metarhyolite	**	Hornblende- chlorite schist	**	Moppin series -chl-amp schists -iron formation -metapelites -meta-argillites -hornblendites -mafic pyroclastic breccia
Picuris Basalts	Picuris basalts -ep amp -pl amp -feldspathic -hbd schist		Basic intrusives*	**		Canada Pueblo Group -metabasites	
Vallecitos Rhyolite	metarhyolite metarhyolite porphyry	Burned Mountain metarhyolite	Burned Mountain metarhyolite	feldspathic schist	teptite (1974) feldspathic schist (1965)		felsic schists -Burned Mountain metarhyolite -m-f-q-schist
Petaca Schist	**	Petaca schist	**	q-m-f schist ky-m schist	muscovitic qtzite	pelitic schists	**
Ortega Qtzite	Ortega qtzite	Kiawa Mountain fm -upper qtzite -amp -lower qtzite -Jawbone Conglomerate -Big Rock Conglomerate Ortega qtzite	Kiawa Mountain fm -qtzite -argillized -graywacke	muscovitic qtzite qtz-pebble conglomerate hbd-chl schist fine-grained qtzite vitreous qtzite	qtzite hbd-chl schist	Ortega Group Ortega qtzite Arroyo Rancho formation	sedimentary rocks -arkoses -pelitic schists -subarkoses -vitreous qtzite

* correlation uncertain
 ** no correlative unit
 amp- amphibolite; b- biotite; chl- chlorite; ep- epidote; f- feldspar; g- garnet; hbd- hornblende; hem- hematite; ky- kyanite; m- muscovite;
 pl- plagioclase; q- quartz; qtzite- quartzite

Table 1. Comparison of Precambrian nomenclature in the Tusas Mountains. No stratigraphy is implied.

several units previously mapped as quartz-muscovite schists in northern New Mexico (within the Tusas Mountains and the Picuris Range) as metarhyolites. They also proposed a return to Just's original stratigraphy in the Tusas Mountains, that is, one quartzite (the Ortega) underlain by a series of sedimentary and mafic volcanic rocks interbedded with rhyolitic volcanic rocks.

Muehlberger (1960) reviewed the Precambrian geology of the Tusas Mountains retaining Barker's terminology, and Bingler (1974) correlated Precambrian divisions of previous workers using non-genetic metamorphic terminology in place of stratigraphic names. He also presented descriptions of the major rock units and reviewed problems of recent interpretations with those units.

Bingler (1965) presented evidence for three episodes of deformation in the Precambrian of the LaMadera (7 1/2-minute) quadrangle and Treiman (1977) attempted to simplify the structural interpretation to one period of deformation with two overlapping episodes. Evidence for as many as four thermal and/or tectonic "events" is provided by geochronologic data. Published U-Pb dates include 1715 to 1765 m.y. for the Burned Mountain metarhyolite (Barker and Friedman, 1974), 1670 to 1715 m.y. for a syntectonic granodiorite (Barker and Friedman, 1974), and 1625 m.y. for the Tusas granite (Maxon, 1976a,b). (All zircon dates have been corrected by -2% to correct for new decay constants.)

Metamorphic and/or tectonic events at 1250 m.y., 1350 m.y., 1425 m.y., and 1625 m.y. are defined by whole rock and mineral Rb-Sr isochrons and K-Ar dates from the Tres Piedras granite (Maxon, 1976), hornblende-chlorite schists (Gresens, 1975), feldspathic schists (Long, 1972; Gresens, 1975) and pegmatites (Long, 1972; Gresens, 1975). Dates for other Precambrian rocks of north-central New Mexico have been reported by Barker and others (1974) and summarized by Brookins (1974).

Based on available geochemical, isotope, trace element, and regional studies, various tectonic models for the Precambrian rocks of northern New Mexico have been proposed. Barker and others (1973, 1976) and Condie (1975) have proposed variations of a subduction zone tectonic setting. Condie and Budding (1979) have proposed a north-trending multiple rift system extending throughout the southwestern United States between 1.9 and 1.2 b.y.

Methods of Investigation

A twelve square-mile area in the Tusas Mountain area was selected for detailed study of the Precambrian rocks. The area was mapped on a scale of 1:12,000 during the summer months of 1978 and 1979. Eighty thin sections were studied and 12 samples were analyzed for major elements and a few trace metals. The chemical analyses were done by the atomic absorption method at the New Mexico Bureau of Mines and Mineral Resources by Lynn Brandvold.

TERTIARY ROCKS IN THE TUSAS MOUNTAIN AREA

Tertiary rocks which unconformably overlies the Precambrian are the only Phanerozoic rocks present in the Tusas Mountain area. They have been studied in detail by Butler (1946, 1971) and Barker (1954, 1958). Within this study area, the only Tertiary unit mapped by Barker (1958) is the rhyolite member of the Los Pinos formation. However, along either side of the mouth of Cleveland Gulch (sec. 33, T.28N., R.8E.) an approximately 30 meters thick conglomerate similar to the Ritito Conglomerate (described by Barker) unconformably overlies the Precambrian. The clasts in this unit are slabby fragments (gravel size to over one meter in longest dimension) of mostly gray, vitreous Precambrian quartzite and minor amounts of Precambrian schists and vein quartz. The unit is poorly sorted, poorly bedded, poorly consolidated and weathers to bold cliffs and rubble-covered slopes. Overlying this unit in the Cleveland Gulch area and along the northern slopes of Spring Creek is a poorly-sorted, poorly- to moderately-consolidated but well-bedded conglomerate (Barker's rhyolite member of the Los Pinos formation). It is composed predominantly of intermediate to rhyolitic volcanic clasts; basalt and Precambrian clasts are minor. The clasts are rounded to subrounded and range from gravel size to over one meter in length. Cross bedding can be observed in the finer layers and graded bedding can be

observed on two scales: on a gross scale of about 3 meters and on a finer scale of less than 0.5 meter in the sandy layers. On steep slopes this unit forms bold cliffs but on gentle slopes outcrops are rare and only gravels and cobbles cover the ground.

Vitreous Quartzite Boulder Fields

Along the ridgetops north of Spring Creek and Deer Park are large areas covered solely with Precambrian vitreous quartzite boulders. These areas were previously included by Barker (1958) as part of the Precambrian (as the upper Kiowa Mountain formation). The lack of outcrops within these boulder fields, together with the understanding of Precambrian stratigraphy gained from this study, suggest that these areas are not actual outcrops of vitreous quartzite. These boulder fields are difficult to interpret but are most probably of Tertiary age and may represent either paleotalus deposits or paleo-valley fill. The quartzite clasts range from gravel size to an excess of 6 meters; most commonly they are less than one meter in diameter.

PRECAMBRIAN GEOLOGY

General Precambrian Geology of the Tusas Mountains

Precambrian rocks are exposed in the Tusas Mountains from Ojo Caliente to north of the Brazos Box. Intrusive rocks of several different compositions are part of the Precambrian terrane. Trondjhemites crop out north of the Brazos Box; their absolute age and their relationship to the rest of the Precambrian rocks is not known at the present time. Early- to syn-tectonic granodiorites are found in the Tusas Mountain area and westward; late- to post-tectonic granites intrude the entire terrane. Quartz veins are common everywhere and pegmatites are abundant south of Kiowa Mountain. A mafic volcanic-volcaniclastic supracrustal succession crops out in the Tusas Mountain area and extends westward. Felsic volcanic rocks crop out throughout the Precambrian terrane and are known collectively as the Burned Mountain metarhyolite. A thick sedimentary rock series consists of quartzites, arkoses, and pelitic schists. Quartzites are found throughout the Tusas Mountains; arkoses are concentrated mainly in the Brazos Box and to a lesser extent in the Tusas Mountain area; and pelitic schists which are minor in and to the north of the Tusas Mountain area are abundant to the south.

The entire Precambrian terrane (except for the post-tectonic granites) has undergone at least one episode of isoclinal folding and possibly as many as three (Bingler, 1965). Deformation has removed original sedimentary features in the south (Bingler, 1965) but bedding and pebble horizons have been preserved in the arkoses in the Tusas Mountain area.

Metamorphic grade ranges from greenschist facies to lower amphibolite facies; there is a general increase in metamorphic grade from northwest to southeast. Reported mineral assemblages characteristic of greenschist metamorphism include albite-chlorite-muscovite and muscovite-albite-biotite-epidote in the south (Bingler, 1965), and chlorite-albite in the north (Barker, 1958). Andesine-hornblende (Bingler, 1965) and muscovite-oligoclase-stauroilite-kyanite-garnet (Barker, 1958) represent the amphibolite metamorphic assemblages.

Field relations show that in the Tusas Mountain area the majority of the mafic volcanism preceded the onset of felsic volcanism. Radiometric dating places felsic volcanism between 1715 m.y. and 1765 m.y. (Barker and others, 1974), syn-tectonic plutonism between 1670 and 1715 m.y. (Barker and others, 1974) and post-tectonic granite emplacement at 1625 m.y. (Maxon, 1976a,b). Pegmatite emplacement was sporadic but probably peaked around 1450 m.y. (Gresens, 1975). Other thermal events have been reported between 1350 and 1250 m.y. (Gresens, 1975; Long, 1972).

Stratigraphy of the Tusas Mountain Area

Bingler (1965) warned that unraveling a stratigraphic sequence in Precambrian rocks of the southern Tusas Mountains (La Madera quadrangle) is virtually impossible due to multiple episodes of deformation including an S-1 episode of mass movement and unrestricted flowage. Because of such extensive deformation, Bingler suggested that the present compositional banding in that area bears no resemblance to the original stratigraphy. Bingler based his conclusions on detailed structural analyses and on the absence of preserved sedimentary features. The presence of Bingler's S-1 event in the Tusas Mountain area has not been conclusively demonstrated at this time. Bedding and other original sedimentary features have been preserved in a few outcrops, and individual units can be traced along strike for several miles. Therefore, a stratigraphic sequence for the Tusas Mountain area will be proposed. If, however, subsequent work demonstrates that the Moppin series has experienced two episodes of folding, the stratigraphy presented below will require substantial revision.

In this study, metamorphic rock names are not used when there is little doubt as to the original nature of the rock unit. Thus, a metavolcanic rock will be referred to as a volcanic rock, a meta-pebble conglomerate as a pebble conglomerate. When metamorphic names are used, the minerals

will be listed in increasing order of abundance. A muscovite-biotite-quartz schist implies biotite is more abundant than muscovite and quartz is more abundant than biotite. Foliation will be used as a general term to designate a planar element in a metamorphic rock, whereas schistosity will imply the alignment of platy minerals to such an extent that weathering and breakage easily follows the planes of alignment.

The stratigraphy for the supracrustal succession in the Tusas Mountain area is presented in figure 3 and table 2. The lowermost portion of the supracrustal section crops out along Cunningham Gulch (north of this study area), however the base of the supracrustals is not exposed. Within this study area the lowermost unit is a porphyritic basaltic andesite (Barker, 1958, p. 16). Overlying this porphyry is a series of volcanoclastic sedimentary rocks, mafic volcanic rocks, and hematite-quartz ironstones. Near the top of this mafic section, on the east side of the area, is a series of felsic to intermediate phyllites with minor greenschists which are interpreted as volcanoclastic shales, mudstones, ironstone, and minor mafic igneous rocks. Mafic igneous and volcanoclastic rocks overlie the phyllitic section. This entire sequence is the equivalent of Just's (1937) Hopewell series and Picuris basalts and Barker's (1958) Moppin metavolcanic series. In this report this section is referred to as the Moppin series.

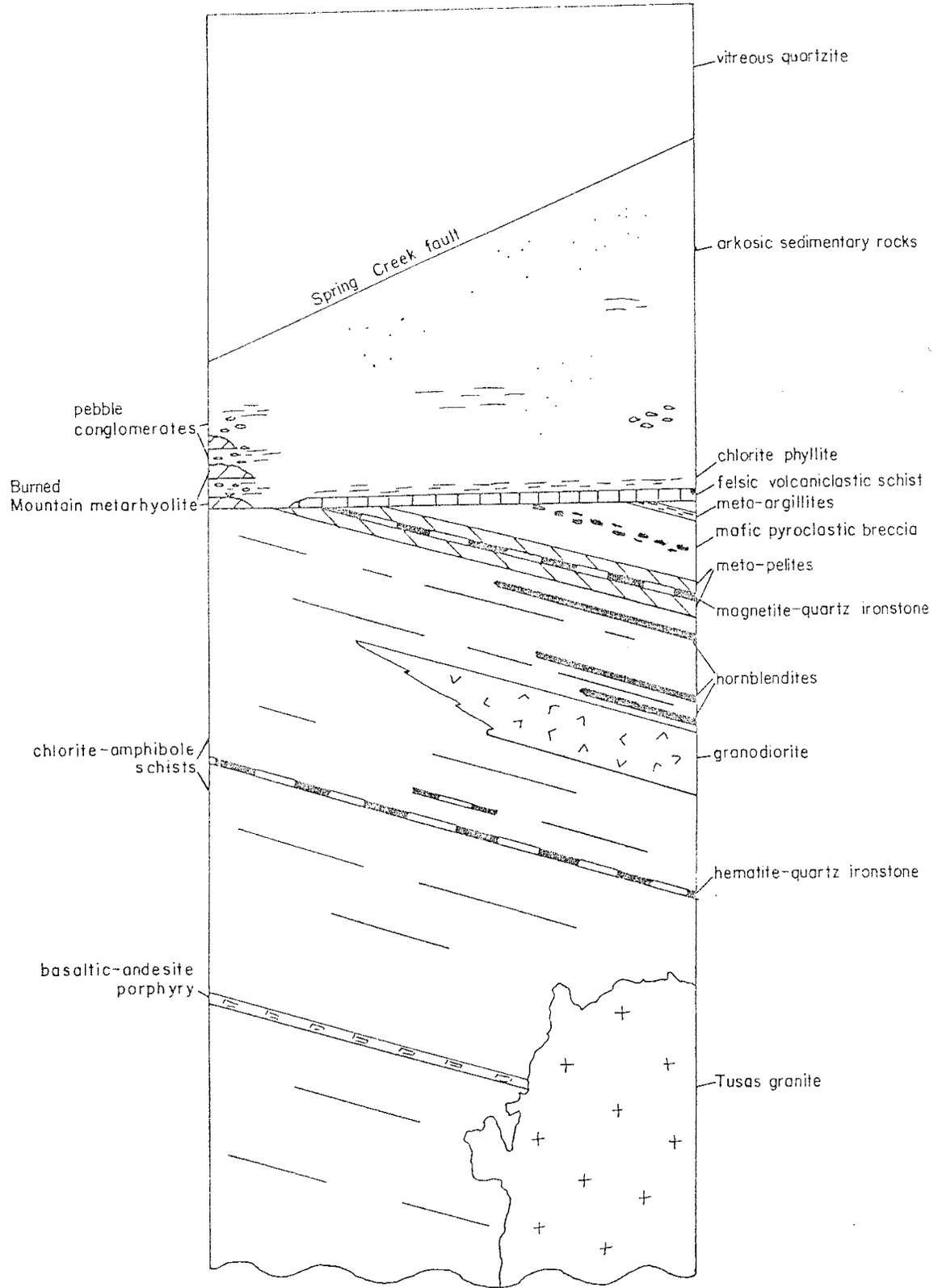


Figure 3. Diagrammatic stratigraphic column for the Tusas Mountain area.

Rock Unit	Description	Estimated thickness of exposure	Percent of Precambrian
Moplin series chlorite-amphibole schists	greenschists, amphibolites, and gneisses; an estimated 65% are volcanoclastic rocks, 35% igneous rocks	2250 meters	60
hornblendites	dark-green, massive chlorite-hornblende schists	3*	
iron formation	banded magnetite-quartz and hematite-quartz ironstones	3*	
metapelites	intermediate to felsic phyllites and schists; interbedded with minor amounts of mafic schists, hornblendites, and iron formation	245	
mafic pyroclastic breccia	contains poorly-sorted, round or irregularly shaped amphibolite clasts in a fine-grained amphibolite matrix	8	
meta-argillite	green and gray, banded biotite-quartz (+/- feldspar, muscovite) schist	30	
Felsic schists			5
Burned Mountain metarhyolite	brick red, massive rhyolite tuff with conspicuous quartz and potassium feldspar phenocrysts; relict eutaxitic textures	30-150	
muscovite-feldspar-quartz schist	bleached-white or satiny-yellow 'quartz-eye' porphyry	8	
Sedimentary rocks			35
arkoses	orangish to gray, muscovite-feldspar-quartz schists, fine- to coarse-grained, well bedded, poorly to well sorted, rarely cross bedded	60	
pelitic schists	gray and white mica-quartz (+/- garnet, kyanite, staurolite) schists interbedded with hornblende gneisses and amphibolites	500	
subarkoses	light gray muscovite-quartz (+/- feldspar) schists, well bedded, commonly cross bedded, pebbly horizons	1000	
vitreous quartzite	light purplish gray, crossbedded, pebbly horizons	> 600	

Table 2. Summary of Precambrian supracrustal rock units in the Tusas Mountain area.
*indicates outcrop thickness of individual exposures.

Overlying the Moppin series with angular unconformity are interbedded pebble conglomerates, arkoses, and felsic volcanic rocks. Just included the pebble conglomerate and arkoses within the Hopewell series and he called the felsic volcanic rocks the Vallecitos rhyolite. Barker named the pebble conglomerate the Jawbone conglomerate, included the arkoses as part of the upper quartzite member of the Kiowa Mountain formation, and renamed the felsic volcanic rocks the Burned Mountain metarhyolite. In this study the pebble conglomerates are part of the arkosic sequence and the distinctive felsic volcanic rock in the western part of the area will continue to be called the Burned Mountain metarhyolite. A mappable felsic schist in the eastern part of the study area which is tentatively identified as a distal facies of the Burned Mountain metarhyolite is referred to as the muscovite-feldspar-quartz schist. Coarse arkosic sandstones and felsic volcanoclastic rocks predominate in the east whereas the pebble conglomerate and the Burned Mountain metarhyolite are confined to the western edge of the area. A major fault along Spring Creek valley juxtaposes subarkose, pelitic schist, and vitreous quartzite on the south (Just's Ortega, Barker's upper quartzite member of the Kiowa Mountain formation) against the arkosic sequence on the north. The relative age relationship between the rocks on either side of the Spring Creek fault is not known at this time.

Supracrustal Rocks

Moppin series

Mafic volcanoclastic and volcanic rocks comprise the oldest portion of the supracrustal succession in the Tusas Mountain area (although it is possible that the vitreous quartzite may ultimately prove to be older). The minimum age of this series is constrained by a date on the overlying Burned Mountain metarhyolite of between 1715 and 1765 m.y. (Barker and Friedman, 1974). The mafic succession consists of chlorite-amphibole phyllites, schists, and gneisses, meta-pelites, iron formation (both hematite-quartz and magnetite-quartz ironstones), hornblendites, a mafic pyroclastic breccia, and meta-argillite. Large scale cyclicity typical of many Precambrian greenstone terranes (eg., repeated successions of mafic volcanic rocks overlain in turn by felsic volcanic rocks, clastic sedimentary rocks, and iron formation or cherts) has not been recognized in the Moppin series.

The chlorite-amphibole schists comprise approximately 73 percent of the Moppin series, an estimated 65 percent of which is interpreted to be volcanoclastic sedimentary rocks. Chlorite +/- biotite +/- actinolite assemblages in the northwestern part of the Moppin grade southeastward into hornblende +/- biotite assemblages.

Sedimentary rocks can be distinguished from volcanic rocks in the field by their lithological and textural inhomogeneity, highly schistose nature, and compositional banding. Volcanic rocks are typically more massive and are characteristically phenocryst-bearing. The field distinctions are supported by thin section examination: rocks interpreted as volcaniclastic sedimentary rocks in the field are typically more quartz-rich whereas igneous rocks are characterized by chlorite + plagioclase or amphibole + plagioclase mineral assemblages.

Several coarse-grained hornblendite units (averaging less than 3 or 4 meters in thickness) are ultramafic in composition (table 3, p. 64). They are mappable horizons in the eastern half of the map area but their distinctive character is lost westward along strike as the schists become dominated by greenschist metamorphic facies. They are stratigraphically concordant but no evidence was observed as to whether these rocks were extrusive or intrusive.

Ironstones within the Moppin series are of two types: the older units, interbedded with amphibolites and gneisses, are hematite-quartz ironstones; younger units, interbedded with fine-grained clastic sedimentary rocks, are magnetite-quartz ironstones. The hematite-quartz ironstones form unbanded massive units which are traceable up to 7 kilometers along strike. Outcrops characteristically have a

sugary texture and display mottled colors (due to brecciation?) of light to dark pinkish brown, black, brown, and grayish white. Rounded "clasts" of sugary quartz grains of one color are surrounded by sugary quartz of another color. Veins of hematite and chlorite are discontinuous and irregular. Magnetite-quartz ironstones form dark purplish-black, banded outcrops in which magnetite-rich bands (less than 1 mm to several centimeters thick) alternate with quartz-rich bands and lenses (1 to 10 mm in width). In addition, the magnetite-quartz banded portions are interlayered with felsic and chlorite phyllites and schists (metapelites) which commonly contain disseminated magnetite octahedra. A detailed discussion of the iron formation is given in a following section.

The metapelites are typically chlorite phyllites, biotite-feldspar-quartz schists, muscovite-feldspar-quartz schists, and biotite-quartz schists. These schists are well foliated and color bands and lenses are common. They are interbedded with the magnetite-quartz ironstones and a few mafic schists and hornblendites. Magnetite octahedra are common and infrequently form thin stringers and rectangular outlines. The chlorite phyllites (and in places, biotite-feldspar-quartz and biotite-quartz schists) contain abundant white spots flattened in the plane of foliation that may be retrograded cordierite.

A mafic pyroclastic breccia occurs near the top of the Moppin series and contains poorly-sorted, angular, rounded, or irregularly-shaped clasts (from very small to 20 cm in length) in a medium-grained amphibolite matrix. The clasts are of three rock types: fine-grained plagioclase-rich amphibolite, fine-grained plagioclase-poor amphibolite and porphyritic amphibolite. The western outcrops also contain quartzitic clasts. Bedding and channel scours are preserved in some outcrops.

A horizon of meta-argillites forms green and gray outcrops in which color bands are generally less than 1 cm thick. The bands are composed of varying amounts of biotite, muscovite, quartz, feldspar, epidote, and chlorite.

Felsic schists

Two felsic schist units overlie the Moppin series. The Burned Mountain metarhyolite is associated with pebble conglomerates in the arkosic sedimentary rock section on the western edge of the study area. The second felsic schist is east of, but apparently at the same stratigraphic level as, the Burned Mountain metarhyolite. It is mapped as a muscovite-feldspar-quartz schist and it is interpreted as a distal, slightly reworked, facies of the Burned Mountain metarhyolite. A zircon age for the Burned Mountain metarhyolite is between 1715 m.y. and 1765 m.y. (Barker and others, 1974).

The Burned Mountain metarhyolite and the muscovite-feldspar-quartz schist appear to overlie the Moppin series with an angular unconformity. The unconformable contact between the felsic schists and the Moppin series cannot be observed directly in outcrop; however, two lines of evidence support an unconformable relationship between the two: 1) several horizons in the Moppin series are apparently truncated by the felsic schists (eg., the mafic pyroclastic breccia, the iron formation and its enclosing metapelites, and the meta-argillites) suggesting that uplift and erosion preceded the deposition of the felsic schists; 2) the association of interbedded pebble conglomerates with the Burned Mountain metarhyolite further suggests an extended period of nearby uplift and erosion which both preceded and accompanied felsic volcanism.

Three horizons of Burned Mountain metarhyolite are separated by pebble conglomerates and other arkosic sedimentary rocks in sec. 22, T.28N., R.7E. These horizons of Burned Mountain may be either three stratigraphic horizons or a single unit repeated by isoclinal folding. Additional structural work now underway (T. Gibson, 1980, personal commun.) should help to identify the correct interpretation. The Burned Mountain metarhyolite forms bold outcrops with brick red or grayish-pink weathered surfaces; schistosity is usually poorly developed and samples break with a conchoidal fracture. The Burned Mountain is characterized by

approximately equal numbers of quartz and potassium feldspar phenocrysts and an aphanitic silicic groundmass. Some samples contain light-gray aphanitic stringers and lenses of reddish to grayish lenticular aggregates of fine-grained material that are interpreted to be flattened pumice fragments.

Four lines of evidence support a largely extrusive origin for the Burned Mountain metarhyolite within and to the west of this study area. 1) There are no signs of intrusion such as intrusive breccias, xenoliths of country rock, or discordant contacts. 2) The metarhyolite displays a sharply restricted stratigraphic distribution, occurring only within the arkoses. 3) Cobbles of the Burned Mountain metarhyolite have been found within the overlying pebble conglomerates (T. Gibson, 1979, personal commun.) which indicates that the metarhyolite was available for erosion soon after its emplacement. 4) Eutaxitic textures and flow banding are prominent in the outcrops on Burned Mountain (sec. 8, T.28N., R.7E.) and to a lesser extent in the outcrops of Sheep Gulch. All of the above features are consistent with an extrusive origin.

The muscovite-feldspar-quartz schist forms bold, highly schistose outcrops. The schist is usually white and forms satiny yellow, reddish-brown, or greenish-white weathered surfaces which have prominent quartz phenocrysts in relief. This schist is characterized by 5 to 10 percent

prominent quartz phenocrysts (1 to 3 mm in length), and 0 to 10 percent microcline phenocrysts (2 to 4 mm in length) and 10 to 20 percent muscovite. Also present are 2 to 4 mm pyrite cubes, 1 mm magnetite octahedra, and 0.5 to 1 mm biotite flakes. Lenticular shapes suggestive of relict pumice fragments (up to 5.0 mm in length) are present in outcrops above the Moppin Ranch (sec. 26, T.28N., R.7E.).

The muscovite-feldspar-quartz schist differs from the Burned Mountain metarhyolite in appearing to be less silicic (although no chemical analyses for the muscovite-feldspar-quartz schist are available), in having a much higher muscovite content, having fewer microcline phenocrysts, and in being more schistose. These differences could be explained in three ways: 1) the muscovite-feldspar-quartz schist could be a reworked epiclastic equivalent of the Burned Mountain metarhyolite; 2) the muscovite-feldspar-quartz schist could be a sheared equivalent of the Burned Mountain metarhyolite; or 3) the muscovite-feldspar-quartz schist and the Burned Mountain metarhyolite could be two distinct units. There is very little evidence for shearing and it seems very improbable that two units would occur in the same stratigraphic horizon and not be somehow related. Therefore, this author favors the first explanation although it is not an entirely satisfactory explanation. Mapping by T. Gibson (in progress) suggests the source for the Burned Mountain metarhyolite and

the pebble conglomerates is west of this study area which is consistent with the muscovite-feldspar-quartz schist and chlorite phyllites being a distal, more easterly, facies of those units. It is puzzling, however, that lithologic and textural gradations between the two assemblages are missing.

Sedimentary rocks

The sedimentary rock section consists of four map units: arkoses, pelitic schists, subarkoses, and vitreous quartzite. The relationships between the sedimentary section and the Moppin series, as well as among the various sedimentary members, are not always clear and various interpretations have been made in past studies.

Arkosic sandstones and shales, subarkoses, and minor quartzites and mafic schists between sec. 22, T.28N., R.7E. and sec. 32, T.28N., R.8E. and in secs. 33, 34, T.28N., R.8E. are associated with the felsic schists and appear to unconformably overlie the Moppin series. They are predominantly fine- to medium-grained, finely to moderately laminated, poorly-sorted arkoses. Rounded quartz, feldspars, and opaques are generally less than 1 mm in size, and are contained in an orangish-yellow, mica-rich matrix; biotite is usually minor or absent. Less common rock types include coarsely recrystallized gray quartzite with minor interstitial muscovite; biotite-quartz (+/- feldspar) (sometimes phyllitic) schists in alternating dark and light

bands, greenish-gray phyllitic schists which may contain visible rounded quartz grains, and gray spotted phyllites with rhombohedral cavities (weathered-out siderite?).

Pebble conglomerates on the west side of the field area contain highly flattened clasts (up to 16 cm in length) primarily of black hematitic chert, vein quartz, and felsic schists with rare clasts of red chert. Clasts of mafic schists are very rare. The matrix is typically purplish gray and composed of fine-grained quartz and muscovite. Pebble conglomerates which contain only angular clasts of quartz (1 to 3 cm in length) crop out on the east side of 8820 hill (sec. 34, T.28N., R.8E.).

Subarkoses, composed primarily of quartz with varying amounts of feldspar and muscovite, form bold light-gray outcrops. Typical outcrops are well bedded and contain preserved pebble horizons. Subarkosic rocks are isoclinally infolded with the pelitic schists along the Spring Creek roadcuts (sec. 33, T.28N., R.8E.).

The pelitic schists map unit contains four major rock types: hornblende gneisses, amphibolites, mica-quartz (+/- garnet) schists, and kyanite-garnet-staurolite-mica schists. The hornblende gneisses are thinly interbedded with the mica-quartz schists and are banded, well-foliated mica-quartz-hornblende gneisses. The amphibolites are more massive and thicker than the hornblende gneisses. They are green and white, medium- to coarse-grained, and contain

hornblende blades up to 15 mm in length which are randomly oriented within the planes of foliation. Mica-quartz (+/- garnet) schists are light gray, highly schistose and friable, and well banded due to varying proportions of biotite and muscovite. Red euhedral garnets up to 20 mm in length are abundant in some layers. Plagioclase grains, 2 to 7 mm long, also occur in some of the bands. Kyanite-garnet-staurolite-mica schists form distinctive light reddish-gray outcrops which appear conglomeratic due to a bumpy surface created by staurolite porphyroblasts (1 to 4 cm in diameter).

The pelitic schists and subarkoses exposed in the Spring Creek roadcuts (sec. 33, T.28N., R.8E. are problematic. First, they are dissimilar to the other units in the map area although similar subarkoses crop out to the west in Rock Creek (sec. 21, T.28N., R.7E.) and pelitic schists are reported to the south in the Las Tablas quadrangle by Ritchie (1969). Secondly they are isoclinally folded on a scale that is not present in the rocks to the north and northwest. Folds along the Spring Creek road have wave lengths of several meters to several 100's of meters, display axes that plunge steeply to the SSE, and have apparently repeated the section several times. Small quartz and pegmatite veins occupy the axial portion of each fold. Thirdly, they contain a suite of metamorphic alumino-silicate minerals not present elsewhere. For these reasons, it is suggested that these rocks may not be correlable with the

rocks in the rest of the study area, and may be, instead, faulted into their present position.

Vitreous quartzite reported by Bingler (1965) to be 99 percent quartz has been variously interpreted in the past as the youngest and oldest of the stratified rocks. The quartzite is in fault contact with the rest of the supracrustal succession and its stratigraphic relationship cannot be determined within the Tusas Mountain area. It forms massive, poorly-schistose outcrops which weather to purplish-gray, smooth, rounded outcrops which break with conchoidal fracture. Large- to fine-scale crossbedding, defined by purple or black bands, is commonly highly contorted.

Iron Formation

Introduction

Iron formations are known to occur in a number of Proterozoic supracrustal successions of New Mexico and Arizona, for example, in the southern Sangre de Cristo Mountains (Robertson and Moench, 1979; Riesmeyer, 1978), the Pedernal Mountains (Dale Armstrong, 1979, personal commun.), the Picuris range (J.M. Robertson, 1979, personal commun.), and in the Yavapai series of Arizona (Anderson and others, 1971). Although Barker (1958) made no mention of the occurrence of ironstone in the Tusas Mountains, the presence of iron deposits in the Cleveland Gulch area and on Iron Mountain (to the west of this study area) has been noted by several authors (Bertholf, 1960; Harrer and Kelly, 1963; Harrer, 1965; Bingler, 1968; McLeroy, 1970). There has been no recorded production from these deposits. The presence of ironstone in the Moppin series is of interest not only from an economic standpoint but also for the constraints it provides on interpretations of the depositional environment of part of the Moppin series. Because iron formations are relatively rare, of minor extent, and very poorly understood in the middle Proterozoic rocks of southwestern United States, careful attention was paid to the stratigraphic relationships and the textures preserved in the iron formation of the Moppin series.

In this report, the terminology for iron formation will, in part, follow that of Kimberley (1978). Ironstone will be used for a chemical sedimentary rock which contains greater than 15 percent iron; iron formation is a mappable rock unit which is composed mainly of ironstone. Although the quartz present in the ironstone is interpreted as recrystallized chert, quartz, rather than chert, will be used. This terminology will be followed because of the difficulty in distinguishing between original detrital quartz grains and recrystallized chert.

Distribution and stratigraphic relationships

Two types of ironstones are found interbedded with the volcanic and sedimentary rocks of the Moppin series. One type (in this study, the banded ironstone) is a very fine-grained banded magnetite-quartz unit intimately interlayered with the metapelites (described above) in the upper portion of the Moppin series. The second type (in this study, the hematite-quartz ironstone) is a sugary-textured, unbanded, hematite (+/- magnetite)-quartz unit which is interbedded with amphibolites and amphibolitic gneisses in the lower part of the Moppin series. The hematite-quartz units have not been previously recognized in the Tusas Mountains.

The banded ironstone forms a distinct stratigraphic horizon which extends from the eastern flanks of hill 9180

(sec. 29, T.28N., R.8E.) westward to the middle of sec. 25, T.28N., R.7E. The contact relations between the ends of the ironstone and the surrounding schists are not exposed. A small outcrop of banded ironstone was observed at about the 8800 foot elevation in the NW 1/4, NE 1/4, sec. 32, T.28N., R.8E. The area mapped as banded iron formation on Plate I is a horizon in which ironstone layers are discontinuously interbedded with other rock types. The aggregate thickness of the banded ironstone, excluding phyllitic interlayers, is generally 1.5 to three meters. The iron formation was mapped primarily on the basis of float as outcrops are poor. An s-shaped fold mapped in sec. 29, T.28N., R.8E. is diagrammatic; abundant ironstone colluvium on the eastern side of the valley hinders an exact structural interpretation. Trenches along this horizon show the ironstone as: 1) stratigraphically concordant with the surrounding sedimentary rocks (figure 4); 2) made up of lenticular bodies which thin from a couple meters to several centimeters across the width of a trench exposure; and 3) consistently associated with chlorite or felsic volcaniclastic phyllites (the metapelites). Although the iron formation does not seem to occur at the same stratigraphic level throughout the phyllitic unit (see Plate I), it is not discordant to the enclosing rocks at any point along strike. It would appear, therefore, that lensing of units within the phyllitic section or unrecognized

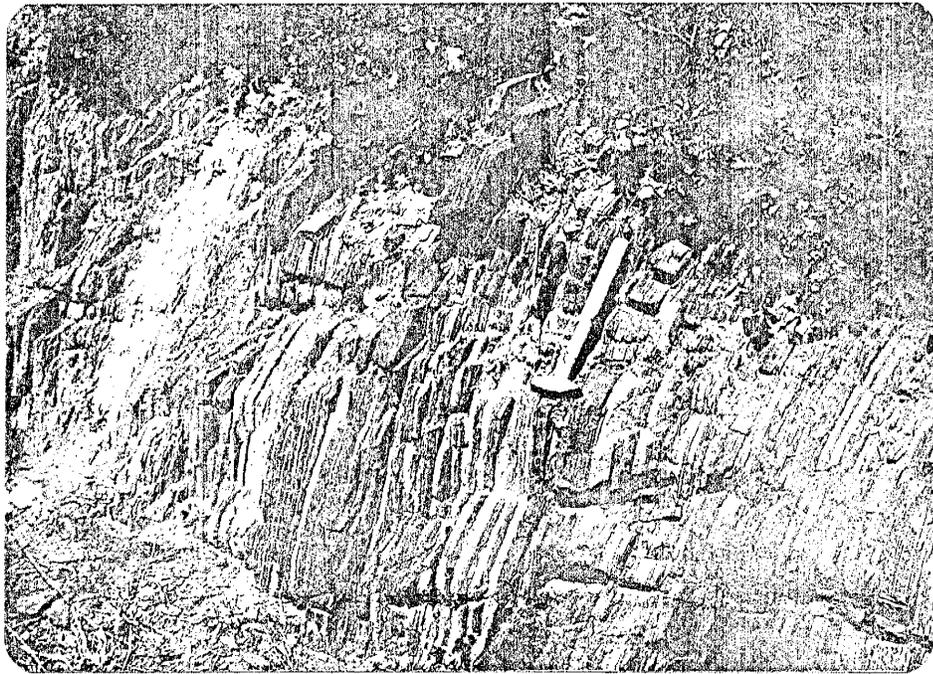


Figure 4. Photograph of banded iron formation. Dark bands are hematite-quartz ironstone, lighter bands are chlorite and felsic phyllites (the metapelites).

small-scale faulting within the depositional basin are the probable causes of the apparently discordant relationship on Plate I.

Unbanded hematite-quartz ironstone forms several stratigraphic units. A major horizon extends at least 7 kilometers along strike from sec. 28, T.28N., R.8E. to the middle of sec. 23, T.28N., R.7E. Two smaller horizons are found north of the major one in secs. 20 and 21, T.28N., R.8E. and one (or possibly two) to the south in secs. 24 and 29, T.28N., R.8E. The unbanded hematite-quartz ironstones appear to abruptly end laterally. These horizons form bold outcrops, appear to be concordant with enclosing metavolcanics, and in places are interbedded with phyllitic chlorite schists. The thickness of the beds varies from less than 3 meters to approximately 15 meters.

Textures and lithology

The banded iron formation outcrops are subdued, blocky, and a distinctive dark purplish-black color. Magnetite-rich bands (less than 1 mm to several centimeters thick) alternate with quartz-rich bands and lenses (1 to 10 mm in width). In addition, the quartz-magnetite banded portions are interlayered with felsic and chlorite phyllites and schists (figure 5) which commonly contain disseminated magnetite octahedra. Quartz forms either pinch-and-swell textures or pods completely enclosed within the



Figure 5. Photograph of quartz-magnetite ironstone illustrating the fine banding between magnetite-rich and quartz-rich layers as well as banding between ironstone and the metapelites.

magnetite-rich matrix. Laminae in the iron-rich material bend around the quartz bodies. Quartz bands infrequently form ptygmatic folds with axes that parallel the foliation. Other features developed in the quartz-magnetite banded units are brecciation and possible soft-sediment slumping; cross bedding or graded bedding were not observed.

In thin section, magnetite grains (0.01 to 0.20 mm) and magnetite aggregates form bands in which quartz is subordinate (10 to 20 percent) and hematite occurs in trace amounts (figure 6). The quartz forms concave boundaries with the magnetite, a texture that may represent silica filling of original pore space (Dimroth, 1976). These bands alternate with quartz-rich bands in which the quartz grains are polygonal and larger (0.08 to 0.16 mm in diameter) than in the magnetite-rich bands; magnetite is less than 25 percent and hematite and muscovite occur in trace amounts. Very small minor folds can be seen in thin section (defined by 'z-ing' of the magnetite-rich bands) as well as round to oval, 0.2 to 0.4 mm, ragged quartz grains which are usually rimmed with hematite. Some thin sections also contain thin (less than 0.25 mm in width) discontinuous carbonate-magnetite layers. Interbedded chlorite schists are composed principally of chlorite and quartz with hematite as the main accessory and only minor magnetite.

The unbanded hematite-quartz ironstone forms bold, nonfoliated outcrops which typically display mottled (due to



Figure 6. Photomicrograph of quartz-magnetite ironstone illustrating the fine banding between magnetite-rich and quartz-rich layers.

brecciation?) colors of light to dark pinkish brown, black, brown, and grayish white. Banding is only poorly developed and, if present, discontinuous and irregular. A granular or sugary texture is characteristic of all specimens. Rounded "clasts" of sugary quartz grains of one color are surrounded by sugary quartz of another color. Veins of hematite and chlorite are discontinuous and irregular.

In thin section this unit is composed almost entirely of anhedral quartz grains (0.08 to 0.71 mm in diameter). Chlorite, hematite, and chlorite + hematite form discontinuous, nonparallel (0.03 to 0.61 mm in width) veinlets with hematite usually in the core of the vein and chlorite along the edges. The chlorite crystals are subparallel within each vein, but neither the veins nor the chlorite crystals in different veins parallel each other. In some cases the chlorite is perpendicular to the direction of the vein and these chlorite crystals feather into the quartz adjacent to the veins; in other cases the chlorite parallels the direction of the vein. When only chlorite occurs in a veinlet, the chlorite forms aggregates rather than distinct crystals.

Behind a small cabin in the upper part of Spring Creek (sec. 23, T.28N., R.7E.) an atypical outcrop of the hematite-quartz ironstone is white and regularly banded with chlorite. In thin section, calcite makes up to 50 or 60 percent of the rock. Bands or lenses of polygonal (0.08 to

0.15 mm in width) quartz grains, with traces of carbonate and opaques, alternate with bands of muscovite + carbonate + quartz and bands of intergrown (up to 0.7 mm in length) carbonate with only minor quartz. In the quartz-rich bands, carbonate is approximately the same size as the quartz grains and occurs only at corners of the polygonal grains. In the carbonate-rich layers, however, quartz is much smaller than the carbonate and occurs both within and along grain boundaries of the carbonate. Chlorite and muscovite parallel the layering in the rock as well as the long direction of the carbonate grains.

Discussion

Iron formations are poorly understood rock types. Carbonate rocks are apparently the closest comparable rock type, and indeed recent studies document the similarities between carbonate and iron formation textures and structures (Dimroth, 1975, 1976). Although cherty iron formation is almost universally accepted to be a chemical rather than detrital sediment (Dimroth, 1975, 1976; Goodwin, 1962; Gross, 1972; Eichler, 1976) several problems continue to be enigmatic. Briefly stated the most major of these problems are:

(1) The source of iron and silica. The vast amounts of iron and silica present in iron formations must be accounted for. Most speculations center around three main

alternate sources: contemporaneous volcanic rocks, continental weathering products, or normal to abnormal concentrations in sea water.

(2) Transport and deposition. Major problems concern the solubility of iron under geologically realistic conditions, and the characterization of a physical and geochemical environment in which large amounts of iron and silica are deposited to the almost complete exclusion of other elements such as calcium and magnesium.

(3) Layering. No totally satisfactory explanation has been proposed to account for the rhythmic banding which occurs in nearly all iron formations whereby silica and iron are concentrated in separate bands on a scale from less than 1 mm to several centimeters.

Until recently, most Precambrian iron formations were classified as either Superior or Algoma types (Gross, 1965). Algoma types were typically thought to be associated with volcanic successions whereas Superior types were associated with continental-shelf sedimentary rock assemblages. This classification has proven unsatisfactory, however, because iron formations are now known to occur with many different rock associations and in various tectonic settings. Kimberley (1978) suggested a classification scheme based on the paleoenvironmental interpretation of the ironstones and their rock associations. He described six paleoenvironmental types of iron formation. The deposits in

Cleveland Gulch appear to most closely resemble Gross's Algoma type deposits and Kimberley's Shallow-Volcanic-Platform Iron Formation (SVOP-IF). Not only do they occur within a volcanic pile (the Moppin schists) but they are thin and areally limited as are the classic Algoma and SVOP-IF types. In addition they are delicately laminated, exhibit soft sediment deformation, and intraformation brecciation; they lack cross-bedding and oolitic textures.

The banded ironstones in Cleveland Gulch were thought by McLeroy (1970) to be hydrothermal replacement products of muscovite and chlorite schists by iron-rich fluids associated with the Tusas granite. McLeroy based his conclusions primarily on his interpretation of the field relationships between the Moppin series, the iron formation, and the Tusas granite. His interpretations include: 1) the iron formation is cross-cutting to the enclosing Moppin schists; 2) the Moppin series is primarily a continental volcanic sequence; and 3) the granitic intrusives are the source of the iron. Close study of these deposits during this investigation makes a hydrothermal replacement interpretation unlikely in light of observable field relationships and preserved textures.

(1) Cross-cutting relationships. McLeroy claimed that somewhere in Cleveland Gulch the strike of the iron formation changes 15 degrees over a strike length of 100 feet

whereas an amphibolite directly "below" (stratigraphically or topographically?) maintains a constant strike over the same distance. He used this fact as evidence for an apparent discordance between the iron formation and the enclosing schists. Whether this can be used as proof of discordance is doubtful due to the fact that broad, low amplitude folds as well as tight to isoclinal folds are present in this area of the Moppin series with wavelengths ranging from a few centimeters to over 30 meters. Due to discontinuous outcrops in this area, one unit could conceivably appear to maintain a constant strike while another appear to change strike. In addition, strike measurements on magnetite-bearing rocks, unless carefully taken, may contain errors easily in excess of 15 degrees.

Magnetite layers in the quartz bands and the quartz-magnetite bands in the phyllitic schists are, without exception, conformable both on an outcrop scale and in thin section. If the iron deposits were hydrothermal replacement products, it would be expected that cross-cutting relationships could be observed. Instead, contacts between layers are sharp and closely follow what appears to be the original bedding (see figure 35). The probability that the hypothetical replacement fluids followed bedding planes on a scale of less than 1 mm in all cases is extremely remote.

(2) Continental volcanic sequence. McLeroy interpreted the Moppin series to be largely a sequence of

continental volcanics. He based his interpretations on: a) the presence of a rhyolitic ash-flow tuff (the muscovite-feldspar-quartz schist) which overlies the Moppin schists and on b) identifying the metapelites interbedded with the ironstone as tuffs and flows. This study, however, interprets the metapelites to be finely laminated shales and mudstones based on the absence of preserved igneous features. This study also presents evidence for an episode of uplift and erosion between the Moppin series and the deposition of the felsic schists which suggests that there may have been different depositional environments for the felsic volcanics and the iron formation. Although conclusive textural evidence has not been preserved to suggest a subaqueous environment (such as relict pillows in the mafic volcanic rocks), the presence of a thick phyllitic (shale) horizon associated with the iron formation suggests a subaqueous depositional environment.

(3) Granite source. McLeroy proposed that the hydrothermal replacement fluids emanated from the Tusas granite. Aside from the fact that the Tusas granite is leucocratic (biotite is very minor) and does not appear to be a good source for iron, another problem exists with this idea. If the granite were the source for the iron it would be expected that the iron formation would halo the intrusives. Instead, the banded iron formation appears on both sides of a small granite intrusive with no increase in

the width of the deposit next to the granite. The iron deposits are "associated" only with a very small pluton of the granite (sec. 25, T.28N., R.8E.) and no iron association exists with the much larger, mineralogically identical pluton to the north.

For these reasons a hydrothermal replacement origin for the Cleveland Gulch deposits is considered to be unlikely and it is proposed instead that the banded iron formation is part of a sedimentary sequence deposited during a pause in volcanic activity (it is commonly assumed that volcanism and chemical precipitation are antipathetic, see Dimroth, 1976). The quartz and magnetite layers are interpreted to be recrystallized chemical precipitates on the basis of comparison with other iron deposits and because of the difficulty in forming alternating bands of pure quartz and pure iron by mechanical processes. The close association with chlorite and felsic phyllites (volcaniclastic shales) suggests quiet deposition of fine-grained clastic sediments alternated with intervals of chemical precipitation. The unbanded hematite-quartz deposits associated with the amphibolites to the north of the Cleveland Gulch deposits are also interpreted to represent volcanic lulls. The possibility that these deposits are mechanically eroded quartzites rather than recrystallized cherts is rejected because of the difficulty in depositing an essentially pure quartzite between mafic volcanic horizons.

Intrusive Rocks

Granodiorite

Two types of granodiorite are present within the study area. A distinctive lineated granodiorite is Barker's type Maquinita; in this study it is referred to as the lineated granodiorite. This granodiorite is gray and white and poorly foliated. It is distinguished by prominent, near-vertical lineations of biotite/chlorite knots and 1 to 5 mm plagioclase phenocrysts; quartz phenocrysts are generally absent. The granodiorite is apparently concordant to the enclosing Moppin schists and contacts with the Moppin are sharp and parallel the foliation of the schists. Large inclusions of the country rock are only rarely found along the border of the granodiorite although felsic veins are common in the country rock within 30 meters of the contact. A cataclastic texture is prominent in thin sections.

A second granodiorite, best exposed along the upper reaches of Spring Creek (sec. 27, T.28N., R.7E.), is referred to as the granodiorite of Spring Creek. The contact with the lineated granodiorite is both gradational and sharp, but the exact chronologic relationship between the two has not been determined. Foliation in the granodiorite of Spring Creek parallels that of the country rock; greenschist inclusions (which are common) also parallel the regional strike although their dips are often very different from the near vertical

ones generally found in the Tusas Mountain area. This granodiorite is pinkish gray and highly schistose (a phyllitic sheen is common on weathered surfaces). Rounded, slightly bluish, quartz phenocrysts (up to 3 mm) and broken plagioclase phenocrysts are prominent on weathered surfaces. Biotite/chlorite knots are generally absent. In thin section this granodiorite differs from the lineated granodiorite in that microcline, carbonate, and plagioclase are not present in the groundmass which are present in the lineated granodiorite.

Tusas granite

The Tusas granite is discordant with the country rock on its western margin and concordant on its southern margin. The contacts with the Moppin series are sharp although felsic material interfingers with the country rock within 150 meters of the contact. Fluorite is commonly found within the granite and greenschists along the contact. Xenoliths are absent except for a single large lens of mafic schists, possibly a roof pendant, exposed in a mine cut south of Tusas Peak (sec. 24, T.28N., R.7E.). The granite is only poorly foliated, if at all. Three mineralogic variations of Tusas granite were noted within the major pluton which are described in Appendix I (p. 135, ff.). All three variations are intergradational.

Another small exposure of the Tusas granite in the southeast corner of sec. 25, T.28N., R.8E. appears to discordantly intrude the lower arkoses and upper portion of the Moppin series. This area is covered solely by granite boulders although no actual outcrops could be found. It is possible that this exposure represents a paleotalus deposit similar to the quartzite boulder fields described earlier rather than another intrusive outcrop.

The Tusas granite is interpreted as a late- to post-tectonic granitic intrusive. This interpretation is based on the sharply discordant contacts along the western edge of the granite and the general lack of foliation within the body of the pluton. The concordant foliation of the porphyritic basaltic andesite immediately adjacent to the western border of the granite suggests that forceful intrusion of the granite locally created parallel schistosity in the country rock. This feature discounts Gresens and Stensrud (1974, Table 1) suggestion that the Tres Piedras granite is the possible basement to the Moppin supracrustal sequence. The general lack of schistosity in the Tusas granite suggests that it post-dated the major period of deformation in the Tusas Mountains. This interpretation is supported by a zircon date for the syn-tectonic lineated granodiorite between 1670 m.y. and 1715 m.y. (determined by L.T. Silver, in Barker and Friedman, 1976) whereas the Tusas granite zircon age is 1625 m.y. (Maxon, 1976a,b).

Late intrusives

Five varieties of late intrusives cross-cut the Precambrian rocks of the Tusas Mountains. The most abundant types are quartz veins (and pegmatites), aplites, and amphibolite dikes. One occurrence of a granodiorite dike and a granitic dike are also found. All except the aplites are considered to be Precambrian in age based on metamorphic textures observed in the field and in thin section. The aplites are considered to be Precambrian but because they lack obvious metamorphic fabrics, it is possible they are Tertiary.

Structure

Two opposing structural interpretations have been made in the past for the Tusas Mountains. Barker (1958) proposed that the Precambrian rocks of the Tusas Mountains have been folded by a single deformational event. He mapped two major folds in the Las Tablas 15-minute quadrangle: the Kiowa syncline underlying Quartzite Peak and Kiowa Mountain, and the Hopewell anticline to the north of the Kiowa syncline. Bingler (1965) presented evidence for three deformational episodes in the La Madera (7 1/2-minute) quadrangle (south of the Las Tablas 15-minute quadrangle): the first episode produced northeast-trending isoclinal folds, the second, overturned northwest-trending isoclinal folds, and the third, broad west-trending folds. Although Bingler acknowledged that few remnants of the first deformation have been preserved, he described the episode as a penetrative deformation with unrestricted flowage of material; he concluded that the first deformation destroyed the original stratigraphic sequence.

Foliations

The Precambrian schists in the Tusas Mountain area are generally well foliated and strike N80W to N80E in the eastern portion of the map area and N60W to N80W in the western half. In sec. 23, T.28N., R.7E. and in sec. 33, 34, T.28N., R.8E. the foliations are more northerly.

Foliation appears to parallel compositional banding in many exposures but clearly cross-cuts bedding in a number of outcrops. The arkoses best illustrate the parallel relationship because preserved sedimentary structures allow easy identification of the original bedding surfaces. Although the Moppin schists retain few primary sedimentary and igneous features, schistosity closely parallels the contacts between different lithologic units. In addition relatively thin and lithologically distinctive units (such as the quartz-hematite ironstones, the quartz-magnetite ironstones, and the hornblendites) can be traced for several miles along strike suggesting that bedding or original compositional layering is preserved. Two foliations at high angles to each other are present in the subarkoses along the Spring Creek roadcuts (sec. 33, T.28N., R.8E.). The most readily observable foliation (striking in a northerly direction) represents transposed compositional layering which is a result of isoclinal folding of these rocks. Pseudo-crossbedding occurs in this foliation. A second foliation (less easily identified from the road, but easily seen looking down on these outcrops) strikes in a northwesterly direction and represents the foliation associated with original bedding. True crossbedding can be found in this direction. Foliation is poorly developed in the vitreous quartzites and where present does not always parallel bedding; this is especially true on Quartzite Peak.

Throughout the area the beds are generally steeply dipping or dip moderately to the northeast. crossbedding and graded bedding in the arkoses indicates that the beds face to the south but dip to the northeast and therefore are overturned to the southwest.

Lineations

Several types of lineations are present in the Tusas Mountain area. The alignment of plagioclase phenocrysts and hornblende crystals are the most common types of lineations in the Moppin schists. Crinkling around magnetite grains in the muscovite-feldspar-quartz schist and the chlorite phyllite associated with it, and stretched pebbles in pebble conglomerates form other distinctive types of lineations. Megascopeic fold axes form fairly common lineations in the Moppin and intersecting foliations form the rarest type of lineation. The lineated granodiorite has abundant biotite knots aligned down the dip direction of the foliation. Most lineations are vertical or plunge steeply to the northwest.

Minor folds

Megascopeic folds are common in the amphibolites, the muscovite-feldspar-quartz schist and associated chlorite phyllite, the iron formation, the hornblendites, and in

biotite-quartz schists of the Moppin series. Tight to isoclinal folds are the most common type, but open asymmetric types also occur. These folds have wavelengths from less than 30 cm to 30 m. The axes of these folds trend between N70W and north and plunge steeply. A few of the larger s-shaped folds in the hornblendites and muscovite-feldspar-quartz schist are shown on Plate I.

Faults

Evidence for faulting in the Tusas Mountain area consists mainly of stratigraphic offsets; brecciation and alteration are observed only along the southern branch on the eastern-most portion of the fault following Deer Park and Spring Creek.

The vitreous quartzite is mapped in this report in fault contact with the arkoses and subarkoses for five reasons: 1) The style of deformation in the quartzites is apparently different than that developed in the arkoses. 2) The pronounced lithological differences between the two units. The quartzite is composed of essentially pure quartz whereas the arkoses are not only poorly sorted and immature but are interlayered with mafic schists. 3) The contact is in all cases hidden by Tertiary or Quaternary gravels. The deposition of these young units may have utilized a previous fault zone. 4) Prospect pits and quartz veins are common along the contact in the southern portion of sec. 33, T.28N.,

R.8E.). Prominent fault scarps are present on the north side of the fault (between Sullivan and Deer Parks) northwest of this study area.

Two major faults in the Tusas Mountain area juxtapose vitreous quartzite, arkoses, and subarkoses. The two faults are postulated to merge in sec. 32, T.28N., R.8E. and follow Spring Creek and Deer Park to the western edge of the study area.

Discussion

The deformational history of the Tusas Mountain area is probably quite complex, but interpretation of the structure is greatly hindered by limited exposures and an incomplete knowledge of the stratigraphic relationships. One of the most important structural questions in the Tusas Mountain area is whether the Moppin schists have been deformed more than once. The limited structural data obtained in this study does not allow a definitive answer to this question but structural work to the west by T. Gibson (1980, in progress) should yield a more complete interpretation.

At present and even with limited structural data, it appears that Barker's original structural interpretation is too simple and that the rocks of the Tusas Mountain area may have been complexly deformed. Foliations almost everywhere parallel original bedding, suggesting at least one

episode of isoclinal folding. Mineral lineations in the schists are consistently near vertical. If these lineations are b-lineations, this would suggest nearly vertical fold axes. Nearly vertical fold axes are present in megascopic folds in many of the eastern exposures of the Moppin schists. If the isoclinal folding event that produced the present northwesterly foliation in the schists also developed the near-vertical mesoscopic fold axes, then the deformational history of this area is more complex than that suggested by Barker. Either a single complex stress field or two separate folding events has deformed these supracrustal rocks.

Several units in the supracrustal rocks are apparently repeated in the map area: the Burned Mountain metarhyolite, the pebble conglomerates, the hornblendites, and the hematite-quartz schists. Although these units may be separate beds and represent some sort of cyclical deposition in portions of the supracrustal succession, they may also be single beds that have been repeated by isoclinal folding. In addition to displaying repeated beds, these units are also geographically restricted; the Burned Mountain and the pebble conglomerates occur only in the western portion of the map area whereas the hornblendites and the hematite-quartz schists are present only east of the Burned Mountain and pebble conglomerates. Two explanations for the geographic restrictions of these units can be suggested. These units may never have been deposited outside of where they now

appear in the supracrustal succession. This could be the result of separate source areas for these rocks, one to the west (for the felsic rocks) and one to the east (for the mafic rocks). Alternatively the apparent geographic restrictions could be the result of complex folding. Support for this interpretation, however, requires the collection of much more structural data from the field.

Metamorphism

The Precambrian rocks in the Tusas Mountain area have undergone regional dynamothermal metamorphism. Metamorphic grade increases from northwest to southeast across the map area. Although most of the map area is dominated by low-grade assemblages, rocks in the eastern and southeastern areas contain assemblages characteristic of the low temperature end of medium-grade metamorphism (Winkler, 1974).

The felsic schists do not contain mineral assemblages that reflect changing metamorphic conditions. The arkoses are also poor reflectors of metamorphic grade although they do show a general trend for quartz + Kspar + chlorite in the west to be replaced to the east by quartz + Kspar + biotite. The Moppin series, however, contains mineral assemblages which are diagnostic for determining metamorphic grade. The majority of the Moppin schists are metamorphosed volcanoclastic rocks with interlayered volcanic rocks. The mineral assemblages consistent with low-grade metamorphism in the volcanoclastic rocks are:

chlorite + biotite + quartz
muscovite + biotite + chlorite
plagioclase + quartz + biotite
biotite + Kspar + quartz
chlorite + quartz + Kspar

In thin section, zoisite/clinozoisite, epidote, and carbonate are commonly present either in trace amounts or as major

constituents. Quartz is generally very abundant as is biotite when it is present. Chlorite is usually a minor constituent, but is always present in at least trace amounts. Plagioclase, although usually highly altered, has a composition in the oligoclase range. These observations are consistent with low-grade metamorphism of sedimentary rocks. The composition of the plagioclase and the common presence of biotite suggests that metamorphism was generally within the upper temperature range of low-grade conditions. These assemblages provide little constraint, however, on the metamorphic pressure.

The low grade mineral assemblages for the mafic volcanic rocks is chlorite + hornblende-actinolite + plagioclase +/- quartz, epidote, clinozoisite, garnet. Quartz is always minor, garnet is very rare, and epidote and clinozoisite occur either in trace amounts or as major constituents. The plagioclase composition is in the oligoclase range. Because hornblende + oligoclase occurs on both sides of the transition between low-grade and medium-grade metamorphism, it is usually the presence or absence of actinolite and/or chlorite + muscovite which determines the metamorphic grade of mafic volcanic rocks, rather than the composition of the plagioclase.

The medium-grade assemblages in both the sedimentary and volcanic rocks of the Moppin series are the same as in the low-grade but are marked by the absence of

chlorite, and in the sedimentary rocks, by the presence of hornblende. The plagioclase composition does not appear changed. The temperature at which actinolite and/or chlorite + muscovite disappears (eg., the boundary temperature between low- and medium-grade metamorphism) is estimated by Winkler to occur at about 500 C. The metamorphic pressure cannot be defined by these assemblages because the transition between chlorite and hornblende is essentially independent of pressure. However, kyanite is present in a small exposure in sec. 27, T.28N., R.8E. which suggests that, at a minimum temperature of 500 C, the pressure would exceed 4 kbar (Winkler, 1974, fig. 15-3) in that area.

The pelitic schists south of the Spring Creek road (sec. 33, T.28N., R.8E.) contain representative mineral assemblages for the southeastern portion of the study area. In thin section the assemblage is staurolite + garnet + biotite + quartz + muscovite. Two flakes of chlorite remain in contact with the staurolite in one sample which suggests that the metamorphic conditions were close to the reaction line between (chlorite + muscovite)-out and staurolite-in. In addition, although kyanite is not in the thin sections studied from these exposures, staurolite and kyanite appear in outcrop at about the same place. The simultaneous appearance of these two minerals, plus the very small remnants of chlorite, indicates the nearness of the metamorphic conditions to those necessary for the formation

of both staurolite and kyanite. The point at which the two reaction lines cross each other in a pressure-temperature diagram suggests that the metamorphic temperature was approximately 550 C and the pressure was approximately 5.2 kbar according to Winkler. Holdaway's (1978) determinations for the stability field of the alumina silicates suggests a much lower pressure for this assemblage, around 3.3 kbar. Both sets of stability curves are illustrated in figure 6.

Contact metamorphism associated with the intrusion of the Tusas granite and the granodiorites is negligible. Although epidote veins as well as felsic veins are common in the Moppin series within 150 meters of the granite contact, hornfels are not developed and common contact metamorphic minerals such as diopside and tremolite are absent. This suggests that the granitic intrusives were probably at equal or lower temperatures than the surrounding schists at the time of intrusion.

A second episode of metamorphism is suggested by the common presence of micas developed at high angles to the foliation. Thin sections for all rocks in the Tusas Mountain area contain poikiloblastic, euhedral to ragged biotite flakes or small muscovite blades generally at high angles to the foliation. In addition, porphyroblasts of staurolite are randomly oriented throughout the pelitic schists, the metapelites of the Moppin series, and the chlorite phyllite in the arkoses.

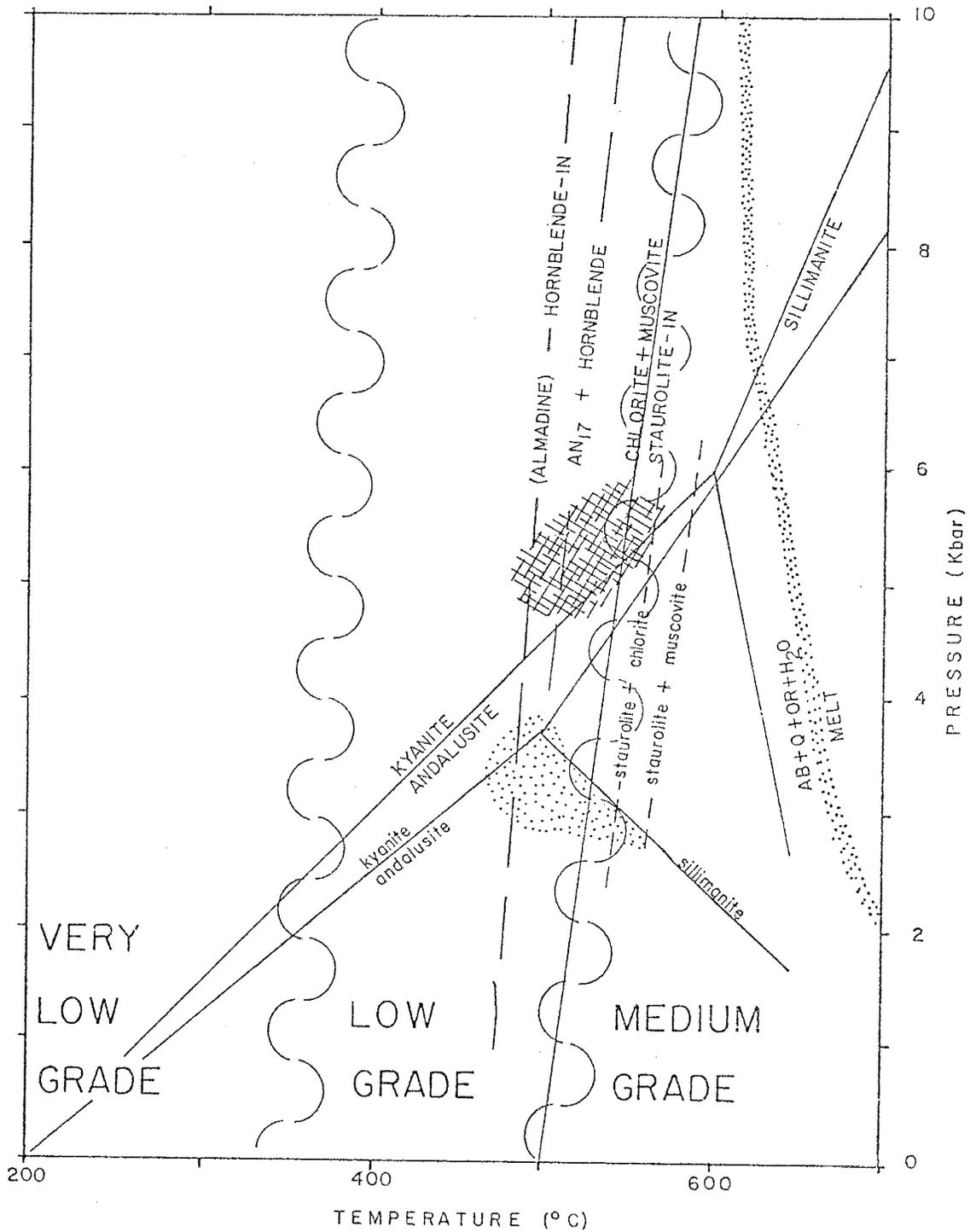


Figure 7. Petrogenetic grid illustrating the approximate pressure-temperature environment for some of the metamorphic rocks in the Tusas Mountain area. Mineral stability curves taken from Holdaway (1978) (in lower case letters) and from Winkler (1974). The P-T conditions for metamorphism in the Tusas Mountains is dotted using Holdaway's data and ruled using Winkler's.

Retrograde metamorphism does not appear to be of major importance in the Tusas Mountain area. The only evidence of retrograde metamorphism is preserved in several outcrops which contain what appear to be amphibole porphyroblasts that have been pseudomorphously replaced by chlorite-actinolite (see Appendix I, p. 96). However, evidence for retrograde metamorphism has not been observed in any other rock type within the study area.

Geochemistry

Chemical analyses of 12 rock samples collected from the Tusas Mountain area are presented in tables 3 (mafic schists) and 4 (granodiorites) and CIPW norms in table 5. Sample descriptions are reported in Appendix I. Table 6 presents chemical analyses from other studies on Precambrian rocks in the Tusas Mountains.

Supracrustal rocks

Five samples were collected from the Moppin series: a typical greenschist of probable igneous origin (sample 99197), a hornblendite (99191), a medium-grained amphibolite (90051), a porphyritic amphibolite (99293), and matrix material from the mafic pyroclastic breccia (99292). The greenschist compares with Nockold's (1954) average tholeiitic basalt whereas the amphibolites (90051, 99293, 99292) compare more closely to Nockold's average andesite. The porphyry (99293) is slightly more silica-rich and magnesium- and sodium-poor compared to the other two amphibolites. The hornblendite (99191) is very similar to Nockold's average hornblendite except in having a higher magnesium content and lower calcium, sodium, and potassium contents.

Four samples were taken from the Spring Creek road cuts: two are probably of igneous origin (99252, 99195), and two are sedimentary rocks (a hornblende gneiss, 99251, and a

pelitic schist, 99253). The igneous rocks from the Spring Creek roadcuts differ from the igneous rocks in the Moppin series in having lesser amounts of silica, aluminum, sodium and potassium, and greater amounts of magnesium and calcium.

One sample of amphibolite dike (98262) compares closely to Nockold's average gabbro. The amphibolite in the Spring Creek road cut (99252) is very similar to the composition of this dike and Nockold's average gabbro suggesting that perhaps it is a sill.

The volcanic rocks are classified by Irvine and Baragar's classification method (figures 8a, b, c). They are also plotted (along with other published analyses from the Tusas Mountains) on a AFM diagram and a Jenson cation plot (figure 9).

Irvine and Baragar's classification method reveals that the volcanic rocks from the Tusas Mountain area are subalkaline (figure 8a), calc-alkaline (figure 8b), and basalts (figure 8c). A silica content vs. frequency histogram (figure 10) reveals that the rocks from the Tusas Mountains have a bimodal distribution: except for one sample, intermediate rocks (SiO₂ content between 56 and 71 percent) are absent. The bimodal distribution is also evident in the AFM diagram and the Jenson cation plot. The majority of these rocks are basalts and rhyolites (figures 8c and 9). Irvine and Baragar's normative plagioclase vs. Al₂O₃ plot suggests that the rocks of the Tusas Mountain area are

calc-alkaline. They also appear calc-alkaline on the Jenson cation plot, although rocks from other studies in the Tusas Mountains do not appear as such. The Pearce discrimination diagram (figure 11) suggests the majority of the rocks in the Tusas Mountain area were emplaced within a continental environment.

A compilation of geochemical data from other Proterozoic terranes in the southwestern United States reveals that the rocks of the Tusas Mountain area are anomalous; calc-alkaline mafic rocks are not present elsewhere. Figures 12, 13, and 14 demonstrate the tholeiitic nature of the rocks in northern New Mexico-southern Colorado, central New Mexico, and Arizona, respectively. Precambrian volcanic rocks from the northern New Mexico-southern Colorado area are strongly bimodal (basalts and rhyolites), the basalts are consistently tholeiitic, and the rhyolites are both calc-alkaline and tholeiitic. Arizona rocks are unique in having an almost complete absence of calc-alkaline rocks and in displaying a complete continuum from basalts through andesites, dacites, and rhyolites. The rocks from central New Mexico appear to be somewhat bimodal although intermediate rocks are present in the Pedernal Mountains. Calc-alkaline rhyolites are present but they are not as abundant as in northern New Mexico-southern Colorado terranes.

	99197	a	99191	b	90051	99293	99292	c	99252	99195	99251	99253	98262	d
SiO ₂	50.25	50.83	42.61	42.00	49.79	54.52	53.56	54.20	49.46	44.16	51.74	54.99	48.14	48.36
TiO ₂	0.88	2.03	0.82	2.86	1.11	0.96	0.90	1.31	1.04	1.35	1.22	0.84	0.82	1.32
Al ₂ O ₃	16.27	14.07	12.34	11.39	18.14	18.52	18.50	17.17	16.25	17.78	16.23	21.52	17.27	16.84
Fe ₂ O ₃ (T)	10.88	12.95	11.60	16.71	11.01	9.86	10.28	9.50	10.29	15.48	11.14	8.85	9.86	
MgO	6.28	6.34	17.06	12.35	4.90	2.35	3.38	4.36	6.36	7.77	6.06	2.05	6.40	8.06
CaO	9.92	10.42	8.80	11.31	7.05	8.56	7.63	7.92	10.32	8.07	8.95	3.97	10.07	11.07
Na ₂ O	3.16	2.23	0.62	1.80	4.07	4.01	3.43	3.67	2.77	2.80	1.61	1.58	3.10	2.26
K ₂ O	0.23	0.82	0.22	0.80	1.48	0.39	1.64	1.11	0.40	0.41	0.49	3.60	0.61	0.56
MnO	0.15	0.18	0.17	0.24	0.16	0.14	0.15	0.15	0.19	0.36	0.23	0.18	0.17	0.18
P ₂ O ₅	0.12	0.23	0.43	0.33	0.18	0.20	0.11	0.28	0.14	0.30	0.12	<0.05	0.09	0.24
LOI	1.94		3.86		1.29	0.55	0.72		0.93	1.73	1.92	2.15	2.56	
TOTAL	100.08		98.53		99.18	100.06	100.30		98.15	100.21	99.71	99.73	99.09	
Cu	72		10		166	104	58		34	48	110	66	34	
Ni	114		280		50	88	46		94	140	84	62	88	
Pb	20		80		120	100	140		60	80	120	80	200	
Zn	98		114		208	138	150		132	166	174	126	130	
Co	130		124		140	94	96		88	154	124	84	100	

analyst: L.A. Brandvold, New Mexico Bureau of Mines and Mineral Resources, by atomic absorption

99197, typical greenschist in Moppin series; 99191, hornblende in Moppin series; 90051, amphibolite in Moppin series; 99293, porphyritic amphibolite in Moppin series; 99292, matrix of mafic pyroclastic breccia; 99252, amphibolite from pelitic schists; 99195, mafic schist from subarkoses; 99251, hornblende gneiss from pelitic schists; 99253, pelitic schist; 98262, dike cross-cutting lineated granodiorite, a, average tholeiitic basalt; b, average hornblende; c, average andesite; d, average gabbro (from Nockolds, 1954).

Table 3. Chemical analyses of mafic schist in the Tusas Mountain area.

	99192	99000	*	**
SiO ₂	62.35	66.60	61.86	66.88
TiO ₂	0.40	0.27	0.58	0.57
Al ₂ O ₃	18.05	17.76	16.91	15.66
Fe ₂ O ₃ (T)	5.08	3.34	5.24	4.20
MgO	2.52	1.08	0.96	1.57
CaO	2.76	0.32	2.54	3.56
Na ₂ O	5.27	5.01	5.46	3.84
K ₂ O	3.13	2.07	5.91	3.07
MnO	0.07	0.06	0.11	0.07
P ₂ O ₅	0.18	0.11	0.19	0.21
LOI	0.52	1.73		
H ₂ O			0.53	0.65
TOTAL	100.33	98.35		
Cu	136	46		
Ni	48	10		
Pb	100	80		
Zn	100	70		
Co	96	60		
Analyst: L.A. Brandvold, New Mexico Bureau of Mines and Mineral Resources, by Atomic Absorption.				
99192, three-sample composite of lineated granodiorite				
99000, three-sample composite of granodiorite of Spring Creek				
*Nockold's (1954) average alkali syenite				
**Nockold's (1954) average granodiorite				

Table 4. Chemical analyses of granodiorites in the Tusas Mountain area.

	99192	99000	99197	99191	90051	99293	99292	99252	99195	99251	99253	98262
q	8.58	26.72			6.13	3.58	0.20			9.56	18.25	
c	0.94	6.89									7.78	
or	18.34	12.64	1.19	1.22	9.05	2.39	2.43	2.40	2.40	3.00	21.52	3.61
ab	44.89	43.08	27.32	5.23	35.37	34.15	24.40	24.05	24.05	13.73	13.69	26.74
an	13.91	1.52	29.79	31.21	28.00	31.63	31.64	35.35	35.35	36.05	20.07	32.13
wo			8.46	5.76	3.69	4.77	8.76	2.27	2.27	3.86		7.91
en	6.23	2.79	15.35	20.39	1.96	6.03	16.42	2.39	2.39	15.54	5.29	7.71
fs	3.10	2.14	10.00	5.33	1.62	9.02	9.69	1.77	1.77	10.34	4.82	4.48
fo			0.34	16.48	7.35			12.15				5.98
fa			0.24	4.75	6.69			9.90				3.83
mt	2.76	1.92	3.66	4.04	3.70	3.37	3.58	5.30	5.30	3.82	4.84	4.00
il	0.76	0.58	1.73	1.57	2.13	1.92	1.96	2.70	2.70	2.31	1.54	1.55

Table 5. Normative compositions of some granodiorites and mafic schists in the Tusas Mountain area.

	BMR*	BMR*	TG*	TG*	BMR**	MS**	MS**	MAQ**
SiO ₂	79.69	76.31	73.16	77.25	75.47	44.44	54.57	68.13
TiO ₂	0.01	0.00	0.00	0.14	0.40	1.84	1.50	0.26
Al ₂ O ₃	11.67	11.75	13.40	11.56	11.42	15.04	16.27	16.81
FeO				0.92	11.42	9.94	6.88	1.56
Fe ₂ O ₃				0.55	2.26	3.19	3.48	1.19
Total Fe	2.45	1.88	0.62					
MgO	0.16	0.00	0.31	0.11	0.30	7.34	2.50	1.28
CaO	0.00	0.02	1.36	0.27	0.55	8.70	6.93	1.88
Na ₂ O	3.04	3.45	3.33	2.65	3.29	1.51	3.54	5.00
K ₂ O	5.62	4.72	4.51	5.59	4.90	0.13	2.11	2.62
MnO				0.03	0.07	0.23	0.16	0.05
P ₂ O ₅				0.06	0.05	0.35	0.61	0.13
CO ₂				0.00	0.00	2.30	0.00	0.00
TOTAL	102.64	98.13	96.69	99.66	99.47	100.26	99.62	99.93
BMR: Burned Mountain metarhyolite								
TG : Tusas granite								
MS : Moppin series								
MAQ: Maquinita granodiorite (lineated granodiorite)								
*Hovorka (1978)								
**Barker (1958)								

Table 6. Published chemical analyses of rocks in the Tusas Mountain area.

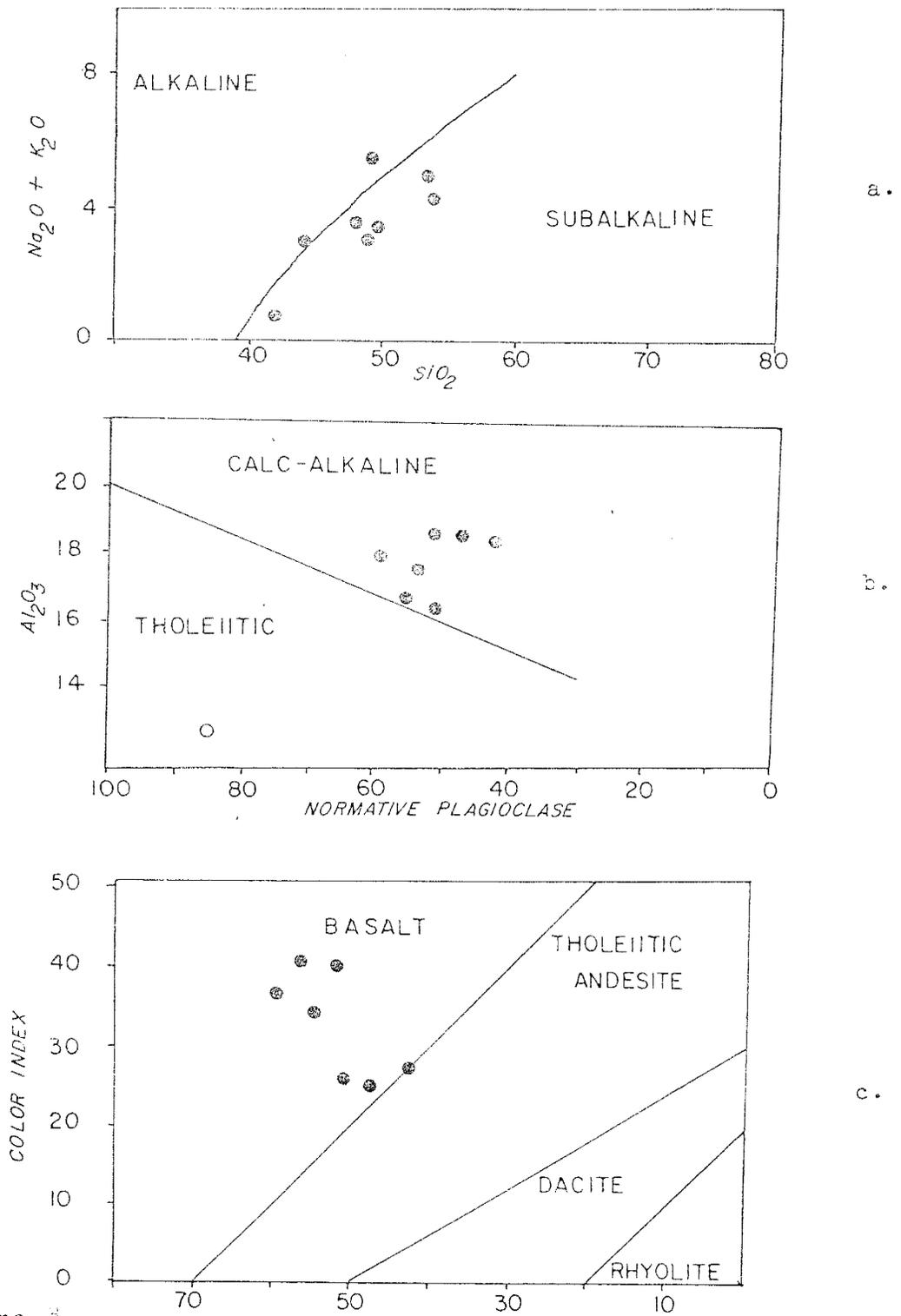


Figure 5. Classification of volcanic rocks from the Tusas Mountain area (after Irvine and Baragar, 1971). a) $\text{Na}_2\text{O} + \text{K}_2\text{O}$ vs. SiO_2 ; b) Al_2O_3 vs. normative plagioclase; c) color index vs. normative plagioclase.

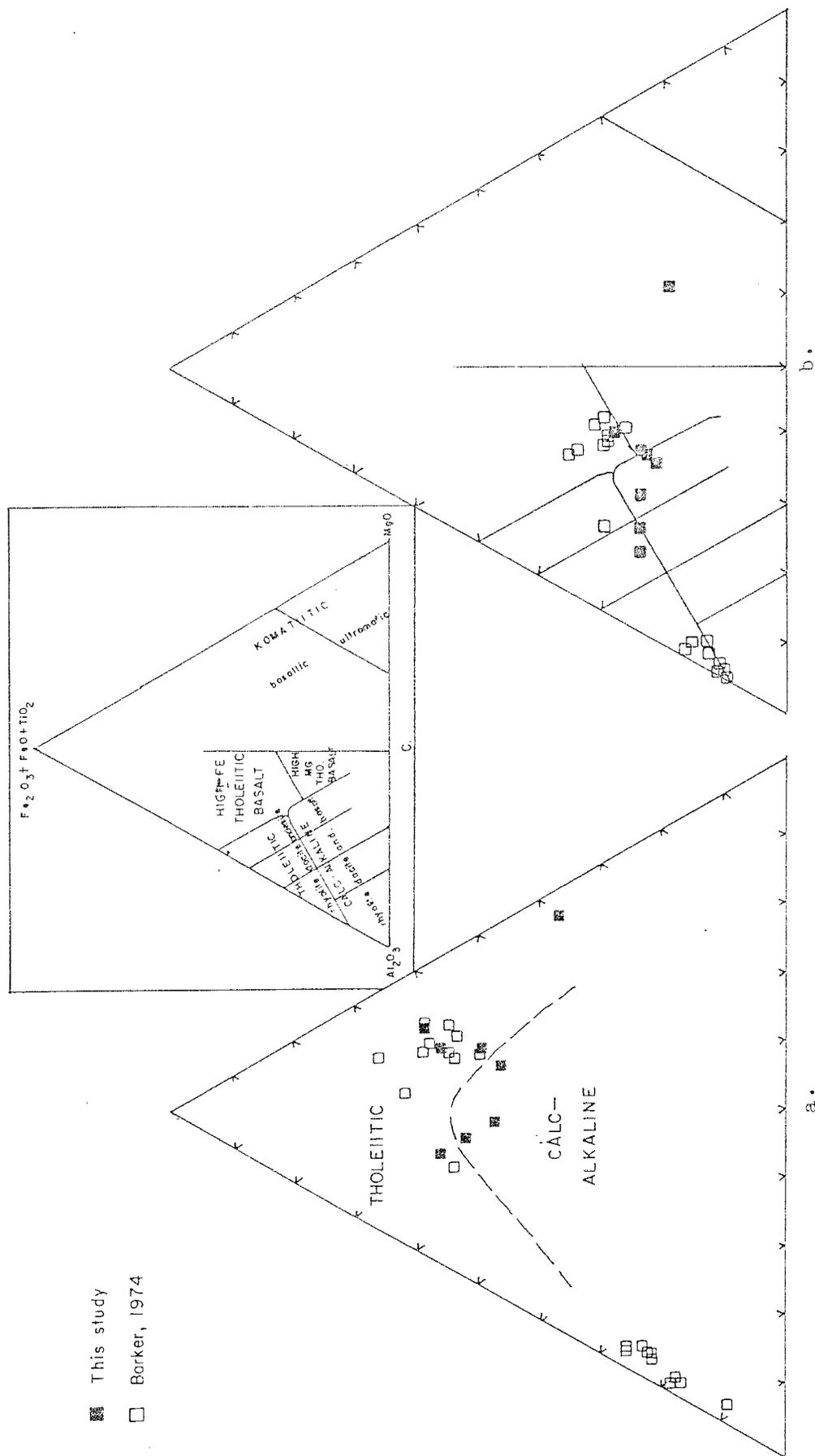


Figure 9. AFM (a) and Jenson cation diagram (b) for volcanic rocks from the Tusas Mountains. Inset (c) labels the subdivisions for the Jenson cation diagram.

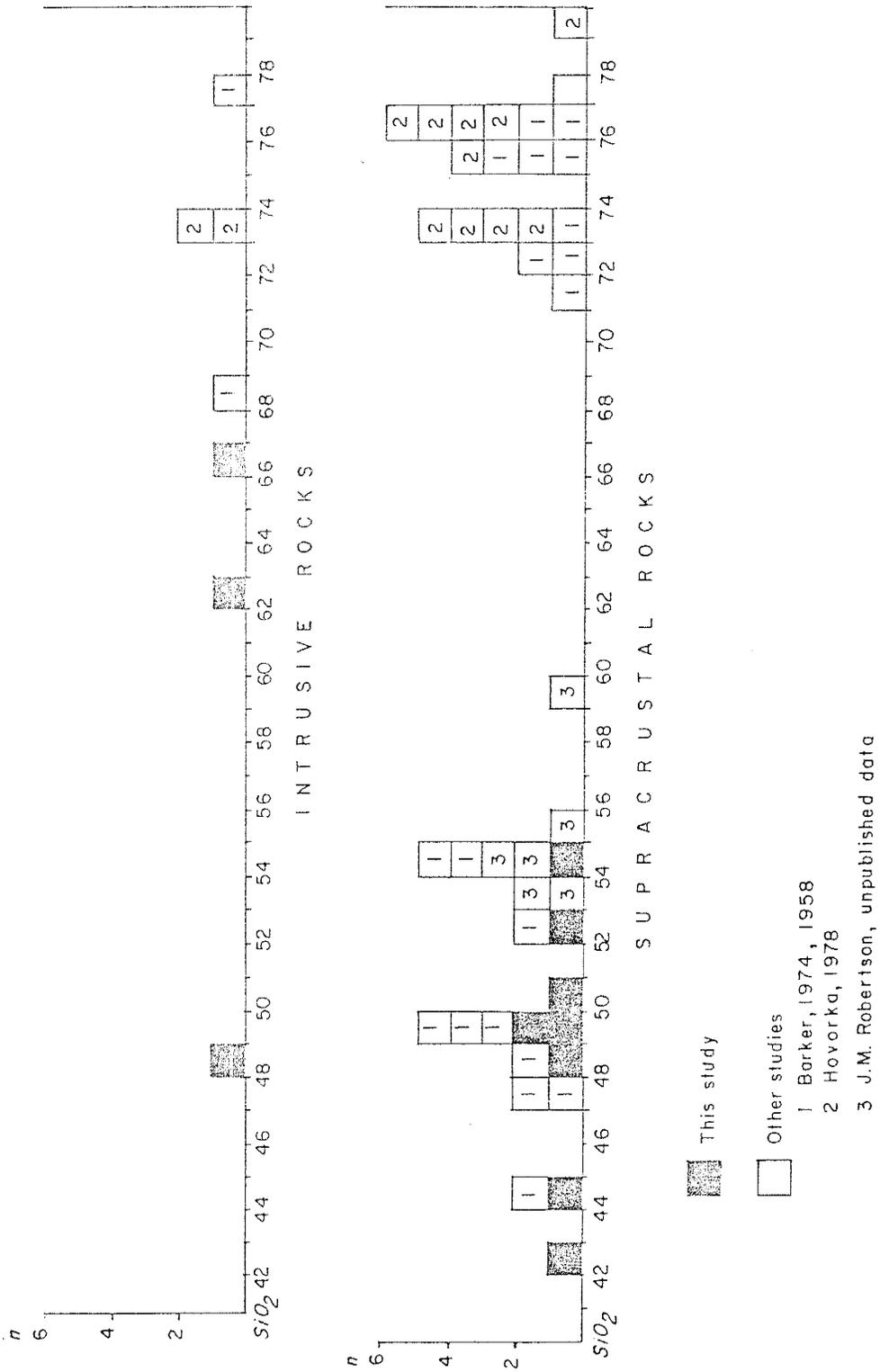


Figure 10. Frequency vs. SiO₂ content histogram for rock from the Tusas Mountains a) for intrusive rocks; b) for extrusive rocks.

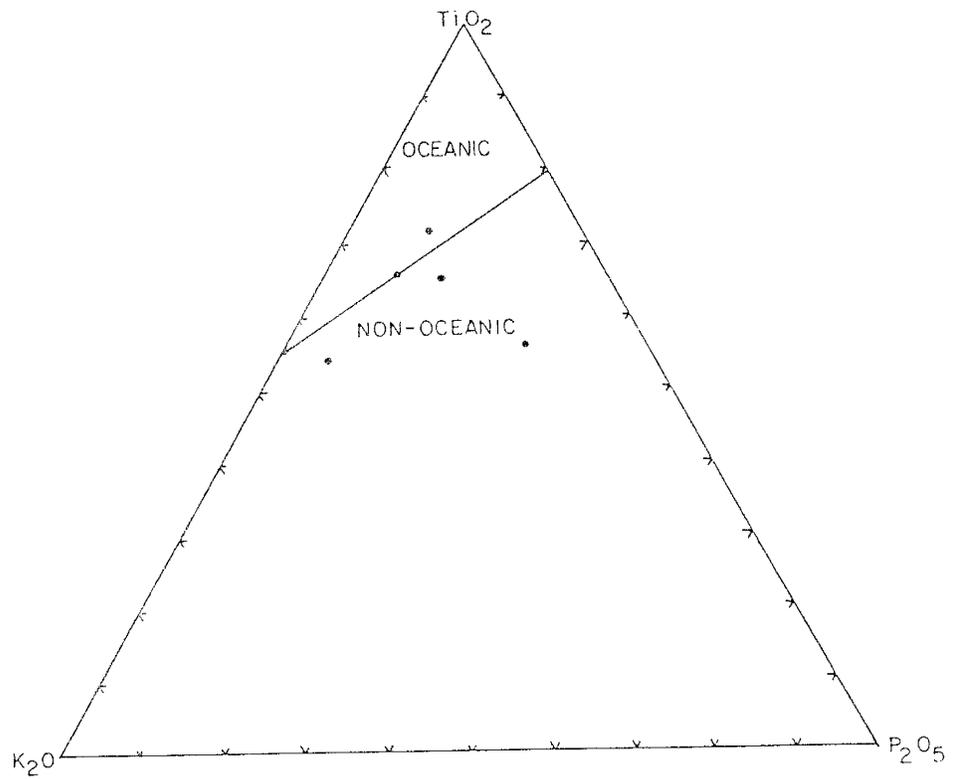


Figure 11. Pearce discrimination diagram for rocks from the Tusas Mountain area.

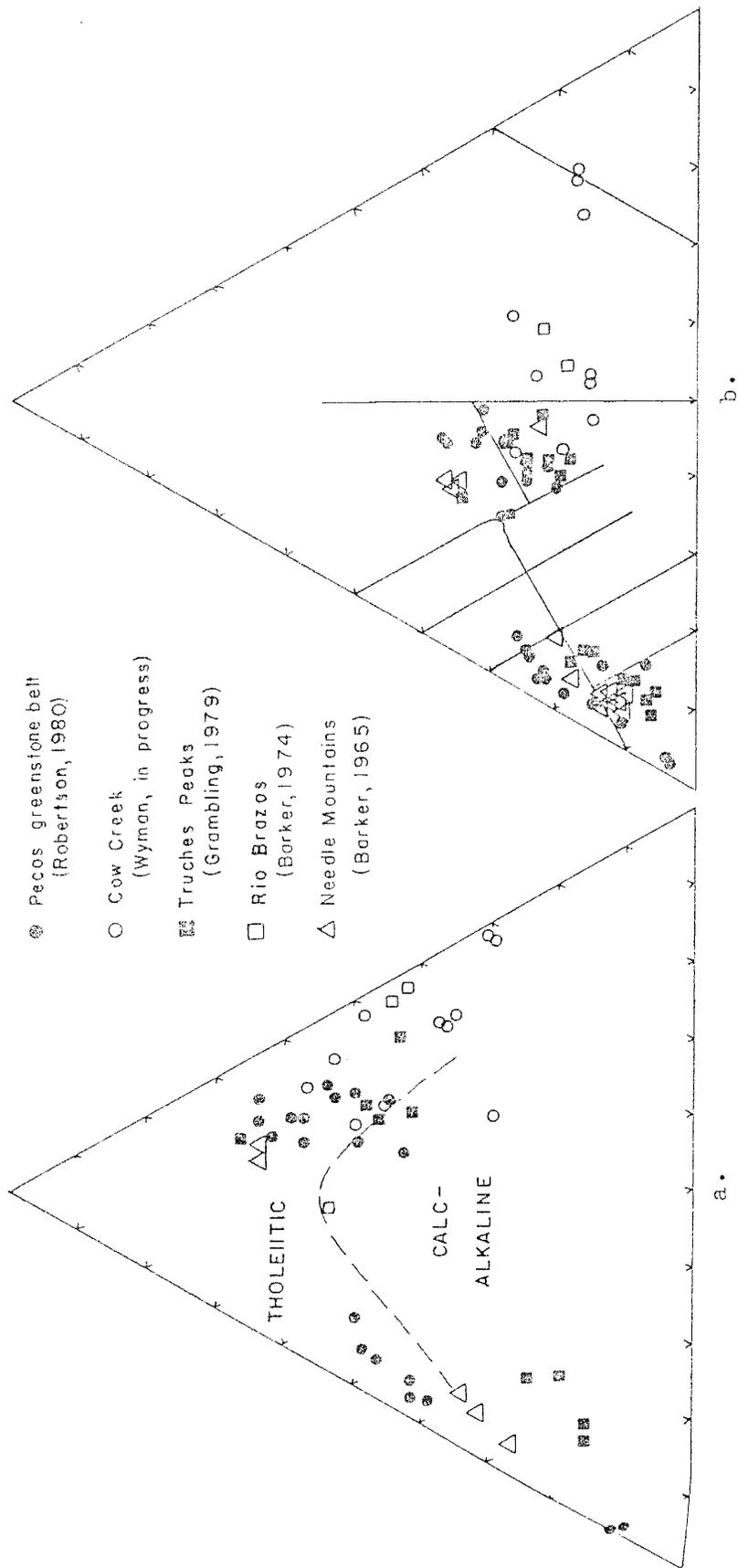
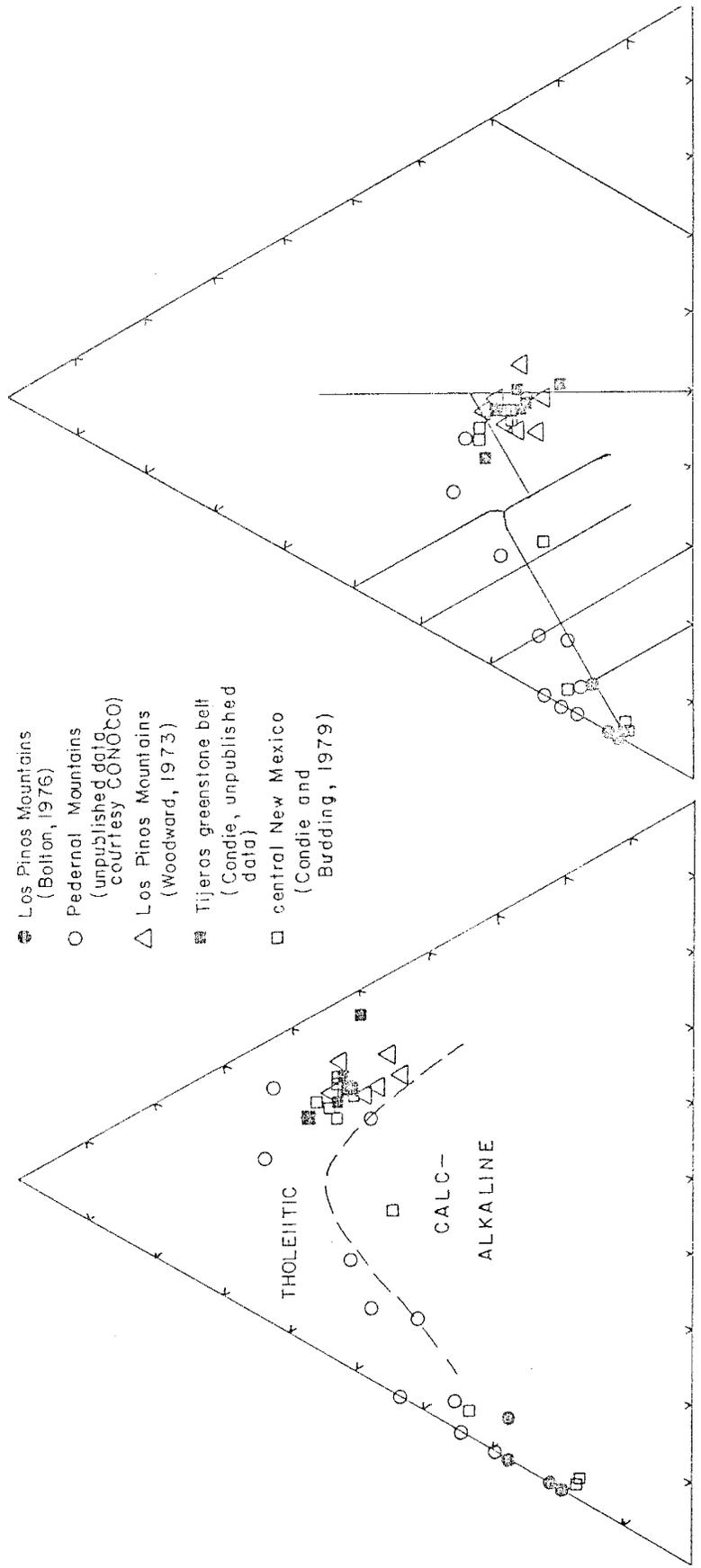


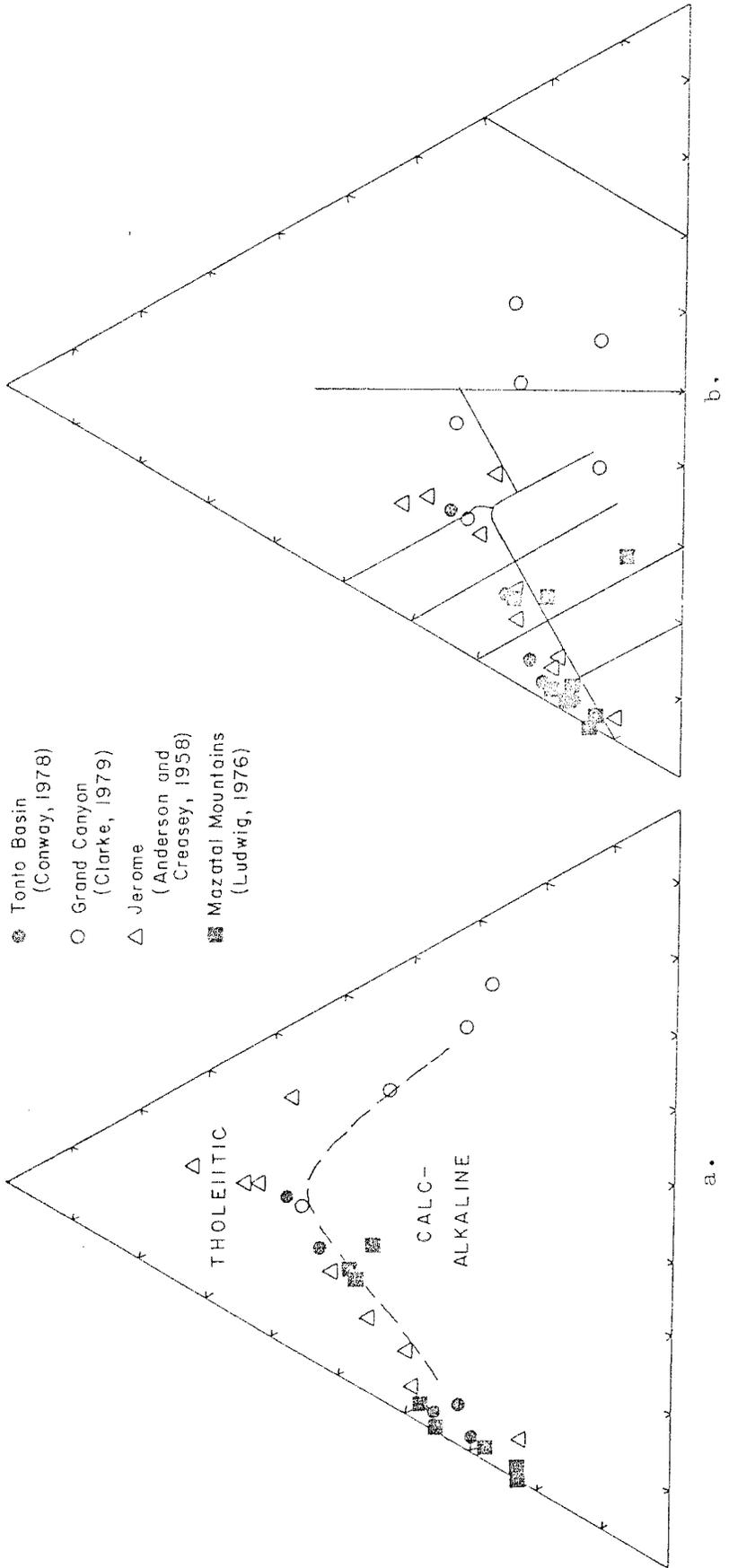
Figure 12. AFM (a) and Jenson cation (b) diagrams for rocks from northern New Mexico-southern Colorado.



a.

b.

Figure 13. AFM (a) and Jenson cation (b) diagrams for volcanic rocks from central New Mexico.



a. b.
 Figure 14. AFM (a) and Jenson cation (b) diagrams for volcanic rocks from Arizona.

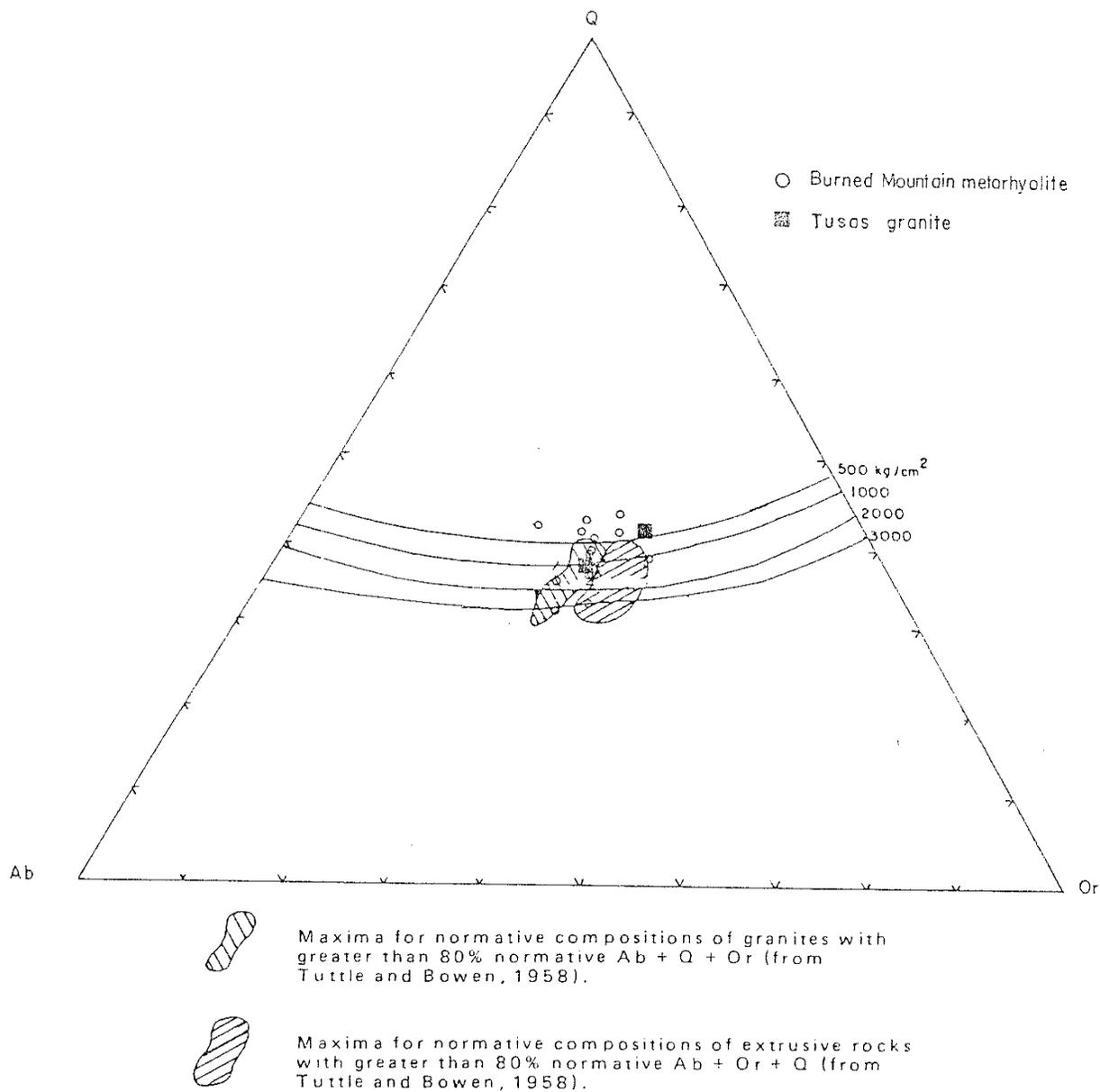


Figure 15. Normative quartz-albite-orthoclase ternary diagram for Burned Mountain metarhyolite and Tusas granite.

Intrusive Rocks

Composite analyses of three samples each from the lineated granodiorite (99192) and the granodiorite of Spring Creek (99000) are given in table 4. These analyses show that the granodiorite of Spring Creek compares closely to the average granodiorite of Nockold's (1954), whereas the lineated granodiorite more closely resembles the average alkali syenite.

Published chemical analyses for the Tusas granite are given in table 6. Both Hovorka (1978) and Maxon (1976b) suggest that the Tusas granite is either the source for or is a remobilized equivalent of the Burned Mountain metarhyolite based on similar chemical composition and the presence of "recrystallized bipyramidal quartz" in the Tusas granite. Published chemical analyses of the Burned Mountain metarhyolite are given in table 6. Although several varieties of the rhyolite can be distinguished in the field based on textural criteria, all these samples, as well as the Tusas granite are chemically indistinguishable on the basis of major elements. All of the samples plot within the rhyodacite field of a normative alkali feldspar-plagioclase-quartz ternary diagram (Hovorka, 1978). However any direct genetic relationship between the Burned Mountain metarhyolite(s) and the Tusas granite is unlikely because: 1) Relative and absolute age relationships. The Burned Mountain metarhyolite was determined to have a zircon

age between 1715 and 1765 m.y. (Barker and Friedman, 1974) whereas Maxon (1976a,b) reported a zircon age for the Tres Piedras granite as 1625 m.y. Field relations show that the extrusion of Burned Mountain metarhyolite was followed by an extensive period of arkosic and quartzite deposition, an episode of widespread deformation, syn-tectonic intrusions of granodiorite and, finally, late- to post-tectonic intrusion of granites. The Tusas granite is clearly not time equivalent to the Burned Mountain metarhyolite, and it is difficult to envision a single magma chamber persisting, unchanged, for over 100 m.y. between Burned Mountain time and the intrusion of the Tusas granite. 2) Metamorphic effects. Any metamorphic event strong enough to remobilize a rhyolitic tuff should produce high-grade metamorphism in the adjacent mafic schists of the Moppin series. The lack of high-grade metamorphism in the area adjacent to the Tusas granite argues against the possibility that the Tusas granite is a remobilized flow or tuff. 3) Chemical composition. Both the Burned Mountain metarhyolite and the Tusas granite analyses plot near the minimum melting composition of granites in a normative quartz-albite-orthoclase ternary plot (figure 15). A variety of source rocks could yield similar compositions upon initial melting which weakens attempts to correlate the two units based on major element chemistry.

Discussion

Tectonic setting

constraints

At the present time, discussion of tectonic models for any or all of the Proterozoic terrane in the southwestern United States must be done with the realization of several limitations. First of all, correlation between isolated blocks of Precambrian can only be conjecture at best. Secondly, much of the Precambrian has only been studied on a regional scale, very few detailed studies have been completed. Deformation, metamorphism, geochemistry, and relative (as well as absolute) age relationships are known only in a general sense; much more detailed work is needed before there can be a thorough understanding of the tectonic environment. Thirdly, the applicability of Phanerozoic-type plate tectonic processes to the Precambrian is disputed by many workers although Phanerozoic-type tectonic processes have been documented in some Proterozoic terranes. Despite the fragmentary Precambrian record preserved in northern New Mexico-southern Colorado and the incomplete study of these rocks, it does appear that many of the exposures have similar rock assemblages, plutonic and tectonic dates, and deformational styles. Any model which might be proposed for middle Proterozoic rocks of the northern New Mexico-southern

Colorado region must take into consideration certain constraints. These constraints include the following:

1) Presence of subalkaline volcanic rocks.

Published chemical data for rocks of middle Proterozoic terranes in the southwestern United States reveals that the volcanic rocks are primarily subalkaline. Alkaline rocks, for the most part, have not been reported.

2) Bimodal volcanism. Andesitic rocks are

noticeably absent in northern New Mexico-southern Colorado (figure 12) and to a certain extent throughout all of New Mexico (figure 13). Volcanic rocks are predominantly tholeiitic basalts and calc-alkaline rhyolites (in northern New Mexico) or tholeiitic rhyolites (in southern and central New Mexico). Dacites are reported in the Pecos greenstone belt and probably occur elsewhere in the Sangre de Cristo Mountains (Robertson, 1980, personal commun.).

3) Isotopic ratios from plutonic rocks. The

plutonic rocks of central and southern New Mexico typically have low Sr $^{87}/^{86}$ initial ratios (between .701 and .703, Condie and Budding, 1979). If these intrusives were produced by the partial melting of lower crust then such low initial ratios demand that the crustal source was not in existence for more than 200 m.y. before melting began. Initial ratios of the post-tectonic plutons in northern New Mexico, however, have high initial ratios often exceeding .710 (Fullagar and

Shiver, 1973; Maxon, 1976a,b) suggesting crustal sources (at least in part) for these rocks. Rb-Sr data for syn-tectonic intrusives in northern New Mexico have not been reported in the literature.

4) Quartzite source. Both a source as well as a source area for the thick sequences (up to several 1000 meters) of quartzites, present in all of the northern New Mexico Proterozoic terranes and to a lesser extent in central New Mexico, has not been recognized. Possible methods for producing essentially pure quartzites have been discussed by Donaldson and Ojakangas (1977) who list the following sources: recrystallized cherts, erosion of vein quartz, quartz phenocrysts from felsic volcanics, and granitic rocks. Although these beds may represent recrystallized chert beds, the common occurrence of crossbedding, pebble horizons, and graded bedding strongly suggests that these rocks represent clastic sediments rather than chemical precipitates. The erosion of quartz veins and pegmatites may be another source for quartz sands; however the relatively small volume of vein material typically present in most igneous terranes makes this alternative unlikely. Reworking of felsic tuffs commonly concentrates quartz phenocrysts originally present in a tuff and could possibly produce essentially pure quartz sands after several cycles of reworking. The problem with this alternative is that quartz phenocrysts generally make up less than 20 or 25 percent of felsic volcanic rocks. The

volume of felsic volcanics needed to produce the volume of quartzite present in northern New Mexico would be extremely large. The weathering of granitic rocks could possibly produce the amount of quartz present in the quartzite beds; however a problem remains in that large amounts of feldspars and micas need to be removed from the source. This can be done by intense weathering, recycling, or local reworking in either tidal or beach environments (or alternatively by aeolian processes).

Detrital zircons separated from the quartzite in the Picuris Mountains form a chord which intercepts the concordia at 1795 m.y. (Maxon, 1976a,b). This is an important constraint in that the source area was not the Archean craton; in fact, the source is very close in age to that of the volcanic successions in northern New Mexico. Barrett and Kirschner (1979) suggest that the quartzites from the Rinconada Formation in the Picuris Mountains were derived from the north, thin to the south, and accumulated in a trough that paralleled a paleo-coast.

5) Relationship between the quartzites and the volcanics. A tectonic model for northern New Mexico must be able to account for the stratigraphic and/or structural juxtaposition of apparently thick sequences of stable continental shelf-type sediments with mafic and felsic volcanic rocks in all the stratigraphic sections between the Needle Mountains of southern Colorado and the Pecos

greenstone belt in northern New Mexico. The relationships between these two apparently distinct depositional environments and tectonic settings must be carefully evaluated and explained.

6) Age belts. Syntheses of published radiometric data for Proterozoic terranes in the United States (Van Schmus and Bickford, in press; Condie, 1980, personal commun.) suggest that a) Proterozoic orogenic age belts can be defined, b) some belts can be correlated between the southwestern United States and the mid-continent region; and c) the belts progressively young outward (south and southeastward) from the Archean craton. Two of the belts trend northeastward across northern New Mexico: a 1690 to 1780 m.y. belt and a 1610 to 1680 m.y. belt (figure 16). The rock assemblages in both belts are similar: quartzites and shales, bimodal volcanic rocks, syn-tectonic intrusive rocks of intermediate compositions, and late- to post-tectonic granites. The boundary between the two belts is not well defined in New Mexico. It may be gradational, or just appear to be gradational due to lack of published zircon dates. In addition, the younging southward trend is only documented for plutonic rocks; few volcanic rocks have been dated. Whether the volcanic rocks show the same younging trend as the plutonic rocks must be resolved by further radiometric studies.



Figure 16. Map of the two major age belts in the southwestern United States (from VanSchmus and Bickford, in press)

7) Basement to the supracrustal succession.

Archean rocks have not been recognized in New Mexico. This presents a problem as to what the supracrustal succession was deposited upon: oceanic or continental crust?

8) Precambrian deformation. Multiple episodes of

deformation, typically including overturned isoclinal folding, are recorded in the Tusas Mountains (Barker, 1958; Bingler, 1965; this study), the Taos region (Condie, 1979), the Truchas Peaks area (Grambling, 1979), the Picuris (Nielson and Scott, 1979), and at least one episode of isoclinal folding in the Pecos greenstone belt (Reismeyer, 1978; Robertson and Moench, 1979).

9) Precambrian metamorphism. Metamorphic grade in

the northern New Mexico terranes range between greenschist facies and middle amphibolite facies. In the Tusas Mountains, staurolite-kyanite assemblages lie south of chlorite-oligoclase and chlorite-albite assemblages. In the Truchas Peaks area, geothermal gradients are steeper to the south (Grambling, 1979), as they are also suggested to be in the Picuris (Grambling, 1979).

10) Presence of trondjemites. Exposures of

trondjemitic intrusives are found north of the Rio Brazos in New Mexico and in southern Colorado (Barker and others, 1976). In Phanerozoic tectonic settings, trondjemites are found in ophiolite suites and along continental plate margins.

11) 170 m.y. orogenic cycle. The quartzite source apparently has an approximate age of 1795 m.y. and the post-tectonic Tusas granite has a U-Pb date of 1625 m.y. During this 170 m.y. time span occurred the erosion and deposition of thick quartzite sediments, mafic volcanism, uplift and erosion, felsic volcanism, a second interval of quartzite deposition, multiple episodes of isoclinal folding, syntectonic intrusions, and post-tectonic granite intrusions.

possible tectonic setting(s)

In speculating on the possible tectonic setting for the Tusas Mountains and the rest of the Proterozoic in southern Colorado-northern New Mexico, it appears best to consider first the supracrustal rock association. In the Tusas Mountains, the volcanic rocks appear to be bimodal (basalts and rhyolites), mainly tholeiitic but in part calc-alkaline, and the basalts apparently older than the rhyolites. Arkoses, subarkoses, pelitic schists, and quartzites are the major sedimentary rock types although mafic volcanoclastic sedimentary rocks are interbedded with the mafic volcanic rocks. In all of the Proterozoic exposures of northern New Mexico-southern Colorado, similar supracrustal rock associations are present; sedimentary rock terranes (consisting in large part of arkoses and quartzites) are consistently associated with bimodal (apparently tholeiitic) greenstone terranes. In some places the two

terrane stratigraphically overlies (and/or underlies) one another (the Tusas Mountains, Truchas Peaks, Taos area), in others they are in fault contact (Tusas Mountains), and in the Pecos region they may interfinger with each other (Robertson and Moench, 1979).

An extensional environment (a continental rift, an aulacogen, or a back-arc basin) is the most satisfactory environment for a supracrustal succession of bimodal volcanics and arkosic-quartzitic sedimentary rocks to accumulate. There is, however, a notable absence of alkaline-peralkaline igneous activity and extensional faulting. Such a setting, also does not satisfactorily account for any of the activity which post-dates the deposition of the supracrustal rocks such as complex deformation (often including overturned isoclinal folds), high-pressure (kyanite grade) metamorphism, and the intrusion of trondjhemitic bodies. In addition, because there is very little evidence for an older continental crust in northern New Mexico and southern Colorado, the question arises as to what was being rifted as the supracrustal succession was being deposited.

Although a compressional environment (a convergent plate boundary?) is the most satisfactory setting for producing the events post-dating the deposition of the supracrustal sequence (the deformation, metamorphism, etc.), it is a completely unsatisfactory environment for generating

the supracrustal succession itself: convergent plate boundaries produce calc-alkaline volcanic suites (especially with abundant andesites) and graywackes, not bimodal volcanics and arkoses-quartzites. Convergent plate boundaries are also typified by ophiolite suites and thrust faulting, neither of which have been recognized in northern New Mexico (that is not to say they are not present). This suggests that the tectonic setting was either complex and involved more than one process or perhaps a tectonic process, unique to the Precambrian, was in operation. Perhaps a model in which extensional and compressional events alternate would be more satisfactory in accounting for the data presently available in northern New Mexico-southern Colorado. It may be that this region occupied a back-arc basin setting. Continental sediments and bimodal volcanics could be juxtaposed in such a setting and compressional events could follow the deposition of the supracrustal rocks. Although this may be the best tectonic setting for the northern New Mexico-southern Colorado exposures, there does not appear to be any nearby island-arc assemblages (including andesites and graywackes) to accompany the possible back-arc basin assemblages.

Table 7. Summary of Precambrian Geologic History of the Tusas Mountains

A Precambrian geologic history for the Tusas Mountains is briefly summarized below. Reference to figure 17 illustrates the time relationships between the events as provided by published radiometric data and field evidence.

1) Deposition of clastic (and minor chemical) sediments accompanied by sporadic mafic volcanism; uplift and erosion of this sedimentary-volcanic succession. There is no maximum age limit for the deposition of these rocks but a minimum age is provided by a date for the Burned Mountain metarhyolite which overlies the mafic sedimentary-volcanic pile.

2) Deposition of quartzite conglomerates and extrusion of felsic volcanics (a). Field evidence establishes that deposition of the quartzite conglomerates and felsic volcanism was essentially contemporaneous.

3) Deposition of several 1000 feet of arkoses and mature sandstones. A maximum age for these sediments is provided by the age of the felsic volcanics but the duration of sedimentation cannot be determined by either radiometric dating or field evidence.

4) Intrusion of early or syn-tectonic granodioritic plutons (a).

5) Deformation and metamorphism. An approximate maximum age limit for deformation and metamorphism (possibly involving more than one episode) is provided by the age of the syn-tectonic granodiorite and a lower age limit by the intrusion of late- to post-tectonic granites.

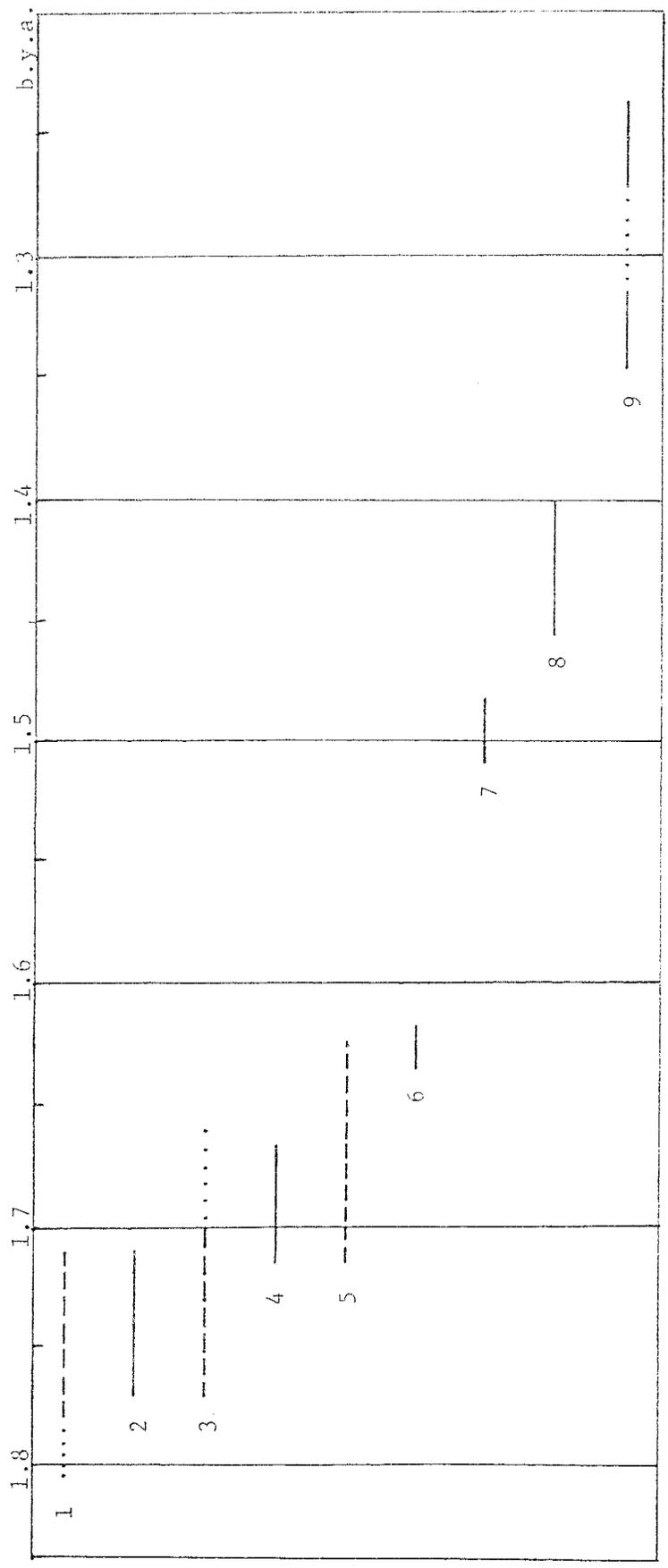
6) Intrusion of late- to post-tectonic granites (b).

7) Post-tectonic thermal event (b).

8) Pegmatite emplacement and metasomatism (c,d).

9) Late thermal event (c,d) (+/- local deformation).

 (a) zircon dating by Silver (in Barker and Friedman, 1974)
 (b) zircon, Rb-Sr dating by Maxon, 1976 (c) Rb-Sr, K-Ar dating by Gresens, 1975 (d) Rb-Sr, K-Ar dating by Long, 1972



- determined by published radiometric dates
- - - - limited by published radiometric dates and field evidence
- age limits unknown

Figure 17. Compilation of published radiometric dates from northern New Mexico.

Summary and conclusions

The oldest group of rocks in the Tusas Mountain area is the Moppin series. The Moppin series is primarily a succession of volcanoclastic sedimentary rocks (chlorite-quartz-biotite +/- plagioclase and hornblende schists), calc-alkaline basaltic volcanics and mafic pyroclastic rocks (hornblende-oligoclase +/- chlorite schists), and minor magnetite-quartz and hematite-quartz ironstones interpreted as chemical precipitates. A minimum age for this succession is defined by the age of the Burned Mountain metarhyolite (between 1715 and 1765 m.y.). Overlying the Moppin series with angular unconformity are basal quartzitic conglomerates, felsic volcanic and volcanoclastic rocks, and a thick sequence of arkosic sedimentary rocks. Also present in the Tusas Mountain area is a thick sequence of mature quartzites. The relationship between these quartzites and the other supracrustals is unclear. They have been interpreted both as older and younger than the Moppin and the controversy remains unresolved at the present.

The unconformity between the felsic volcanic rocks and the Moppin series marks a fundamental change in the sedimentary source area(s) and the depositional environment. The Moppin series appears to be derived mainly from a mafic source area and to have been deposited in a low energy

environment. This interpretation is supported by the lack of cross-bedding and the presence of chemical precipitates. The source area for the Moppin appears to be to the east. This is suggested by a greater percent of volcanic rocks in the eastern portion of the Moppin than in the western portion, whereas the volume of sedimentary rocks appears to increase westward. The arkosic sandstones, on the other hand, were deposited in a high energy depositional environment as suggested by the presence of conglomerates and cross-bedding. The arkosic sediments are derived partly from a felsic volcanic terrane (supported by the presence of abundant rounded quartz and feldspar phenocrysts and felsic volcanic clasts) and probably, in part, from continental crust. These rocks appear to be derived from a westerly source (T. Gibson, 1979, personal commun.).

Following the deposition of the quartzites, at least one episode of folding deformed the rocks of the Tusas Mountain area. This episode produced tight to isoclinal folds with northwest-trending axes. Metamorphism at this time reached a maximum of lower middle-grade conditions (staurolite-garnet-muscovite-biotite-kyanite assemblage). Accompanying the deformational period was the intrusion of several "granodioritic" plutons. The age of deformation is constrained by radiometric dates between 1715 m.y. (the age of a syntectonic intrusive) and 1625 m.y. (the age of a post-tectonic granite). Near the end of the regional

deformation late-tectonic granites (the Tres Piedras granite) as well as a post-tectonic granite (the Tusas pluton) were intruded.

A northeast-trending belt crosses central Arizona, northern New Mexico, and southern Colorado which contains similar supracrustal and plutonic rock assemblages and orogenic dates. A parallel belt immediately to the southeast contains similar rock assemblages but slightly younger orogenic and plutonic dates. The process(es) responsible for the volcanism, deformation, and plutonism are not understood at this time. Much more work is required to fully understand the significance of the Tusas terrane within the larger framework of Proterozoic geologic history in the southwestern United States.

APPENDIX I. Rock descriptions.

Moppin Series

The Moppin series crops out in a northwest-trending belt between Tertiary cover on the east to as far as the southeastern flanks of Jawbone Mountain (northwest of the study area). A lineated granodiorite concordantly intrudes the central portion of the Moppin series and the Tusas granite discordantly (in part) intrudes the lower exposed portion. The base of the Moppin series is not exposed; interbedded felsic schists and arkosic sedimentary rocks unconformably overlie the Moppin series.

The rocks of the Moppin series generally crop out poorly except along valley walls. The dense vegetative cover in this area makes tracing even distinctive units very far along strike impossible and viewing contacts between successive units is equally difficult. A further complication in tracing units is an increase in the grade of metamorphism from northwest to southeast. The western portion of the study area is dominated by chlorite +/- biotite +/- actinolite schists whereas the eastern portion is composed mainly of biotite- and hornblende-bearing schists. Metamorphism has removed many of the original textures of the Moppin series and these rocks are instead dominated by metamorphic fabrics.

Chlorite-amphibole schists

sedimentary rocks The main constituents of the chlorite-bearing volcanoclastic rocks are quartz, plagioclase, and chlorite with variable amounts of carbonate, biotite, actinolite, and muscovite. Magnetite and pyrite are fairly common. Weathered out rhombohedral holes filled with brown staining (after siderite?), generally less than 2 mm in length, is another feature common to many of the greenschists. In thin section chlorite generally comprises 20 to 45 percent of the rock, plagioclase 25 to 40 percent, and quartz 10 to 30 percent. The chlorite grains are usually parallel whereas large (generally less than 2.0 mm in length) plagioclase grains (if present) are typically randomly oriented. The plagioclase occurs as anhedral laths or fragments, often highly sericitized or epidotized, which contain well developed twins that often are broken or bent. Layering is commonly defined by chlorite layers (which drape around the large plagioclase grains when present) and lenses and layers of fine-grained intergrown quartz and plagioclase. Carbonate and clinozoisite are usually concentrated and aligned along distinct layers. If biotite is present, it is almost invariably ragged, poikiloblastic, and randomly oriented. Although the biotite may occur in chlorite-rich layers and thus appear to parallel the foliation, the crystallographic orientations are actually random. Some biotites contain pleochroic haloes, suggesting that small zircons or monazite grains may be present.

One distinctive rock type found in many places throughout the Moppin series (especially along the ridge bordering the contact with the Tusas granite, the NE 1/4, SW 1/4, and the NW 1/4, SW 1/4 of sec. 23, T.28N., R.7E., and in various outcrops along the south side of Cow Creek, sec. 21, 29, 28, T.28N., R.8E.) contains abundant chlorite-actinolite pseudomorphs after hornblende (?). This rock type was also described by Barker (1958). The pseudomorph-bearing rocks are restricted in their vertical and lateral extent and are usually confined to a single outcrop. Within an outcrop, pseudomorph-bearing lenses and pods alternate with pseudomorph-absent areas. The oval or hexagonal pseudomorphs are dark green in a lighter green matrix. In thin section this rock is 40 to 45 percent chlorite, 15 to 20 percent actinolite, 15 to 20 percent epidote, and 15 to 20 percent plagioclase with opaques and carbonate as accessory minerals. Shreddy intergrowths of approximately equal amounts of chlorite and actinolite form the pseudomorphs, and epidote typically outlines the hexagonal grains (figure 18). A few pseudomorphs appear to be cut by fractures along which a differently oriented chlorite formed. The pseudomorphs are contained in a matrix of chlorite, actinolite, epidote, and plagioclase. The chlorite in the matrix is randomly oriented, the epidote forms either granules or euhedral crystals, and the plagioclase occurs as irregular masses and grains which have sutured and indistinct boundaries and are

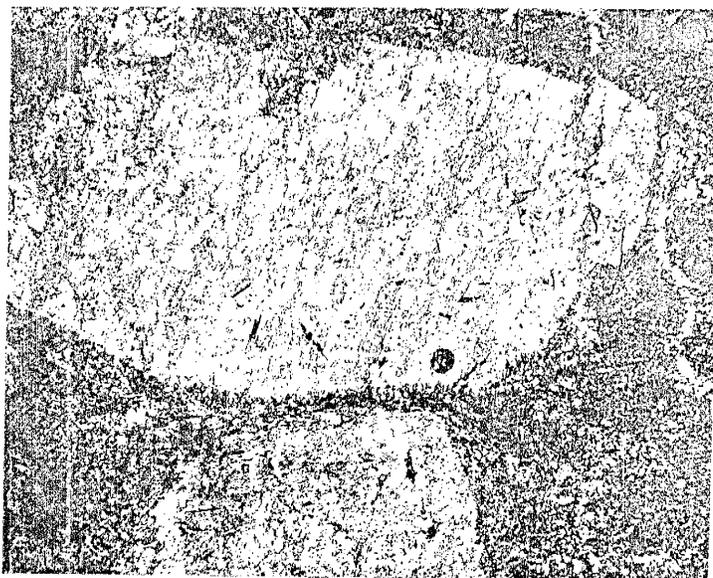


Figure 18. Photomicrograph of chlorite-actinolite pseudomorph after hornblende. The lighter portions within the pseudomorph are actinolite whereas the darker areas are chlorite.

infrequently twinned. It is difficult to interpret the original hornblende crystals as phenocrysts in an igneous rock because they occur in unconnected pods and lenses within single outcrops rather than being uniformly distributed. This suggests they were porphyroblasts formed by an earlier metamorphic episode in a heterogeneous mafic metasediment. Barker also suggested they were original porphyroblasts (Barker, 1958, p. 15).

Thin sections which have biotite as a major constituent contain between 35 to 70 percent biotite, 30 to 40 percent plagioclase and 50 to 60 percent quartz and up to 45 percent epidote with carbonate, opaques, chlorite, and hornblende in minor amounts. The biotite is generally unoriented but may occur along layers in which the biotite flakes continue to be crystallographically unoriented. In those cases the biotite is generally fairly large, ragged, and associated with vestiges of chlorite. More commonly the biotite is small (generally less than 0.6 mm) and randomly oriented. Plagioclase occurs both as large (up to one centimeter in length) anhedral laths and/or fragments or as small (less than 0.5 mm in length) grains in a mosaic texture. The large grains are generally broken and bent, sericitized, and epidotized. Epidote is especially abundant in rocks adjacent to the Tusas granite. Poikiloblastic garnets (0.08 to 1.50 mm in diameter) as well as hornblende are also found in trace amounts in outcrops adjacent to the

Tusas granite. The hornblende laths are 0.3 to 0.8 mm in length and diamond-shaped cross-sections are 0.15 to 0.6 mm in longest direction. They are generally euhedral (except for poorly terminated ends on the longitudinal sections) and sharply cut the biotite. They are distinctly pleochroic: indigo blue, green, and yellowish green.

igneous rocks Unfoliated rocks are either 1) porphyritic with plagioclase phenocrysts or 2) fine-grained green and white, or black and white, speckled rocks. Plagioclase phenocrysts (up to 15 mm in length) in the porphyries are either lineated parallel to the dip direction of the foliation or are randomly oriented. Porphyritic rocks typically become increasingly well-foliated along strike while the phenocrysts become increasingly smaller or broken and lineated biotite-chlorite knots form parallel to the dip direction. Eventually the porphyritic units grade into well-foliated schists with dark-green lineations and small fragments of feldspar.

In thin sections of the phenocryst-bearing schists, the plagioclase phenocrysts (oligoclase in composition) are twinned, highly resorbed along the grain boundaries, and extensively altered to sericite and/or epidote. The groundmass is a very fine-grained intergrowth of randomly oriented biotite and feldspar with clinozoisite as an accessory mineral. In thin sections which also contain hornblende in the mineral assemblage, the hornblende is small

(less than 0.4 mm in length) and defines a crude foliation. Biotite-hornblende aggregates in roughly lenticular shapes (up to 1 cm in length) were noted in some sections.

In thin sections of unfoliated green and white speckled rocks, amphibole comprises approximately 40 percent of the rock, plagioclase 35 percent, epidote 15 percent, and quartz 10 percent. The plagioclase occurs as relict phenocrysts (which are extensively altered to epidote and contain relict twins) or are fine grained and intergrown with quartz in the rock matrix. The amphibole is predominantly actinolite with small cores of hornblende.

The westernmost appearance of hornblende and garnet is within 60 meters south of the 9730 foot road intersection in sec. 14, T.28N., R.7E. (just north of this study area). Thin sections from these rocks contain minor amounts of poikiloblastic, pleochroic (blue green, grass green, yellowish green), euhedral, less than 2.0 mm in length, hornblende and 0.6 to 1.5 mm euhedral, poikiloblastic garnets. The remainder of the rock is a fine-grained assemblage of biotite, quartz, plagioclase (with a few relict rounded phenocrysts) and magnetite and clinozoisite as accessory minerals.

Amphibole-bearing rocks, concentrated mainly between Cleveland Gulch and the eastern edge of the Precambrian exposures, are intermixed with biotite-bearing rocks. The rocks along Cow Creek are very poorly exposed on

the south side of the valley but outcrops of rocks with chlorite pseudomorphs after hornblende (like those described above), amphibolites, hematite-quartz ironstones (described in the section on iron formation), and other mafic schists can be found. On the north side of Cow Creek are coarse-grained gneisses, fine-grained phyllitic schists, and fine-grained schists with thin (1 to 2 mm thick) pink stringers of felsic material. The ridge directly south of the Tusas granite in sec. 19 and the western portion of sec. 20, T.28N., R.8E. and westward to the ridge south of the Tusas granite in sec. 24, T.28N., R.7E., contains gneissic to almost migmatitic amphibolites. Light and dark bands, usually less than 30 cm in width, alternate. The gneissic outcrops are generally less than 15 meters wide and are intermixed or interbedded with fine-grained chlorite schists and hematite-quartz ironstones.

Hornblendites

Hornblendites form several mappable horizons in the eastern portion of the map area. They form blocky outcrops which are generally less than 3 meters wide and are poorly foliated. Weathered surfaces have a dark green sheen with radiating crystals of hornblende (up to 1 cm in length). Freshly broken surfaces have a characteristic surface which is the result of closely intergrown, randomly oriented hornblende needles. Typically they are interbedded with

amphibolites but in places they occur with plagioclase-biotite-quartz schists. In sec. 30, T.28N., R.8E., massive chlorite outcrops appear to possibly be the greenschist equivalents of the hornblendites. However, lack of outcrops prevents tracing the hornblendites directly into the chlorite schists.

In thin section the hornblendites consist of 55 percent chlorite and 45 percent hornblende and opaque accessories. The hornblende forms unoriented, ragged, poikiloblastic crystals which cut a well-foliated chlorite aggregate (figure 19). Quartz inclusions in the hornblende are aligned parallel to the foliation formed by the chlorite. Many of the hornblende crystals are rimmed with iron staining.

Metapelites

A horizon comprised mainly of felsic to intermediate phyllites and schists but with minor mafic schists crops out between sec. 26, T.28N., R.7E. and sec. 28, T.28N., R.8E. The lower (northern) contact is defined by a distinctive feldspar-quartz (+/- muscovite) phyllitic schist. The best exposures of this schist are found on hill 9180 (sec. 29, T.28N., R.8E.) but it can be traced discontinuously along almost the entire length of the phyllitic section. This unit weathers to bold, highly schistose outcrops with a yellowish or reddish weathered surface and a characteristic



Figure 19. Photomicrograph of hornblendite. Randomly oriented hornblende sharply cuts a well-foliated matrix of chlorite.

dull sheen. Ragged biotites and traces of magnetite are randomly distributed. The upper (southern) boundary of the phyllitic section is tentative and is bounded by the last occurrence of phyllites in the east and the muscovite-feldspar-quartz schist (the felsic volcanoclastic rock) in the west. The metapelites appear to pinch out to the west due to an erosional episode (see the section on the felsic schists) and the eastern boundary is extremely tentative as the distinctly phyllitic nature is lost and mafic schists become increasingly important. Two or three hornblendite units (described above) occur within this unit as well as the banded iron formation (described in another section) and other mafic schists.

The metapelites are typically chlorite phyllites, biotite-feldspar-quartz schists and biotite-quartz schists. These schists are well foliated and color banding and lensing are common. Good exposures of these phyllites and schists are on the hill east of hill 9180 (sec. 29, T.28N., R.8E.) and in trenches along the iron formation horizon. Magnetite octahedra are common and infrequently form thin stringers and rectangular outlines. Many of the chlorite phyllites and reddish felsic schists contain abundant white spots which are flattened in the plane of foliation. In thin sections cut parallel to the foliation, the white spots are contained in a fine-grained (less than 0.1 mm grain sizes) matrix consisting almost entirely of quartz with tourmaline, hematite,

magnetite, muscovite, and chlorite in trace amounts. The spots are composed of amorphous material (and tourmaline) which is always rimmed by hematite staining which is usually rimmed in turn by muscovite (figure 20). Because the white spots are found in many different schists, it appears that they were originally metamorphic minerals (cordierite?) which have since been altered to a fine-grained material.

Mafic pyroclastic breccia

A mafic pyroclastic rock crops out in a thin band between sec. 32, T.28N., R.8E. and the very southwestern corner of sec. 29, T.28N., R.8E. This unit forms dark reddish-brown, rounded outcrops which are poorly foliated. It is underlain by a porphyritic amphibolite which contains randomly oriented, euhedral, epidotized plagioclase phenocrysts (generally less than 8.0 mm, but up to 2.0 cm long) and cavities (possible relict vesicles) lined or filled with carbonate. In thin section the porphyritic amphibolite also contains rounded or lenticular (less than 5 mm in length) polygonal quartz aggregates which sometimes contain epidote. Hornblende occurs both as small, subhedral needles and as large poikiloblastic anhedral fragments. Stratigraphically overlying the pyroclastic unit is a medium-grained amphibolite similar to the matrix material of the pyroclastic unit. This amphibolite contains in thin section 35 percent hornblende-actinolite, 35 percent

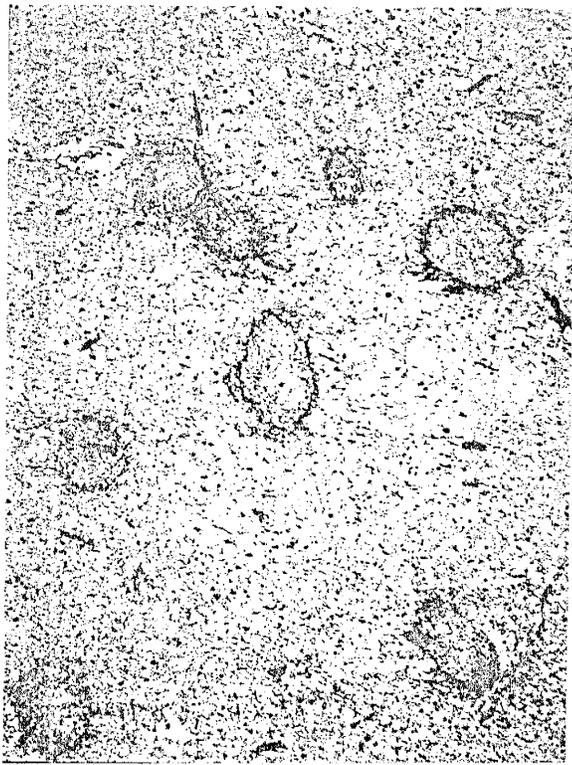


Figure 20. Photomicrograph of white spots in the metapelites (the thin section is cut parallel to the foliation). Each spot is composed of fine-grained material rimmed with hematite staining which is generally rimmed with muscovite.

plagioclase of unknown composition, 15 percent epidote, 15 percent biotite, and a small amount of quartz. The hornblende forms ragged, unoriented fragments usually with actinolite cores. Biotite appears to be both replaced by and replacing hornblende. The plagioclase is highly altered to epidote and overgrown by amphibole and biotite.

The pyroclastic breccia contains poorly-sorted angular, rounded, or irregularly shaped clasts (from very small to 20 cm in length) in a medium-grained amphibolite matrix (figure 21). The clasts are elongated parallel to the dip of the foliation and generally weather in relief. Three rock types comprise the clasts: fine-grained plagioclase-rich amphibolite, fine-grained plagioclase-poor amphibolite, and porphyritic amphibolite similar to the rocks directly underlying the breccia. These fragments predominate in the eastern outcrops whereas the western outcrops also contain quartzitic clasts. Bands of rocks similar to the (stratigraphically) underlying porphyritic amphibolite are also interbedded or intermixed with the pyroclastic unit. On the slope south of hill 9180 (sec. 32, T.28N., R.8E.) fragment-bearing portions of the mafic pyroclastic are crossed at high angles to the foliation by fragment-absent amphibolitic portions (figure 22). The presence of rip-up structures and small-scale channels in this outcrop suggests that the fragment-bearing portions and the fragment-absent portions represent sedimentary layering which is at a high



Figure 21. Photograph of mafic pyroclastic breccia. The clasts are irregularly shaped and contained in a medium-grained amphibolite matrix.



Figure 22. Photograph of bedding preserved in the mafic pyroclastic breccia which is at high angles to the foliation. A possible rip-up structure within the finer-grained portion suggests that stratigraphic up is towards the top of the picture. This is in agreement with a small channel scour also found in this outcrop.

angle to the foliation. These structures also indicate that stratigraphic up is to the southwest. These outcrops suggest that both westward and stratigraphically upward the pyroclastic breccia has been increasingly reworked by water.

In thin section the matrix of the pyroclastic breccia is composed of 30 percent plagioclase, 20 percent hornblende, 15 percent biotite, 15 percent epidote, and 20 percent quartz. The plagioclase and quartz form a fine-grained matrix in which larger unoriented plagioclase phenocrysts, rock fragments, and chert (?) fragments are contained. Hornblende occurs both as small, aligned, subhedral needles and as large unoriented poikiloblastic subhedral crystals. Some crystals have small cores of actinolite. The biotite appears as aligned anhedral flakes which have been replaced in part by amphibole.

Meta-argillite

Well-banded biotite/chlorite-quartz-plagioclase schists form a thin horizon near the top of the Moppin series in sec. 32, T.28N., R.8E. This unit forms subdued, blocky, well-foliated outcrops with prominent green, black, gray, and greenish-gray bands (figure 23). Layers within this unit are generally less than 1 cm and differ mainly by the amount of biotite in each band.

In thin section the mineral assemblages differ between color bands. The major assemblages are 1) biotite +

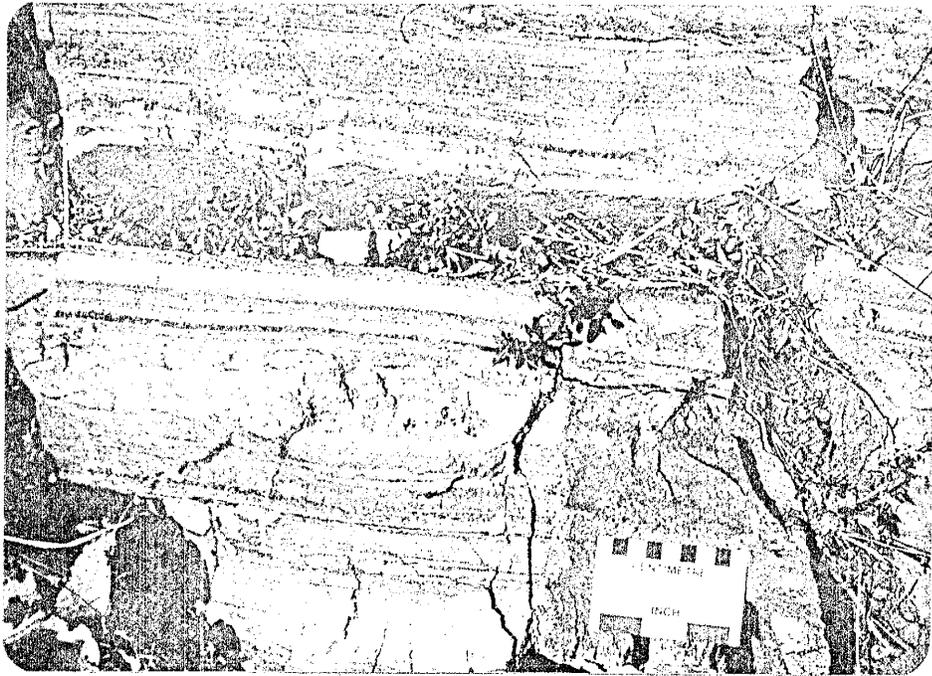


Figure 23. Photograph of the meta-argillite.

muscovite + quartz + feldspar + epidote + chlorite; 2) biotite + quartz + feldspar + chlorite; 3) quartz + feldspar with trace amounts of biotite and muscovite; 4) quartz + biotite + muscovite; 5) biotite + quartz + epidote; 6) biotite + quartz; 7) quartz + biotite + epidote + chlorite. Clinzoisite and magnetite are present in all the layers. The boundaries between each assemblage is sharp and usually defined by a marked change in biotite or quartz grain size. Chlorite is always minor and forms a foliation parallel to the color banding; biotite is present in all bands and sometimes comprises 40 percent of an individual band. It is always poikiloblastic, forms ragged flakes or fragments, and sharply truncates the chlorite. Quartz is the major constituent of all the bands and in some bands appears as very well-sorted, polygonal grains, and in other bands it has irregular grain boundaries and forms a poorly sorted matrix with feldspar. In layers with muscovite, the quartz forms straight-edged, elongated grains whereas in layers without muscovite the quartz grains are more equidimensional.

Felsic Schists

Burned Mountain metarhyolite

Burned Mountain metarhyolite is exposed on the western edge of the study area along Sheep Gulch (sec. 22, T.28N., R.7E.). In this area three horizons of the rhyolite are associated with pebble conglomerates and phyllites of the arkosic sedimentary rock section. A thin interval of the Burned Mountain metarhyolite also crops out within the arkoses in sec. 37, T.28N., R.7E. The Burned Mountain metarhyolite appears to be stratigraphically conformable in all cases with the enclosing arkoses, although contacts are poorly exposed.

The Burned Mountain metarhyolite forms bold blocky outcrops and extensive talus; schistosity is usually poorly developed and samples break with a conchoidal fracture. The rhyolite is typically gray or pink and has brick red or grayish-pink weathered surfaces. On weathered surfaces, quartz phenocrysts stand in relief. All samples of Burned Mountain metarhyolite are characterized by approximately equal numbers of quartz and potassium feldspar (now microcline) phenocrysts (generally less than 4 mm in length) in an aphanitic silicic groundmass which comprises 65 to 80 percent of the rock. Portions of the outcrops in sec. 22 (T.28N., R.7E.) contain abundant parallel light-gray aphanitic stringers which impart a foliation to the rock and

are gradational to portions which lack any megascopic foliation. A relict eutaxitic texture is preserved in some outcrops. Small (less than 2.0 cm in length) black, greenish, and pink lenticular bodies, often with flared ends, are suggestive of preserved flattened pumice fragments. The southernmost horizon of Burned Mountain metarhyolite on the west side of Sheep Gulch is non-foliated and lacks obvious relict pumice, or light-gray stringers. In these outcrops the microcline phenocrysts are much larger, generally 5 to 10 mm and range up to 2.0 cm in length. In sec. 36, T.28N., R.7E. the Burned Mountain metarhyolite appears to have been slightly sheared and chloritized. It is greenish gray with prominent 1 to 2 mm stretched microcline crystals and minor (less than 5 percent), small (1.5 mm in length) lenticular quartz phenocrysts. Thin greenish-gray streaks resemble the light-gray stringers in the Sheep Gulch outcrops.

The quartz phenocrysts are rounded to subhedral, 0.5 to 5.0 mm in diameter, and dark gray or slightly bluish. In thin section the quartz phenocrysts are strained and have sharp grain boundaries, or have resorbed edges with deep embayments of the groundmass, or (very rarely) are recrystallized to polygonal aggregates. Fractures filled with sericite and small bleb-like inclusions of sericite are common in the quartz phenocrysts. The microcline phenocrysts are pink, generally 1 to 4 mm in length, and euhedral to slightly rounded or lenticular. The lenticular microclines

are elongated parallel to the plane of foliation. In thin section microcline always exhibits gridiron twinning and the grain boundaries are highly resorbed and rimmed with opaque dust. Included in the microcline grains are opaques, (rare) carbonate, and sericite-filled fractures. The aphanitic groundmass is composed of fine (less than 0.01 mm) intergrown grains of quartz, microcline, muscovite, and plagioclase, and minor opaque minerals. The muscovite grains have a preferred orientation defining a fabric which in many thin sections drapes around the quartz and microcline phenocrysts.

Abundant, light-gray, parallel stringers are 0.2 to 1.0 mm in width and 0.5 to 2.0 mm apart forming a distinctive fabric in some samples. In thin section the stringers are concentrations of muscovite and quartz or opaques and quartz. The gray stringers may represent flow bands of a lava, flattened pumice of a very densely welded ash-flow tuff, or a metamorphic texture. Black bands, which are prominent in some outcrops, can exceed 30 cm in length and either parallel or the fabric defined by the gray stringers or form curved bands resembling solution fronts. Hovorka (1978) X-rayed this material in an unsuccessful attempt to identify the mineralogy of the bands and could conclude only that the banding was due to concentrations of very finely disseminated opaque material.

Muscovite-feldspar-quartz schist

A muscovite-feldspar-quartz schist forms a thin but continuous horizon over four miles in length between sec. 26, T.28N., R.7E. and sec. 33, T.28N., R.8E. although it is concealed by Tertiary cover in two places (sec. 25, T.28N., R.7E. and sec. 32 and 33, T.28N., R.8E.) and intruded by a small pluton of Tusas granite (sec. 25, T.28N., R.7E.). The muscovite-feldspar-quartz schist forms bold, highly schistose outcrops. The schist is usually white and forms satiny yellow, reddish-brown, or greenish-white weathered surfaces which have prominent quartz phenocrysts in relief.

The muscovite-feldspar-quartz schist is characterized by 5 to 10 percent prominent quartz phenocrysts (1 to 3 mm in length), and 0 to 10 percent microcline phenocrysts (2 to 4 mm in length). Also present are 2 to 4 mm pyrite cubes, 1 mm magnetite octahedra, and 0.5 to 1 mm biotite flakes. Lenticular shapes suggestive of relict pumice fragments (up to 5.5 cm in length) are present in outcrops above the Moppin ranch (sec. 26, T.28N., R.7E.). The outcrops in secs. 33 and 34, T.28N., R.8E. exhibit pronounced, closely spaced slip cleavages at high angles to the foliation.

In thin section the quartz megacrysts are square or rounded and are typically deeply embayed with groundmass material. They commonly contain blebs of sericite and fractures filled with the groundmass, sericite, or quartz and

microcline. The microcline always exhibits gridiron twinning and usually contains inclusions of, and is bordered with, opaques. The groundmass consists of very-fine-grained (less than 0.01 mm) intergrowths of polygonal quartz and feldspar with aligned muscovite and accessory apatite and opaques. The aligned muscovite flakes and short stringers of larger (0.06 mm) polygonal quartz form a prominent foliation which drapes around the phenocrysts of quartz and microcline. The phenocrysts are typically rotated with respect to the foliation.

Sedimentary Rocks

The sedimentary rocks form a northwest-trending belt of continuous exposure along the southern portion of this map area to as far as Eureka Canyon, sec. 1, T.28N., R.6E. (T. Gibson, 1979, personal commun.). They are also exposed further to the northwest in the Brazos Box.

Arkosic sedimentary rocks

arkoses Metamorphosed arkosic sandstones, shales, subarkoses and minor quartzite form the lower 600 meters of the sedimentary rock section. The sedimentary rocks between the NW 1/4, sec. 32, T.28N., R.8E. and sec. 22, T.28N., R.7E. crop out only poorly and are generally feldspar-muscovite-quartz schists. They are predominately

fine- to medium-grained, finely to moderately laminated, poorly sorted arkoses. Rounded quartz, feldspar, and opaques are generally less than 1 mm in size, and are contained in an orangish-yellow, mica-rich matrix; biotite is usually minor or absent. North of the Moppin Ranch the arkoses are green and appear to contain appreciable amounts of chlorite and magnetite. Less common rock types throughout the section are coarsely recrystallized gray quartzite with minor interstitial muscovite, biotite-quartz (+/- feldspar) (sometimes phyllitic) schists in alternating dark and light bands, and greenish-gray phyllitic schists which may contain visible rounded quartz grains. The quartzites and feldspar-muscovite-quartz schists typically contain thin (less than 1 mm) dark laminae of mafic minerals which define bedding and rare crossbedding. Close to the bottom of the sedimentary rock section (eg., at the 9200 foot elevation, sec. 25, T.28N., R.8E.) is a gray, micaceous, spotted, phyllitic schist that contains numerous, rhomb-shaped cavities that appear to have formed by the weathering-out of 2 mm siderite (?) grains.

In thin sections the arkoses range between 40 to 70 percent quartz, 20 to 30 percent mica, and 10 to 20 percent feldspar, with rock fragments, chlorite, epidote-clinozoisite, hematite, magnetite, and other opaques commonly present in minor amounts. The quartz grains vary widely in shape, size, and intergranular boundaries. The

quartz grains range in size from 0.03 to 0.3 mm in phyllites and up to 2 mm in other schists. They form equant, elongated or stretched grains with straight or sutured boundaries. Both strained and unstrained quartz grains occur within individual thin sections. Both plagioclase and microcline are present, commonly in equal amounts. The plagioclase can be either fresh or extensively altered to sericite. The quartz and feldspar grains are angular to subrounded and very rarely display a crude grading. The muscovite consistently defines a distinct foliation which is commonly transected at high angles by biotite flakes if present. The opaque minerals are commonly concentrated in distinct layers suggesting they are probably detrital grains.

The arkoses in sec. 33 and 34, T.28N., R.8E. appear to be stratigraphically continuous with the arkoses in sec. 32, T.28N., R.8E. based on the presence of the felsic volcanoclastic schist (the muscovite-feldspar-quartz schist) at the base of the section. These arkoses form bold outcrops which are generally whitish gray and weather to a dark gray or pinkish gray. They are well bedded, but crossbedding and graded bedding are rare. Bedding cannot be distinguished from schistosity in the southeastern and eastern edges of this exposure; the schistosity strikes in a northerly rather than a westerly direction. Mafic schists parallel the northerly foliation and are similar to mafic schists in the arkoses north of Spring Creek (described below). These mafic

schists crop out only within drainages and are generally poorly exposed. They are medium-grained and have approximately equal proportions of dark and light minerals. In thin section, the schists contain hornblende, quartz, and chlorite with minor amounts of plagioclase, clinozoisite, and opaque minerals. The hornblende is randomly oriented and forms 0.3 to 1.2 mm longitudinal sections with poorly terminated ends and 0.15 to 0.5 mm cross-sections which are generally euhedral. Chlorite occurs in unoriented aggregates which are crossed by hornblende euhedra. Because bedding and schistosity in the arkoses cannot be distinguished, it is difficult to tell whether the mafic schists cross-cut or parallel the bedding. Foliation in the mafic schists follow a similar trend as the amphibolite dikes that intrude the lineated granodiorite suggesting that the northerly trend of the foliations in this area may represent a secondary foliation along which dikes were intruded. On the other hand, the mafic schists are petrographically very comparable to the mafic schists interbedded in other parts of the arkosic section. It is therefore suggested that these mafic schists parallel bedding which has been folded.

pebble conglomerates The pebble conglomerates in the middle of sec. 22 form light-gray, jagged or rounded, highly schistose outcrops whereas the exposures in the southern portion of section 22 are less schistose and form dark-gray blocky outcrops. Gradations between conglomerates and

fine-grained phyllites occur in some of the outcrops. Crossbedding and small-scale channel scours can be identified in many outcrops. All the clasts (except vein quartz) are highly flattened in the plane of foliation (figure 24) and form a prominent lineation (figure 25). The majority of the clasts in the Sheep Gulch exposures are black hematitic chert, vein quartz, and felsic schists with rare clasts of red chert. No clasts of mafic schists were found. Clasts are moderately to poorly sorted and generally range between 1 and 2 cm, but clasts up to 16 cm were found. The matrix is typically purplish gray and composed of fine-grained quartz and muscovite.

In a thin section with clast size averaging approximately 1.5 cm, 90 percent is quartz and quartzitic rock fragments. The rock fragments are predominantly cherts, polygonal quartz aggregates, and sutured quartz aggregates with minor fine-grained felsic schists. The clasts are generally lenticular shaped. Five to 15 percent is a well-foliated matrix composed of feldspar, muscovite, quartz, and opaques in which thin stringers of opaques and/or muscovite rim many of the clasts. Subrounded to angular grains of quartz and plagioclase (up to 1.5 mm) are also included in the matrix. In thin sections of progressively finer grained outcrops of the pebble conglomerate, the percent matrix increases, a decussate texture develops, and the clasts increasingly lose distinct boundaries.

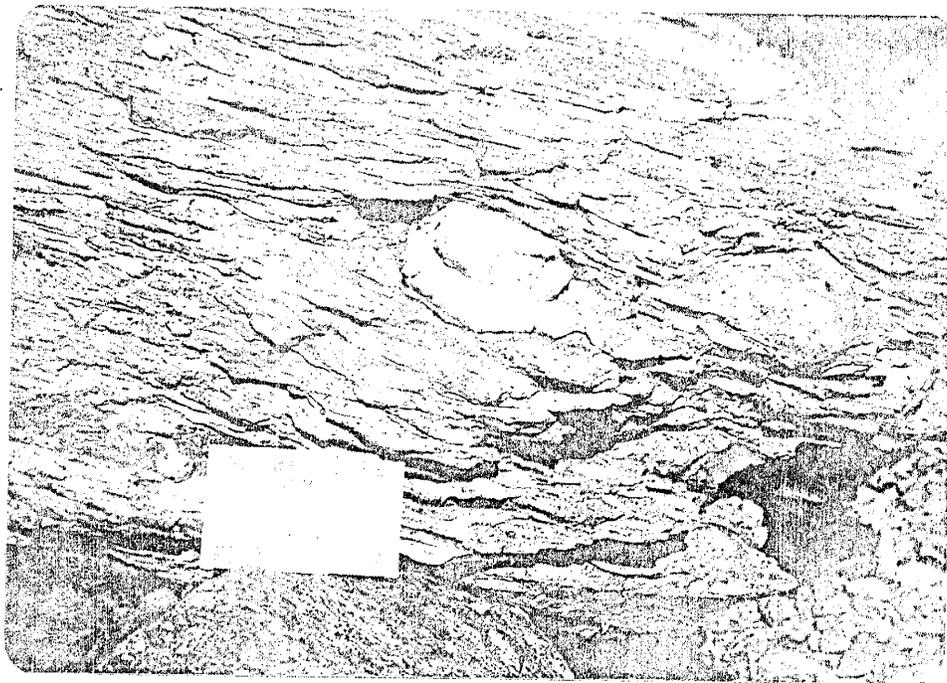


Figure 24. Photograph of pebble conglomerate taken perpendicular to foliation. Note the extreme flattening of all the clasts except for clasts of vein quartz.

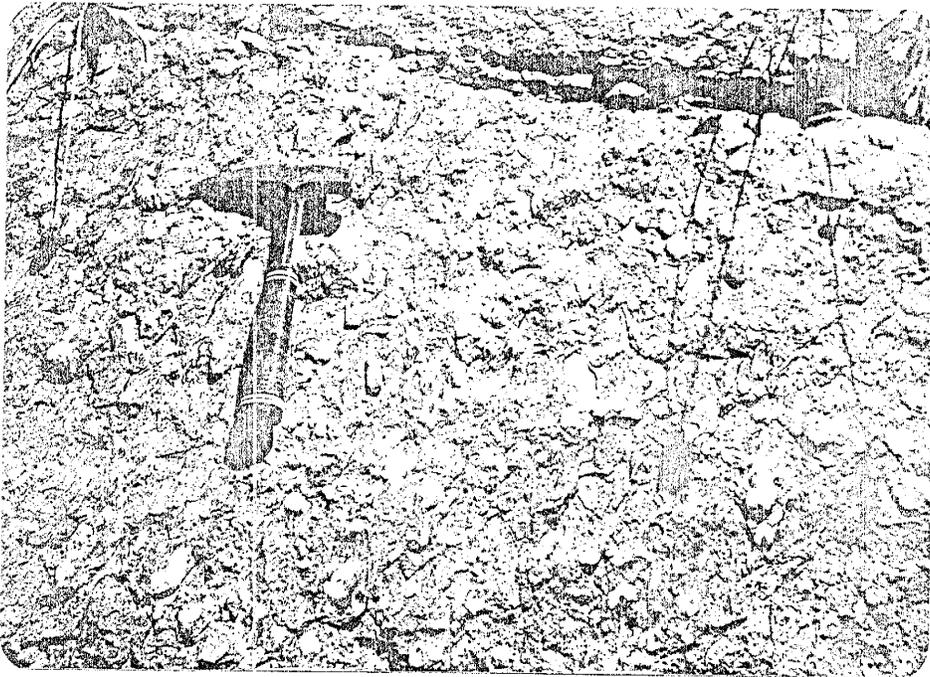


Figure 25. Photograph of pebble conglomerate parallel to the foliation. Note the prominent lineation formed by the stretched pebbles.

Pebble conglomerates also crop out on the east side of 8820 hill, sec. 34, T.28N., R.8E. but there they contain only angular clasts of quartz. These outcrops are whitish gray and weather to dark gray. They are poorly to moderately well sorted, and contain a variable amount of matrix material; at some localities very little matrix material is present. The matrix is composed primarily of quartz, muscovite, and biotite. The clasts range between 0.1 and 3.0 cm and are moderately stretched in the direction of dip.

chlorite phyllite At the base of the arkosic section and directly overlying the muscovite-feldspar-quartz schist (the felsic volcanoclastic schist) is a distinctive chlorite phyllite. It forms a continuous horizon between sec. 26, T.28N., R.7E. and sec. 33, T.28N., R.8E. although it is covered by Tertiary rocks in two places (sec. 25, T.28N., R.7E. and sec. 32 and 33 T.28N., R.8E.) and intruded by a small pluton of the Tusas granite (sec. 25, T.28N., R.7E.). It has a phyllitic dark- to light-green weathered surface and typically forms bold, highly schistose, blocky outcrops.

The phyllite commonly contains magnetite octahedra and in sec. 25, T.28N., R.7E. it also contains staurolite porphyroblasts. Thin sections of the staurolite-bearing phyllite contains 45 to 50 percent muscovite, 40 to 45 percent quartz, 5 to 10 percent biotite, 5 percent chlorite, 5 percent staurolite, and magnetite and other opaques as accessory minerals. The quartz grains (less than 0.1 mm in

length) are elongated parallel to the foliation displayed by the mica grains. Porphyroblasts of staurolite laths (up to 5.0 mm long) sharply cross the foliation and appear to have been subsequently rotated (figure 26). The staurolite has straight edges, ragged terminations, and is poikiloblastic to the quartz-mica matrix. It also appears to be replaced in part by a chlorite that is pleochroic light yellowish green, grass green, and light blue.

pebbly horizon Approximately 100 to 125 meters from the base of the arkoses between sec. 32, T.28N., R.8E. and sec. 26, T.28N., R.7E. is a pebbly horizon which was mapped and used as a stratigraphic marker within the arkoses. This bed is 1.5 to 3 meters thick and forms blocky, schistose outcrops with a characteristic reddish-brown, bumpy weathered surface. The outcrops are well bedded with relict crossbedding and graded bedding only rarely preserved. Reddish-brown banded chert stringers parallel the bedding in some outcrops. An outcrop below the saddle just west of hill 9021 (sec. 31, T.28N., R.8E.) contains a 30 cm long chert "nodule" and veins or stringers of black and reddish-brown banded cherty material which cross the foliation.

The pebbly horizon is composed of square to round feldspar grains (less than 5 mm in diameter), rounded quartz grains and lenticular felsic schist fragments which are generally less than 1 cm in diameter, and a highly schistose, muscovitic groundmass. In thin section, 25 to 30 percent of



Figure 26. Photomicrograph of staurolite in the chlorite phyllite (arkosic sedimentary rock section). The staurolite porphyroblast sharply cuts the well-developed foliation formed by a muscovite-quartz matrix.

the rock is rounded quartz, rock fragments, microcline, and plagioclase grains (in descending order of abundance) and 70 to 75 percent matrix. The microcline exhibits gridiron twinning and has indistinct grain boundaries; the plagioclase exhibit bent albite twins and are highly sericitized. The matrix is 60 percent polygonal quartz, less than 0.4 mm in diameter, and 40 percent muscovite. The quartz and muscovite of the matrix forms well-foliated lenses and layers which drape around the larger grains. Chlorite, clinozoisite, hematite, and opaque minerals occur in trace amounts. Clinozoisite and the opaque minerals are highly concentrated within a few layers of the matrix.

mafic schists Fairly continuous mafic schists 1 to 15 meters thick were also mapped for stratigraphic control within the arkosic sedimentary rocks. The mafic schists are distinctive in being dense, poorly to moderately foliated, and form orangish-red and dark brown (or green) rounded outcrops and boulders. They are generally very fine grained which makes mineral identification difficult in hand specimen; the major mineral constituents in coarser grained varieties are chlorite and plagioclase (+/- magnetite). Relict plagioclase phenocrysts (up to 1 1/2 cm in length) commonly weather in relief on the surface; generally they are broken and in varying stages of alteration to epidote and sericite. Some outcrops contain numerous thin (less than 1 mm) calcite stringers and nodules (less than 1 cm) flattened in the plane of foliation.

Subarkoses

Interbedded subarkoses and muscovite-quartz schists which are exposed along the Spring Creek road cuts weather to greenish-gray or grayish-white exposures. These rocks are isoclinally infolded with the pelitic schists.

These sedimentary rocks are predominately composed of quartz with varying amounts of feldspar and muscovite. The quartz grains are usually well sorted except in pebbly horizons. In the pebbly horizons, clasts of clear or milky quartz and feldspar up to 6 mm in length are stretched along strike. Mica-rich layers often show 5 mm euhedral biotite flakes developed at high angles to the foliation defined by the muscovite.

In thin section the subarkose contains 70 to 80 percent quartz, 10 to 15 percent feldspar, and 10 to 15 percent muscovite; clinozoisite, biotite, microcline, and carbonate occur in trace amounts. The quartz and feldspar grains display both straight and sutured grain boundaries. The quartz grains are generally 0.3 to 0.5 mm but range to almost 1.0 mm and are highly strained. The feldspars are 0.15 to 0.3 mm, highly sericitized, and exhibit albite or Carlsbad-albite twins. The muscovite exhibits a slight tendency to be oriented in one of two directions, approximately 70 to 80 degrees apart.

Pelitic schists

A section of pelitic schists and interbedded amphibolites is exposed along and south of the Spring Creek roadcuts (sec. 33, T.28N., R.8E.). This map unit contains four major rock types: hornblende gneisses, amphibolites, mica-quartz (+/- garnet) schists, and kyanite-garnet-staurolite-mica schists.

hornblende gneisses These beds are commonly thinly interbedded with the mica-quartz schists. The hornblende gneisses are banded and well foliated with a dark:light minerals ratio of about 1:1. In thin section the gneiss contains approximately 65 percent hornblende, 25 percent quartz, 10 percent muscovite, and small amounts of plagioclase, clinozoisite, and opaque minerals. The hornblende forms anhedral fragments which are strongly foliated and poikiloblastically encloses quartz and muscovite. The quartz forms bands and lenses of polygonal, unstrained grains. Muscovite is anhedral and is not aligned.

amphibolites The amphibolites have dark:light mineral ratios of about 3:1 and are thicker and more massive than the hornblende gneisses. These rocks are green and white, medium- to coarse-grained, and contain hornblende up to 15 mm in length which is randomly oriented within the planes of foliation. In places the hornblende forms a lineation down the dip of the schistosity. In thin section the amphibolite

is 50 to 70 percent hornblende, 25 to 30 percent plagioclase, 0 to 20 percent quartz, and 5 percent biotite with carbonate, clinozoisite, and opaques as accessory minerals. The hornblende is 0.3 to 5.0 mm in length and is well aligned although a few blades cut across the foliation. Biotite flakes are ragged and are sharply overgrown by the hornblende. The plagioclase grains are generally 0.1 to 0.5 mm, have sutured boundaries, and are generally fresh although some grains are slightly sericitized. The plagioclase are commonly twinned according to the albite and Carlsbad twin laws. Quartz grains are 0.1 to 0.5 mm in length, and also sutured and unstrained. The carbonate is strained and exhibits polysynthetic twinning as well as anomalous interference colors.

mica-quartz (+/- garnet) schists These schists are light gray, highly schistose and friable, and well banded due to varying proportions of biotite and muscovite. Red euhedral garnets up to 20 mm in diameter are abundant in some layers. Plagioclase grains, 2 to 7 mm in length, also occur in some of the bands. In thin section the major minerals are quartz (50 to 55 percent), muscovite (20 to 25 percent), and biotite (20 to 25 percent). Garnet, tourmaline, and opaque minerals are accessory minerals. The quartz grains are equigranular, generally less than 0.25 mm, and only slightly strained. Large biotite flakes (up to 1.4 mm in length) and smaller (0.15 to 0.5 mm) muscovite grains define a decussate texture.

The tourmaline is generally less than 0.12 mm in length and dark green pleochroic. In thin sections which contain plagioclase, the plagioclase grains range up to 1.0 mm in length, exhibit Carlsbad-albite twins, and are extensively altered to muscovite. In many cases the muscovite appears to form pseudomorphs after the plagioclase.

kyanite-garnet-staurolite-mica schists This unit directly underlies the subarkoses. Between the Spring Creek road and Spring Creek it forms distinctive light reddish-gray outcrops which appear conglomeratic due to a bumpy surface created by staurolite porphyroblasts (1 to 4 cm in diameter). Following strike southward the amount and size of kyanite in the rock increases and the amount of garnet decreases. Biotite flakes are at high angles to the foliation defined by muscovite. Along Spring Creek are outcrops of massive blue kyanite with blades 10 to 15 cm long. In a thin section without kyanite, the rock is composed of 60 to 70 percent quartz, 20 to 25 percent muscovite, 5 to 10 percent biotite; staurolite, chlorite, and garnet are each less than 5 percent. The quartz forms an equigranular mosaic texture with an average grain size of approximately 0.3 mm. Lenses of elongated quartz and aligned muscovite and biotite impart a distinctive fabric which is transected by both staurolite and chlorite. The biotite, staurolite, and garnets are poikiloblastic to the quartz.

Vitreous quartzite

A small area of vitreous quartzite was mapped between sec. 33, T.28N., R.8E. and sec. 27, T.28N., R.7E. The quartzite is assumed to be in fault contact with the arkoses and subarkoses for reasons discussed in the structure section of this report.

The vitreous quartzite forms massive, poorly schistose outcrops which weather to purplish-gray, smooth, rounded outcrops which break with conchoidal fracture. Jointing is generally conspicuous. Quartz veins of various sizes commonly form ptigmatic folds that cross the bedding of the quartzite. Detailed descriptions of the vitreous quartzite have been made by Barker (1958) and Bingler (1965). The reader should refer to these authors for more detailed descriptions of the quartzite.

Intrusive Rocks

Granodiorite

lineated granodiorite Outcrops of the lineated granodiorite form a broad wedge-shaped band between the western corner of sec. 27, T.28N., R.8E. and the south-central portion of sec. 23, T.28N., R.7E. This rock also crops out in the west-central portion of sec. 23, T.28N., R.7E. and the east-central portion of sec. 22, T.28N., R.7E. The eastern edge of the granodiorite is covered by Tertiary gravels and the western edge interfingers with the schists of the Moppin series. The contacts with the Moppin series are sharp and parallel the foliation of the schists.

The lineated granodiorite is gray and white and commonly weathers to grayish-white or pale-green or rarely to a bleached white color. It is moderately to very poorly schistose and typically forms large blocky outcrops. Elongated, 0.5 to 3.0 cm long, biotite-chlorite knots and 1 to 5 mm plagioclase phenocrysts are conspicuous in all outcrops and define the lineation; quartz phenocrysts are generally absent. In thin section the plagioclase phenocrysts and biotite-chlorite knots are seen to be contained in a fine-grained matrix of quartz, plagioclase, microcline, and biotite with carbonate, clinozoisite, epidote, and opaque minerals as accessories. The grain

boundaries of the minerals in the groundmass are sutured and irregular. The plagioclase phenocrysts are anhedral with irregular grain boundaries, highly sericitized, and are generally twinned according to Carlsbad-albite twin laws. Inclusions of calcite, quartz, epidote and clinozoisite are common within the phenocrysts. The biotite lineations consist of intergrown biotite (+/- chlorite) with small amounts of epidote, clinozoisite, and carbonate. Specimens which contain quartz phenocrysts are seen in thin section to have muscovite as a major component and the quartz phenocrysts are recrystallized to polygonal aggregates.

granodiorite of Spring Creek The granodiorite of Spring Creek crops out between sec. 25, T.28N., R.7E. and sec. 22, T.28N., R.7E. The western edge of the granodiorite body ends abruptly at Sheep Gulch and the eastern edge is gradational with the lineated granodiorite.

The granodiorite of Spring Creek is pinkish gray and weathers to a distinctive pinkish-orange color. The outcrops are very schistose and do not form good exposures. Much of the area mapped as granodiorite of Spring Creek is covered by colluvium.

Rounded, slightly bluish, quartz phenocrysts (up to 3 mm) and broken plagioclase phenocrysts are prominent on weathered surfaces. One to six millimeter pyrite cubes are found in some outcrops. The groundmass is highly schistose

due to abundant muscovite and chlorite and is typically pitted with elongate holes. In the field the granodiorite of Spring Creek can be distinguished from the lineated granodiorite by its prominent quartz phenocrysts, a distinctive pinkish-orange color, a phyllitic sheen on weathered surfaces, and the scarcity of biotite-chlorite knots.

In thin section, the granodiorite of Spring Creek consists of large sericitized plagioclase phenocrysts and rounded quartz grains in a groundmass of strained quartz and muscovite with opaque accessories. Biotite, chlorite, and epidote occur in varying amounts. Unlike the lineated granodiorite, microcline, carbonate, and plagioclase are not present in the groundmass. Carlsbad-albite twins are the predominant type of twin planes in the plagioclase. Bent twin planes, broken feldspars, and sericitized fractures in the plagioclase are evidence of cataclastic deformation.

Tusas granite

The Tusas granite is exposed over a two to three square mile portion of the northeastern flank of Tusas Mountain. The granite is discordant with the country rock on its western margin; a porphyritic igneous rock (reported by Barker, 1958, to be basaltic andesite in composition) exposed in the northeast corner of sec. 23, T.28N., R.7E. ends abruptly against the granite. Immediately adjacent to the

western border of the granite, however, the foliations of the granite and the porphyritic basaltic andesite parallel each other. The southern boundary of the granite is concordant with the enclosing greenschists and amphibolites. The contacts with the Moppin series are sharp although felsic material interfingers with the country rock within 150 meters of the contact. Fluorite is commonly found within both the granite and greenschists along the contact.

Three mineralogic variations of Tusas granite were noted within the major pluton; all three types are intergradational. The most common type of granite found throughout the pluton is poorly to moderately foliated and weathers to a reddish-orange, gray, or bleached white surface with phenocrysts of clear quartz and pink or white microcline in relief. Phenocrysts commonly comprise 30 to 50 percent of the rock (30 to 50 percent of the phenocrysts are quartz, and 50 to 70 percent are microcline) although samples were noted where the phenocryst content was in excess of 90 percent. In such cases quartz represents only 15 to 25 percent of the total phenocryst content. Rounded quartz phenocrysts are commonly 3 to 6 mm long (although lengths reach 1 cm; one phenocryst observed was 2 1/2 cm long), slightly bluish, and appear to have originally been bipyramidal. The microcline phenocrysts are pink or white, euhedral, 3 to 21 mm long (most average 3 to 5 mm), and contain mafic inclusions. The groundmass is very fine grained, pink or white, and typically

comprises 50 to 70 percent of the rock. Biotite is very rare.

A second distinctive type of granite occurs in a seemingly random manner throughout the pluton and is commonly, but not always, associated with the Moppin series contacts and with quartz veins. Good exposures are found along the ridge which trends east-southeast from 9809 hill in sec. 24, T.28N., R.7E. This granite is very dark red and consists of muscovite, quartz phenocrysts, and a fine-grained groundmass. In these rocks, quartz phenocrysts may comprise as much as 40 percent of the total rock, essentially no feldspar phenocrysts can be seen, and muscovite plus a variable amount of groundmass make up the remainder of the rock. In thin section the red color can be seen to be a result of hematite staining on muscovite grains. The quartz phenocrysts are recrystallized aggregates of strained quartz.

A third type of granite is very minor and is most abundant along the westernmost edge of the exposed portion of the pluton. This granite is deeply weathered, biotite is a major component (up to 10 or 15 percent of the rock), and schistosity is non-existent. Prominent quartz phenocrysts and only minor feldspar phenocrysts are contained in a white, fine-grained matrix which comprises between 65 and 80 percent of the rock.

Late intrusives

amphibolite dikes Amphibolite dikes intrude the eastern portion of the lineated granodiorite (sec. 28, T.28N., R.8E.) but poor exposures and their similarity to the greenschists and amphibolites prevented their identification within the Moppin series. The dikes could not be traced into the country rock surrounding the lineated granodiorite and lack of exposures prevented tracing them on the north slopes of the east-west ridge formed by the granodiorite. The dikes trend approximately N30W and cut the foliation of the granodiorite at high angles. The dikes form poor exposures, but are easily traced by abundant float. They are fine- to medium-grained, equigranular to slightly schistose amphibolites, with plagioclase:hornblende ratios between 1:1 and 1:2. Some specimens contain 3 to 5 mm feldspar blebs which may represent relict phenocrysts.

In a thin section with approximately equal amounts of hornblende and oligoclase, longitudinal sections of intimately intergrown hornblende (0.9 to 1.3 mm in length) are often bent and broken. The edges of cross-sections are ragged and resorbed and the ends of longitudinal sections are poorly terminated. The pleochroism for the hornblende is pale yellow, olive green, and dark green. The oligoclase grains are fresh, (0.06 to 0.31 mm in length) and equigranular with uneven extinctions.

granodiorite dike A granodiorite dike intrudes the Moppin series in sec. 20, T.28N., R.8E. The dike trends approximately N70W and although contacts with the country rock cannot be observed due to lack of outcrops, the trend of the dike forms a 20 degree angle with the foliation. The relationship between the dike and the Tusas granite cannot be determined because of poor exposures. Exploration pits are common on many of the exposures. This dike and the lineated granodiorite are not in contact with each other so that their relationship is not clear.

The granodiorite dike forms distinctive light-gray, massive outcrops which are discontinuously exposed along strike. The dike consists of 1 to 5 mm plagioclase phenocrysts in a quartz, plagioclase, biotite groundmass. In thin section, chlorite, clinozoisite-epidote, and carbonate are present in trace amounts. The highly sericitized and saussaritized plagioclase shows complex Carlsbad, Carlsbad-albite, and albite twinning. Biotite occurs throughout the groundmass in knots often associated with carbonate.

granitic dike In sec. 28, T.28N., R.8E. a north-trending granitic dike crosses the lineated granodiorite and the amphibolites of the Moppin series. Contacts with the enclosing rocks are not exposed along strike of the dike or at either end.

This unit is light gray, weathers to dark-gray surfaces, and forms distinctive blocky outcrops. The rock is very fine grained and contains 5 to 15 mm white oval spots which grade into white streaks and which usually have pyrite at their centers. The spotted portions of the rock can grade into unspotted portions in a seemingly random manner.

In thin section this rock contains 45 to 50 percent microcline, 5 to 10 percent plagioclase, 20 to 25 percent quartz, 5 to 10 percent muscovite and 5 to 10 percent biotite in a fine-grained (less than 0.15 mm) mosaic texture. Opaques (including pyrite) and clinozoisite are accessory. The white spots with or without pyrite centers consist of larger (up to 1 mm) grains of matrix material. When pyrite is present it is often intimately intergrown with biotite and clinozoisite. Muscovite and biotite form a well developed foliation.

pegmatites and quartz veins The only occurrences of pegmatite veins are found in cuts along the Spring Creek road (sec. 33, T.28N., R.8E.). These veins are small (less than 1.5 meters wide) and have a northerly trend. They cross-cut and contort the schists immediately adjacent to the veins. The mineralogy of the pegmatites is quartz, microcline, muscovite, and epidote; copper oxide staining is common.

Quartz veins intrude all the Precambrian rocks including the Tusas granite and an amphibolite dike in the

lineated granodiorite. They form sharp, usually discordant, contacts with the country rock, and do not appear to follow any consistent trend. Generally the veins are less than 3 meters wide and rarely exceed 60 meters in length.

Prospect pits can be found on a majority of the quartz veins, especially within the Bromide Mining District. The Bromide Mine (an abandoned silver mine in sec. 26, T.28N., R.7E.) is situated on a vertical quartz vein containing siderite and chalcopryrite (Lindgren, Graton, and Gordon, 1910). The enclosing schists are reported to contain stringers and lenses of tetrahedrite, calcite, and chalcopryrite.

aprites Pods and lenses of granitic aprite intrude the lineated granodiorite and the Moppin series in sec. 25, T.28N., R.7E. Equigranular, fine-grained (less than 1 mm) quartz, plagioclase, and microcline anhedral are contained in a pink or white aphanitic groundmass. In thin section the composition of the aprite is approximately 25 percent plagioclase phenocrysts, 25 percent quartz phenocrysts, 10 percent microcline phenocrysts and 40 percent groundmass. The plagioclase phenocrysts are 0.12 to 0.40 mm, very highly sericitized and cut by sericitized fractures. They show Carlsbad-albite, Carlsbad, and albite twins. The quartz phenocrysts are square or rounded, less than 0.3 mm in length, partially resorbed, and exhibit undulose extinction.

The groundmass shows a mosaic texture and consists of fine-grained (less than 0.01 mm) feldspar, quartz and muscovite with trace amounts of epidote and biotite.

APPENDIX II. Sample locations and descriptions.

- 99251 (SE 1/4, sec. 33, T.28N., R.8E.) Quartz-hornblende schist. Black and white, well-foliated schist; 65% hornblende, 25% quartz, and 10% muscovite. Basalt in Church's (1975) classification; compares with Nockold's (1954) central basalt except for a lower alumina and alkali content in this schist.
- 99252 (SE 1/4, sec. 33, T.28N., R.8E.) Amphibolite. Black and white schist; 50% hornblende, 30% plagioclase, 25% quartz, 5% biotite. Basalt in Church's classification; compares with Nockold's normal tholeiite basalt except for a slightly higher alumina content in this schist.
- 99195 (SW 1/4, sec. 33, T.28N., R.8E.) Amphibolite. Green and white foliated schist; 70% hornblende, 30% plagioclase. Basalt in Church's classification; compares to Nockold's average gabbro except for being slightly less calcium-rich.
- 98262 (NE 1/4, sec. 28, T.28N., R.8E.) Amphibolite dike. Fine-grained, black and white, poorly-foliated schist; 50% hornblende, 50% plagioclase. Basalt in Church's classification; compares to Nockold's average gabbro.
- 99197 (NW 1/4, sec. 23, T.28N., R.7E.) Greenschist. Medium-grained, green and white, unfoliated greenschist; 45% actinolite, 40% plagioclase, 15% epidote. Basalt in Church's classification; compares with Nockold's average tholeiite basalt.
- 99191 (SE 1/4, sec. 28, T.28N., R.8E.) Hornblendite. Coarse-grained, dark-green, poorly-foliated hornblendite; 55% chlorite, 45% hornblende. Basalt in Church's classification; compares closely with Nockold's average hornblendite except for a higher magnesium and lower calcium, sodium, and potassium content in this schist.
- 99253 (SE 1/4, sec. 33, T.28N., R.8E.) Pelitic schist. Gray and black, friable, well-foliated mica-quartz +/- garnet schist; 50% quartz, 25% muscovite, 25% biotite.
- 91051 (NE 1/4, sec. 32, T.28N., R.8E.) Amphibolite. Black and white, medium-grained, homogeneous amphibolite; 35% hornblende, 15% epidote, 15% biotite, small amount of quartz. Within both the basalt and andesite field of Church's classification; compares to Nockold's average andesite except it has a lesser amount of silica.

- 99292 (NE 1/4, sec. 32, T.28N., R.8E.) Matrix of mafic pyroclastic. Medium-grained, reddish-brown schist; 30% plagioclase, 20% hornblende, 15% biotite, 15% epidote, 20% quartz. Within both the basalt and andesite field of Church's classification; compares to Nockold's average andesite.
- 99293 (NE 1/4, sec. 32, T.28N., R.8E.) Porphyritic amphibolite. Black and white amphibolite with epidotized, unoriented plagioclase phenocrysts; 40% hornblende, 40% plagioclase, 15% epidote, 5% quartz. On border between andesite and basalt in Church's classification; most closely compares to Nockold's average andesite.
- 99192 (composite of three samples) Closely resembles composition of Nockold's average alkali syenite.
- 99000 (composite of three samples) Closely resembles composition of Nockold's average granodiorite except it is alkali-poor, especially in calcium.

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This dissertation is accepted on behalf of the faculty of the
Institute by the following committee:

James M. Ritt
Adviser

David L. Thomas

W. P. Lindsey

July 16, 1980
Date