BULLETIN 45

Precambrian and Tertiary Geology of Las Tablas Quadrangle, New Mexico

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1958

STATE BUREAU OF MINES AND MINERAL RESOURCES
NEW MEXICO INSTITUTE OF MINING & TECHNOLOGY
CAMPUS STATION SOCORRO, NEW MEXICO

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Abstract

Las Tablas quadrangle is in Rio Arriba County, northern New Mexico, and lies between north latitudes 36°30' and 36°45' and west longitudes 106°00' and 106°15'. Its center is about 33 miles west-northwest of Taos and about 65 miles north and slightly west of Santa Fe. Its principal geographic features include the Jawbone Mountain-La Jarita Mesa highland, which trends diagonally from northwest to southeast across the area; the valleys of the southeastward flowing Rio Tusas and Rio Vallecitos, which flank the highland area; a northeastern area that slopes gently eastward and is a part of the Taos Plateau; and a southwestern area that also slopes gently eastward and is a part of the highland that lies east of the Chama River valley. The Jawbone Mountain-La Jarita Mesa highland is underlain chiefly by Precambrian rocks, and the other areas mainly by Tertiary rocks.

The oldest exposed rocks in the quadrangle are Precambrian metasedimentary and metavolcanic rocks that comprise the Ortega quartzite, of which a 14,000- to 20,000-foot section is exposed; the Moppin metavolcanic series, which consists of metamorphosed basaltic rocks with minor intercalated metasedimentary rocks and is from 1,000 to several thousand feet thick; and the Kiawa Mountain formation, which is composed of five members. These members are the Big Rock conglomerate, which is from 50 to 100 feet thick and overlies the Ortega quartzite; the Jawbone conglomerate, which overlies the Moppin series in the northwestern part of the area, and is more than 1,000 feet in maximum thickness, and pinches out to the southeast; the lower quartzite member, which overlies the Big Rock conglomerate member, and is several hundred feet thick; the amphibolite member, which consists of one to seven thin layers of amphibolite and intercalated beds of quartzite, and is from 35 to 2,000 feet thick; and the upper quartzite member, which is from 4,000 to 8,000 feet thick and is the youngest Precambrian rock in this area.

These strata were compressed during Precambrian time into two large overturned folds that trend and plunge northwest. These are the Hopewell anticline and the Kiawa syncline. Two subsidiary and similarly oriented folds, the Poso anticline and the Big Rock syncline, lie on the southwest flank of the Kiawa syncline. Numerous internested minor folds, ranging from a fraction of an inch to several thousand feet in flank-to-flank dimension, occur on the larger folds.

Sills of Precambrian metarhyolite were injected into the sedimentary rocks prior to the folding. Three plutons of granodiorite were emplaced during the folding, and four bodies of granite were intruded during a late stage of the folding or after the folding ceased. Many bodies of granitic pegmatite, at least 36 of which are of some commercial importance, lie in muscovitized quartzite and metarhyolite on La Jarita Mesa.

Regional metamorphism was essentially synchronous with the folding of the Precambrian rocks. Basalt, the only widespread rock in the area that is sensitive to changes in metamorphic grade, was progressively metamorphosed to chlorite-muscovite-albite greenschist; to chlorite-biotite-albite greenschist; to chlorite-biotite-oligoclase greenschist; to oligoclase-biotite-hornblende amphibolite; and finally to hornblende-andesine amphibolite. These rocks represent the greenschist and amphibolite facies. The regional metamorphism has involved breakdown of unstable minerals, migration of atoms along grain boundaries, nucleation of new phases by statistical fluctuations of concentration, and grain growth by accretion of atoms to surfaces of nuclei.

Kyanite is present in all the vitreous quartzite, and is associated with both metamorphic facies. It occurs along bedding planes, in hematite-rich laminae, and in quartzose veins. Bodies of quartz-kyanite rock, which are oval in plan, occur in quartzite and metarhyolite at and near Big Rock on La Jarita Mesa.

The La Jarita pegmatites are surrounded by an aureole of pegmatitic-hydrothermal metamorphism that postdates the regional metamorphism. In this aureole the quartzose and feldspathic rocks have been muscovitized, and amphibolite has been partly converted to chlorite and quartz. Locally the amphibolite has been wholly replaced by muscovite, biotite, garnet, epidote, and quartz. The net material changes in this process have been addition of K, Al, and H_2O, and loss of Ca, Mg, Si, and a little Na. Fluids from pegmatitic magmas undergoing second boiling are believed to have caused the metasomatism.

Terrestrial sedimentary and volcanic rocks of Tertiary age underlie much of the quadrangle. The general stratigraphic relations are summarized in the following table:

Several thin beds with the cherty, feldspathic, quartzose sandstone lithology of the Santa Fe formation are present within the Cordito member of the Los Pinos formation.

Quaternary alluvium lies along the bottoms of the larger creeks, and some material of probable eolian origin underlies parts of the upper Tusas Valley.

Fault zones extend along the Rio Tusas and Rio Vallecitos valleys. The Tusas zone consists of northwest trending main normal faults connected by crossfaults that are subnormal to the main faults; the general displacement is east side down on the main faults. The Vallecitos zone is defined by northwest to west trending main faults, only some of which are joined by crossfaults; rocks west of the fault zone have been lowered relative to the rocks that lie east of the fault zone. Thus the Jawbone Mountain-La Jarita Mesa highland has been elevated relative to the Tertiary rocks to. the southwest, and depressed relative to the Tertiary rocks that are part of the Taos Plateau on the northeast.

TABLE 1. GENERALIZED STRATIGRAPHIC SECTION OF THE TERTIARY IN LAS TABLAS QUADRANGLE

Series	Formation	Character	Thickness (feet)
Pliocene(?)	Dorado basalt	Flows	40-100
	Unconformity		
	Cisneros basalt	Flows	10-30
	Unconformity		
	Los Pinos formation		
	Cordito member	Rhyolite-fragment conglomerate, tuff, and sandstone	600*
	Unconformity		
	Jarita basalt	Flows, disconnected	50*
	Biscara-Esquibel member	Conglomerate with fragments of andesitic to latitic composition, tuff, and sandstone	1,000*
Pliocene(?) or Miocene(?)	Biscara member	Conglomerate with fragments of andesitic to latitic composition, tuff, sandstone, andesite flow breccia, and andesite dikes	700*
	Unconformity		
Miocene(?)	Conejos(?) and Treasure Mountain formations (in Tusas Valley)	Conglomerate of Precambrian rock fragments, sandstone, tuff, conglomerate of felsite fragments; interlayered rhyolite welded tuff, from 10 to 18 feet thick	150-400*
	Ritito conglomerate (In Vallecitos Valley)	Conglomerate of Precambrian rock fragments; correlative with Conejos(?) or Biscara conglomerate	400*

* Maximum.

Introduction

GENERAL STATEMENT

Las Tablas quadrangle, in northern New Mexico, is underlain by Precambrian and metamorphic rocks, and by Tertiary and Quaternary sedimentary and volcanic rocks. Tusas Mountain, which extends diagonally across the area from northwest to southeast, is a part of the southeastern extension of the San Juan Mountains of Colorado into northern New Mexico. Gently tilted Tertiary rocks in the northeastern part of the area lie along the western edge of the Taos Plateau.

The main purpose of this investigation was to determine the stratigraphy, structure, and metamorphism of the Precambrian rocks. The sequence of folding, regional metamorphism, and pegmatitic-hydrothermal metamorphism constitutes a particularly complex problem in the area. Another major problem is the correlation of the exposed Precambrian rocks with those of the Ojo Caliente district, 15 miles south of the quadrangle; the Picuris Range, 35 miles to the southeast; and the Needle Mountains, which are a part of the San Juan Mountains of Colorado, about 95 miles to the northwest.

The major problem of the Tertiary rocks of the area is the relationship of the San Juan Mountains sequence with that of the Rio Grande Valley. The origin of the Abiquiu tuff of Smith (1938) is related closely to that of the Los Pinos formation of Atwood and Mather (1932, p. 92-101) and later workers.

LOCATION AND ACCESS

Las Tablas quadrangle lies between 36°30' and 36°45' N. lat. and 106°00' and 106°15' W. long. in Rio Arriba County, New Mexico (fig. 1). The center of the quadrangle lies about 33 miles west-northwest of Taos and 65 miles north-northwest of Santa Fe. The town of Tres Piedras lies 2 miles east of the eastern boundary of the quadrangle.

Several graded and graveled roads provide access to the area. State Highway 111 extends westward from Tres Piedras to Tusas, continues southwestward to Canon Plaza, and thence southward to Vallecitos. From Tusas a road extends 8 miles northwestward up the Tusas Valley to Deer Trail Junction and continues past Hopewell into the Brazos River drainage area beyond the west border of the quadrangle. A branch road connects Deer Trail Junction with San Antone, and also with Antonito, Colorado, via U. S. Highway 285. State Highway 111 and the El Vallecito and T-Bone ranches are joined by 8 miles of ungraded road. Cañon Plaza and Jaramillo's ranch, in Escondida Canyon, are connected by a graded gravel road. The Cafion del Agua road extends from U. S. Highway 285, south of Tres Piedras, to Las Tablas, Petaca, and La Madera. La Jarita Mesa is served by the Jarita Mesa

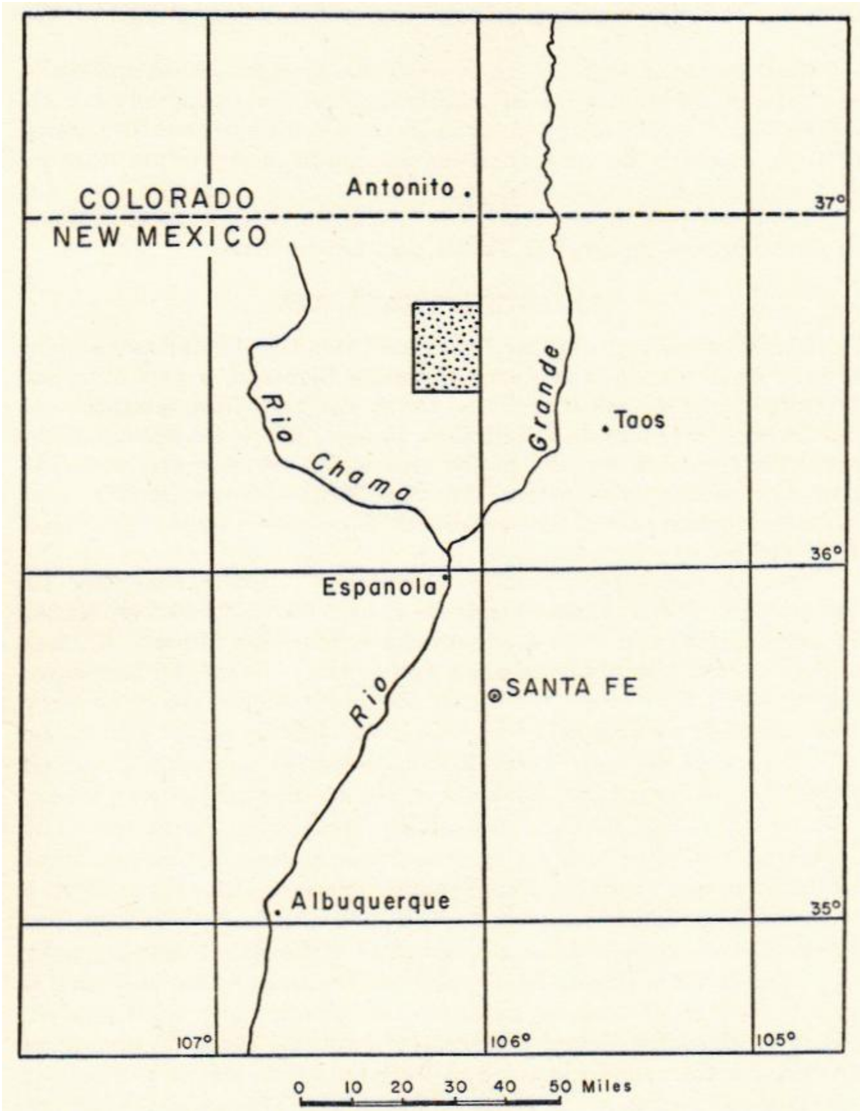

Figure 1.

INDEX MAP OF PARTS OF NEW MEXICO AND COLORADO, SHOWING LOCATION OF LAS TABLAS QUADRANGLE.

road, which connects Petaca with Route 111 at Spring Creek. A branch road joins Vallecitos with the Jarita Mesa road near Big Rock. A segment of the Vallecitos-Canjilon road lies in the southwest part of the area.

Many narrow, ungraded roads, truck trails, wagon roads, and trails serve the quadrangle. Vehicles with four-wheel drive generally can be driven within a mile of any point in the area during dry weather. Parts of all the roads in the quadrangle become muddy and difficult to travel during periods of very wet weather.

About 400 people live in the Las Tablas quadrangle, most of them in the villages of Petaca, Las Tablas, and Cañon Plaza.

PHYSICAL FEATURES

The dominant geographic features of the Las Tablas quadrangle are the La Jarita Mesa-Jawbone Mountain highland, a part of Tusas Mountain that trends diagonally across the area from southeast to northwest; the Tusas and Vallecitos Valleys, which bound the highland; the northeastern part of the area which slopes gently eastward and is the westernmost part of the Taos Plateau in this latitude; and the southwestern part of the area, which slopes gently toward the Vallecitos Valley.

The La Jarita Mesa-Jawbone Mountain highland comprises La Jarita Mesa, which ranges in altitude from 8,000 to 9,000 feet; Kiawa 'Mountain, shown in Plate 4, a monadnock that rises almost 1,000 feet above La Jarita Mesa to the south and to more than 1,500 feet above Spring Creek Canyon on the north; Tusas Mountain, shown in Plate 5A-B, another monadnock, which is 10,100 feet in height and is the culmination of the high Tusas Mountain-Burned Mountain-Hopewell ridge; the hills north and northeast of Hopewell; and Jawbone Mountain, whose east peak alone lies within Las Tablas quadrangle. The highland is covered with an open to heavy growth of timber, except in the meadows, which constitute about 10 percent of La Jarita Mesa, 5 percent of the Tusas Mountain-Burned Mountain-Hopewell ridge, and more than 50 percent of the area north of Hopewell. The topography of La Jarita Mesa, Burned Mountain, and the area around Hopewell is gently rolling. Most of the small streams flow in steep-sided gulches, which areally form about 20 percent of La Jarita Mesa, 10 percent of the ridge at Burned Mountain, and 5 percent of the area in the vicinity of Hopewell and south of Jawbone Mountain. This general highland area is crossed by only one stream, Spring Creek, which has cut a steep-sided canyon, as much as 700 feet deep, north and northwest of Kiawa Mountain. Some eastward flowing streams have deeply incised the eastern edge of the highland. These are Apache, La Jarita, Cow, Cunningham, Maquinita, Duran, and Buckhorn creeks, whose courses and the pattern of whose tributaries are partly controlled by the schistosity, joints, and lithologic variations in the underlying rocks.

The soft Tertiary rocks in the northeastern part of the quadrangle have been cut by shallow valleys, a few hundred feet deep, that trend northeastward to eastward. This is the western edge of the Taos Plateau, which slopes gently eastward and southeastward to the Rio Grande.

The area southwest of the Rio Vallecitos also slopes gently eastward from a high divide west of the quadrangle. It consists of east-northeast trending valleys and canyons about 800 feet deep, and of southeast to south trending shallow canyons, separated by rounded ridges. Altitudes in this part of the quadrangle range from about 7,500 to 9,500 feet. Only a small segment of the Rio Vallecitos-El Rito River divide is in Las Tablas quadrangle.

The Rio Tusas, the largest creek in the area, heads just northeast of Jawbone Mountain and flows into the quadrangle in a canyon that opens out into a wide valley (see pl. 5C), which extends southeast from Deer Trail Junction to Tusas. This valley has a steep northeast side and a gentle southwest side. One mile south of Tusas the creek enters a canyon, 700 feet deep, that extends south and southeast to Las Tablas. There Callon del Agua joins the Rio Tusas, which thence flows in a narrow-bottomed valley southward to Petaca and beyond. In this report the term upper Tusas Valley is applied to the part that extends northwest from Tusas, and the term lower Tusas Valley is applied to the part between Las Tablas and Petaca. Tusas Canyon lies between these two segments of the valley. Three large, south flowing tributaries join the upper part of the Rio Tusas. These are Guido Canyon, Biscara Canyon, and the nameless creek whose mouth is at Deer Trail Junction.

The Rio Vallecitos enters Las Tablas quadrangle at its western boundary about 2 miles west-southwest of Hopewell. It flows along the open upper Vallecitos Valley for about 4 miles, but just south of the T-Bone ranch it enters a canyon that continues southeast to Felipito Creek, with one break at El Vallecito ranch, where it is joined by Rock Creek. The Rio Vallecitos flows in a narrow-bottomed valley from Felipito Creek to Cañon Plaza, where the valley floor broadens out into what can be conveniently termed the lower Vallecitos Valley, part of which is shown in Plate 6B.

The only cultivated areas and open grasslands are in the bottoms of the major valleys, in the northeast corner of the area, and in the vicinity of Hopewell. The forest growth varies with altitude. Piñon is the dominant tree below about 8,000 feet, ponderosa pine is dominant from 8,000 to 9,000 feet, and spruce flourishes at altitudes above 9,000 feet. Aspen occurs in large stands above 8,500 feet where there is relatively moist ground. Scrub oak grows on many of the dry south slopes below an altitude of 9,000 feet.

PREVIOUS GEOLOGIC WORK

The earliest recorded geologic work in Las Tablas quadrangle is the brief description of several pegmatite bodies near Petaca by Holmes (1899). Graton (Lindgren, Graton, and Gordon, 1910, p. 124-133), briefly discussed the general geology of the Hopewell and Bromide districts and described more than 20 mines, prospects, and placer deposits. He recognized the Precambrian granite and dioritic rocks, amphibolite, quartzite, conglomerate, schists, and slate, as well as the Tertiary conglomerate, but did no general field mapping.

The Tertiary rocks of the Tusas Valley were described and mapped on a scale of 10 miles to the inch by Atwood and Mather (1932, p. 95-97), who noted the occurrence of the San Juan peneplain on Tusas Mountain (p. 23).

The first general geologic study of the area was made by Just (1937), who mapped in reconnaissance a large area extending from just north of Jawbone Mountain to a point south of Ojo Caliente, which lies about 15 miles south of Petaca. Despite his brief visit, Just's concepts of the geology, stratigraphy, and structure of the area, though much generalized, are remarkably accurate.

The Tertiary rocks of all but the southwestern part of the quadrangle were mapped by Butler (1946) on a scale of 1 inch to the mile. Butler traced the Tertiary and Quaternary rocks from the Colorado boundary to a point 10 miles south of Petaca, and thus was able to make the first definite correlation between these rocks and the well-known Cenozoic section of the San Juan Mountains in Colorado.

Smith (1938) described the Tertiary rocks of Abiquiu quadrangle, a part of which adjoins the southern boundary of Las Tablas quadrangle.

All the then economically important pegmatites on La Jarita Mesa were mapped in detail during 1943 and 1944 by Jahns and assistants (1946). Jahns also gave brief descriptions of the predominant rock types in that area.

The kyanite deposits at, and north of, Big Rock, on La Jarita Mesa, were studied and mapped in 1951 and 1952 by Corey (1953).

FIELD AND LABORATORY WORK

The writer devoted a total of eight and one-half months to field work in 1952 and 1953. Eleven months were spent in laboratory investigations and in preparation of this report.

Field mapping was done on air photographs at a scale slightly smaller than the 1:31,680 scale of the Soil Conservation Service planimetric sheet that was used as a base map. The map of the small area at Poso Spring (pl. 3) was made by stadia methods, with plane table and telescopic alidade.

ACKNOWLEDGMENTS

The author is greatly indebted to Dr. Richard H. Jahns, who suggested the Las Tablas area as one that merited detailed study, assisted at various stages of the work, and critically reviewed the manuscript.

Thanks are due the New Mexico Bureau of Mines and Mineral Resources for financial support for field expenses, chemical analyses, and thinsections. Dr. Eugene Callaghan, formerly director of the bureau, was especially helpful in obtaining support from that organization.

The manuscript was reviewed carefully by the late Dr. Robert Balk, also of the New Mexico Bureau of Mines and Mineral Resources.

Mr. Robert V. Shull, student at the New Mexico Institute of Mining and Technology, provided excellent nontechnical field assistance in July and August of 1953.

The writer also wishes to acknowledge the numerous personal kindnesses of Dr. and Mrs. A. F. Corey, of Yellow Springs, Ohio; Messrs. Joseph Veseley and F. Hawkins, of El Vallecito ranch; and Messrs. E. Wilton and L. Rush, of the U. S. Forest Service ranger station at Tres Piedras.

Stratigraphy

PRECAMBRIAN ROCKS

GENERAL STATEMENT

Within Las Tablas quadrangle are three major stratigraphic units, the Ortega quartzite, the Moppin metavolcanic series, and the Kiawa Mountain formation, comprising 5 members (table 2). The base of the lowest unit, the Ortega quartzite, and the top of the uppermost unit, the quartzite member of the Kiawa Mountain formation, are not exposed in the quadrangle; hence the total thickness of the Precambrian strata is not known. Their estimated present thickness, however, is between 21,000 and 32,000 feet.

TABLE 2. PRECAMBRIAN STRATIFIED ROCKS OF
LAS TABLAS QUADRANGLE

FORMATION	CHARACTER	THICKNESS (*feet*)
Kiawa Mountain formation		
Upper quartzite member	Light bluish-gray vitreous massive kyanitic quartzite, with tabular crossbedding, hematite laminae, and pebbly beds	5,000-10,000
Amphibolite member	Interbedded layers ofamphibolite and quartzite	35.2,000
Lower quartzite member	Light-gray, commonly vitreous, massive quartzite	Several hundred
Jawbone conglomerate member	Quartz-pebble conglomerate and gray quartzite; pinches out to the southeast	2,000(?)*
Big Rock conglomerate member	Quartz-pebble conglomerate, partly with interbedded quartzite	50-200
Moppin metavolcanic series	Greenschist and amphibolite, with minor conglomerate, phyllite, gneiss, and schist	Several thousand
Ortega quartzite	Light-gray to pink vitreous quartzite; with tabular crossbedding, hematite-ilmenite laminae, and pebbly beds	14,000-20,000

* Maximum.

ORTEGA QUARTZITE

Definition

The Ortega quartzite, as originally defined by Just (1937, p. 43), included all the Precambrian quartzite and intercalated conglomerate in Tusas Mountain. As redefined here this formation includes only the quartzite that is stratigraphically equivalent to the quartzite in the Ortega Mountains (Just's type area) and to overlying quartzite that extends up to, but does not include, the Big Rock conglomerate.

Distribution

The Ortega quartzite in Las Tablas quadrangle is exposed in two general areas separated by alluvium of the Vallecitos Valley. One area is formed mainly by the southwest slope of La Jarita Mesa; the other includes the rounded hills west of the Vallecitos Valley, from the latitude of Canon Plaza to the south boundary of the quadrangle.

Lithology

The lithology of this formation is slightly different in the two areas, but the differences are not marked enough to justify a subdivision into two members. In the area west of the Rio Vallecitos and immediately south of Cañon Plaza, the quartzite contains 90-95 percent of quartz, with accessory kyanite, hematite, ilmenite, muscovite, and rutile. It is light gray to pink, vitreous, dense, and generally massive in outcrop. Most of this rock is conglomeratic, and layers of rounded quartz pebbles are common. These layers are typically lenticular and range from one-half inch to several inches thick. The quartzite has a seriate mosaic texture of 0.12 mm average grain size, and contains quartz granules and pebbles as much as 10 mm in diameter, many of which are single xenomorphic crystals. The ultimate origin of the quartz in these clasts is not known.

Tabular crossbedding is common in this part of the Ortega quartzite, and is the only primary sedimentary feature that can be used to determine tops and bottoms of beds. Individual beds are outlined clearly by hematite-ilmenite laminae generally less than 1 mm thick.

Kyanite occurs in three general ways: (1) Along bedding planes in light-colored iron-oxide-free quartzite, (2) with hematite in original sedimentary laminae, and (3) in veinlets with quartz. Where irregularly disseminated along the bedding surfaces, the kyanite is randomly oriented, and ranges in grain size from 0.06 to 10 mm. Kyanite in this form may amount to as much as 5 percent of the rock. The kyanite is irregularly distributed in the hematite laminae, and commonly is concentrated as sheaves along the axes of tiny drag folds. It is also disseminated irregularly in unfolded laminae as individual crystals or as sheetlike aggregates.

Quartz veins are common in the Ortega quartzite, and generally are parallel or subparallel to bedding. They vary in size, and the largest of those observed is about 1 foot thick and 4 feet long. Only a few of them contain visible kyanite, which is disseminated in the quartz as prisms and in rosettes. Its grain size is variable and ranges from less than one-sixteenth of an inch to 1 inch or more. Kyanite in vitreous quartzite and conglomerate of the Kiawa Mountain formation is described below.

Ortega quartzite on the southwest slope of La Jarita Mesa is in part similar to the relatively "clean" quartzite described above, but much of it contains abundant muscovite. The beds beneath the Big Rock conglomerate and above the largest drag-folded sill of metarhyolite in secs. 20 and 21, T. 27 N., R. 8 E., consist of clean, commonly pebbly quartzite that is characterized by laminae rich in iron oxides. In contrast, the beds beneath the sill, which extend southeastward along the slope to the cap of Jarita basalt northeast of Vallecitos, consist of dominantly muscovitic, slightly feldspathic quartzite. This quartzite is light to dark gray, greenish gray, buff, and purple, has a sugary texture, and is well bedded. It commonly is laminated and comprises hematite- and leucoxene-rich laminae from 1 mm to about 5 mm thick alternating with quartz-muscovite layers from 2 mm to 5 cm thick.

In thinsection the muscovite quartzite is an equigranular mosaic, and the grain size ranges in different specimens from 0.08 to 0.12 mm. Specimens with seriate texture and grain size as large as 1 mm are common. Muscovite in flakes, 0.06 to 0.12 mm in diameter, that impart a fair to good schistosity to the rock, is present to the extent of 5 to 15 percent in most of the quartzite. A little sodic plagioclase is scattered as equant grains in the quartz mosaic. Hematite and leucoxene are present in approximately equal amounts in the dark laminae, in which they form tiny grains, single or in aggregates. Anhedral epidote, most of which is partly cloudy, is present in the more muscovitic quartzite in amounts of 3 percent to 5 percent. Kyanite has not been observed in this part of the Ortega quartzite.

Amphibolite

Seven thin layers of amphibolite are present in the Ortega quartzite in secs. 21, 29, 30, and 33, T. 27 N., R. 8 E., along the east slope of La Jarita Mesa, and in secs. 26 and 36, T. 27 N., R. 7 E., one-half mile west, and DA miles south-southeast of Cañon Plaza. This rock consists of hornblende, oligoclase, epidote, and chlorite, and probably is metamorphosed basalt. Six of the layers can be traced for short distances only. They appear to be parallel to the bedding of the quartzite, and hence are flows or sills.

Amphibolite, 1.5 miles south-southeast of Canon Plaza, which separates the large sill of metarhyolite from the quartzite to the west, reaches a maximum thickness of about 200 feet at the northwest end of its exposure. It pinches out to the southeast. The exposures at the northwestern end of this body are not complete enough to show whether the amphibolite has cut across the bedding of the quartzite. If this amphibolite is older than the metarhyolite, as seems likely by analogy with the relations of similar rocks elsewhere in the area, it was basalt (and not amphibolite) when the rhyolite was intruded. The basalt along its contact with the quartzite may have been much less resistant to intrusion by the rhyolite than the quartzite — thereby partly controlling the location and shape of the sill. The somewhat abrupt southward decrease in thickness of this amphibolite mass may be due to stoping by the metarhyolite. The lithology of this amphibolite is discussed farther on, in connection with the Moppin metavolcanic series.

Thickness

The true thickness of the Ortega quartzite, as exposed in the Las Tablas quadrangle, could not be measured, owing to indeterminable changes in thickness caused by drag folding and normal faulting. The present thickness of the Ortega quartzite from Cañon de Los Posos to the base of the Big Rock conglomerate in the NW' of section 2, T. 27 N., R. 8 E., was determined graphically to be about 25,400 feet, assuming no duplication or omission of strata by folding, faulting, or other means.

Beds of the upper unit of micaceous and vitreous quartzite face northeast almost entirely, as determined by crossbedding. Only a few drag folds are observable. The most elusive factor in determining the thickness of this unit is the proportion of metarhyolite sills in the section. The rocks are poorly exposed and contain muscovite of metasomatic origin for about 1 mile of outcrop breadth southwest of the Big Rock syncline, so that their lithology is not fully known. The large sills of metarhyolite northwest and southeast of this area of poor exposures, if they can be projected along their strikes into this area, suggest that one-half to two-thirds of the area must be underlain by metarhyolite. If this speculation is correct, the present observed thickness of the Ortega quartzite from the edge of the alluvium on the east side of the Vallecitos Valley to the Big Rock conglomerate must be 5,000 to 7,000 feet.

Bedding in the lower part of the Ortega quartzite is locally intensely folded. In most exposures, such as those on the northwest trending ridge that rises west from the Rio Vallecitos at the southern boundary of the area, the quartzite has relatively few drag folds; crossbedding indicates that the tops of the beds are to the northeast. The total thickness, corrected for dip, of the section from the edge of the alluvium on the east side of the Vallecitos Valley to the westernmost exposed quartzite in Canon de los Posos is approximately 17,400 feet, exclusive of any allowance for duplication by faulting and drag folding. The true thickness of this section is estimated to be from 9,000 to 13,000 feet. Thus, the total estimated thickness of the Ortega quartzite is between 14,000 and 20,000 feet.

Origin

The Ortega quartzite is very similar in nature to the Kiawa Mountain formation, and the origin of these two is discussed together in the section on the Kiawa Mountain formation.

MOPPIN METAVOLCANIC SERIES

Definition and Distribution

Greenschist, amphibolite, schist, and other, much less abundant, rock types, are exposed in a northwest trending belt extending from Hopewell to Cow Creek, American Creek, and in part, the southeast portion of Kiawa Mountain. These rocks were called the Hopewell Series by Just (1937, p. 42). That name, however, had been applied previously to another formation, and the group of rocks described above in this report is named the Moppin metavolcanic series, after the excellent exposures in upper Spring Creek just north of the Moppin ranch.

Greenschist and amphibolite are also exposed in a 2-square-mile area along the upper reaches of Buckhorn Gulch and the west fork of Duran Creek. Amphibolite also crops out in Apache Canyon (sec. 2, T. 26 N., R. 8 E.); just above the mouth of Biscara Canyon (sec. 5, T. 28 N., R. 8 E.); and on the northeast side of Tusas Valley, opposite the mouth of Maquinita Canyon (sec. 31, T. 29 N., R. 8 E., and sec. 36, T. 29 N., R. 7 E.). Several other occurrences of amphibolite have been listed in the discussion of the Ortega quartzite. All these scattered metavolcanic rocks, although they lie in different stratigraphic horizons, are similar in composition. Whether these rocks are flows, and hence of the same ages as the enclosing strata, or sills, and hence younger than the enclosing beds, is a question that is discussed below in connection with the origin of the Moppin metavolcanic rocks.

Lithology of Greenschists and Amphibolites

The Moppin metavolcanic series is largely greenschist and amphibolite, but thin layers of schist and gneiss occur locally. The series is intruded by sills and dikes of Burned Mountain metarhyolite and by sills, dikes, and plutonic masses of Maquinita granodiorite and Tres Piedras granite. The sills and dikes are present from points near Burned Mountain to Hopewell, and along upper Buckhorn Gulch, mostly as bodies too small to be mapped on the scale of Plate 1.

Typical chlorite-albite-epidote-calcite greenschist[1] is exposed in Spring Creek, 0.15 mile north of Moppin ranch, and northward from that area. The uniformly fine-grained green or dark-green rock weathers to jagged outcrops in which fractures follow the schistosity.

In thinsection this greenschist shows well-developed schistosity, which is given by 0.25-mm diameter flakes of chlorite, 0.12-mm elongate

[1]Throughout this report, the mineral modifiers in rock names are listed in order of decreasing abundance.

anhedra of albite, and 0.25-mm lenticular grains of calcite. Colorless epidote occurs as equant anhedra, many of which are grouped in irregular clusters. For the chlorite $\gamma - \alpha = 0.003 \pm 0.001$; X = Y = green, Z = pale yellowish green. The albite is clear, untwinned, and generally is zoned with the cores slightly more sodic than the rims. Leucoxene is present, but neither hematite nor magnetite was seen. Evidently the iron in the rock is wholly contained in the chlorite and epidote. About 15 percent of saussurite occurs as equant to irregular grains and clusters of grains.

Similar (Moppin) greenschist is exposed in Sheep Gulch, Rock Creek, and in areas farther northwestward to Hopewell. Aggregates of chlorite, interpreted as being retrograde after porphyroblasts of amphibole, are common in much of the greenschist. These relicts were found in the greenschist from several hundred yards west of the contact between greenschist and Tres Piedras granite at the west end of Tusas Mountain westward and northwestward to Hopewell. The aggregates of chlorite are elongated parallel to the dip of the schistosity. They are flattened in the plane of the schistosity, and their intermediate axes are parallel to the strike of the schistosity. Some are prismatic, showing oval to round cross-sections normal to the long axes.

Under the microscope the aggregates comprise chlorite grains that range from flaky to chunky and have an average length of about 0.05 mm. Most are parallel to the schistosity of the rock, but some of the aggregates are formed of randomly oriented grains. The chlorite is similar to that disseminated through the rock. Its birefringence is about 0.003, and its pleochroism is X = green, Y = pale green, Z = pale yellow to colorless. A few tabular grains of magnetite and equant crystals of epidote are present in many of the aggregates. Sericite forms as much as 20 percent of these aggregates; it appears only in sulfide-bearing greenschist, and is probably of hydrothermal origin. Several of the aggregates are about one-half albite.

The present shapes of most of the chlorite knots appear to be different from those of the original(?) amphibole grains. Deformation after development of the c-axes are essentially parallel, and give a lineation easily seen in hand specimen. The hornblende is pleochroic, with X = straw yellow, Y = green, Z = deep greenish blue, with $X < Y < Z$; $Z \wedge c = 22°$.

Iron and titanium oxides were not seen in thinsection, and these components are assumed to be contained in the biotite, hornblende, and epidote.

The amphibolite layers in the Kiawa Mountain formation and the Ortega quartzite that are exposed along the west edge of La Jarita Mesa are only slightly different from specimen 36-B-37 (table 3, no. 2). A specimen of a 5-foot layer from the edge of La Jarita Mesa, 2¼ miles north-northeast of Vallecitos (NE¼SW¼ sec. 33, T. 27 N., R. 8 E.) contains 45 percent hornblende, 30 percent calcic oligoclase, 10 percent

TABLE 3. CHEMICAL ANALYSES OF METAMORPHIC ROCKS IN LAS TABLAS QUADRANGLE

(In percent)

	1*	2*	3*	4*	5
SiO_2	44.44	54.57	52.53	48.78	51.3
TiO_2	1.84	1.50	1.41	2.28	1.9
Al_2O_3	15.04	16.27	14.22	14.82	14.0
Fe_2O_3	3.19	3.48	4.33	6.15	3.3
FeO	9.94	6.88	7.72	8.52	10.1
MnO	0.23	0.16	0.25	0.28	0.3
MgO	7.34	2.50	5.49	5.02	5.5
CaO	8.70	6.93	9.00	8.67	9.8
Na_2O	1.51	3.54	2.05	3.46	2.8
K_2O	0.13	2.11	0.56	0.26	0.7
P_2O_5	0.35	0.61	0.45	0.64	0.3
H_2O+	5.21	1.07	1.58	0.79	—
H_2O-	0.04	0.00	0.00	0.00	—
CO_2	2.30	0.00	0.00	0.00	—
Totals	100.26	99.62	99.59	99.67	100.0

1. Greenschist. From the SW¼NE¼ *sec.* 26, T. 28 N., R. 7 E. The composition of this schist is similar to that of many basalts and olivine basalts, except for the high content of water and carbon dioxide and the relatively low content of alkalis. Formation of calcite in vesicles and other openings after crystallization of the basalt accounts for the presence of the carbon dioxide and part of the lime. Water, in which chlorite and other hydrous minerals were formed, may have been added during weathering, or it may have been added during metamorphism. Water that migrated during metamorphism from the amphibolite that lay eastward and beneath(?) the greenschist, and from the original pores of the quartzite member of the Kiawa Mountain formation that lay to the south, may have been incorporated into the basalt. The low alkali content of the greenschist may be an original characteristic, or it may be due to leaching during weathering. The content of Na_20, especially, may have been lowered during weathering of original calcic plagioclase, in which a soluble sodium compound and calcite were formed.
2. Specimen 36-B-37. From the NE¼NE¼ sec. 23, T. 28 N., R. 7 E., just west of Tusas Mountain. The composition of this rock is intermediary to that of the average andesite and the average quartz basalt of Daly (1936), except for the low ratio of MgO/FeO, which is 2.50/6.88. This ratio may be due to separation from the original melt of early formed crystals rich in magnesia. The original basaltic rock appears to have undergone little change in composition during metamorphism.
3. Typical amphibolite. Specimen 36-D-2. From SW¼SE¼ sec. 9, T. 27 *N.*, R. 8 E.
4. Amphibolite with knots of chlorite. Specimen 36-D-60. From SE¼SE¼ sec. 14, T. 27 N., R. 8 E.
5. Average Deccan basalt (Walker and Poldervaart, 1949, p. 649).

* Analyst: H. B. Wiik, Helsinki, Finland.

epidote, 10 percent chlorite, 4 percent magnetite-ilmenite, and 1 percent apatite.

The two relatively thick, east trending units of the Moppin metavolcanic series, southeast of Tusas Mountain and north of Spring Creek, are largely hornblende-oligoclase amphibolite, with minor amounts of epidote, magnetite-ilmenite, and chlorite.

The most northerly of these two amphibolite units, which underlies Cow Creek, is composed almost wholly of very fine-grained hornblende and oligoclase. The grains of chlorite have altered the original pseudomorphic(?) shapes. The writer knows of no term that is applicable to deformed pseudomorphs or pseudomorphic aggregates. Perhaps they should be called metapseudomorphs(?). However, the author did not find partially formed pseudomorphs, nor other direct proof that these chlorite clusters formed from amphibole porphyroblasts. The possibility that they are porphyroblastic aggregates that formed from smaller disseminated grains of original pyroxene or amphibole, or from chlorite amygdules must be recognized.

True pseudomorphs of chlorite were noted only in chlorite-albite-sericite-carbonate greenschist from upper Rock Creek (NW¼SW¼ sec. 15, T. 28 N., R. 7 E.). These pseudomorphs comprise 75-90 percent of chlorite and 25-10 percent of magnetite; a few also contain sericite, albite, and epidote. The grains of magnetite commonly form clusters both within and along the margins of the pseudomorphs. Those within the pseudomorphs tend to occur in two crudely linear sets that are visible only on sections cut normal to the long axes of the pseudomorphs, whose angular relationships are similar to those of the cleavage directions of amphiboles. The magnetite is believed to be distributed along such an original cleavage. This is the best evidence that these true pseudomorphs and the much more abundant metapseudomorphs(?) have formed from amphibole grains.

Aggregates of chlorite are well developed in the thick chlorite-albiteepidote-calcite(?) greenschist at the head of Sheep Gulch. A layer of rock of ½-mile outcrop breadth contains clusters as much as 30 mm long, 4-5 mm wide, and 1-1.5 mm thick. In thinsection this schist is similar to the one from Spring Creek, 0.15 mile north of Moppin ranch, except for well developed metapseudomorphic(?) chlorite, calcite lenses as much as 20 mm long that show a marked lineation down the dip of the lineation, and the presence of poikiloblastic magnetite.

Relict phenocrysts of plagioclase are present in much of the greenschist of the Moppin metavolcanic series. They are especially well developed just north of Burned Mountain, in sec. 5, T. 28 N., R. 7 E., and near the head of Buckhorn Gulch in sec. 30, T. 29 N., R. 7 E. At the first locality the greenschist contains 25-30 percent of subhedral to anhedral relict phenocrysts of plagioclase, from 0.5 to 5 mm in maximum dimension. At the second locality the euhedral to subhedral phenocrysts form 25 to 30 percent of the rock and range from 1 to 12 mm in length. All of the relict phenocrysts have been moderately to completely altered to saussurite, sericite, chlorite, calcite, and secondary albite in various combinations. The least altered phenocrysts range from oligoclase to albite. The plagioclase of the original basalts, which is discussed below, appears to have been altered to sodic plagioclase and to the calcium-bearing saussurite and calcite, and partly replaced by sericite and chlorite.

A few of the relict phenocrysts probably have been rotated during the folding, as suggested by the wrapping of the schistosity around such grains. None appears to have grown during the metamorphism, and many have been fractured and veined by later metamorphic albite, chlorite, calcite, and sericite. A few have been granulated and the fragments rolled out along the schistosity.

The greenschists of the Moppin metavolcanic series one to 1½ miles north of Hopewell are similar to the chlorite-muscovite types described in the preceding paragraphs. Most of the greenschist in this area contains relict phenocrysts of plagioclase. Oligoclase-albite-chlorite-epidote-sericite-calcite schist from the SD/$_4$ NEN sec. 30, T. 29 N., R. 7 E., contains 25-30 percent of sericitized and saussuritized phenocrysts of oligoclase, mostly euhedral, chunky to lathlike in shape, and 1 to 15 mm in length. A relict ophitic groundmass is suggested by many relict(?) plagioclase laths, 0.05-0.20 mm long, that are dispersed among the relict phenocrysts. The secondary albite in this schist is rounded to polygonal in outline. Twinning is rare, and most of the grains are more calcic in their margins than in their cores.

Small amounts of biotite are present in some of the chlorite-albite-epidote greenschists. At Rock Creek in the SE¼SE¼ sec. 16, T. 28 N., R. 7 E., the schist shows biotite grains 0.12 mm in diameter scattered or interlayered in larger chlorite grains. The birefringence of the biotite ranges from 0.030 to 0.005, and there appears to be a complete transition from biotite to chlorite. Pleochroism in the biotite ranges from X = Y = pale brown, Z = dark brown in the least altered(?) material to X = Y = pale olive, Z = olive brown to X = Y = pale green, Z = green in the material that is almost totally altered to chlorite. The chlorite that appears to have formed from biotite is partly and irregularly stained olive to dark brown, probably by ferric oxide. This implies that the ferric iron content of the biotite is higher than that of the coexisting chlorite, which, in such greenschists, is probably less than 1 percent (Wiseman, 1934, p. 362). The altered biotite is found only in sericite-bearing rocks, and most of the sericite and biotite is believed to have formed by hydrothermal alteration, in conjunction with sulfide mineralization.

All of the chlorite- and muscovite-bearing greenschist described above represents the muscovite-chlorite subfacies of the greenschist facies, as discussed in the section on regional metamorphism.

PLATE 4

KIAWA MOUNTAIN

VIEW FROM THE NORTHEAST. TUSAS CANYON IS IN THE FOREGROUND, WITH VERTICALLY JOINTED CORDITO CONGLOMERATE FORMING ITS WESTERN RIM.

A B C D

PLATE 5

TUSAS MOUNTAIN AREA

A. TUSAS MOUNTAIN FROM KIAWA MOUNTAIN, LOOKING NORTH-NORTHWEST. CLEVELAND GULCH IS IN RIGHT MIDDLE BACKGROUND. SPRING CREEK EXTENDS LEFT TO RIGHT ACROSS THE PHOTOGRAPH.

B. TUSAS MOUNTAIN FROM THE EAST ousT SOUTH OF mE DIVIDE ON THE TUSAS-TRES PIEDRAS ROAD).

C. VIEW OF THE UPPER TUSAS VALLEY FROM SEC. 15, T. 27 N., R. 7 E., LOOKING SOUTHEAST. TUSAS MOUNTAIN IS ON THE RIGHT SKYLINE.

D. VIEW DOWN THE. TUSAS VALLEY. PART OF LAS TABLAS IS VISIBLE ON THE RIGHT. PETACA MESA RISES IN THE LEFT AND MIDDLE DISTANCE.

A

PLATE 6

ORTEGA QUARTZITE

A. CLIFF OF ORTEGA QUARTZITE. ON THE NORTH SIDE OF CAÑADA JACQUES, IN SEC. 35, T. 27 N., R. 7 E.

B. VIEW SOUTHWESTWARD ACROSS VALLECITOS VALLEY FROM THE WEST EDGE OF LA JARITA MESA, SEC. 33, T. 27 N., R. 8 E. THE HILLS IN THE MIDDLE DISTANCE ARE UNDERLAIN BY ORTEGA QUARTZITE.

PLATE 7

PRECAMBRIAN ROCKS

A. CONTORTED MUSCOVITE - OLIGOCLASE -STAUROLITE-KYAN ITE-MAGNETITE SCHIST, INTERLAYERED IN THE MOPPIN SERIES. SW¼SW¼ SEC. 27, T. 28 N., R. 8 E. PORPHYROBLASTS OF STAUROLITE AND KYANITE ARE VISIBLE.

B. JAWBONE CONGLOMERATE ON JAWBONE MOUNTAIN. THIN DARK LAMINAE OF HEMATITE DIP TO THE. RIGHT, PARALLEL TO THE BEDDING. LOOKING SOUTHWEST

A

B

PLATE 8

BISCARA CONGLOMERATE AND PEGMATITE
IN MAQUINITA GRANODIORITE

A. CONGLOMERATE, WITH FRAGMENTS OF PRECAMBRIAN ROCKS, IN THE. BISCARA MEMBER OF THE LOS PINOS FORMATION, SE¼NE¼ SEC. 30, T. 27 N., R. g E.
B. SMALL PEGMATITE BODIES IN THE MAQUINITA GRANODIORITE, NW¼ SEC. 27, T. 28 N., R. 8 E. THE PEGMATITES ARE PARALLEL TO THE FOLIATION OF THE GRANODIORITE AND SHOW A LINEATION PARALLEL TO THE STEEPLY PLUNGING LINEATION IN THE HOST ROCK.

A

B

PLATE 9

CORDITO CONGLOMERATE

A. CONTACT OF CORDITO RHYOLITE-BOULDF.R CONGLOMERATE AND MOPPIN AMPHIBOLITE. AND SCHIST IN THE SPRING CREEK CANYON ROADCUT, SW¼SE¼ SEC. 33, T. 28 N., R. 8 E. THE PRECAMBRIAN ROCK DIPS MODERATELY TO THE. LEFT.

B. CORDITO CONGLOMERATE, DIPPING GENTLY TO THE NORTHEAST, 1.3 MILES SOUTH OF TUSAS. THE RIO TUSAS IS AT BOTTOM OF SLOPE.

PLATE 10

TRES PIEDRAS GRANITE

A. TRES PIEDRAS GRANITE ON THE WEST RIM OF TUSAS CANYON, NW¼ SEC. 2, T. 27 N., R. 8 E. WEATHERING OF THE GRANITE HAS BEEN PARALLEL TO THE LEFT-DIPPING FOLIATION. QUARTZITE PEAK IS VISIBLE ON THE MIDDLE LEFT SKYLINE; TUSAS PEAK ON THE EXTREME RIGHT.

B. CONTACT OF THE TUSAS MOUNTAIN PLUTON OF THE TRES PIEDRAS GRANITE WITH AMPHIBOLITE, WEST OF TUSAS MOUNTAIN, NE¼NE¼ SEC. 23, T. 28 N., R. 7 E.

PLATE 11

RIG ROCK CONGLOMERATE

FOLDS IN PEBBLY QUARTZITE OF THE BIG ROCK CONGLOMERATE, SW¼ SEC. 23, T. 27 N., R. 8 E. LOOKING NORTHWEST.

The Moppin greenschists are slightly biotitic in the eastern part of the exposure between Buckhorn Gulch and Duran Canyon (sec. 28 and the eastern part of sec. 29, T. 29 N., R. 7 E.). Oligoclase-chlorite-epidote-biotite schist from the SW¼NE¼ sec. 29 contains about 6 percent biotite, most of which occurs with chlorite in flattened ellipsoids oriented down the dip of the schistosity. The biotites, pleochroic from pale brown to dark greenish brown, are tabular to chunky, and as much as 2 mm long; a few are chloritized at their edges. These biotitic schists are grouped with the biotite-chlorite subfacies of the greenschist facies.

The facies boundary between the greenschists and the amphibolites of the Moppin metavolcanic series extends from just west of Tusas Mountain, south and east to the head of Cleveland Gulch. It extends from there southwest into the Kiawa Mountain formation, and lies beneath the Tertiary rocks west of the amphibolite layers exposed at the junction of Vallecitos and Escondida creeks.

The amphibolite exposed just west of the contact between the Moppin series and the Tusas Mountain unit of Tres Piedras granite lies a few hundred yards east of the greenschists. Specimen 36-B-37 of this amphibolite, taken 25 feet west of the granite contact, is an oligoclase-epidote-biotite-hornblende amphibolite, composed of 55 percent plagioclase, 30-35 percent epidote, 6-8 percent biotite, 4-5 percent hornblende, and a trace of apatite. This rock shows a crude schistosity subparallel to the granite contact, and a lineation of hornblende grains plunging 15 degrees in a S. 15° W. direction. The rock weathers to jagged, irregular outcrops.

Relict phenocrysts of andesine, now considerably epidotized, fractured, and veined by all of the metamorphic minerals, constitute about 15 percent of this rock. The younger metamorphic plagioclase, a median oligoclase, is present as equant to slightly elongate grains, 0.01-0.12 mm across, that gives a mosaic texture. The coarser-grained oligoclase forms irregular patches, lenses, and veinlets, mostly associated with the relict phenocrysts of andesine. Epidote occurs as clusters or as single colorless grains, 0.02 mm across. Its birefringence is 0.025, and its Fes + :Al ratio, therefore, may be about 1:6. Chunks or flakes of biotite with pleochroism pale olive to dark greenish brown give the amphibolite its schistosity.

Subhedral to euhedral prisms of hornblende, from 0.12 to 1 mm long, are irregularly distributed in the thinsection, but regularly scattered in the rock. Their hornblende generally occurs as stubby subhedral prisms, which are 0.10 mm in average length and have well-alined c-axes that give a clearly defined lineation. Hornblende forms 60-75 percent of the amphibolite, and is present as felted masses and as separate grains set in equigranular mosaic aggregates of oligoclase. The oligoclase is interstitial to the felted grains of hornblende, and also is present as lenticular and irregular aggregates or clusters. The oligoclase ranges in grain size from 0.01 to 0.25 mm, but most grains are close to he average size of 0.06-0.10 mm. Slight to moderate zoning is common, in which the cores are more sodic than the marginal portions of the grains. Magnetite-ilmenite forms 1-5 percent of the amphibolite. Epidote is absent or is present in trace amounts. An unusual epidote-rich part of this amphibolite mass, from exposed rock near Cow Creek (SE¼SW¼ sec. 21, T. 28 N., R. 8 E.), 50 feet south of the Tres Piedras granite, contains irregular veins of epidote and quartz. A thinsection showed about 60 percent epidote, 25 percent oligoclase, 15 percent hornblende, and a trace of fluorite. The epidote is believed to have formed mostly from original hornblende. Several relict phenocrysts of plagioclase are present in the section, and these contain only a few grains of epidote. Hydrothermal solutions associated with the adjacent granite probably effected this change from typical amphibolite to this epidotic variety.

The most southerly of these two amphibolite units, which underlies parts of American Creek and Cleveland Gulch, is largely amphibolite, but does contain some gneiss and schist of intermediate composition.

The amphibolite in the central part of this mass differs from that of the mass to the north in that it is coarser grained (especially the hornblende) and commonly is slightly to moderately epidotic. Near the head of a tributary of Cleveland Gulch, in the NW¼SW¼ sec. 29, T. 28 N., R. 8 E., this rock is composed of 55 percent hornblende, 25 percent oligoclase, 15 percent epidote, 4 percent magnetite-ilmenite, and traces of chlorite and quartz. The hornblende subhedra are porphyroblastic, and have an average length of about 1 mm; about one-half of them are alined, and the remainder are randomly oriented. The mineral is pleochroic, with X = pale greenish yellow, Y = sea green, Z = blue green, and the absorption $X < Y < Z$. The extinction Z Λ C = 21° and (—) 2V = ca. 80°. These properties are typical of the hornblende in the Moppin amphibolite. The other minerals in this amphibolite are about 0.05 mm in average grain size. The few chlorite grains present are xenomorphic toward hornblende.

The grain size of these amphibolites increases to the east, where they are well exposed on the west side of Tusas Valley just west of their contact with the Cordito member of the Tertiary Los Pinos formation. Hornblende-oligoclase amphibolite and hornblende-oligoclase-epidote amphibolite are interlayered here with gneiss and schist described below. A number of quartz and quartz-epidote veins, a few inches to a foot thick, cut these amphibolites and schists. They are rodlike to tabular in shape, and evidently owe their form to close control by the steep hornblende lineation and the east-west trending schistosity. An epidotized amphibolite from the bed of Aveta Creek consists of 55 percent hornblende, 30 percent epidote, 10 percent oligoclase, and 5 percent vein quartz. The hornblende is irregularly bleached from the typical blue-green (along Z) type to a colorless tremolite or soda tremolite. Porphyroblasts of hornblende transect the probably older hornblende, which is well lineated. Both varieties of the hornblende are similar in appearance. The secondary epidote and the possibly secondary hornblende in this rock differ from that described above for the amphibolite at Cow Creek, in that here the plagioclase evidently has been destroyed and the original hornblende has remained unaltered.

The coarsest amphibolites in the quadrangle are in the western part of sec. 34, T. 28 N., R. 8 E., and along Spring Creek. In thinsection, the dark-green, slightly schistose amphibolite from the NW¼ SW¼ sec. 34 shows 50 percent euhedral to subhedral, poikiloblastic prisms of hornblende that range from 0.25 to more than 10 mm long; 30 percent equant, anhedral grains of calcic oligoclase that average 0.12 mm in size; 10 percent chlorite in sheaflike clusters and single platy irregular grains; 7 percent equant to irregular grains of epidote; and 3 percent magnetite-ilmenite and apatite. This rock is crudely laminated, and consists of hornblendic and plagioclase-rich layers that pinch, swell, and converge. It could not be determined whether this feature is the result of layering of the original basaltic rock or of metamorphic diffentiation. The anhedral, irregularly shaped grains of chlorite are interstitial to the larger euhedral and subhedral prisms of hornblende, a relationship which suggests that the hornblende formed later than the chlorite. Thus the reaction between chlorite and epidote to form hornblende did not go to completion.

Lithology of Other Rocks

The Moppin metavolcanic series contains a number of thin, discontinuous beds of conglomerate, phyllite, gneiss, and schist. These rocks are not extensive enough to warrant their treatment as members or as separate formations. They underlie parts of Rock Creek, Sheep Gulch, Spring Creek, Cleveland Gulch, American Creek, and Aveta Creek.

Conglomerate and phyllite are exposed on the most northerly tributary of Rock Creek, about 1 mile southeast of Burned Mountain. The conglomerate contains pebbles of quartz, dark-red slate, and buff felsite(?) set in a gray to purple micaceous matrix. This vertically dipping unit ranges in thickness from 250 to 300 feet. Greenschist underlies the conglomerate on the north, and sheared granodiorite bounds it on the south. About 150 yards south of the conglomerate, a silvery-gray pebbly sericite-quartz-magnetite phyllite crops out in the narrow meadow that extends along the creek bottom. The outcrop breadth of this phyllite layer ranges from 30 to 50 feet. A steeply dipping schistosity makes high angles with the faintly preserved bedding, and hence the true thickness is probably much less than that implied by the outcrop breadth. Neither the conglomerate nor the phyllite layer could be traced far into the forests that bound the meadow.

Near the head of Sheep Gulch, in the N½ sec. 22, T. 28 N., R. 7 E., a schistose, chloritic and sericitic greenish-gray to silvery-gray quartzite overlies the greenschists to the north, shows a vertical schistosity, and contains many lenses or sheared veinlets of quartz elongate down the dip of the schistosity. The true thickness of this quartzite probably is less than its outcrop breadth, which is about 200 feet; but the bedding is not well enough preserved to permit an estimate.

The chloritic and sericitic quartzite is overlain by interlayered quartzose pebble conglomerate and quartz-sericite-plagioclase-magnetite schist of probable subgraywacke composition. The outcrop width of this vertically dipping unit is 500 feet. The schist is dark gray to brown, shows crossbedding deformed by folding, and has steeply dipping schistosity and lineation like those in the quartzite to the north. The conglomerate layers contain quartz pebbles whose long axes are well alined down-dip in the plane of the schistosity. Vitreous quartzite, which contains at least one bed of conglomerate and metamorphosed subgraywacke similar to the foregoing unit of 500-foot outcrop breadth, overlies these beds.

Grayish-green quartz-plagioclase-sericite-chlorite schist crops out in the canyon north of the Moppin ranch. It immediately underlies the Kiawa Mountain quartzite and is interlayered with the greenschists described above. Under the microscope it is a mosaic of quartz and plagioclase streaked with well-foliated sericite and chlorite that commonly wrap around lenticles of the equigranular minerals. The average grain size is 0.12 mm. Porphyroblasts of quartz and plagioclase as much as 0.5 mm in diameter are scattered throughout the rock. This schist originally may have been a tuff of dacitic composition, or a tuffaceous sandstone.

On the north side of Cleveland Gulch, about 1½ miles from its mouth (SW¼SW¼ sec. 29, T. 28 N., R. 8 E.), a green laminated quartz-sericite phyllite is exposed. It has an outcrop breadth of about 200 feet. The quartz grains in this rock are about .06 mm in average diameter, and the sericite flakes are .01-.05 mm long. The laminae are 0.5 mm thick and show well-developed graded bedding similar to that in varved siltstones. Porphyroblasts of magnetite, 1 mm average diameter, appear in the coarser parts of the laminae. They are adjoined by quartz with well developed pressure shadows.

Rocks of the Moppin series along the west side of the Tusas Valley from American Creek south to Aveta Creek include several types of schist and gneiss.

On the north side of American Creek, in the NE¼ SE¼ sec. 28, T. 28 N., R. 8 E., a 10-foot layer of laminated quartz-plagioclase-biotite-epidote-muscovite schist is intercalated in the amphibolite. This rock may be another metamorphosed tuff of intermediate composition.

The section exposed from the first creek south of American Creek to just south of Aveta Creek comprises approximately 20 feet of oligoclase-biotite-epidote-microcline schist that may be another metamorphosed tuff; 20 feet of amphibolite; 200 feet of oligoclase-quartz-biotite-chlorite-hornblende gneiss that may be a metamorphosed dacitic tuff or flow, or, less probably, a graywacke; 10 to 30 feet of coarse-grained muscovite-oligoclase-staurolite-kyanite-magnetite schist that is a metamorphosed layer of pelitic sediments; and several hundred feet of interlayered oligoclase-quartz-biotite-microcline schist and amphibolite. Amphibolite crops out from just north of Aveta Creek south to the pebbly conglomerate.

A 12-foot bed of quartz-muscovite-biotite-plagioclase-garnet schist is interbedded with amphibolite in the Spring Creek Canyon roadcut 40 yards west of where the Cordito member crosses the road. Dark-red garnets 4 or 5 mm in diameter form slightly flattened crystals with quartz inclusions parallel to the flattening. The garnets have been slightly rotated about axes subparallel to the strike of the schistosity. The average grain size of the schist is about 0.25 mm.

Thickness

The thickness of the Moppin metavolcanic series could not be determined because of repetition by folding and omissions or displacements at boundaries of adjacent intrusive rocks. The minimum apparent thickness, as measured between the bounding Kiawa Mountain formation and the Tres Piedras granite (omitting the included Maquinita pluton) in secs. 24 and 25, T. 28 N., R. 7 E., is about 3,000 feet. The relatively large areal extent of this part of the Moppin series implies a true thickness of at least several thousand feet.

The outcrop breadth of the Moppin series and included intrusive rocks from Duran Canyon to Buckhorn Gulch is about 1 mile. These rocks are close to the nose of the Hopewell anticline, and have been closely folded. The true thickness of this part of the Moppin series is probably from one to several thousand feet, assuming that a relatively minor portion of the original section has been displaced by intrusive rocks.

Origin

The greenschists and amphibolites of the Moppin metavolcanic series are believed to be metamorphosed volcanic rocks, because their chemical composition is close to that of many flow basalts and because they exhibit, in part, relict phenocrysts of plagioclase set in matrices with much finer-grained and ophitic(?) textures.

The accordant contacts between the metabasalts and the intercalated and overlying metasedimentary rocks indicate that the former are either flows or sills. The association of interlayered metabasalt, conglomerate, and schists that may be metamorphosed tuff suggests that the metabasalt formed as flows. The entire Moppin series, from Aveta Creek to Buckhorn Gulch, may well be part of a locally thick pile of volcanic rocks that thins abruptly to the south and southwest. The relationship of the Moppin metavolcanic series to the amphibolite member of the Kiawa Mountain formation is discussed below.

KIAWA MOUNTAIN FORMATION

Definition

The Ortega quartzite of Just (1937, p. 43) included the quartzite that underlies Kiawa Mountain, Quartzite Peak, and parts of the upper Vallecitos Valley. This unit of quartzite has been found in the present study to be the youngest Precambrian metasedimentary rock in the area. The quartzite including and overlying the Big Rock conglomerate and overlying the Moppin metavolcanic series from Cleveland Gulch to Jawbone Mountain is here defined as the Kiawa Mountain formation, after the excellent exposures on Kiawa Mountain. The formation can be divided into five members: The Big Rock conglomerate, the Jawbone conglomerate, an amphibolite member, and two quartzite members.

Big Rock Conglomerate Member

Definition and distribution. The Big Rock conglomerate extends along its strike from the NE¼ sec. 20, T. 27 N., R. 8 E., to the Big Rock syncline (0.2 mile northwest of Big Rock) and north to the Poso anticline. A poorly exposed layer of similar conglomerate was found east-northeast of Poso Spring in the SE¼ SW¼ sec. 23, T. 27 N., R. 8 E. The Big Rock conglomerate overlies the Ortega quartzite, and in this locality is the basal member of the Kiawa Mountain quartzite.

Lithology. The Big Rock conglomerate consists of dark to medium-gray quartz-pebble conglomerate with intercalated slightly feldspathic quartzite. Most of the pebbles are light-gray quartz; a few are red and black quartz, and probably were originally ferruginous chert or jasper. Crossbedding is common, and current bedding is commonly well defined by laminae of iron oxide as well as by layers of differing grain size. The pebbles are mostly in the 1- to 5-inch range, are well to poorly sorted and highly rounded, and are now of elongate shapes that give a marked down-dip lineation.

The easternmost part of the Big Rock conglomerate, associated with the Petaca schist, has been slightly muscovitized.

On the flanks of folds, the schistosity of the conglomerate is parallel to the bedding; but at fold axes the two planar features are markedly discordant. Schistosity is especially well developed in the central half of the W½ sec. 22, T. 27 N., R. 8 E.

The Big Rock conglomerate member contains a higher percentage of quartzite and pebbly quartzite in its eastern portion than it does in its western portion.

Thickness. The westernmost part of the Big Rock conglomerate is about 50 feet thick. In the Big Rock and Poso folds the thickness could not be directly measured, but it is estimated to be between 100 and 200 feet, to include many more intercalated finer grained beds, and to be slightly less conglomeratic than the rocks to the west. The member appears to pinch out east of Poso Spring.

Jawbone Conglomerate Member

Definition and distribution. The Jawbone conglomerate underlies Jawbone Mountain, in the extreme northwest corner of the quadrangle. This member lies between the Moppin series and the Kiawa Mountain quartzite from about 1½ miles west-northwest of Burned Mountain northwestward for several miles to the west boundary of the quadrangle.

Lithology. The Jawbone conglomerate is mostly quartz-pebble conglomerate with varying amounts of interlayered gray quartzite, and resembles the quartz-pebble beds in the Ortega and Kiawa Mountain quartzites. The pebbles are of light-gray, red, and black quartz, range from 4 to 25 mm in size, are moderately well sorted, very well rounded, and mostly ovoid in shape. Granules and coarse grains of sand are about as abundant as the pebbles. The matrix of the conglomerate is mostly blue-gray, vitreous, fine-grained, kyanitic quartzite, with hematitic layers. Crossbedding is widespread. Some of the conglomerate is dark gray and slightly micaceous. These rocks show a clearly defined cleavage that commonly makes moderate to high angles with the bedding.

Thickness. The thickness of the Jawbone conglomerate is about 500 feet in the canyon 1 mile southwest of Hopewell. From there southeastward, this member appears to finger out into the Kiawa Mountain quartzite member. On Jawbone Mountain, the conglomerate is at least 500 feet thick, and its consistent northwesterly dips and fairly consistent crossbedding suggest a thickness of perhaps 2,000 feet.

Amphibolite Member

Definition and distribution. The amphibolite member of the Kiawa Mountain formation includes the series of amphibolite layers and intercalated quartzite beds exposed along the Rio Vallecitos near its intersection with Escondida Creek (sec. 4, T. 27 N., R. 7 E.); along La Jara Creek one-fourth to one-half mile above its mouth (secs. 3 and 10, T. 27 N., R. 7 E.); in Canada del Oso (sec. 12, T. 27 N., R. 7 E.); on La Jarita Mesa along upper Kiawa Canyon; and on the lower east and northeast slopes of Kiawa Mountain to Spring Creek Canyon.

Lithology. The amphibolite layers exposed in Canada del Oso, La Jara Canyon, and Vallecitos Canyon near Escondida Creek are all similar in mineralogy and texture. The compositional limits of the amphibolite in this area are approximately 30-40 percent chlorite, 12-30 percent oligoclase, 20-25 percent epidote, 5-20 percent hornblende, 2-4 percent magnetite-ilmenite, and small amounts of quartz, biotite, saussurite, apatite, and leucoxene. The content of hornblende is higher, and the contents of oligoclase and chlorite are lower, in the Canada del Oso amphibolite than in the amphibolites of La Jara Canyon and Vallecitos Canyon. Biotite was seen only in the rock from La Jara Canyon. The Canada del Oso rock also is more distinctly lineate and coarser grained, although all the rock is fine grained, with a maximum average grain size of about 0.5 mm.

In the Canada del Oso amphibolite many magnetite poikiloblasts, as much as 1 mm in diameter, show a crude prismatic shape, which suggests, in a rock of this composition, that they were derived from pyroxene.

Epidote occurs as single porphyroblasts and clusters of grains as large as 0.5 mm in diameter. It is pleochroic with X = colorless, Y = greenish yellow, Z = pale yellow, and absorption X < Z < Y. The birefringence is 0.027. The Fe^{3+} :Al ratio therefore may be about 1:4.

The hornblende is similar to that in specimen 36-B-37 described above. The textural relations of euhedral and subhedral prisms of hornblende against smaller interstitial anhedral grains of chlorite suggest (but, of course, do not prove) that the hornblende formed later than the chlorite.

The amphibolite five-eighths of a mile south and east of Kiawa Lake, forms a part of the layer that outlines the Kiawa syncline, and consists of 21 percent hornblende, 35 percent oligoclase, 22 percent chlorite, 2 percent quartz, 7 percent epidote, and 12 percent magnetite-ilmenite. The amphibolite 1 mile east of this exposure lacks chlorite and epidote.

Amphibolite at the roadcut on the north side of Spring Creek Canyon, just west of the contact of the Cordito member, consists of 65 percent poikiloblastic porphyroblasts of hornblende as much as 8 mm long, 30 percent calcic oligoclase, and 5 percent ragged biotite, magnetite-ilmenite, and apatite. The sheets of biotite partly finger into and cut across prisms of hornblende; they apparently have replaced this mineral. The oligoclase shows zoning in which the cores are more sodic than the rims, and some has been largely altered to hydromuscovite(?) (sericite with low birefringence). Both alteration to biotite and hydro-muscovite(?) may well be associated with dikes of pegmatite several feet thick along this side of the canyon.

The lithology of the amphibolite layer that delineates the Kiawa syncline is summarized as modes of 11 specimens, 36-D-53 through 36-D-63, in Table 4. The three layers of amphibolite east and south of Kiawa Lake overlie this more extensive layer and are very similar to it.

These amphibolites are dark green to black in hand specimen, uniformly fine grained, and commonly lineate. They are exposed in irregular, jagged, toothlike outcrops in which most fractures are controlled by the steeply plunging lineation and by a fault, steeply dipping planar schistosity. A few thin, contorted quartz veins and several pegmatite dikelets a few inches thick penetrate the amphibolite.

The 11 thinsections, representing specimens 36-D-53 to 36-D-63, of typical amphibolite from the extensive layer are uniform in texture and

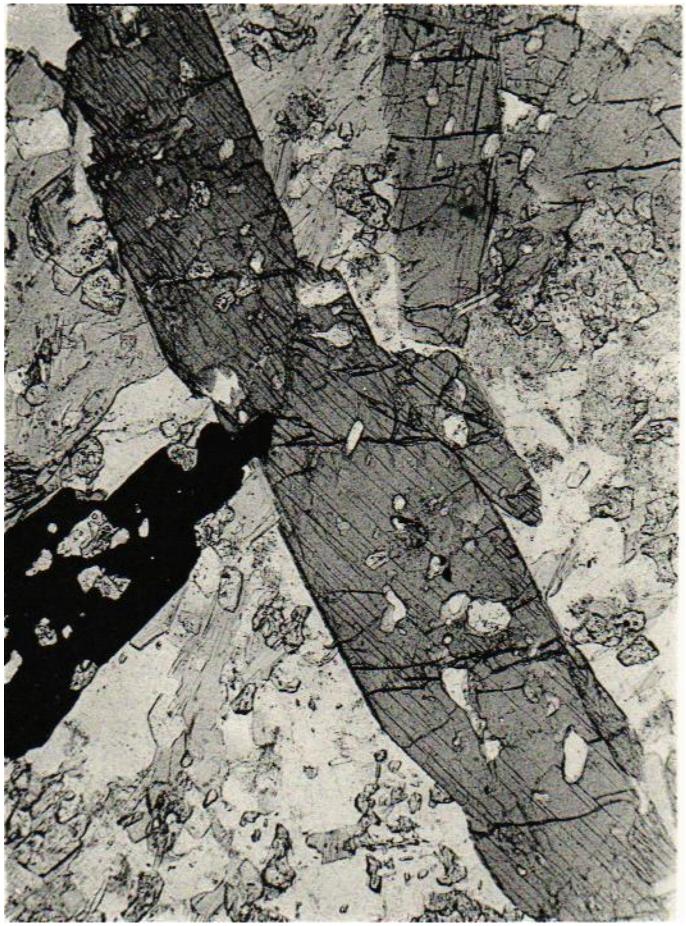

Plate 12
PHOTOMICROGRAPH OF AMPHIBOLITE

Large subhedral and euhedral grains of hornblende and an anhedral grain of magnetite-ilmenite set in a matrix of irregular grains of chlorite, clear to cloudy andesine, equant to euhedral epidote, and apatite. Specimen 36-D-2. Plane-polarized light. X 70.

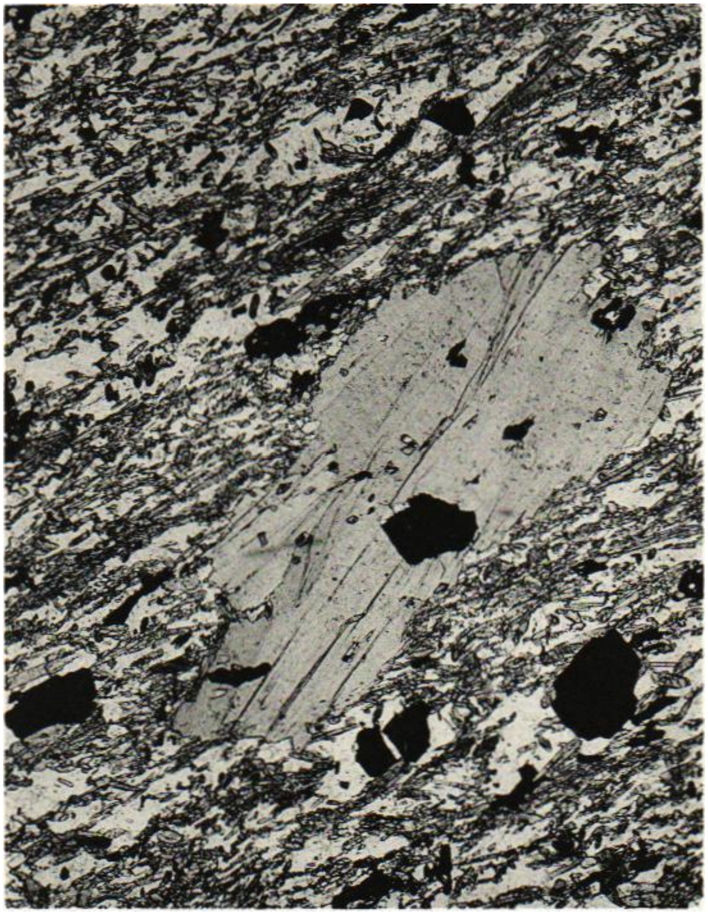

Plate 13

PHOTOMICROGRAPH OF CHLORITIZED AMPHIBOLITE

A knot of chlorite, containing several smaller grains of ' magnetite-ilmenite, in a matrix of hornblende, andesine, and magnetite-ilmenite. The knot transects the lineation of the amphibolite, which extends diagonally across the photo. Plane-polarized light. X 30.

composition. The compositional limits, as given in Table 4, are 44-75 percent hornblende, 13-32 percent plagioclase, 2-12 percent chlorite, 0-3 percent epidote, 3-6 percent magnetite-ilmenite, 2-30 percent quartz and about 1 percent apatite.

TABLE 4. VOLUMETRIC MODES OF KIAWA MOUNTAIN AMPHIBOLITE MEMBER

SPECIMEN NO.	HORN-BLENDE	PLACID-CLASE	CHLO - RITE	QUARTZ	MAGNETITE AND ILMENITE	En-DOTE	APA-TITE
36-D-2*	54	21	5	5	4	10	1
36-D-53	70	20	4	2	4	—	tr
36-D-54	54	20	2	20	3	—	1
36-D-55	61	25	3	6	4	—	1
36-D-56	63	13	12	7	4	—	1
36-D-57	70	17	2	5	4	—	2
36-D-58	62	17	4	9	4	3	1
36-D-59	44	28	7	14	5	1	1
36-D-60*	55	32	6	5	6	tr	1
36-D-61	75	13	5	3	3	—	1
36-D-62	60	20	5	30	4	—	1
36-D-63	59	28	4	2	3	3	1

* Modes of 36-D-2 and 36-D-60 were determined by counting 1,000 points; the remainder *were* made with a Leitz integrating stage.

Nearly all the prismatic and needle-shaped hornblende crystals have their c-axes within 5 degrees of one another, giving a marked and characteristic down-dip lineation to the rock. The subhedral to euhedral grains are 0.75-1 mm in average length, and their widths and thicknesses are about 0.12 and 0.08 mm, respectively. The optical properties of the hornblende are: y = 1.679 ± 0.002, X < Y < Z, X = straw yellow to pale yellowish green, Y = green, Z = blue green, Z A C = 22°, and (—)2V = 65°-80°. The composition of hornblende with these optical properties would be approximately femaghastingsite (Winchell and Winchell, 1951, fig. 325, p. 434, and Billings, 1928, p. 292).

Plagioclase is interstitial to the hornblende, and forms anhedral grains, 0.12-0.25 mm average diameter, and lenticular, mosaiclike aggregates. Most grains are slightly elongate parallel to the hornblende lineation. The plagioclase ranges in composition from calcic oligoclase to sodic andesine. Moderate to faint zoning, with sodic cores, is common. Almost all the plagioclase is clear; alteration effects are very slight in all the specimens that were observed.

Chlorite grains, 0.12-5 mm across, form knots or lenticular aggregates as shown in Plate 13. The longest dimensions of the knots are parallel to the lineation of the amphibolite. In contrast to the amphibolites of lower metamorphic grade to the north and west, in which the chlorite is interstitial to and xenoblastic toward hornblende, the knots of chlorite in this amphibolite layer either sharply truncate euhedral grains of hornblende or they finger into hornblende grains at rough, irregular interfaces. The knots range in length from 0.12 to 15 mm, with an average length of about 2 mm. The c-axes of the grains are commonly normal to the long axes of the lenticles. Some optical properties of the chlorite are $\beta = 1.601$, $\gamma - \alpha = 0.004$, (+) 2V = ca. 10°, X = Y > Z, X = pale green, Y = pale green, Z = very pale yellow green. The data most closely correspond to those cited for a variety of chlorite called rumpfite by Winchell and Winchell (1951, fig. 261, p. 383).

Epidote, when it occurs, is present only in small amounts in this amphibolite layer. It occurs as equant or irregular anhedral grains, 0.08 mm in average size, and commonly is associated with plagioclase rather than hornblende.

The magnetite-ilmenite grains are variable in shape; many are irregularly xenoblastic and skeletal, some are tabular, octahedral, or roughly elliptical, and a few suggest crude cubic forms. Almost all of the tabular and elongate grains are oriented with their longest axes parallel to the lineation and schistosity. The diameter of the magnetiteilmenite often ranges from 0.12 to 0.5 mm in a single section, but few larger or smaller grains were seen.

Apatite is the only other common accessory mineral in the amphibolite. It forms prisms, most of which are less than 0.06 mm long, and equant anhedra, many of which are 0.06 to 0.1 mm in size.

Relict amygdules of epidote, hornblende, and quartz were observed in the amphibolite three-fourths of a mile northeast of Poso Lake, in the NE¼NE¼ sec. 15, T. 27 N., R. 8 E. Some are composed of epidote and hornblende, others of quartz with a little epidote. Those of the epidote and hornblende are spheroidal, and have sharp boundaries. The others are roughly ovoid to spherical, and have somewhat irregular boundaries. Both types range from 1 to 10 mm in diameter. The matrix of this rock is approximately 75 percent epidote, 15 percent hornblende, 9 percent plagioclase, and 1 percent magnetite-ilmenite. The original rock is assumed to have been a basalt flow that contained amygdules of quartz, calcite, and perhaps epidote.

The only outcrop of layered amphibolite in the Kiawa syncline is on the south side of Spring Creek Canyon. Light-colored layers, with a hornblende:plagioclase ratio of about 1:3, alternate with dark layers, in which the ratio is about 3:1. The hornblende prisms in the dark laminae are 2 mm in average length, whereas those in the light-colored laminae are 0.25 mm in average length. The layering is probably a relict feature of an original basaltic tuff.

The quartzite beds interlayered with the three amphibolite layers along the upper reaches of Kiawa Canyon are light gray, vitreous, and partly pebbly. The quartzite gradually changes eastward to a micaceous, schistose, intensely folded rock. This increase in muscovite content is a fringe effect of the metasomatized aureole that surrounds the La Jarita pegmatites.

The quartzite beds of the amphibolite member in La Jara and Vallecitos Canyons are mainly light gray, vitreous, coarse grained, slightly pebbly, and massively bedded. Some beds are slightly micaceous.

Thickness. The amphibolite member with its associated quartzite zones varies widely in thickness. To the east and southeast of Kiawa Mountain the member consists of one layer of amphibolite, whose thickness ranges from 35 to 50 feet. Much of this layer is duplicated three or more times by isoclinal folding. South and east of Kiawa Lake the member comprises three layers of amphibolite and two layers of quartzite; as many minor folds are present, its thickness is much less than the 2,400 feet that the outcrop breadth indicates. The single layer of amphibolite that crops out in Caflada del Oso is approximately 50 feet thick. The amphibolite member is about 1,900 feet thick, of which 300 feet are amphibolite. The member is 2,000 feet thick in Vallecitos Canyon; 7 amphibolite layers are present, and their aggregate thickness is about 550 feet.

Origin. The composition of the amphibolite layers is similar to that of the Moppin metabasalt to the north. Analyses of two specimens, one-fourth of a mile to 2 miles east of Kiawa Lake, and from the extensive layer that outlines the Kiawa syncline, are given in Table 3. The analyses closely resemble the average composition of Deccan basalt.

All the observed contacts between the amphibolite layers and the enclosing quartzite beds are concordant. Thus, the original basalt layers were emplaced as flows or sills. The relatively uniform thickness of these layers over distances of several miles suggests that they are flows. The relict amygdules found in the epidotic amphibolite three-fourths of a mile northeast of Poso Lake are also suggestive of an origin as a flow. However, direct evidence of the origin of these metabasalt layers is lacking.

Relation to Moppin metavolcanic series. The relation of the amphibolite member of the Kiawa Mountain formation to the Moppin meta-volcanic series is an important stratigraphic problem in this area. Three relatively thin layers of amphibolite are exposed in Spring Creek Canyon. Tertiary conglomerate covers the north rim of that canyon. To the north of this conglomerate are amphibolites 750 feet in outcrop breadth; pebbly, micaceous quartzite 1,300 feet in outcrop breadth; and the amphibolite and schist of the Moppin series. The relations of the amphibolite layers of Spring Creek Canyon to those both south and north of the pebbly, micaceous quartzite are critical, for they may indicate a fold or an abrupt stratigraphic pinchout.

The amphibolite exposed in Spring Creek Canyon may be strati-graphically connected to the amphibolite exposed one-half mile north of the rim of the canyon by the nose of an anticlinal fold that may lie beneath the Tertiary rocks. Similarly, the latter unit of amphibolite actually may be joined to the metavolcanics of Aveta Creek by the nose of a synclinal fold that may be concealed beneath the Tertiary rocks to the east. Thus, the pebbly, micaceous quartzite should show a synclinal structure. Drag folds were not seen in this quartzite; so the hypothesis cannot be directly substantiated.

Another possibility is that the amphibolite member of the Kiawa Mountain formation extends up Cleveland Gulch under the Tertiary conglomerate to join the metavolcanic rocks exposed in the upper parts of the gulch. This would necessitate that the pebbly, micaceous quartzite abruptly pinch out westward under the Tertiary conglomerate, and that the lower quartzite member of the Kiawa Mountain formation pinch out northwestward between the amphibolite layers in Spring Creek Canyon and the amphibolite mass that lies one-half mile north of the rim of the canyon. The writer prefers the fold hypothesis to the pinchout hypothesis, but both would be adequate to explain the distribution of the rocks.

Upper and Lower Quartzite Members

Definition. The lower quartzite member of the Kiawa Mountain formation is defined to include strata overlying the Big Rock conglomerate member and underlying the amphibolite member. The upper quartzite member overlies the amphibolite member; its top is not exposed in Las Tablas quadrangle.

Distribution. The lower quartzite member is exposed in Spring Creek Canyon, Tusas and lower Kiawa Canyons, on La Jarita Mesa, in Canada del Oso, and at the mouth of La Jara Canyon. The upper quartzite member underlies most of Kiawa Mountain, parts of del Oso, La Jara, Vallecitos, and Spring Creek Canyons, Quartzite Peak, and parts of the upper Vallecitos Valley.

Lithology. The part of the lower quartzite member that extends from the longitude of Kiawa Lake to La Jara Canyon is similar in lithology to the light-gray, commonly vitreous, massive quartzite that characterizes the upper member of the same formation. The part of this member that lies east of the longitude of Kiawa Lake on La Jarita Mesa, and along Spring Creek, Tusas, and Kiawa Canyons is muscovite bearing. Some of the muscovite is of normal sedimentary origin; some of it, in amounts increasing south- and southeastward, probably is of hydrothermal origin and is genetically related to the granitic pegmatites. The quartzite from the mouth of Spring Creek to a point about 1 mile west of Las Tablas is transitional into the muscovitized quartzite that is described farther on, in connection with the Petaca schist.

Parts of the quartzite members of the Kiawa Mountain formation that lie directly above and beneath the extensive amphibolite layer that delineates the Kiawa syncline contains varying amounts of hornblende, plagioclase, epidote, and chlorite. These two layers of mafic quartzite, found immediately adjacent to the amphibolite only, range in thickness from a small fraction of an inch to 20 feet. Exposures are so poor, however, that these figures should be regarded as approximate.

The quartz in this rock is of the same grain size, 0.06 to 0.25 mm, as that in the clean quartzite, and it has the same granulose mosaic texture. The plagioclase, which is calcic oligoclase, and the hornblende, epidote, and chlorite, are all optically similar to the corresponding minerals in the amphibolite. In a specimen of the mafic quartzite that overlies the amphibolite on the north side of Kiawa Canyon, the hornblende occurs as well-lineated, slightly poikiloblastic prisms 1-30 mm long, oligoclase is in equant anhedra, 0.06-0.5 mm across, and the chlorite forms elongate, well-alined knots 1-10 mm in length. In the layer underlying the amphibolite at the nose of the Kiawa syncline, the hornblende is concentrated in dark layers in the form of flat prisms 0.25-2 mm in length; their long axes lie parallel to the bedding. Epidote is evenly disseminated in this rock. A specimen from three-eighths of a mile northwest of the Kiawa mine, from the altered quartzite overlying the amphibolite, contains faintly alined prisms of poikiloblastic hornblende, which have been slightly altered to biotite; the biotite, in turn, has been slightly chloritized.

The upper quartzite member of the Kiawa Mountain formation is dense, vitreous, light bluish-gray quartzite that contains irregularly distributed pebbly layers. This rock is similar to the Ortega quartzite exposed west of the Rio Vallecitos; indeed, these two quartzites are essentially indistinguishable in outcrop.

In thinsection the typical quartzite of this member shows scattered granules or small pebbles of quartz set in a much finer grained equigranular mosaic in which individual grains of quartz are about 0.1 mm in average diameter. Small amounts of hematite, muscovite, and kyanite are commonly present. Kyanite is absent if the muscovite content is more than 10 percent. Similarly, muscovite is absent if the kyanite content is more than about 5 percent.

Kyanite is present in the upper quartzite member of the Kiawa Mountain formation as in the Ortega quartzite, i.e., along bedding planes; with hematite in original sedimentary laminae; and in veinlets with quartz. Seen in thinsection most of the light-colored vitreous quartzite contains several percent of kyanite, randomly distributed among the quartz grains and pebbles. Kyanite is associated with almost all of the dark laminae of hematite in the quartzite. Most of them are less than 3 mm thick; kyanite occurs as grains, not over 0.5 mm across, disseminated in the laminae, and in thin, kyanite-rich laminae either adjacent to the hematitic laminae or separated from them by a few millimeters of much less kyanitic quartzite.

The hematite-kyanite laminae in the quartzite at Quartzite Peak are generally coarser grained, and contain bladed crystals of kyanite as much as 10 mm long. The bounding surfaces of these laminae are jagged in detail, pinch abruptly, swell, and even terminate here and there; and their kyanite content varies from almost nothing to more than 50 percent of the laminae. These features suggest postdepositional transfer of both the iron oxide and aluminous material.

The quartz-kyanite veins in this member are tabular, lenticular, or irregular in shape, and are roughly parallel to the bedding. The largest vein observed is about 4 feet long and 1 foot thick. The kyanite in these veins is commonly colorless or light orange, forms bladelike crystals 1 or 2 inches in maximum length, and forms partial or complete rosettes. The kyanite content generally is between 25 and 50 percent of the rock, and local concentrations of interlocked prisms are present in many veins.

Intensely folded hematite-kyanite laminae, 10-20 mm thick, are present in the low knob of quartzite about seven-eighths of a mile northwest of Kiawa Lake. The quartz, hematite, and kyanite in the axial regions of the tightest folds have been partly replaced by small irregular masses of quartz.

Thickness. The thickness of the lower quartzite member of the Kiawa Mountain formation is not determinable, owing to intense minor folding. This unit may be several hundred feet thick.

The upper quartzite member of the Kiawa Mountain formation is estimated to be 5,000-10,000 feet thick. No accurate estimate of the thickness can be made, owing to the absence of marker beds or clearly delineated folds in this member. The same quartzite member is exposed several miles northwest of Las Tablas quadrangle, in the Brazos River canyon, so that the total thickness may be markedly greater than the thickness of that part of the member exposed in the Las Tablas quadrangle itself. Top-and-bottom relations, as shown by crossbedding and drag folds, are not consistent in this quartzite, implying the presence of numerous minor folds.

Origin. The origin of the Kiawa Mountain quartzite is discussed below, in connection with that of the other members of the Kiawa Mountain formation and the Ortega quartzite.

ORIGIN OF THE ORTEGA QUARTZITE AND KIAWA MOUNTAIN FORMATION

Krynine (1941) distinguishes first-cycle and second-cycle orthoquartzites. The first-cycle types "generally follow prolonged and intense chemical decay in peneplaned regions," and the second-cycle quartzites are formed by reworking of quartzose sediments. The quartzites in Las Tablas quadrangle have been too severely recrystallized to permit determination of whether they are first- or second-cycle type, or both. Neither their great thickness nor their contained minor minerals definitely suggest one origin rather than the other. The black and red quartz pebbles, probably recrystallized jasper, suggest at least partial derivation from preexisting sedimentary rocks.

The author was unable to determine whether the hematite in the quartzites was originally deposited as such, or was derived later from magnetite or other minerals. Similarly, no hint was found as to the nature of the mineral(s) that were metamorphosed to kyanite. The kyanite may have been developed from kaolinite, the bauxite minerals, or from other sources. The close association of kyanite with the laminae of hematite suggests that some bauxite may have been present prior to metamorphism.

The quartzose sediments are inferred to be near-shore, shallow-water deposits, perhaps in part subaerial, and their origin may have been somewhat similar to that of the Cambrian coarse-sand facies of the Grand Canyon area as inferred by McKee (1945, p. 47-51). The Big Rock conglomerate probably was a beach gravel. The similarity of the Ortega quartzite to the Kiawa Mountain quartzite, and the general homogeneity of these two units, suggest that the conditions of sedimentation remained similar throughout the period of their deposition, with the rate of subsidence of the material already deposited very nearly equal to the rate of sedimentation of new material.

Many investigators in the field of sedimentary tectonics have agreed that quartzose sandstones are typically thin (less than 1,000 feet), and commonly have formed as transgressive sea deposits (Krynine, 1943, p. 5; Dapples, 1947, p. 93; Pettijohn, 1949, p. 454-455; Krumbein and Sloss, 1951, p. 360; Kay, 1951, p. 86-88). The stable shelf association of Krumbein and Sloss (1951, p. 360) and the foreland facies of Pettijohn (1949, p. 454) include quartzose sandstones. Conversely, these same geologists agree that graywacke and shales of graywacke composition are typical of thick basin fillings, with associated volcanic rocks in eugeosynclines and without igneous rocks in miogeosynclines (Pettijohn, 1949, p. 447; Kay, 1951, p. 86; Krumbein and Sloss, 1951, p. 367). Krumbein and Sloss (p. 367) state that abnormally thickened shelf sandstone may occur in miogeosynclinal associations. The Precambrian strata of Las Tablas quadrangle could be reconciled with these tectonic schemes by changing the definition of eugeosyncline to a thick basin filling of either graywacke or quartzose sandstone with intercalated igneous rocks, or by changing the definition of a miogeosyncline to include igneous rocks. With such modified definitions, the sedimentary rocks of Las Tablas area can be classified as occupying a eugeosyncline of the quartzose sandstone type, or as occupying a miogeosyncline of abnormally thickened shelf sandstone which includes volcanic rocks. The writer prefers to classify these rocks with the miogeosynclinal type, but both of the altered definitions of geosynclines are of doubtful value. The tectonic environment in which such thick sections of quartzose sandstone are accumulated deserves much more study.

PETACA SCHIST

Definition and Distribution

The quartz-muscovite schist that underlies much of La Jarita Mesa was believed by Just (1937, p. 43), to be a muscovitized variant of the Ortega quartzite. He named this variant the Petaca schist. The same name is herein used for this rock, which represents altered parts of the Ortega quartzite, part of the lower Kiawa Mountain formation, and many layers of Burned Mountain metarhyolite. The Petaca schist was mapped as such in the field, where bleached and muscovitized metarhyolite, muscovitized quartzite, and slightly feldspathic, pink to bleached muscovite are essentially indistinguishable from each other. Masses of the original rocks are recognizable in many outcrops, but are enclosed in the bleached, highly altered rock, and cannot be traced from one point to another; thus accurate mapping of metarhyolite and quartzite within the metasomatized area is impossible on a scale as small as 1:31,680.

Lithology

Typical Petaca schist is a light-gray, schistose, slabby to massive, slightly muscovite quartzite. Crossbedding and iron-oxide-rich laminae, typical of the unaltered quartzite, are rare. In thinsection the quartz shows a granulose mosaic texture, with average grain sizes from one-eighth to one-fourth mm. Muscovite occurs as well-alined platy to tabular grains, about 0.25 mm across, and generally parallel to the axial planes of the minor folds. The muscovite content of the quartzite varies from only a few percent to more than 40 percent. Variations from the lower amount to about 15 percent are common one-quarter mile or more from granitic pegmatites. Accessory minerals include biotite, epidote, garnet, plagioclase, microcline, magnetite-ilmenite, and hematite. Biotitic and biotitic-garnetiferous varieties of the muscovitic quartzite are common. They appear to be irregularly distributed in the Petaca schist, generally within a few hundred yards of exposed pegmatites, or large quartz veins.

In feldspathic quartzite near Poso Spring (pl. 3), pebbles and granules of quartz and microcline are common, but where well-formed pebbles are absent it is very difficult and often impossible, with hand lens only, to distinguish the rock from metarhyolite porphyry, especially where both rocks have been bleached from their originally pink hues to tints of light gray or greenish gray.

The more muscovitic varieties of quartzite, i.e., those with 15 to more than 40 percent of muscovite, are closely associated with exposed bodies of pegmatite. The wall-rock alteration of quartzite at the pegmatite boundaries has been described in some detail by Jahns (1946, p. 52-54). Figure 2 is a diagrammatic sketch after that given by Jahns (1946, p. 52), showing the typical relations of the pegmatites to the muscovitized and feldspathized quartzite. This alteration is discussed below, on p. 62ff and 87ff.

Two exposures of gray quartz-pebble conglomerate were found in the Petaca schist. A layer of 30 feet outcrop breadth appears in the SW¼SW¼ sec. 27, T. 27 N., R. 8 E., and a 20-foot layer appears in the N¼S½ sec. 35, T. 27 N., R. 8 E. These two conglomerates are quite similar. Both contain flat to ovoid pebbles of quartz, many of which show a marked lineation directed almost downdip. Some of the flat pebbles have been folded into canoe or S shapes, with fold axes parallel to the axes of drag folds. A small percentage of muscovite is present. Some is wrapped around drag folds; some is parallel to the axial planes. The latter type is thought to be of metasomatic origin.

Amphibolite, or metasomatized equivalents of amphibolite, occur in the Petaca schist near Poso Spring; in the saddle immediately west of

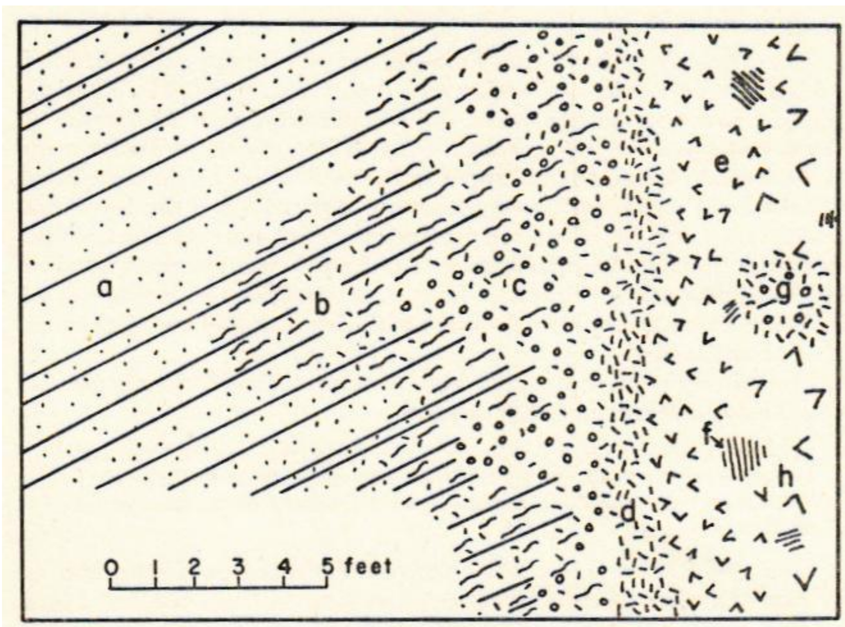

Figure 2.

DIAGRAMMATIC SKETCH SHOWING TYPICAL GRADATIONAL CONTACT RELATIONS BETWEEN PEGMATITE AND QUARTZITE COUNTRY ROCK (AFTER JAHNS, 1946).

a. Slabby micaceous quartzite. b. Quartzite with muscovite-rich partings and disseminated muscovite flakes. c. Muscovite-impregnated quartzite, often with metacrysts of microcline and albite-oligoclase. d. Fine-grained mica-rich contact zone of pegmatite. e. Medium-grained microcline-quartz-albite-oligoclase-albite-muscovite pegmatite. f. Large book of muscovite. g. Inclusion of altered quartzite (lithologically similar to that in unit c). h. Coarse-grained microcline-quartz pegmatite.

Twin Peaks; in Apache Canyon (sec. 2, T. 26 N., R. 8 E.); and on the south side of Cañada de la Jarita about 1 mile east of Big Rock.

Four intensely folded layers of biotite-epidote-quartz-oligoclase schist are exposed just southeast of Poso Spring (pl. 3). This schist probably was originally an amphibolite. Its composition differs from that of any normal sedimentary or igneous rock, and it is present in a sequence that contains many layers of amphibolite. However, no relict minerals or structures characteristic of amphibolite were seen in this rock. Its genesis is discussed farther on, in connection with pegmatitic-hydrothermal metamorphism.

A 5-foot layer of muscovite-biotite-garnet-quartz skarn is exposed one-quarter mile east of Poso Spring. The chemical analysis and genesis of this rock also are considered below.

TERTIARY ROCKS

PREVIOUS WORK

The Tertiary rocks of the Tusas Valley were mapped by Atwood and Mather on a scale of 10 miles to 1 inch (1932, p. 92-101). They divided the strata on the east side of the Tusas Valley, from the vicinity of Tusas to Petaca, into a lower unit, 500-600 feet thick, correlative with the Conejos andesite, and an upper unit, 300-500 feet thick, correlative with the Los Pinos gravel, a conglomerate exposed along the Los Pinos River. The Treasure Mountain formation, which overlies the Conejos andesite in the Conejos quadrangle of Colorado, was not recognized in the Rio Tusas area. The basalt flows that underlie the mesas east of the lower Tusas Valley were grouped with the Hinsdale and post-Hinsdale volcanic series.

Preliminary work in the area from Abiquiu to La Madera indicated to Atwood and Mather (1932, p. 98) that

> . . . only a small part, approximately the upper third, of the Santa Fe formation [of the Rio Grande Valley] is really of Los Pinos age. The rest is believed to be principally the southeastward extension of the Conejos formation.

Cross and Larsen (1935, p. 94-100) regarded the Los Pinos gravel as the basal member of the Hinsdale formation. As such, the Los Pinos member was thought to have been deposited on the San Juan peneplain, a finding in agreement with the conclusions of Atwood and Mather. The Los Pinos gravel was estimated to be of late Pliocene age.

On the basis of work in the Abiquiu area, Smith (1938, p. 944-958) was able to separate the Santa Fe formation from an unconformably underlying tuff and conglomerate, which he named the Abiquiu tuff. He correlated the Abiquiu tuff with tuff of the Conejos age, and the lower part of the Santa Fe formation with the upper part of the Conejos formation. Smith believed that the Los Pinos gravel is about 50 feet thick in the Abiquiu area, that it unconformably overlies the Santa Fe formation, and that it is overlain by basalt flows (1938, p. 957-958).

The relationship of the Tertiary rocks of the San Juan Mountains of Colorado to those of the Rio Grande Valley of New Mexico was studied by Butler (1946). He mapped in detail the Tertiary rocks from the Colorado-New Mexico boundary to the Tusas Valley south of Petaca, and mapped in reconnaissance the Tertiary rocks in a part of the Vallecitos Valley. His final map is on a scale of 1 inch to the mile.

Butler (1946, p. 5-6) concluded:

> (1) that the Los Pinos formation, as here redefined, is largely equivalent to the Abiquiu tuff of Smith; (2) that a little of the upper part of the formation is equivalent to part, probably the lower part, of the Santa Fe formation; (3) that some of the basalt previously included in the "Hinsdale formation" is, instead, a member of the Los Pinos; (4) that the Los Pinos formation as well as the Santa Fe formation is separated from the Hinsdale volcanic series by an unconformity, which may correspond to the San Juan peneplain. . . .

The stratigraphic nomenclature used in this report is generally similar to that proposed by Butler.

GENERAL FEATURES

The Tertiary rocks of Las Tablas quadrangle are herein divided into six formations, largely following the work of Butler (1946). These are the Conejos, Treasure Mountain, Ritito (new name), and Los Pinos formations, and the Cisneros and Dorado basalts. These rocks lie along and east of the Rio Tusas, and partly along and west of the Rio Vallecitos. Smaller, unconnected patches of Tertiary rocks crop out on La Jarita Mesa and in the area between Spring Creek and Vallecitos Creek.

These six formations consist largely of graywacke and arkosic, commonly tuffaceous, sandstone, with interlayered conglomerate, tuff, and welded tuff, as well as flows of basalt, rhyolite, latite, and breccia. They are cut by several masses of intrusive andesite porphyry. These rocks, which form a veneer over the Precambrian terrane, are of variable thickness, with a maximum of about 1,500 feet.

The stratigraphy of the Tertiary rocks of Las Tablas quadrangle is schematically shown in Table 5.

CONEJOS(?) AND TREASURE MOUNTAIN FORMATIONS

Summary Statement

The Conejos andesite was named by Cross and Larsen (1935, p. 69) after exposures in the canyon of Conejos River, Conejos County, Colorado. Butler (1946, p. 22-23) changed the name to "Conejos formation" because this unit contains a variety of volcanic rocks, as well as much fluviatile material. He tentatively correlated (1946, pp. 28-30) the Tertiary rocks of the upper Tusas Valley that underlie the Treasure Mountain formation with the lithologically different Conejos rocks to the north. Like Butler, the writer uses the name Conejos with a query, to denote the tentative nature of this correlation.

TABLE 5. TERTIARY ROCKS OF LAS TABLAS QUADRANGLE

Series	Formation	Character	Thickness (feet)
Pliocene(?)	Dorado basalt	Flows of quartz basalt; only east of lower Tusas Valley	40-100
	Unconformity		
	Cisneros basalt	Disconnected remnants of basalt flows; in eastern part of area only	10-30
	Unconformity		
	Los Pinos formation		
	Cordito member	Rhyolite-fragment conglomerate, tuff, and sandstone, with minor rhyolite flows; most extensive Tertiary rock unit in this area	600*
	Local unconformity		
	Jarita basalt	Flows of basalt; widely separated, disconnected single and multiple flows	50*
	Biscara-Esquibel member	Conglomerate with fragments of andesite to latite, tuff, and sandstone	1,000*
Miocene(?)	Biscara member	Conglomerate of andesite to latite fragments, tuff, sandstone, andesite flow breccia and dikes	700*
	Unconformity		
	Treasure Mountain and Conejos(?) formations (in Tusas Valley)	Conglomerate of Precambrian rock fragments, sandstone, tuff, and felsite-fragment conglomerate, with interlayered rhyolitic welded tuff, 10-18 feet thick; only along east side of upper Tusas Valley	150-400
	Ritito conglomerate (in Vallecitos Valley)	Conglomerate of Precambrian rock fragments; only in the lower Vallecitos Valley and tributary canyons; correlative with either Conejos(?) or Biscara conglomerates	400*

* Maximum.

The Treasure Mountain formation was named by Cross and Larsen (1935, p. 68) from exposures at Treasure Mountain, Summitville quadrangle, Colorado. Welded tuff and associated strata of the upper Tusas Valley were mapped as the Treasure Mountain formation by Butler (1946, p. 30).

The only mappable layer in the Conejos(?) and Treasure Mountain formations of the upper Tusas Valley is a bipartite welded tuff unit. The poorly exposed conglomerate and sandstone that overlie the welded tuff and underlie the Biscara-Esquibel member of the Los Pinos formation have produced float that is indistinguishable from that derived from strata that underlie the welded tuff. Thus, these two formations have been mapped as one unit wherever the welded tuff unit is not present.

Lithology

The Conejos(?) and Treasure Mountain formations are exposed only along the tributary to the Rio Tusas that extends north-northeastward from Deer Trail Junction. Along this tributary, in the SE% sec. 13, and the NE¼ sec. 24, T. 29 N., R. 7 E., the exposed rock is conglomerate with rounded to subangular pebbles of Precambrian rocks and Tertiary volcanic rocks of intermediate composition. The pebbles of Precambrian rocks are mostly quartzite, granodiorite, granite, and amphibolite. The pebbles of volcanic rocks have been weathered until they crumble when handled, and hence only the relatively fresh fragments of Precambrian rocks appear in the float. The matrix is gray to buff, tuffaceous (now partly bentonitic), poorly sorted arkose, and graywacke. Butler (1946, pp. 28-29) notes "maroon, pebbly bentonitic arkose and shaly arkose" intercalated with conglomerate from this locality. This unit is poorly exposed; much tuff and sandstone could be interbedded with the conglomerate without clear indication of its presence from the float, and hence the relative proportions of the sediments that are present cannot be estimated.

The remainder of the Conejos(?) and Treasure Mountain formations, on the east side of the upper Tusas Valley, is evidenced only by their characteristic float, in which cobbles and pebbles of Precambrian rocks are present almost to the exclusion of volcanic rock fragments. An extremely coarse conglomerate or fanglomerate of Tres Piedras granite is implied by float 1 mile southeast of the mouth of Biscara Canyon. The fragments are as much as 8 feet in diameter. This mass, which is about 200 feet thick, appears to thin abruptly to the northwest and southeast.

A similar but finer grained conglomerate with fragments of Tres Piedras granite in a coarse-grained arkosic matrix rests directly upon granite in secs. 35 and 36, T. 28 N., R. 8 E. The relief along the contact between the hill of Precambrian rock and the overlying Tertiary conglomerate is about 250 feet along the present line of exposure. Pebbles of blue-gray vitreous quartzite and micaceous quartzite are present in variable amounts within this conglomerate. Butler considered this rock to be a part of the Biscara member of the Los Pinos formation (1946, p. 56). The lithology of this conglomerate does not, in itself, indicate whether the rock is part of the Conejos(?) and Treasure Mountain formations or the Biscara member of the Los Pinos formation. It may be stratigraphically equivalent to conglomerate of similar nature in the Conejos(?) and Treasure Mountain formations, or it may be the correlative of the conglomerate with clasts of Tres Piedras granite that is part of the Biscara member and is exposed about 2¼ miles north-northwest of Las Tablas. Thus its assignment to either the Biscara member or to the underlying formations is arbitrary.

Thickness

The Conejos(?) strata that underlie the Treasure Mountain welded tuff in the large, most northerly tributary of the Rio Tusas are estimated by Butler (1946, p. 29) to be at least 200 feet thick. The base of this unit is not exposed in this locality. The Biscara-Esquibel member overlies the welded tuff here.

A thickness of 150-200 feet of Conejos(?) strata overlies the Precambrian amphibolite and underlies the Treasure Mountain welded tuff in Biscara Canyon. About 400 *feet* of conglomerate with cobbles of Precambrian rocks and intercalated sandstone overlies the welded tuff in this same canyon. The Conejos(?) and Treasure Mountain formations appear to pinch out to the northwest.

Origin

Much of the sediment in the Conejos(?) and Treasure Mountain formations was derived from local exposures of Precambrian rocks, and the remainder derived from eroded sediments and flows of earlier Tertiary age. Possibly some also was contributed by contemporaneous ash falls.

TREASURE MOUNTAIN WELDED TUFF

Introductory Statement

Butler found that the Treasure Mountain formation extends far southward from Colorado, and reaches the area of the upper Tusas Valley. The only mappable unit of this formation in Las Tablas quadrangle is the bipartite layer of welded tuff, and thus, in this report, it is termed the Treasure Mountain welded tuff, rather than the Treasure Mountain formation.

Distribution

The Treasure Mountain welded tuff is exposed near the head of the Rio Tusas, in secs. 16 and 21, T. 29 N., R. 7 E., along the north-northeast trending large tributary that joins the Rio Tusas at Deer Trail Junction, and in the lower part of Biscara Canyon.

Lithology

The lower layer of welded tuff is a dark-gray to black, mostly vitreous rock, in which both lithic and crystal fragments are faintly to markedly parallel to the gross layering. Platy, flow-banded fragments of black obsidian, as much as several inches long, were seen in the Biscara Canyon exposure. Crystal grains of feldspar and biotite ranging from a fraction of a millimeter to several millimeters in size can be seen easily by the unaided eye in all the rock.

In thinsection, the matrix, which forms 85-90 percent of the tuff and appears as a mass of partially devitrified shards, has been markedly flattened so that the shapes accommodate one another and the pore space been virtually eliminated. The tabular shards, about 0.25 mm long, have rounded edges, but Y- and S-shaped shards are also common. A few unbroken bubbles with fibrous interiors are scattered throughout the rock. The shards are solidly welded together. The rims of the shards are clear glass, but the interiors are aggregates of tiny crystallites set in a glassy matrix.

The crystal fragments are mostly labradorite and andesine, lath shaped, about 1 mm in average length, normally zoned, and commonly with resorbed boundaries. The shards of glass have been strongly bent around these plagioclase grains. Biotite occurs as tabular to chunky hexagon-shaped subhedra, 1 mm in average diameter, though many have partly resorbed boundaries. The mineral is pleochroic, with X = yellow, Y = red brown, and Z = dark red brown. Other minerals present as crystal grains include hornblende, augite, quartz, apatite, and magnetite.

The upper layer of welded tuff is pink to brick red or olive brown, is dense and homogeneous, and fractures with a crude conchoidal pattern. Fragments of pumice as much as 3 inches long are irregularly scattered in it. In the outcrop in Biscara Canyon the uppermost 2-3 inches of the welded tuff layer are brick red instead of olive brown, probably due to subaerial weathering that took place before deposition of the overlying stratum.

The red to brown welded tuff is similar to the underlying black layer in thinsection except for the magnetite grains, which have been partially oxidized to hematite in the upper layer. Because weathering most likely affected both tuffs equally or involved a gradational or progressive oxidation, the oxidation is probably related to the formation of the rock. Unusually long exposure to air during formation may have resulted in the oxidation.

Thickness

In Biscara Canyon the dark-colored layer of welded tuff is about 7 feet thick and the red to brown layer is about 3 feet thick. Four miles to the northwest, along the large tributary to the Rio Tusas, the thicknesses are 10 feet and about 8 feet, respectively. Near the head of the Rio

Tusas the dark layer is not well enough exposed to permit measurement of its thickness, but it appears to be less than 15 feet thick. The red layer either is absent or is not exposed here.

The Treasure Mountain welded tuff attains a thickness of about 100 feet in T. 32 N., R. 7 E., and thins westward and southward (Butler, 1946, p. 33). Butler estimates that it covers at least 160 square miles and that its volume is not less than 1 cubic mile (1946, p. 33, 40).

Origin

The origin of this welded tuff has been discussed by Butler (1946, p. 36-41), who concludes that the material may have been deposited from a large nuee ardente, similar to that envisioned by Williams (1942, p. 79-81) for the Mount Mazama pumice flow. The latter, however, did not develop into a welded rock. Butler postulated that the Treasure Mountain nuee ardente fragments must have retained sufficient heat to weld them together. If exothermic devolatilization kept the nuée ardente at a temperature sufficient for welding of the deposited fragments (Mansfield and Ross, 1935, p. 312-314), then perhaps the difference between the Mount Mazama and Treasure Mountain nuées ardentes was that the glass fragments in the latter contained more dissolved volatile constituents — enough to result in strong welding and partial resorption of the plagioclase.

A second possible mode of origin of this rock is that it could be the lower portion of a greatly compacted pumice flow. If a pumice flow 100 or more feet thick were compressed in its lower portions by the weight of the overlying material, and all but the lower 10-40 feet stripped off by erosion, the resultant layer might well be similar to the Treasure Mountain welded tuff in this area. This mode of origin has been proposed by Dr. George Kennedy (oral communication, February, 1954) for certain welded tuffs in Yellowstone National Park and southeastern Idaho.

RITITO CONGLOMERATE

Definition and Distribution

Conglomerate with gravel-size fragments of Precambrian rocks only was found to lie directly upon Precambrian rocks along the lower Vallecitos Valley. This formation is herein named the Ritito conglomerate, after excellent exposures in Ritito Canyon, in secs. 11 and 14, T. 27 N., R. 7 E. Other exposures are 2 miles southeast and 1 mile west to south of Cañon Plaza, as well as along the northeast side of the Vallecitos Valley from Cañon Plaza to Jarosita Creek, and along parts of Escondida and Felipito Canyons.

Lithology

Rounded to subangular pebbles and small boulders of fine-grained to pebbly quartzite, amphibolite, and metarhyolite are the common clasts in the Ritito conglomerate. Most of the amphibolite fragments are weathered and very friable, and hence are rare in the float that almost everywhere mantles this formation. The matrix of this conglomerate was observed only in the roadcut 250 yards northwest of the mouth of Ritito Canyon, and at the mouth of Escondida Canyon. Like the larger fragments in the rocks, the matrix is quartzose and is composed of poorly sorted, subangular to subrounded grains. Generally the rock is weakly cemented and typically has a medium-gray color.

Thickness

The Ritito conglomerate reaches its maximum known thickness in this quadrangle on the east side of Ritito Creek, and in Escondido Canyon, where about 400 feet of strata are exposed. The formation thins in Canada del Oso, where it lies upon a hill of Precambrian rock, and it pinches out entirely a mile to the east. It also thins out in the upper parts of La Jara and Vallecitos Canyons, and is absent northward from Jarosita Canyon and Quartzite Peak. The thickness of the Ritito conglomerate west to southeast of Cañon Plaza could not be measured, but 100-200 feet is exposed here.

Origin

The Ritito conglomerate probably was deposited as an alluvial mantle of varying thickness on an irregular surface of Precambrian rock. It is older than the period of widespread Tertiary vulcanism. The Ritito conglomerate is probably correlative with the Conejos(?) formation. It may be equivalent, however, to similar conglomerate of the Biscara member exposed near Las Tablas.

LOS PINOS FORMATION

General Statement

The Los Pinos gravel was first described by Atwood and Mather (1932, p. 92-101), who ascribed its name to Cross and Larsen and who defined it as a waterlaid gravel, sand, and siliceous tuff that was deposited on the San Juan peneplain and subsequently covered by Hinsdale basalt. The type locality is the Los Pinos River Canyon near San Miguel, New Mexico. Cross and Larsen (1935, p. 95) later included these rocks as part of the Hinsdale formation. Butler (1946) clarified the stratigraphic relations of the Los Pinos formation, and in the Tusas Valley separated it into four units, the Biscara, Esquibel, Jarita basalt, and Cordito members. He redefined the formation to include all rocks that overlie the Treasure Mountain formation and unconformably underlie the Cisneros basalt.

Butler's terminology is used in this report, but the Biscara and Esquibel members are grouped together northwestward from a point about 3 miles southeast of Tusas.

Biscara Member

Definition. The basal member of the Los Pinos formation, characterized by fragments of andesite and dark quartz latite, was named after the exposure in Biscara Canyon about 1-5 miles from its mouth (Butler, 1946, p. 53). These strata of the type locality, however, are here included in the Biscara-Esquibel member.

Distribution. The Biscara member, as mapped by the author, lies in the lower Tusas Valley east of Tusas Canyon, from 1.3 miles north of Petaca to a point about 3.5 miles north-northwest of Las Tablas.

Lithology. The Biscara member consists of interlayered tuffaceous sandstone, tuff, conglomerate, and volcanic flow breccia. Conglomerate, with fragments of gray, brown, and dark-red andesite and quartz latite, graywacke, and tuff, forms most of the unit. Exposures generally are lacking, so that this member was mapped mostly by means of its float of dark volcanic rocks.

A flow breccia crops out 2.5 miles north of Las Tablas, in secs. 6 and 7, T. 27 N., R. 9 E. This rock is an andesite porphyry with angular fragments similar in composition to the matrix, whose phenocrysts comprise about 30 percent andesine, 10 percent hornblende, and 5 percent biotite. The groundmass is mostly plagioclase in small laths, and magnetite is the chief accessory constituent. The 0.01- to 0.1-mm grains of plagioclase in the groundmass show a fluidal structure around the phenocrysts.

The Biscara member along the Rio Tusas from about 1 mile north of Las Tablas to 2 miles south of this village comprises two units. The lower unit consists of about 30 feet of gray rhyolitic(?) tuff, which contains rounded pebbles of volcanic rock, probably waterlaid, and interbedded tuffaceous conglomerate. The upper unit consists of 30 feet of poorly stratified and sorted conglomerate (pl. 8A), composed of sub-angular to subrounded fragments of quartzite, metarhyolite, pegmatite, and other Precambrian rocks set in a clean arkosic matrix. The Jarita basalt, an associated gray tuff, and sediments of the Cordito member successively overlap the conglomerate from south to north.

Thickness. The Biscara member was deposited around and upon hills of Precambrian rocks whose relief Butler (1946, p. 54) estimated to be as much as 600 feet. The maximum observed thickness of the Biscara member on the west side of Canon del Agua is 650 to 700 feet. This section, however, may be in part disturbed by small faults. The Biscara member disappears southward and eastward beneath younger rocks, pinches out westward and northwestward on Precambrian rocks, and grades northward into the Biscara-Esquibel member.

Origin. The conglomerate with fragments of precambrian rocks undoubtedly was derived from local sources. The flow breccia is similar to the dike of late Biscara age that crops out in Cañon del Agua, described below on p. 64, and probably was extruded from a nearby vent.

Biscara-Esquibel Member

Definition. The Biscara-Esquibel member of the Los Pinos formation is a composite of two members that were named by Butler (1946, p. 53-61) after type areas in Biscara and Esquibel canyons. The Esquibel member was defined by Butler as the conglomerate, tuff, and graywacke overlying the Biscara member and underlying the Jarita basalt and Cordito members. It was identified in the field by means of fragments of conspicuous gray or purple-pink quartz latite containing feldspar phenocrysts as much as 8 mm long (Butler, 1946, p. 62). The contact between the two members is gradational, and was not mapped from east of the center of Biscara Canyon northwest to Broke-Off Mountain (Butler, 1946, pl. I). Part of the Biscara-Esquibel member, as here mapped, is equivalent to Butler's undifferentiated Los Pinos formation in parts of the upper Tusas Valley.

Distribution. The Biscara-Esquibel member is exposed from the southwest corner of T. 28 N., R. 9 E., to the northern boundary and part of the western boundary of the quadrangle.

Lithology. The Biscara-Esquibel member consists of interbedded conglomerate, tuff, and graywacke. The conglomerate contains fragments of andesite and quartz latite. The 680-foot section of the Biscara, Esquibel, and Cordito members exposed westward from near the divide on the Tusas-Tres Piedras road was measured by Butler (1946, p. 63):

	Feet
Cordito member	
Well-indurated breccia-like conglomerate of rhyolitic fragments not measured	
Partly indurated conglomerate of mixed fragments of rhyolitic and pre-Cambrian rocks	40
Esquibel member	
Arkosic sandstone, buff, thin-bedded, poorly indurated, and conglomeratic arkosic sandstone containing fragments of pre-Cambrian rock (top of member is 5 feet below summit of road)	20
Sandy conglomerate and some thin beds of tuff or tuffaceous siltstone, mostly gray, but with some buff beds. Fragments of pre-Cambrian rock abundant at top and predominant in some beds, sparse at bottom. Fragments of coarsely porphyritic quartz latite predominate in others	190
Felsitic tuff, waterlaid, and felsitic tuff, mostly gray; some interbedded graywacke and sparse beds of conglomerate, fragments of coarsely porphyritic quartz latite	390
Total Esquibel member	600
Biscara member	
Graywacke, gray, thin-bedded, and conglomerate with dark-colored fragments of felsite; base not exposed	40
Total thickness measured	680

The Biscara and Esquibel members, here differentiated by Butler, have been mapped by the writer as a single unit, the Biscara-Esquibel member.

Thickness. The Biscara-Esquibel member thins out rather abruptly from the Tusas-Tres Piedras road southward to the boundary of T. 28 N. and T. 27 N. Near the head of the Rio Tusas it is about 300 feet thick, and in the northeastern part of T. 28 N., R. 8 E., about 1,100 feet (Butler, 1946, p. 78). In the vicinity of Hopewell this member laps onto the Precambrian rocks.

Jarita Basalt Member

Definition. The Jarita basalt member of the Los Pinos formation was named by Butler (1946, p. 65) from an elongate exposure along the western rim of La Jarita Mesa northeast of Vallecitos. It overlies his Esquibel member and is unconformably overlain by the Cordito member.

Distribution. The Jarita basalt occurs as scattered flows or series of flows, and is present locally along the unconformity at the base of the Cordito member. Its exposures are in Apache Canyon, along a part of the lower Tusas Valley, in Cañon del Agua, at four scattered localities near the divide west of the Rio Tusas from Canon del Agua to the Tusas-Tres Piedras road, along the upper Rio Tusas east of Jawbone Mountain, and in the northeastern corner of the quadrangle.

Lithology. Butler divided the Jarita basalt into (1946, p. 66-69, 134136) a northern, central, and southern type. He further divided the northern type, found north of the Rio Tusas in secs. 15 and 16, T. 29 N., R. 7 E., and in the northeastern corner of the area, into two varieties. The more abundant of these is characterized by "small phenocrysts of rusty iddingsite, sparse plagioclase, and considerable intergranular pore space" (p. 134), and the other by a few pyroxene phenocrysts, partly altered yellow-brown olivine, secondary calcite, and an aphanitic texture. The petrography of the northern type of Jarita basalt has been described in detail by Butler (1946, p. 134-135):

> NORTHERN SUBDIVISION.—The basalts of the northern subdivision can be divided into two varieties. The more abundant variety has small phenocrysts of rusty iddingsite, sparse plagioclase, and considerable inter-granular pore space. The less abundant variety has some pyroxene phenocrysts, yellow-brown olivine that is only partly altered, and veinlets and amygdules of calcite, and it lacks intergranular pore space.
>
> Most flows of the first variety are fine- to very fine-grained, but a few are medium-grained. Although plagioclase can generally *be* distinguished in hand specimen, olivine is the more common phenocryst. . . . The rock can best be described as sparsely porphyritic.
>
> As seen under the microscope, the rock ranges from partly to wholly crystalline, is fine- to medium-grained hypautomorphic, interscrtal or inter-granular to ophitic . . . and sub-porphyritic to sparsely porphyritic. Phenocrysts form less than 10 percent of the rock and are generally gradational in

size to the grains of the ground-mass. Subhedral, slightly zoned grains of calcic labradorite from 0.4 to 3.0 millimeters in maximum dimension, form sparse phenocrysts in some flows.

The groundmass consists of intermediate labradorite, generally about An63 pyroxene, olivine, magnetite, accessory apatite, and generally a small amount of glass. . . . Sodic plagioclase, probably oligoclase, forms thin borders around the labradorite of some flows. Pyroxene, probably augite, is commonly interstitial to the feldspars, but in some of the coarser-grained flows it is ophitically intergrown with plagioclase and is as much as 1.5 millimeters in diameter. Most of the olivine has been partly or wholly altered to iddingsite. . . . The alteration commonly proceeds outward from the center of the grain so that only the rims of some grains are fresh. However, grains altered on the borders and along the fractures are abundant in some sub-porphyritic, medium-grained flows. Glass amounts to less than 5 percent of most flows. In some flows it is pale brown and clear; in others heavily dusted with microgranular minerals.

....

Under the microscope the less abundant variety of the northern type of rock is seen to differ from the other in the following respects: the olivine is commonly altered inward from the borders and fractures; pyroxene phenocrysts, probably diopside, are more common; and oligoclase commonly forms sheaths around labradorite and makes up some of the groundmass. Orthoclase is probably present in one flow. Glass forms as much as 10 percent of some flows....

TABLE 6. CHEMICAL ANALYSIS, NORM, AND MODE OF JARITA BASALT*

(In percent)

Chemical Analysis†		Norm		Mode	
SiO_2	52.09	Quartz	8.1	Plagioclase (An_{62})	29.3
TiO_2	1.84	Orthoclase	7.8	Clinopyroxene	17.5
Al_2O_3	14.59	Albite	22.0	Olivine and iddingsite	9.6
Fe_2O_3	5.44	Anorthite	24.0	Groundmass and magnetite	36.5
FeO	5.89	Diopside	12.2	Vesicles	7.1
MnO	0.15	Hypersthene	11.9		
MgO	5.64	Magnetite	7.9		
CaO	8.34	Ilmenite	3.5		
Na_2O	2.60	Apatite	0.6		
K_2O	1.32				
P_2O_5	0.29				
H_2O+	1.42				
H_2O-	0.22				
CO_2	0.00				
Total	99.83				

* From north side of upper Tusas Creek, NE¼SW¼, sec. 15., T. 29 N., R. 7 E.

† Analyst: H. B. Wiik, Helsinki, Finland.

The central type of Jarita basalt is exposed in Cañon del Agua and in areas to the northwest along the upper slopes of the Tusas drainage. The basalt at Cañon del Agua contains 10-15 percent of phenocrysts of partially resorbed, subhedral to euhedral labradorite as much as 10 mm long; stubby euhedra of pigeonite; and equant to subhedral olivine with fringes of iddingsite. The groundmass is about 80 percent plagioclase, and the remainder is iddingsite, pyroxene, and magnetite.

The southern type of Jarita basalt is exposed in the lower Tusas Valley from a point 1 mile southeast of Las Tablas to points southwest of Petaca, on La Jarita Mesa northeast of Vallecitos, in Apache Canyon, and on the south side of the Rio Tusas near its head. These basalt flows are somewhat similar to the northern type, but Butler (1946, p. 136) has pointed out certain differences:

> SOUTHERN SUBDIVISION.—The flows of the southern subdivision that have considerable intergranular pore space and resemble the northern flows in texture lack the rusty iddingsite common in the porous northern flows. Flows of the southern type that have phenocrysts of iddingsite also have phenocrysts of plagioclase and some pyroxene and are dense. The characteristic small pale green or yellow-green spots that can be recognized in many of the flows in the field are unrecognized in thin section.
>
> Microscopic examination shows that the southern flows can be divided into ordinary and hypersthene basalt. The hypersthene basalt is fine- to medium-grained, hypautomorphic, intergranular and sub-porphyritic. In most of the rock hypersthene is the chief phenocryst although it amounts to less than 2 percent of the rock. The grains are partly resorbed, and some have reaction rims of clinopyroxene. Iddingsite after olivine is very sparse but generally present. Otherwise the rock of these flows is much like that of the abundant variety in the northern flows.

Thickness. The Jarita basalt flows are of variable thickness, with a maximum of about 50 feet. Thicknesses in different areas are as follows: near the head of the Rio Tusas, 20-30 feet; in the northeastern corner of the quadrangle, 0-50 *feet;* from 1 mile south of the Tusas-Tres Piedras road to Cañon del Agua, 0-30 feet; in Apache Canyon, 0-20 feet; and on the west rim of La Jarita Mesa, 0-50 feet. The flow that extends from the vicinity of Las Tablas to points southwest of Petaca increases in thickness southward from 0 to 40 or 50 feet.

The Jarita flows appear to have much greater ENE.-WSW. dimensions than NNW.-SSE. dimension. The layer in the lower Tusas Valley may be an exception, as its greatest dimension now exposed is along a north-south direction, and it pinches out abruptly against the Precambrian rock to the west. The present extent of these flows is directly related to the dominant north-northeast to northeast trend of the drainage that was developed on much of the older Tertiary section. The vesicles in the Jarita basalt exposed in sec. 19., T. 29 *N.,* R. 9 E., are well alined in a S. 40° W. direction, which implies flowage of this general trend. The author believes that the source probably lay to the northeast.

Cordito Member

Definition. The Cordito member, the uppermost unit of the Los Pinos formation, was named by Butler (1946, p. 70) from exposures in Canyon de Cordito, 4 miles south of Tres Piedras. It overlies disconformably the Jarita basalt and the Biscara and Biscara-Esquibel members of the Los Pinos formation. In about one-third of the quadrangle the Cordito member, where present, rests directly on Precambrian rocks.

Distribution. The Cordito member, largely interbedded conglomerate, tuff, arkose, graywacke, and siltstone, with minor flows of rhyolite, underlies about 80 square miles of Las Tablas quadrangle. The main areas of exposure are in a strip, 1 to 3 miles wide, along the eastern boundary, along the southwest side of the Tusas Valley between Spring Creek and Duran Canyon, in the southwestern corner of the quadrangle, in the area between Kiawa Mountain and La Jara Canyon, and in the area of the Tierra Amarilla Grant. As exposed along the southern boundary of the quadrangle, the Cordito member is equivalent to strata mapped as the Abiquiu tuff by Smith (1938, p. 937.)

Lithology. Most of the fragments in the conglomerate of the Cordito member are rhyolite, which Butler (1946, p. 72) divided into two main types. One is blue and sparsely porphyritic to gray, red, or light gray and markedly porphyritic. Quartz is the dominant phenocryst. The other type is quartz latitic to rhyolitic in composition. It is coarsely porphyritic, with abundant phenocrysts of feldspar and some phenocrysts of biotite, hornblende, and quartz. Cobbles and boulders of Jarita basalt are abundant in the Cordito conglomerates, especially in layers that lie immediately above the basalt flows. The gravel-size fragments in all of the conglomerates are poorly sorted, and clasts commonly 2-6 feet in diameter are scattered among the smaller ones about 6 inches across.

Much of the Cordito member is weakly cemented, and hence does not form good outcrops. Well-cemented cobbly conglomerate is common, however, and gives good exposures such as those along Tusas Canyon just above the junction with Spring Creek (pl. 4). Owing to the general rarity of exposures, the proportions of the various kinds of rocks in this member could not be determined.

Tuff and tuffaceous sandstone are more common in the southern part than elsewhere in the quadrangle. They are gray to light gray, buff, and light greenish gray. Probably all are of rhyolitic composition. The matrices of almost all of the conglomerates are tuffaceous. Mappable tuff beds were found east and southeast of Las Tablas, and in Cation del Agua. All of these layers are less than 15 feet thick.

Well-cemented rhyolite tuff is intercalated with Cordito sedimentary rocks on La Jarita Mesa along a part of Apache Canyon, along the upper part of Cañon del Agua, and on the west bank of the Rio Tusas at the southern boundary of the quadrangle. About 20-50 percent of this rock is composed of crystal fragments of quartz, sanidine, orthoclase, oligoclase, hornblende, biotite, and magnetite of 0.5 mm average grain size. The matrix is largely glassy, and shards are clearly visible in most sections. Tiny crystalline grains are scattered throughout these shards. All of the glass is partially devitrified.

Thickness. The thickness of the Cordito member reaches a maximum of about 600 feet, in T. 28 N. (Butler, 1946, p. 78-79). The Cisneros and

Dorado basalts unconformably overlie the Cordito member, so that the figures given represent only a part (commonly most?) of the sediment that was deposited. This member is about 400 feet thick on the east side of the Tusas Valley 2 miles south of Las Tablas. Northeast of Las Tablas it ranges from about 250 to more than 400 feet thick. At the divide east of the Tusas Valley, in T. 28 N., the Cordito member is about 250 feet thick (Butler, 1946, p. 78).

The highland from La Jarita Mesa to Hopewell is partly fringed with Cordito rocks, which lap against the Precambrian rocks.

Southwest of the Rio Vallecitos the Cordito member is at least 500 feet thick, and may possibly be as much as 1,000 feet.

Origin of the Los Pinos Formation

The origin of the Los Pinos formation in the eastern and northern part of Las Tablas quadrangle is discussed at some length by Butler (1946, p. 81-87), who concludes that the alluvial debris came from essentially contemporaneous volcanic centers located in the present Taos Plateau or in the adjoining San Luis Valley to the north in Colorado. The author concurs with this conclusion, the major evidence for which can be summarized as follows:

1. Direct evidence
   a. Some of the fragments in the conglomerates are similar to the intercalated masses of volcanic rocks.
   b. Volcanic intrusions are present in the southeastern part of the area.
   c. Stratigraphic changes are greater in a north-south direction than in an east-west direction.
   d. The largest boulders in conglomerates of volcanic rock fragments are found from T. 28 N. to T. 26 N.; there is no regular north-to-south gradation in size.
   e. Local thickening occurs in the Biscara member near Las Tablas.
   f. Pebble changes are systematic from andesite to rhyolite.
   g. The fragments in the conglomerates differ from those of pre-Los Pinos volcanic rocks to the north that are within transportable distance.
2. Indirect evidence
   a. The Santa Fe formation (discussed below on p. 51) was derived from the Sangre de Cristo Mountains to the east.
   b. Tertiary volcanic centers are found to the northeast and east; none are known to the west.
   c. The Conejos andesite to the north, which is lithologically similar to some of the pebbles, apparently was covered by the Treasure Mountain formation, which does not appear to have yielded fragments to the Los Pinos formation.

The origin of the Cordito member should be the same in both the southwestern and eastern parts of the quadrangle, as the rocks are similar except for the general westward increase of tuff. If the streams that deposited the Cordito beds flowed southwestward, they must have passed through gaps in the La Jarita Mesa-Jawbone Mountain highland. Such gaps may have existed at Hopewell, Spring Creek, immediately south of Kiawa Mountain, and on La Jarita Mesa west of Petaca. These conditions are similar to those proposed by Smith (1938, p. 949) for the Abiquiu tuff, which is in large part equivalent to the Cordito member, as already noted. The central part of Tusas quadrangle, a 30-minute quadrangle which includes Las Tablas quadrangle and the 15-minute quadrangles to the north, northwest, and west, was designated by Smith (1938, p. 956) as the apex of a large tuff fan that extended southward to Abiquiu. This fan evidently extended much farther north and east, i.e., to Tres Piedras and points north, than Smith described it, and its deposits become much more conglomeratic in those directions.

SANTA FE FORMATION

The Santa Fe formation was mapped by Butler (1946) from Petaca to the southern edge of his map area, about 12 miles south and east. He found (p. 104-109) that the buff to red, clean to silty sandstone of this formation fingers into the Cordito member of the Los Pinos formation. On the east side of the Rio Tusas, across from Las Tablas, the author found the base of the Cordito member to be a 30-foot layer of conglomerate with boulders of Jarita basalt. A 15-foot layer of buff tuff is next, and in turn is overlain by about 30 feet of light-orange sandstone with thin lenses rich in rhyolite pebbles. In thinsection this orange sandstone is a very fine-grained cherty feldspathic sandstone, of 0.1 mm average grain size, and it consists of about 55 percent quartz, 20 percent plagioclase and microcline, 20 percent chert, and 5 percent basalt grains, along with muscovite, zircon, blue fluorite, and basaltic hornblende.

This layer probably is correlative with the more extensive Santa Fe formation to the south, but it is shown on Plate 1 as part of the Cordito member because of the rhyolite pebbles contained in it and because it could be traced laterally for a distance of only 800 yards. Similar, very scantily exposed, pink to light-orange sandstone was found immediately east of Petaca, but could not be adequately mapped. These relations directly support Butler's conclusion that the Santa Fe and upper Los Pinos formations interfinger with each other.

CISNEROS BASALT

Definition

The Hinsdale series was redefined by Butler (1946, p. 110) to include three formations: the Cisneros basalt, the Dorado basalt, and the Servilleta formation. The Cisneros basalt was named (p. 111) from Cisneros

Park, in the NW¼ T. 29 N., R. 8 E. It overlies the Cordito member of the Los Pinos formation with a slight angular unconformity.

Distribution

The Cisneros basalt was found in Las Tablas quadrangle west and north of Cañon del Agua, and in the extreme northeastern corner in secs. 16 and 17, T. 29 N., R. 9 E.

Lithology

The Cisneros olivine basalt that crops out 3.5 miles north of Las Tablas is a dark-gray, slightly vesicular, sedate rock. Laths of calcic labradorite from 0.05 to 1 mm in length form 55-60 percent of the rock. Olivine, colorless, or stained brown, is present as equant to subhedral grains in amounts from 35-40 percent, and the grains have an average size of 0.12 mm. They are interstitial to intergranular. About 5 percent magnetite is present.

The basalt in sec. 17, T. 29 N., R. 9 E., differs from that described above in that it is only slightly porphyritic, has a very fine-grained groundmass, is dense, and shows well-developed fluidal texture contributed by elongate phenocrysts of plagioclase. The chemical analysis, norm, and modes of a similar basalt are given in Table 7.

TABLE 7. CHEMICAL ANALYSIS, NORM, AND MODE OF CISNEROS BASALT*

(In percent)

Chemical Analysis		Norm		Mode	
SiO_2	50.38	Orthoclase	2.78	Plagioclase (An_{55})	44
Al_2O_3	16.55	Albite	24.63	Augite	45
Fe_2O_3	1.45	Anorthite	30.58	Olivine	4
FeO	9.78	Diopside	12.09	Magnetite	3
MgO	7.54	Hypersthene	17.12	Groundmass	4
CaO	9.27	Olivine	8.07		
Na_2O	2.88	Magnetite	2.09		
K_2O	0.50	Ilmenite	1.98		
H_2O-	0.18	Apatite	0.34		
H_2O+	0.44				
TiO_2	1.04				
P_2O_5	0.12				
MnO	0.13				
	100.26				

* Specimen from Buffalo Buttes, NW. rim of NE. depression. Analysis by George Steiger; norm and mode by E. S. Larsen. From unpublished manuscript cited by Butler (1946, p. 150).

Thickness

The flows of Cisneros olivine basalt in the area are not overlain by younger rocks, and their original upper surfaces have been destroyed. Only 10-30 feet of basalt remain in most places.

Origin

The flow in the northeast corner of the quadrangle is part of a relatively large series of flows centered about Buffalo Buttes, northeast of Las Tablas quadrangle. The basalt north of Las Tablas probably also was derived from an eruptive center to the east or northeast.

DORADO BASALT

Definition and Distribution

The name Dorado basalt was applied by Butler (1946, p. 115) to the basalt that caps the Petaca Mesas. This rock is particularly well exposed in Dorado Canyon, northeast of Petaca. The mesa east and northeast of Las Tablas is also capped by Dorado basalt.

Lithology

Butler (1946, p. 139) has described the vesicular Dorado basalt as follows:

> In thin section the rock appears partly to wholly crystalline, fine-grained, and porphyritic. The usual phenocrysts are intermediate labradorite, olivine, which is resorbed and partly altered to serpentine or iddingsite, and sparse clinopyroxene, orthopyroxene, oligoclase, and quartz. Fluidal orientation of the plagioclase tablets of the groundmass, sparse phenocrysts of clinopyroxene and orthopyroxene, partial alteration of some of the olivine to serpentine, and the presence of interstitial glass are the chief characteristics that distinguish the Dorado of the type locality. . . .

Thickness

The mesa cappings of Dorado basalt range in thickness from at least 40 feet to about 100 feet.

QUATERNARY ROCKS

ALLUVIUM

Sheets of alluvial gravels have been deposited along the Rio Tusas and the Rio Vallecitos for most of their extent. Coarse to fine sand, light orange to brown in color, was found east of Tusas and north of Vallecitos. Butler (1946, p. 179) has suggested that the sand east of Tusas may be of eolian origin.

Irregular patches of alluvial gravel and eolian(?) sand, whose larger fragments are mostly of Precambrian rocks, overlie the Cordito member on the west side of Tusas Valley, from Aveta Creek to points north of Cow Creek.

Intrusive Igneous Rocks

PRECAMBRIAN INTRUSIVE ROCKS

BURNED MOUNTAIN

METARHYOLITE Definition

The Precambrian metarhyolite in Las Tablas quadrangle and the area immediately to the south was first recognized by Just (1937, p. 44), who named it the Vallecitos rhyolite. Inasmuch as this name had been previously used, the rock is here renamed the Burned Mountain metarhyolite, after the exposures on the northwest side of Burned Mountain, in sec. 8., T. 28 N., R. 7 E.

Distribution

The Burned Mountain metarhyolite is exposed on La Jarita Mesa, in the area between Vallecitos and Cañon Plaza, in Canada del Oso, in La Jara Canyon, along Vallecitos Canyon near Escondida Creek, from points immediately northwest of Tusas Mountain to Hopewell, and in an area immediately south of Jawbone Mountain near the head of Buck-horn Gulch.

Li thology

The Burned Mountain metarhyolite ranges from brick red to light pink, but generally is reddish orange; relict quartz and microcline phenocrysts and relict, commonly drag-folded flow bands are clearly visible in most outcrops. Relict phenocrysts of albite-oligoclase also are present in some of the metarhyolite. All of the phenocrysts have been rotated about axes parallel to those of the drag-folded flow bands.

In thinsection many of the original phenocrysts still exhibit euhedral to subhedral outlines. The microcline, originally sanidine or orthoclase, and the plagioclase grains range in size from groundmass dimension, or about 0.02 mm, to more than 5 mm, with an average of about 0.5 mm. Similar quartz relicts are 0.5 mm in average diameter. A few of these retain their originally bipyramidal form. The quartz and microcline commonly show partially resorbed boundaries. Many of the original quartz euhedra have been recrystallized to aggregates of smaller anhedra that show only slight changes of external shape.

The relict groundmass of the metarhyolite typically exhibits a seriate, granular mosaic texture of average grain size 0.02 mm. Quartz, microcline, and albite-oligoclase are the three major minerals in the groundmass. Muscovite is the only varietal mineral, and magnetite and apatite are accessories. Tiny, tabular crystals of muscovite are typically well alined.

A 5-foot layer of metarhyolite occurs in greenschist in a branch of Rock Creek, in the SW¼NW¼ sec. 15, T. 28 N., R. 7 E. This layer has been intensely sheared, and sericite and chlorite have developed to form a quartz-feldspar-sericite-chlorite schist. Rounded relict phenocrysts of quartz and faint relict flow layering indicate that the rock originally was a rhyolite porphyry.

TABLE 8. CHEMICAL ANALYSIS AND NORM OF BURNED MOUNTAIN METARHYOLITE• (In percent)

CHEMICAL ANALYSISt		NORM	
SiO_2	75.47	Quartz	34.9
TiO,	0.40	Orthoclase	28.9
A1203	11.42	Albite	28.8
Fe20,	2.26	Anorthite	1.7
FeO	0.31	Diopside	1.2
MnO	0.07	Hypersthene	3.3
MgO	0.30	Magnetite	3.2
CaO	0.55	Ilmenite	0.8
Na20	3.29		
K_2O	4.90		
P202	0.05		
H20+	0.45		
H_2O—	0.00		
CO_2	0.00		
Total	99.47		

• **Specimen** from Caflada del Oso, NEV4SEV4 sec. 14, T. 27 N., R. 7 E. The close compositional similarity of this rock to the Tres Piedras granite is discussed on p. 61. 1ˉ Analyst: H. B. Wiik, Helsinki, Finland.

Field Relations and Origin

The layers of Burned Mountain metarhyolite almost everywhere are concordant with the underlying and overlying strata. An imperfectly exposed forked dike at Burned Mountain, however, appears to be a noteworthy exception; it transects the bedding of the enclosing metavolcanic rocks at low angles. The tabular unit of metarhyolite on the west slope of La Jarita Mesa, in sec. 33, T. 27 N., R. 8 E., and sec. 4, T. 26 N., R. 8 E., has conformable top and bottom contacts with the enclosing Ortega quartzite; its northwestern end, however, fingers abruptly into the quartzite, which suggests that this mass is a sill rather than a flow tongue. The body of metarhyolite that crops out in secs. 21 and 22, T. 27 N., R. 8 E., has a similar blunt termination.

Just (1937, p. 44) believed that these bodies of metarhyolite originally were flows, mainly because of their flow banding, their aphanitic groundmasses, their elongate and lenticular shapes, their parallelism with the stratification of the enclosing metasedimentary rocks, their interlayering with conglomeratic quartzite, and the sequence of crystallization of their quartz and orthoclase. None of these features conclusively demonstrates flow origin. The author believes that the metarhyolite was emplaced as sills, partly as dikes, and possibly in small part as flows. Jahns (personal communication) has observed a breccia of amphibolite fragments set in a matrix of metarhyolite in the top of a metarhyolite mass in the Ojo Caliente district — a relation which proves that at least part of the metarhyolite is hypabyssal in nature.

MAQUINITA GRANODIORITE

Definition and Distribution

Granodiorite is exposed in Las Tablas quadrangle as many small dikes and as parts of several plutons. This intrusive rock is here named the Maquinita granodiorite, after the exposures in Maquinita Canyon, secs. 3 and 4, T. 28 N., R. 7 E. This rock was grouped by Just (1937, p. 45) with the Tusas granite. The plutons lie between American and Cow Creeks, on a portion of the lower north slope of Tusas Mountain, along the west side of the upper Tusas valley southeast and northwest from Maquinita Canyon, and along part of Duran Canyon. Small dikes of granodiorite are intrusive into the Moppin metavolcanic series in the area that extends from just north of Burned Mountain to Hopewell and Buckhorn Gulch.

Small isolated exposures of rock similar to the type Maquinita granodiorite, and tentatively correlated with it, lie on the northeast side of the upper Tusas valley about three-fourths of a mile east of the mouth of Maquinita creek, and at the mouth of Canada Biscara, as well as in the northeast corner of the quadrangle in secs. 13 and 24, T. 29 N., R.8 E.

Lithology

The Maquinita granodiorite is gray to dark gray, homogeneous, well foliated, and strongly lineate. Both the foliation and lineation are defined by the distribution and orientation of biotite knots, 1A-1 inch long. In some exposures the foliation is faint, but the lineation is well marked in all of the granodiorite. The granodiorite has been strongly sheared in the dikes that cut the Moppin series and in the plutons exposed in the area from Deer Trail Creek to American Creek. The feldspar and quartz are granulated, and locally the rock is essentially a flaser (lenticular) granodiorite.

Under the microscope the Maquinita granodiorite is composed of moderately sericitized and saussuritized calcic oligoclase to calcic albite, altered orthoclase and microcline, quartz in grains of very irregular shapes and sutured boundaries, biotite and epidote in irregular lenses or clusters, and accessory magnetite-ilmenite, apatite, and calcite. The least sheared granodiorite is exposed along part of Duran creek, in secs. 32 and 33, T. 29 N., R. 7 E. This rock has been moderately sheared; plagioclase grains, which originally appear to have been subhedral and 1-5 mm in length, have been fractured and ground against neighboring grains so that they now are rounded to very irregular. Quartz occurs either as rounded to angular aggregates, 1-4 mm across, or as much finer interstitial grains. Biotite and epidote are associated in irregular clusters that are mostly intersertal to the larger grains of light-colored minerals.

Alteration of the Maquinita granodiorite is very similar in all of the exposed bodies. Plagioclase and orthoclase have been moderately to thoroughly sericitized and slightly saussuritized. The markedly altered feldspar now appears as a thick network of sericite and saussurite, which contains epidote and rounded, interstitial grains of clear albite or oligoclase. Some of the altered plagioclase also is rimmed with clear albite or oligoclase. All of the albite-oligoclase present may well have been formed during the alteration. Similarly, the disseminated grains of epidote and biotite may be alteration products of hornblende.

In the plutons from Deer Trail Creek to American Creek as much as two-thirds of the granodiorite has been sheared to a fine-grained aggregate that is interstitial to the original grains. This aggregate consists largely of equant to irregular grains of quartz and untwinned feldspar, with lesser amounts of biotite, epidote, muscovite, and saussurite, all of which range from about 0.02 to 2 mm. Even in this highly granulated rock, which almost merits classification as a flaser-granodiorite, most of the biotite and epidote occurs as discrete, well-lineated knots. The amount of shearing is variable, even in the same pluton, and does not appear to increase toward the boundaries; however, too few specimens were gathered to warrant a definite conclusion.

TABLE 9. CHEMICAL ANALYSIS, NORM, AND MODE OF MAQUINITA GRANODIORITE
(In percent)

Chemical Analysis	(1)*	(2)†	Norm		Mode	
SiO_2	68.13	66.13	Quartz	22.0	Albite-oligoclase	57
TiO_2	0.26	0.51	Orthoclase	15.5	Quartz	27
Al_2O_3	16.81	15.50	Albite	42.5	Microcline	1
Fe_2O_3	1.19	1.62	Anorthite	9.5	Biotite	10
FeO	1.56	2.70	Magnetite	1.6	Muscovite	4
MnO	0.05	0.07	Ilmenite	0.5	Epidote	1
MgO	1.28	1.73	Enstatite	3.0	Accessory minerals	tr
CaO	1.88	3.70	Ferrosilite	1.6		
Na_2O	5.00	3.55	Corundum	2.2		
K_2O	2.62	3.17		98.4		
P_2O_5	0.13	0.17				
H_2O+	0.91					
		0.89				
H_2O-	0.11					
CO_2	0.00	0.04				
Total	99.93	99.78				

* Specimen 36-B-36. From NW¼ sec. 30, T. 28 N., R. 8 E. Analyst: H. B. Wiik, Helsinki, Finland.

† Average composition of 80 granodiorites (Johannsen, 1932, p. 344).

The Maquinita granodiorite contains slightly more silica, alumina, and soda, and slightly less iron oxides, magnesia, and lime than does Johannsen's average rock. These differences are apparent in the modal analysis, as shown by the abundant albite-oligoclase and quartz, the moderate amount of biotite, and the absence of hornblende. Such variations from the average probably are due to processes of differentiation that took place before the emplacement of the magma in its present position. Settling of mafic crystals from the magma before injection would account for the low amounts of iron oxides, magnesia, and lime.

Field Relations

The contact of the Maquinita pluton south of Cow Creek with the amphibolites of the Moppin series appears to be sharp and parallel with the schistosity of the wall rock. No septa or other bodies of wall rock were seen in this pluton.

The most southerly boundary of the granodiorite pluton on the lower north slope of Tusas Mountain appears to be closely parallel to the foliation of the granodiorite and also to the faint flow foliation of the younger Tres Piedras granite. What probably is a dike of fine-grained granite is poorly exposed at this contact; it cuts the granodiorite, and constitutes the most direct evidence that the Maquinita granodiorite was emplaced earlier than the Tres Piedras granite. Elsewhere homogeneous granodiorite is in contact with homogeneous granite, except for one small amphibolite xenolith lying in the western part of this contact.

This pluton has a poorly exposed, intermixed contact with the greenschist to the north. The granodiorite has intruded the schist along joints to form a gradational zone of breccia that consists of fragments of greenschist set in a matrix of granodiorite. This zone is as much as 600 feet thick.

The margins of the Maquinita Creek pluton are not exposed. They probably are in part parallel to the schistosity of the enclosing green-schist and in part discordant. The foliation of the granodiorite appears to be parallel to the wall-rock contacts.

The boundaries of the Duran Creek pluton are both parallel and transverse to the flow-foliation of the Tres Piedras granite, as well as to the schistosity of the Moppin metavolcanic series.

Few of the dikes or sills of Maquinita granodiorite in the Moppin greenschists are well exposed. Their emplacement seems to have been guided largely by the schistosity and lineation of the greenschist and, to a lesser extent, by joints in the wall rock.

Postemplacement History

The shearing of the Maquinita granodiorite may have been caused either by a continuation or repetition of the forces that originally caused its intrusion, by the orogenic forces that deformed the entire section of

Precambrian rocks, or by intrusion of the neighboring plutons of Tres Piedras granite. The third possibility seems unlikely, because the rocks adjacent to the Tres Piedras granite from the mouth of Spring Creek to south of Las Tablas have not been similarly sheared. The subparallelism of the lineations in the pluton and in the adjacent wall rock does not negate either of the first two possibilities. The alteration products — epidote, biotite, sodic plagioclase, and sericite — and the foliate textures that imply formation of these minerals during shearing, suggest that the temperature may have been as high as 300°C when the granodiorite was deformed. Perhaps if these plutons were subjected to a continuation of the forces that caused their emplacement during and after their crystallization they would now show a protoclastic texture. The author favors the thesis that the Maquinita granodiorite was emplaced during folding, and that it was sheared by subsequent orogenic forces.

TRES PIEDRAS GRANITE

Definition and Distribution

The granite exposed along the lower Rio Tusas and at Tusas Mountain was called the Tusas granite by Just (1937, p. 44-46), who also included under this name the Maquinita granodiorite and granite at Tres Piedras. The granite exposed at Tres Piedras and along Tusas Canyon and the lower Rio Tusas differs from that of Tusas Mountain; the rock at Tusas Mountain appears to be an atypical, very fine-grained, porphyritic phase of the former type. The granite at Tres Piedras and along the Rio Tusas is here redefined as the Tres Piedras granite, after the excellent exposures in and around that town. Other masses of this granite were found east of Hopewell in sec. 33, T. 29 N., R. 7 E., and on the lower east slope of Jawbone Mountain. Dikes of it cut the Moppin metavolcanic series near Hopewell.

Lithology

The Tres Piedras granite, as exposed in the type locality and along Tusas Canyon, is a pink, flesh-colored, or reddish-orange, faintly to well-foliated, fine- to medium-grained, quartz-microcline-albite-biotitemuscovite granite. Crude laminae and flattened knots of the two micas commonly are well developed; most of the knots show a distinct lineation plunging down the dip of the foliation. The foliation controls the weathering of the granite (pl. 10A).

Under the microscope this granite has an allotriomorphic and seriate texture. The grains of quartz and feldspar are mostly 0.5-5 mm in diameter, and the flakes of mica are 0.5-1 mm in diameter. Quartz and feldspar tend to be roughly equiaxial, but many extremely irregular grains are present. Many of the larger grains of quartz have been fractured.

The Tusas Mountain pluton of the Tres Piedras granite differs from the Tres Piedras-Tusas Canyon variety mainly in grain size and in being massive or only faintly foliated. This rock is markedly porphyritic, with 0.12 mm average grain size, and phenocrysts from 1 to 5 mm in size. The groundmass consists of microcline, quartz, albite-oligoclase, muscovite, and biotite, in order of decreasing abundance. Muscovite and biotite are segregated into faint bands in some of the porphyry. Phenocrysts of patchily perthitic microcline and of crudely bipyramidal, recrystallized quartz are abundant. Phenocrysts of plagioclase are rather scarce. The phenocrysts form about one-half the rock.

A specimen taken about 10 inches from the western margin of the Tusas Mountain pluton (pl. 10B) contains about 10 percent elongated aggregates of quartz whose rough polygonal outlines strongly suggest derivation from phenocrysts of single bipyramidal crystals of quartz. These phenocrysts(?) are assumed to have grown at a temperature above 573°C. These aggregates of quartz have their longest dimensions steeply inclined and well alined parallel to the contact with the wall rock. The remainder of the quartz, microcline, plagioclase, and muscovite are very fine grained, averaging about 0.06 mm in diameter. The texture may well be a composite one that reflects early chilling followed by partial granulation.

The granite along the west side of the Tusas Valley south from Las Tablas is medium-grained, from 1 to 3 mm in grain size, and is equigranular, with a mosaic to markedly allotriomorphic texture characterized by equant polygonal to very irregular anhedral grains. A mode of specimen 36-D-45 is given in Table 10.

The partially exposed pluton that underlies the low hill one-half to 1 mile east of Hopewell is composed of porphyritic granite similar to the Tusas Mountain type. However, the foliation is better developed here. About one-third of this rock is xenoliths of greenschist, from a few feet to a few tens of feet long, most of which are crudely tabular or elongate parallel to the foliation in the granite.

The dark-gray to pink sheared granite of the lower east slope of Jawbone Mountain is compositionally similar to the Tres Piedras granite to the south, but is texturally similar to the flaserlike variety of the Maquinita granodiorite. It may be a sheared variant of the regular Tres Piedras granite, or it may be a wholly different rock. Only 1 square mile of it is exposed. It is tentatively correlated with the Tres Piedras granite.

Analyses of two typical specimens of Tres Piedras granite are shown in Table 10.

The Tres Piedras granite contains more silica and potash, and less alumina, lime, and mafic constituents than the average given by Daly; the soda content is approximately the same. The low amounts of lime, magnesia, and iron oxides in the Tres Piedras granite are notable; they are probably the result of differentiation of the source magma prior to its intrusion to form these plutons. The high silica content may be partly due to assimilation of quartzite. Assimilation of muscovite would

TABLE 10. CHEMICAL ANALYSIS, NORM, AND MODES OF TRES PIEDRAS GRANITE
(In percent)

Chemical Analysis	(1)*†	(2)‡	Norm*		Modes	(1)*	(2)§
SiO_2	77.25	70.18	Quartz	39.5	Quartz	42	44
TiO_2	0.14	0.39	Orthoclase	33.5	Microcline	40	36
Al_2O_3	11.56	14.47	Albite	22.5	Albite	12	15
Fe_2O_3	0.55	1.57	Anorthite	1.1	Muscovite	3	3
FeO	0.92	1.78	Magnetite	0.7	Biotite	3	2
MnO	0.03	0.12	Ilmenite	0.3	Accessory		
MgO	0.11	0.88	Enstatite	0.3	minerals	tr	tr
CaO	0.27	1.99	Ferrosilite	1.1			
Na_2O	2.65	3.48	Corundum	0.6			
K_2O	5.59	4.11		98.6			
P_2O_5	0.06	0.19					
H_2O+	0.49	0.84					
H_2O-	0.04						
CO_2	0.00						
	99.66	100.00					

* Specimen 36-D-31. From Tusas Canyon, SE¼NW¼ sec. 35, T. 28 N., R. 8 E.
† Analyst: H. B. Wiik, Helsinki, Finland.
‡ Daly's (1936, p. 2) average granite.
§ Specimen 36-D-45. From south of the junction of Tusas and Kiawa Canyons, NW¼NW¼ sec. 13, T. 27 N., R. 8 E.

increase the alumina content of the granite as well as the potash content, so assimilation of this mineral does not seem likely.

The analyses of the Tres Piedras granite and the Burned Mountain metarhyolite are remarkably similar. The similarity may be fortuitous, or there may be a genetic relationship between the two rocks. They may be derived from the same parent magma. It is difficult, however, to visualize a magma chamber deep in the crust which would first produce the rhyolite sills, and then remain unchanged through an intense orogeny, during which the Maquinita granodiorite was generated, and which would subsequently produce the plutons of granite. The granite and rhyolite magmas may have been produced by fusion of similar rocks in the lower portions of the crust. There is no evidence to support either speculation.

Relation to Wall Rocks

A clearly defined contact of the Tres Piedras granite with the Kiawa Mountain formation extends from a point near the mouth of Spring Creek to the mouth of Kiawa Canyon. It adjoins relatively unaltered (slightly muscovitized) quartzite, and dips southwestward, parallel to the bedding in the quartzite. A septum of metarhyolite (and muscovite quartzite?) a few hundred feet thick lies within the granite along the northeast side of Tusas Canyon in sec. 2, T. 27 N., R. 8 E. The foliation of the metarhyolite in the septum is parallel, or very nearly so, to the foliation of the granite.

The contact between Tres Piedras granite and Petaca schist from Kiawa Canyon southeastward is poorly defined. This contact has been studied by Just (1937, p. 45-46) and Jahns (1946, p. 22). It trends parallel to the schistosity of the wall rock, ranges from a few feet to a diffuse zone about 2,000 feet in breadth, and consists mainly of irregular sills that form a hybrid zone between granite and Petaca schist. These sills have absorbed varying amounts of muscovitic quartzite and metarhyolite. The septa between the sills are similar to the rock west of the contact. Any feldspathization in this contact zone would be difficult to recognize, as the feldspathized quartzite probably would be similar to the muscovitized metarhyolite. The granite appears to have intruded the Petaca schist parallel to the axial planes of minor folds, mainly as a swarm of sills. The intrusion was followed by hydrothermal reactions between wall rock and granite, which resulted in hybridization of xenolithic material and of granite along the contacts.

The contact of the Tusas Mountain pluton is sharp and lies parallel to the schistosity of the amphibolite along Cow Creek almost to the west end of the pluton, where the trend of the contact makes low angles with the schistosity and in detail truncates the schistosity. This contact is sharply delineated. The contact between the granite and the granodiorite to the north is poorly exposed, but is probably sharp also. The faint foliation of the granite is parallel to its contacts. Few xenoliths of country rock were seen in the granite.

The northern margin of the granite pluton east of Hopewell shows slightly gradational relations with the granodiorite at the west end, but is sharp at the east end next to Duran Creek.

Because of lack of exposures, the boundary of the pluton east of Jawbone Mountain is inferred only. It may truncate the Jawbone conglomerate and Moppin greenschists along its entire length.

PEGMATITES

The granitic pegmatites of Las Tablas quadrangle have been mapped and described in detail by Jahns (1946), whose data have served as the basis for the following paragraphs on general distribution, external structural features, zoning and lithology, and genesis of the pegmatite bodies.

Distribution

Three major groups of pegmatites are present on La Jarita Mesa. These are the Kiawa group, which lies mostly in the W½ sec. 11, T. 27 N., R. 8 E.; the Persimmon Peak-Las Tablas group, which extends from west of Las Tablas almost to Poso Spring; and the La Jarita-Apache group, which extends from about 1 mile east of Big Rock to a point in the lower part of Apache Canyon, west-southwest of Petaca.

External Structural Features

Most of the pegmatite bodies are dikes, sills, pipes, and pods; some are scoop shaped, or have the form of upright or inverted troughs. They vary in size, with a minimum horizontal distance between ends at the present surface of 75 feet, a maximum of 1,430 feet, and an average of 410 feet. These figures include pegmatites in the quadrangle to the south, but are based only upon those pegmatites that have been worked for mica. The average maximum breadth of outcrop is 49 feet, ranging from a few inches to as much as 250 feet.

The sills strike north to northwest and dip west to southwest; the dikes and elongate pods strike west-northwest to west-southwest and dip in a general northerly direction. The axes of the pegmatites plunge gently to moderately northwest to southwest. These shapes have been closely controlled by the schistosity of the Petaca schist, by the plunge of minor folds, and by joints (which are discussed below with the structure of the Precambrian rocks).

Zoning and Lithology

'The pegmatites on La Jarita Mesa commonly are zoned as follows: border zone of fine- to medium-grained microcline and quartz with a little mica; wall zone of coarse microcline and quartz with minor mica, garnet, fluorite, and beryl; intermediate zones (rarely complete) of variable composition such as coarse blocky microcline, coarse graphite granite, or massive quartz with disseminated microcline; cores, commonly of massive quartz, plus or minus scattered microcline. These zones vary widely in degree of development in the various pegmatites. Fracture fillings, generally of quartz with or without albite, samarskite, fine-grained mica, sulfides, and bismuth minerals, are present in many of the pegmatites. Replacement bodies, controlled largely by fractures in preexisting pegmatite and consisting chiefly of albite and muscovite, occur in most of the bodies as veinlets or other secondary masses.

Genesis

The formation of the pegmatites in the Petaca district has been summarized by Jahns (1946, p. 74) as follows:

> The Petaca pegmatites can be treated genetically as a unit. Their primary minerals indicate that they may be classified as members of the potash-rich clan, which contains small amounts of soda in the form of early-stage albite-oligoclase crystals, crystalline masses, and perthitic intergrowths. The original microcline, albite-oligoclase, and quartz were distributed to form structural units or zones in most deposits, and were accompanied by garnet, green fluorite, beryl, mica, and small quantities of magnetite and ilmenite. These accessories appear to have formed during the primary consolidation of the pegmatite bodies, and are most common in the outer zones.
>
> A later stage — characterized by the widespread activity of hydrothermal solutions rich in soda, silica, and alumina, and containing appreciable quantities of columbium, tantalum, beryllium, thorium, uranium, the rare earths,

fluorine, bismuth, copper, and other elements — is represented by the development of abundant albite, muscovite, and many accessory species. These minerals were formed through replacement of the primary pegmatitic constituents by solutions that were guided mainly by fractures. Mineral and zone pseudomorphs were formed in some places, and elsewhere replacing solutions penetrated the walls of joints, fissures, and other openings to attack the earlier minerals and form rosettes, tabular bodies, and less regular masses of new material. Not only were most of the later minerals plainly formed at the expense of pre-existing material, but their structural control demonstrates that this replaced material was sufficiently solid to fracture under stress. Additional beryl, garnet, and fluorite were developed during the hydro-thermal or albitization stage; and the physical and chemical properties of each appear to differ somewhat from those of earlier-formed, primary masses of the same species. Relatively small quantities of the accessory minerals genetically identified with the stage of albitization were probably also formed during the last stage of crystallization of the pegmatite zones.

During the final stage, which was characterized by fracturing and the formation of veins of smoky quartz, replacement processes appear to have become subordinate to simple open-space filling. Uranium-thorium minerals, muscovite, and some of the same accessories developed during the preceding albitization accompanied the vein quartz.

TERTIARY INTRUSIVE ROCKS

BISCARA INTRUSIVE ANDESITE PORPHYRY

An eastward forked dike of andesine-hornblende-biotite andesite intrudes the Biscara member of the Los Pinos formation in Canon del Agua, 2 miles northeast of Las Tablas. The vertical dike trends N. 70° E. It is about 350 feet thick at the western end, where the two forks join. The north fork is about 150 and the south fork about 100 feet thick. The dike consists of breccia in which fragments of tuff, conglomerate, and dark felsite are cemented by the andesite porphyry. It has been described by Butler (1946, p. 59-61), who also has noted its similarity to the breccia exposed 1.4 miles to the northwest.

Structural Geology

GENERAL FEATURES

The Precambrian layered rocks have been folded on a large scale into the Hopewell anticline, the trace of whose axial plane lies just north of Hopewell and the Kiawa syncline, the trace of whose axial plane extends from El Vallecito ranch to Kiawa Mountain and thence south-southeastward on La Jarita Mesa. Two subsidiary folds, the Poso anticline and the Big Rock syncline, lie on the southwest flank of the Kiawa syncline. All the Precambrian strata have been intensely deformed, and numerous minor folds whose flank-to-flank dimensions range from a fraction of an inch to several thousand feet are present. All the folds in the quadrangle plunge gently to steeply from northwest to southwest.

The Tertiary rocks have been gently tilted to the east-northeast, and have been faulted along two zones that extend along the Tusas and Vallecitos Valleys. The La Jarita Mesa-Jawbone Mountain highland has been uplifted along the Vallecitos Valley fault zone relative to the highland area to the southwest, and has been depressed along the Tusas Valley fault zone relative to the Taos Plateau.

STRUCTURE OF THE PRECAMBRIAN LAYERED ROCKS

KIAWA SYNCLINE

The Kiawa syncline underlies La Jarita Mesa, Kiawa Canyon, Kiawa Mountain, Quartzite Peak, and all of the upper Vallecitos Valley. The dips of beds are generally steep along the axial region of this fold, where they range between 70 degrees north and 70 degrees south, but some dips are as low as 26 degrees. Presumably they represent flanks of minor folds.

The northeastern limb of the syncline dips steeply southwestward from the upper Rio Vallecitos to Cleveland Gulch, but from Spring Creek Canyon to the latitude of Las Tablas the dip decreases to an average of 55 or 60 degrees.

In the vicinity of La Jara Canyon the southwestern limb of the fold dips steeply northward. Farther east, south of Kiawa Mountain, and to the southeast, west of Cañon Plaza and the lower Rio Vallecitos, this limb is overturned and has steep to moderate southwest dips.

The axial plane of the Kiawa syncline trends N. 60° W. to N. 65° W. at Kiawa Mountain. On the east side of the mountain the axial plane swings to a strike of N. 40° W., and has moderate southwestern dips. On La Jarita Mesa, west of Las Tablas and Petaca, the axial planes of all the folds strike about N. 30° W., and dip about 40° SW.

The plunges of drag folds suggest that the Kiawa syncline plunges 15-30 degrees in a generally western direction in the area west of Kiawa

Mountain, and 25 and 55 degrees in southwestern to northwestern directions in the muscovitized rock east and southeast of Kiawa Mountain. Drag folds are few in the homoclinal Ortega quartzite; the few that were measured have plunges from 19 to 46 degrees to the northwest.

The upper quartzite member of the Kiawa Mountain formation shows reversals of top-and-bottom relationships and some anomalous low dips that imply the presence of minor folds. The flank-to-flank dimensions of these folds appear to be of the order of several hundred to several thousand feet. As stated above, this member is largely vitreous quartzite of essentially uniform competence, so that one part of it may well be as intensely folded as another part. The absence of marker beds and the lack of continuously exposed transverse sections make it impossible to evaluate definitely the nature and extent of minor folding.

The amphibolite member of the Kiawa Mountain formation and the immediately overlying quartzite, the lower quartzite member of that formation and the Big Rock conglomerate member of that formation, and the Petaca schist have all been intensely folded, mostly isoclinally, on five general scales. The intensity of folding increases rather markedly from the vicinity of Kiawa Lake to points 2-3 miles to the east and southeast. Folds with flank-to-flank dimensions of a few inches, a few feet, a few tens of feet, a few hundreds of feet, and a few thousands of feet are internested, or superposed on each other. The larger minor folds are well shown by the amphibolite layer itself, which changes its outcrop breadth in odd multiples where it is singly to multiply drag folded. In a single isoclinal drag fold, for example, which has a simple S-shaped plan, the outcrop breadth transverse to the flanks is three times the breadth of a single fold limb.

Folds of few-inch to few-foot dimensions commonly have sharply angular crests and troughs, with relatively straight limbs. The schistosity of the rock is parallel to the bedding on the limbs, and is parallel to subparallel to the axial planes of the folds on their crests and troughs. The fold axes show locally uniform plunges; in some areas several hundred feet square they vary less than 5 degrees from the mean. Shear folds *have* been developed in the axial parts of some of the small folds. The shear surfaces are parallel to the axial planes of adjacent flow-folded beds, and they are spaced from one-eighth of an inch to several inches apart.

HOPEWELL ANTICLINE

The Kiawa Mountain formation passes northwestward out of Las Tablas quadrangle 0.8 mile southwest of Hopewell, but its surface trace, which is partly covered by Tertiary rocks, reenters the quadrangle on the opposite side of a large fold at Jawbone Mountain. This fold is a northwestward plunging anticline, here named after the town of Hopewell. The Jawbone conglomerate, on the northeast limb and close to the nose of the fold, dips to the northwest at moderate angles. The tops of its beds face in the same direction. Much of it has a very steep, west-northwestward trending axial-plane cleavage. The trace of this cleavage on the bedding is a line that plunges moderately to the west-northwest, and this line probably is subparallel to the plunge of the Hopewell anticline. Minor folds are common in the Jawbone conglomerate, but, as in the upper member of the Kiawa Mountain formation to the south, they cannot be evaluated quantitatively. Much of the Hopewell anticline is covered and concealed by strata of the Los Pinos formation.

POSO ANTICLINE

The Poso anticline, named after Poso Spring (NE¼NE¼ sec. 27, T. 27 N., R. 8 E.), is a subsidiary fold that lies on the southwest flank of the Kiawa syncline. Its axis trends northwest, and plunges moderately in the same direction. The anticlinal nose of the Big Rock conglomerate underlies the NW¼ sec. 22, T. 27 N., R. 8 E. The fold dies out to the northwest; southwest of Kiawa Lake it merges into a less folded portion of the Kiawa syncline.

The Big Rock conglomerate clearly delineates the Poso anticline. This unit, which is 50-100 feet thick, enters the nose of the anticline from the greatly drag folded and thickened southwestern limb and from the very poorly exposed northeastern limb, which is shown in outcrop in Plate 11.

The Poso anticline is a composite fold consisting of two complex anticlines that are separated by a relatively narrow syncline. This structure is outlined by the Big Rock conglomerate, which has been folded into two anticlines with flank-to-flank dimensions of 1,600 feet and 1,400 feet, and a syncline that is only 250 feet across. These folds are shown on Plate 1. Their average plunge is about 35° WNW.

The southwestern anticlinal nose of Big Rock conglomerate is almost wholly conglomerate, whereas the northeastern nose of the same member is partly conglomerate and partly gray micaceous quartzite. This lithologic contrast is believed to be due to a rapid facies change.

The nature of the folding in the Petaca schist that underlies the Big Rock conglomerate and appears in the Poso anticline is shown in the map of the area just southeast of Poso Spring (pl. 3). This area lies in the axial region of the anticline. The rocks on the flanks of this anticline, however, are folded as intensely as those closer to the axis (pl. 11).

The folds in this area are clearly marked by layers of biotite-epidote-quartz-oligoclase schist, pink to flesh-colored, feldspathic, micaceous quartzite, and micaceous metarhyolite. The folds grade in flank-to-flank dimension from less than an inch to several hundred feet. The smaller folds are nested in the larger ones, as in the Kiawa syncline. All the folds gradually diminish along their axial planes in directions both parallel and normal to their plunge. The smallest folds typically have the most angular crests and troughs, with straight, nonparallel limbs. The angularity of the noses decreases with increasing size of the folds, and the limbs become essentially parallel in folds with flank-to-flank dimensions of 10 feet or more. All of the larger folds are isoclinal.

Thickening and thinning are of about the same magnitude in the dark-colored schist and in the light-colored quartz-mica-feldspar rocks, and, in general, the two rock types have been similar in their response to the folding. The maximum change in thickness along a limb is from 6 inches to 14 feet, as measured on the second lowest layer of dark-colored schist. Thickening along fold limbs, as well as at the noses, appears to have resulted from flowage of material and from superposed drag folding.

The schistosity is parallel to the bedding, both along the limbs and at the noses of the folds. The drag folds plunge consistently to the northwest, with extremes of 25 and 58 degrees, and an average of 39 degrees.

BIG ROCK SYNCLINE

The Big Rock syncline is an overturned fold which, together with the Poso anticline, can be regarded as a large drag fold on the southwest flank of the Kiawa syncline. This small syncline has a northwest trend and plunges moderately to the northwest. The Big Rock conglomerate, which clearly outlines the fold, extends along the edge of La Jarita Mesa (in secs. 20 and 21, T. 27 N., R. 8 E.) with a minor flexure, passes southward beneath the Tertiary rocks with an outcrop breadth of 50 feet, and emerges with a breadth of 400 feet in the nose of the syncline. The upper and lower surfaces of the conglomerate unit are not well exposed in the axial region of the Big Rock fold, except on the northeast side; but the general position shown on Plate 1 is accurate to within 100 feet in most places and to within 200 feet elsewhere.

The Big Rock conglomerate and other strata in the syncline are intensely folded in the same manner as the rocks at Poso Spring (pl. 11). The axes of nearly all drag folds in the syncline plunge 30°-45° NW.

GENERAL DISCUSSION

Development of the Folds

The major folds in Las Tablas quadrangle, the Kiawa syncline and the Hopewell anticline, probably were flexures during early stages of their development. As folding continued and the limbs approached parallelism, the amount of flexure between adjacent beds increased. Minor folds were the main result of this interbed deformation. The trends and plunges of the minor folds are markedly consistent and probably were governed in large part by the trends and plunges of the major folds.

As the limbs of the major folds were pushed closer to each other, the vertical component of the distance between the crests and troughs of the folds tended to become larger. The rocks also were strained as they were folded, mainly in a direction parallel to the dip of the beds, as

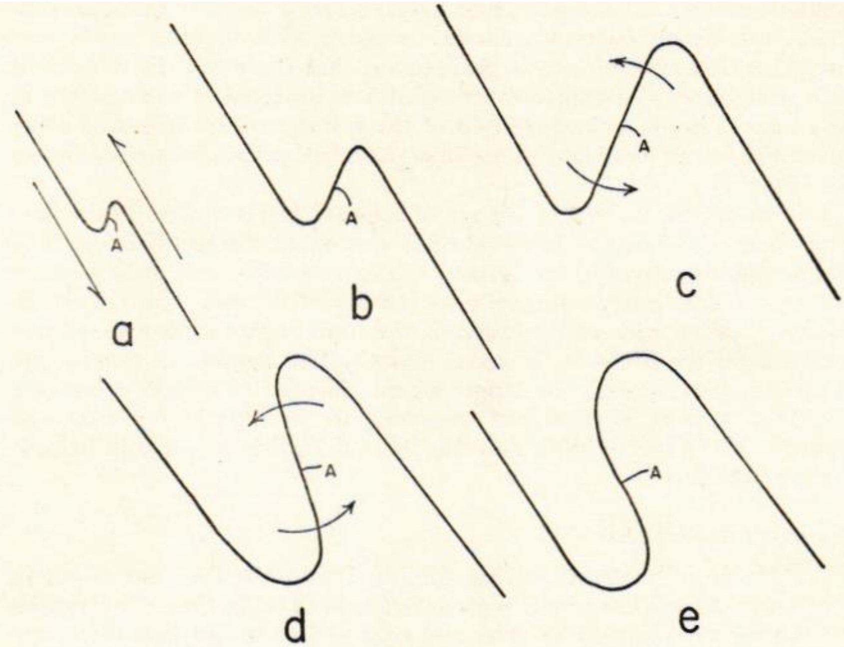

Figure 3.

SCHEMATIC REPRESENTATION OF A CHEVRON-TYPE FOLD, WITH SUBSEQUENT GROWTH INTO AN ISOCLINAL FOLD.

a. Initial bend developed at point A, in response to interbed shear, which is denoted by arrows. b. At a slightly later stage than (a), the crest and trough have migrated away from A, with growth of the dragged limb of the fold. c. The dragged limb has grown larger and has been rotated in a counterclockwise direction. d. Flowage of rock from the flanks to the noses of folds has begun, and the crests and troughs are becoming less angular in cross-sectional pattern. e. Later stage of development of the fold. The dragged limb is larger than in (d); it has been rotated so that it is parallel to the undragged limbs, and the angularity of the crest and trough have been lessened by flowage of rock from the flanks toward the axial regions of the fold.

discussed below. Thus, the vertical component of the distance between adjacent crests and troughs was increased by strain of the material involved. The minor folds, however, shortened the limbs of the major folds, and tended to reduce the vertical dimensions of these folds.

The vertical dimensions of the Kiawa syncline and Hopewell anticline increased with continued folding, for if one assumes an approximately constant cross-sectional area of rock taken vertically and normal to the axial planes; horizontal shortening would necessitate vertical lengthening.

Vertical cross-sections taken normal to the axial planes of all folds that are more than about 10 feet in flank-to-flank dimension show smooth and rounded apices, and parallel limbs. Smaller folds, in contrast, are zigzag in cross-sectional pattern, with angular crests and troughs. One type of fold is gradational into the other. Development of the zigzag or chevron folds probably was initiated at weak points in the beds. The crests and troughs of the resultant folds migrated away from the initial bend or flexure in a given bed, as schematically shown in Figure 3.

Limbs of the chevron folds are of approximately uniform thickness, but those of the larger, isoclinal folds vary markedly in thickness. The larger folds probably were formed as chevron folds, and grew to their present dimensions in response to continued flexure. The change in shape entailed with such growth of the folds is due to flowage of material from the limbs to the axial regions. The amount of flowage appears to increase with the length of the dragged or middle limb of a single drag fold, as measured transverse to the axes of the crest and trough. The dragged limb also was rotated slightly as the fold became larger (fig. 3).

Flowage of Material

Flowage of material during folding is indicated by variations in thickness of beds from axial regions to flanks of folds. Other indications of flowage are elongate pebbles and cobbles in the conglomerates, and well-alined chlorite knots in masses of greenschist.

The elongate pebbles in the Big Rock conglomerate commonly are triaxial ellipsoids with axial ratios of about 1:2:3. The short axis generally is normal to the bedding, the intermediate axis is parallel to the strike, and the long axis is parallel to the dip of the beds. If the pebbles are assumed to have been originally spherical in shape, a pebble with an original radius of unity would have semiaxes of 0.55, 1.10, and 1.65 after folding. If the strain of the entire conglomerate layer is similar to that of the pebbles, the layer is now only 55 percent as thick as it was prior to folding; it has been extended 10 percent parallel to the strike, and 65 percent parallel to the dip. The assumptions as to the original shape of the pebbles and homogeneous strain are, of course, problematical.

The original interstices of the conglomerate have been closed. The change in thickness due to this closure of voids is probably between 5 and 20 percent, but, of course, the degree of cementation of the rock prior to folding is not known.

The strain of the Big Rock conglomerate during folding may have altered the thickness to about one-half of its original value. The rock probably was extended slightly parallel to the strikes of its axial planes, and was stretched along the dip to about one and one-half times its original dimension. Deformation of the Ortega quartzite, the quartzite and Jawbone conglomerate members of the Kiawa Mountain formation, and the Burned Mountain metarhyolite probably was of similar magnitude.

The original shapes and orientations of the knots of chlorite in the greenschists of the Moppin metavolcanic series are not known; hence the amount of their strain can neither be determined nor even estimated. All of the metabasaltic rocks in Las Tablas quadrangle appear to have been strained at least as much as the quartz-feldspar rocks.

Mechanism of Deformation

There appear to be three general means by which the rocks that comprise these folds could have been deformed: (1) gliding, (2) rotation of elongate grains, with or without fracture of individual grains, and (3) solution and redeposition of mineral constituents.

Almost all the grains of quartz in the metamorphic rocks of Las Tablas quadrangle show (Boehm) lamellae, which are thought to be a result of translation-gliding (Fairbairn, 1949, p. 124-129). If these lamellae are a glide phenomenon, the quartz-feldspar rocks may well owe much of their deformation to this general mechanism.

Rotation of grains has occurred in the Burned Mountain metarhyolite, as shown by rolled relict phenocrysts. Some granulation of originally euhedral phenocrysts of quartz also has taken place in the metarhyolite, producing mosaics that show a crude bipyramidal external form. The individual grains in these aggregates are within a few degrees of an average crystallographic orientation, as shown by their extinction positions under the microscope. There is no convincing evidence for deformation by pronounced rotation of nonequidimensional grains, or for splintering and rotation of quartz grains.

Solution and redeposition of material have been of major importance in the deformation of the metabasaltic rocks, as discussed below in the section on metamorphism. Folding and metamorphism of these rocks probably were synchronous, as implied by the well-developed lineation and lack of both pretectonic and post-tectonic textural features, such as granulation, bending of cleavages, distorted crystal outlines, and twin lamellae. The parallelism of the c-axes of the hornblende crystals in the amphibolite layers probably is due to preferential growth parallel to the direction of maximum strain, and to rotation of grains during shear of the rock. Translation-gliding may have occurred in both greenschist and amphibolite, but there is no positive evidence that it did.

STRUCTURE OF THE TUSAS VALLEY AND VALLECITOS VALLEY FAULT ZONES

GENERAL STATEMENT

Fault zones extend along the valleys of the Rio Tusas and the Rio Vallecitos, and the La Jarita Mesa-Jawbone Mountain highland is situated between them. This highland has been raised along the Vallecitos

Valley zone relative to the Tierra Amarilla highland to the southwest, and has been lowered along the Tusas Valley zone relative to the Taos Plateau on the northeast.

TUSAS VALLEY FAULT ZONE

The Tusas Valley fault zone ranges from a single fault to a zone about 3 miles in width, and has been described in detail by Butler (1946, p. 155-160). All the faults southeast of Deer Trail Junction lie northeast of the Rio Tusas; northeast of the junction the zone is wider, and faults cut rocks on both sides of the creek. The zone consists of one to five mappable faults. These are of two types, which have been termed main faults and crossfaults by Butler. The main faults trend N. 26° W. to N. 62° W., and range in mappable length from less than 1 mile to more than 16 miles. The crossfaults are transverse to the main faults; their average trend, which is less regular than that of the main faults, is approximately N. 45° E.

Most of the displacement along the fault zone has taken place along the main breaks. The northeastern side of the zone has been raised relative to its southwestern side. The main faults appear to be of the dip-slip type, but small horizontal components of movement may be represented.

The crossfaults transfer displacement from one main fault to another, and generally involve lowering of the northwest side relative to the southeast side. The movement probably was dip-slip in nature.

The total vertical displacement along the fault zone is variable; approximate values include: 300-400 feet southeast of Las Tablas; 600 feet northeast of Las Tablas; more than 700 feet east of Tusas; about 1,200 feet at Biscara Canyon (Butler, 1946, p. 158); and probably more than 1,200 feet in areas to the northwest. The most prominent main fault extends from upper Cañon del Agua to Biscara Canyon and Deer Trail Junction, and along the upper part of the Rio Tusas beyond the margin of the quadrangle. All movement along the Tusas Valley fault zone, from the Tusas-Tres Piedras road to a point one-half mile southeast of the mouth of Biscara Canyon, has been along this extensive main break.

VALLECITOS VALLEY FAULT ZONE

The Vallecitos Valley fault zone is at least 4 miles wide, and thus is broader than the Tusas Valley zone. Further, the faults within it are less regularly distributed. At Cañon Plaza and along and west of the lower Vallecitos Valley, the zone comprises 6-8 main faults, which have an average trend of about N. 40° W. Several crossfaults range in trend from N. 45° W. through east-west to N. 60° E. Some of the main faults and crossfaults divide blocks of Ortega quartzite from blocks in which Cordito strata appear at the surface; this juxtaposition implies a minimum vertical displacement of about 500 feet.

The minimum vertical displacement of the Vallecitos Valley fault zone, from La Jarita Mesa to the exposures of the Cordito member of the Los Pinos formation west of Cañon Plaza, is from 700-800 feet. The southwestern side of the zone has been lowered relative to the northeastern side. All faults in this zone appear to be of the dip-slip type.

The displacement of the single fault that extends from Cañon Plaza along the Rio Vallecitos to Escondida Creek ranges from 200 feet to at least 400 feet. The fault extending N. 80° W. from upper La Jara Canyon to Jarosita Canyon has at least 500 feet of vertical displacement, with the north block relatively depressed; this movement sense is contrary to that in the remainder of the fault zone. The curved fault along the upper Rio Vallecitos may have a vertical displacement of more than 600 feet, with the southwest block relatively dropped.

OTHER FAULTS

Several minor faults are present in Spring Creek Canyon from sec. 4, T. 27 N., R. 8 E., to sec. 31, T. 28 N., R. 8 E. These faults are both parallel and transverse to the canyon, are mainly very steep, and show vertical displacements of as much as 100 feet.

Other faults are exposed in the Precambrian rocks, but all those seen are too small to be shown on a map with the scale of Plate 1.

STRUCTURE OF THE TERTIARY ROCKS

The Tertiary rocks of Las Tablas quadrangle have been tilted gently northeastward. Most of the beds strike about N. 45° W. and dip at angles of 2-4 degrees (Butler, 1946, p. 152).

If the Tertiary strata of Las Tablas quadrangle were deposited by streams that drained source areas to the east and northeast, as seems likely for reasons given in a preceding section, they originally must have dipped to the west and southwest. The Tertiary rocks, therefore, were tilted after deposition more than their present attitudes suggest.

The present slope of the surfaces underlain by Tertiary formations is toward the east and southeast. The drainage pattern that has been developed in these rocks is composite; it consists of streams that trend northeastward to east-northeastward and flow into southeastward trending streams. The former streams lie upon strata that have not been cut by faults of Tertiary age, whereas the latter, southeastward trending streams lie in the Tusas Valley and Vallecitos Valley fault zones, and appear to have captured the older northeastward flowing streams.

The writer infers that the Tertiary rocks were tilted to the northeast, that parallel, northeastward flowing streams were developed on this sloping surface, that subsequently the Tusas Valley and Vallecitos Valley fault zones were developed, and that streams controlled by the fault zones altered the lower courses of the earlier formed northeastward flowing streams. There was some tilting of the entire area toward the southeast, probably in conjunction with development of the two fault zones.

Metamorphism

GENERAL STATEMENT

The layered Precambrian rocks of Las Tablas quadrangle have been regionally metamorphosed, with attendant development, in rocks of basaltic composition, of mineral assemblages ranging from chloritealbite-epidote-muscovite greenschist to hornblende-andesine amphibolite. Quartzite associated with all of the metamorphosed basaltic rocks contains kyanite. Minerals developed in the metamorphic rocks of non-basaltic composition include microcline, muscovite, albite, oligoclase, biotite, epidote, garnet, staurolite, and chlorite.

In the La Jarita Mesa area, the regionally metamorphosed rocks also were hydrothermally altered, with metasomatic introduction of muscovite, chlorite, biotite, quartz, garnet, and lesser amounts of other minerals. This alteration is spatially and genetically related to masses of granite pegmatite, and in this paper such alteration is referred to as pegmatitic-hydrothermal metamorphism.

Several masses of quartz-kyanite rock of different(?) metasomatic origin also underlie parts of La Jarita Mesa.

GENERAL DISCUSSION OF REGIONAL METAMORPHISM

Definition

Regional metamorphism can be defined as metamorphism that takes place in folded mountain belts in response to increased temperatures and confining pressures, under conditions of high stress. Plutonic rocks commonly are associated with regionally metamorphosed rocks, but there may or may not be a simple relationship between intensity of metamorphism and the known or inferred position of the plutonic rocks.

Equilibrium

In a rock that is formed under a given set of conditions, the components will tend to arrange themselves in an assemblage of minerals that will not change further with time. Such an assemblage thus would be in equilibrium. The number of minerals that are developed in a given rock is less than or equal to the number of components involved, as stated by Goldschmidt (1911, p. 123). The simple oxides of the rock, such as $Si0_2$, CaO, FeO, etc., may be considered as components. A rock with more minerals than components would not be in equilibrium, and would tend to change to one with fewer mineral phases.

Thermodynamically, the free energy of a system tends toward a min-mum value. The Gibbs free energy, F, is defined:

$$F = H — TS,$$

where H is the heat content, T the absolute temperature, and S the entropy. For this equation, pressure is assumed to be constant, and surface, magnetic, gravitational, and other external forces are assumed to be zero.

$$H = \int_0^T C_p dT,$$

where CP is the heat capacity at constant pressure, and

$$S = \int_0^T \frac{C_p dT}{T}.$$

The entropy can be considered as a rough measure of the randomness or disorder of the system. At zero entropy each atom would be in its proper site in a given mineral structure, and there would be no thermal oscillation of atoms or interchange of one specimen of atom with another. Further, at absolute zero the free energy is equal to the heat content, or the lattice energy.

Temperature Effect

As the temperature of a system is raised from zero, the atoms will begin to oscillate about their positions in the lattice, and they may interchange with one another or even diffuse through interstices in the lattice. In other words, the entropy increases, and does so with increasing temperature in a logarithmic manner, as implied by the integral given above. The heat content, in contrast, increases with temperature in approximately linear fashion. Thus, the TS term in the free-energy equation increases more rapidly with temperature than the heat content does. At high temperatures the entropy is very important in determining the free energy, and hence the stability, of the system. The relationship of entropy to crystal structure has been discussed qualitatively by Buerger (1948, 1951), who associates low temperatures with low energy, low entropy, and "collapsed" or compact structures, and associates high temperatures with high energy, high entropy, and open structures.

One important aspect of mineral stability in metamorphic rocks is the variation of crystal structures with metamorphic intensity or grade. The basic structural unit in silicate minerals is a silicon atom surrounded tetrahedrally by four oxygen atoms. In some minerals these tetrahedra are separate, with none of their oxygen atoms shared by adjacent tetrahedra, but in most silicate minerals the tetrahedra are linked into three-dimensional networks, sheets, double chains, single chains, or pairs. Aluminum atoms may substitute for silicon atoms in the tetrahedra, with attendant addition of a univalent cation in the mineral structure to maintain electrical neutrality. Other cations and oxygen atoms are present in many silicates; they serve to bond the tetrahedral groups together.

Buerger has discussed the general effect of temperature on the stability of silicates. He states (1948, p. 116-117) that

> . . . it must be evident that thermal agitation sufficient to disintegrate a structure of linked tetrahedra must leave fragments of simpler linking. Thus a mica sheet could conceivably be disintegrated into amphibole double chains, pyroxene single chains, melilite pairs, or single unshared tetrahedra, all plus a residue. In a similar manner any of the linked structures higher in the series can be disintegrated into fragments of structures having less sharing. Thus with increasing temperature the breakdown sequence is networks, multiple chains, single chains, tetrahedron pairs, and single tetrahedra . . .

An example of a part of such a structural sequence is discussed below in connection with the metamorphism of basalt.

Some atoms can have different numbers of closest neighbors. Aluminum, for instance, has either six oxygen neighbors in octahedral positions, or four, in tetrahedral arrangement. The octahedrally coordinated aluminum tends to occur in minerals formed at low temperatures, and tetrahedrally coordinated aluminum in minerals formed at high temperatures. As Buerger (1948, p. 115) has stated:

> The general tendency for lower coordinations at higher temperatures appears to be a matter of high entropy coupled with lower internal energy. Atoms in lower coordination are freer to wander over larger volumes, and thus have larger entropies. At the same time, if the bond is electrostatic, and the atom can assume either high or low coordination, the low coordination is the one of high energy. In this way the free energy. . . . is minimized by high coordination at low temperature and low coordination at high temperature.

Pressure Effects

The influence of confining pressure on the assemblages that are developed in metamorphic rocks is not well known. Increase of pressure tends to shift equilibria toward mineral assemblages of smaller volume. For instance, the polymorphs of composition Al_2SiO_5 should be increasingly stable with pressure in the sequence andalusite-sillimanite-kyanite. However, much more data are needed to determine the quantitative effects of pressure on silicate systems.

From laboratory studies in the MgO-Al_2O_3-SiO_2-H_2O system, Yoder (1952, p. 617) has given the opinion

> . . . that further experimentation in other systems will demonstrate an insensitivity to pressure for most metamorphic reactions except in the very *lowest* pressure regions. Pressure is significant in determining metamorphic assemblages only where polymorphic transition curves, which are essentially vertical, cross the steeply sloping reaction curves (Bowen's petrogenetic grid).

The pressure of water vapor in a system should increase the stability of hydrous phases, such as micas and amphiboles. Yoder (1952, p. 615) has listed water-deficient assemblages and excess-water assemblages in the system $MgO\text{-}Al_2O_3\text{-}SiO_2\text{-}H_2O$ formed at ca. 600°C. and under 15,000 pounds per square inch of vapor pressure. The various assemblages that he obtained are compatible with the greenschist, epidote amphibolite, amphibolite, pyroxene-hornfels, and sanidinite facies, and possibly with the eclogite facies.

Mechanism of Metamorphism

Metamorphism commonly involves growth of preexisting grains, and nucleation and growth of new grains. The former process implies that some grains grow at the expense of others, and that some atoms diffuse from their former sites to positions on growing crystals. The latter process, nucleation and growth, involves separation of atoms or groups of atoms from preexisting crystals, diffusion of these atoms to sites where nucleation occurs, and growth of nuclei as more atoms reach them. The actual mechanism of metamorphism, therefore, involves four processes: partial or complete breakdown of preexisting minerals, diffusion, nucleation, and grain growth.

Petrographic relationships have been used in this report to infer relative stabilities of minerals in the metamorphic rocks of Las Tablas quadrangle. The sequences of mineral assemblages that were developed during an increase of metamorphic intensity are given farther on.

Diffusion

Diffusion takes place in response to gradients of concentration, temperature, or pressure. In a small volume of metamorphic rock that is an essentially isothermal system, diffusion would be caused by local gradients of pressure. Unstable minerals, which are separating into their constituent atoms or groups of atoms, would be local sources of material. Other gradients of concentration would be generated by resurgent (second) boiling of crystallizing magmas with attendant expulsion of volatile constituents, connate water evolved during progressive metamorphism, and initial inhomogeneities of composition — such as carbonate rocks in contact with quartzose sandstone.

Material that diffuses through a crystalline aggregate of low gross porosity, i.e., a quartzite or greenschist, must move mainly along grain boundaries or through crystal lattices. Examples of transfer of matter through crystals, or intracrystalline diffusion, are the exsolution of homogeneous alkali feldspar into perthite, other order-disorder changes, and possibly the formation of progressively thickened reaction rims. Transfer of matter along grain boundaries, or intercrystalline diffusion, is shown by reactions at the surfaces of grains, such as weathering, replacement, and accretionary growth. Crystal cleavage and lineage, as well as dislocations, microfractures, and other flaws within crystals, also can act as passageways for the transport of matter.

Atoms can diffuse through crystals in three general ways: (1) by interchange with neighboring atoms; (2) by migration between other atoms in the lattice; and (3) by movements in association with coexisting vacancies, or so-called vacancy diffusion. Studies of diffusion in metals have suggested that vacancy diffusion is the dominant mechanism (Seitz, 1951, p. 89). Verhoogen (1952), in determining diffusion rates of lithium, sodium, and potassium along the c-axis of quartz, found that migration takes place by vacancy diffusion, that smaller ions diffuse faster than larger ions of similar charge, and that electric charge is more important than ionic size in governing diffusion rates. Particles with smaller charges were found to move more readily than those with higher charge.

Many petrologists (Niggli, 1954, p. 470-487; Turner and Verhoogen, 1951, p. 403-410; and Ramberg, 1952, p. 76-96) believe that diffusion along grain boundaries in metamorphic rocks is of much greater importance than intracrystalline diffusion. The nature of the surfaces of crystals, and the actual mechanism of transfer through or along these surfaces are only slightly known. The nature of crystal surfaces probably involves: (1) marked departures from the ideal crystal structure; (2) atoms that are not fully coordinated, with attendant unsatisfied bonds and resultant fields of force; and (3) adsorbed water, atoms, or coordinated groups of atoms that are extraneous to the crystal.

Diffusion along grain boundaries in metals has been studied by Smoluchowski (1952), who believes that the vacancy mechanism is dominant. The energy required to force atoms along grain boundaries of certain metals is substantially lower than that required for intracrystalline diffusion. The same appears to be true for silicates, but to a greater degree (Ramberg, 1952, p. 87).

Nucleation and Growth

During metamorphism a great many atoms will be present in the surface phases of crystals and in the adjoining voids that may be present. At the surface of any given crystal there will be a continual breaking and formation of bonds; atoms will be oscillating thermally, some breaking away and others joining the lattice. In other words, there will be some sort of dynamic equilibrium between atoms in the surface phase of the crystal and those in the intergranular interstices. These interstices probably range from several angstrom units to several tens of angstrom units in width. If a particular mineral is unstable, it will tend to break down, probably at its surface, where the free energy is highest and the stability consequently lowest. There will be a local gradient of concentration, and atoms will migrate away from such unstable minerals along grain boundaries. Thus, if chlorite is unstable, there will be many Mg, Fe, Si, Al, and 0 atoms, or partially coordinated groups of atoms, in the interstices close to the disintegrating crystals. These atoms probably will be statistically intermixed.

Nuclei of stable mineral phases, such as biotite or hornblende, will form when the statistical fluctuations of concentration are such that the stoichiometric proportions for the new mineral phase are fulfilled, and when the free energy is below a certain minimal value. The free energy increases with increasing radius of the nuclei to a certain critical value, and thence decreases with increasing radius; hence, each nucleus must pass a certain critical size before it is stable and will undergo spontaneous growth (Smoluchowski, 1951, p. 157).

Nuclei may inherit certain structural characteristics of preexisting minerals. Pseudomorphism of biotite after chlorite would be an example of this. In such a case, most of the mica layers (see Grim, 1953, p. 70-71) would be unchanged, except for some replacement of Si by Al, and the brucitelike layers would either diffuse out of the structure and be replaced by K atoms, or they would remain in the lattice and be joined on both sides of the layer by Si_20_5 sheets that diffuse, tetrahedron by tetrahedron, into the structure. Or, in a less probable instance, small fragments of a structure, such as part of a double chain from an amphibole, may be present in a pore, and may there serve as a basis for nucleation of biotite, chlorite, or other structurally similar mineral.

The growth rate of nuclei is determined mainly by rate of accretion of new atoms to the surface of each nucleus, which in turn is dependent on total supply of atoms, rates of diffusion of these atoms, and the number of nuclei in the system. Other factors would be "form energy" (Turner and Verhoogen, 1951, p. 511), the orientation of the crystal with respect to the stress system of the rock, and the nature of its immediate neighbors. Form energy would be the tendency of a mineral to form euhedrally and as porphyroblasts. Nuclei favorably oriented with respect to the stresses in a rock may grow at the expense of less favorably oriented nuclei. Nuclei commonly grow by replacing one or more of their immediate neighbors, which may show a variable resistance to such replacement.

In summary, metamorphism of a rock is an extremely complicated process that is controlled by temperature, confining pressure, internal vapor pressure, stress, and composition. It takes place by breakdown of preexisting minerals, diffusion, nucleation, and growth, with or without fracturing, gliding, rotation of grains, and other physical processes.

REGIONAL METAMORPHISM

GENERAL FEATURES

Metabasalt is the only rock in Las Tablas quadrangle that is both sufficient in areal extent and sensitive enough to changes in metamorphic intensity to define clearly the regional metamorphism. Most of the following discussion of regional metamorphism therefore concerns these rocks.

Layers of pelitic schist are present at Aveta Creek and on the north side of Spring Creek Canyon, but these are the only localities in the quadrangle where the mineral assemblages developed in pelitic rocks can be compared with those developed in metabasalt. The kyanitebearing quartzites, the slightly micaceous feldspathic quartzites, the conglomerates, and the metarhyolite are only briefly discussed below, because they are relatively insensitive to changes in metamorphic intensity.

METAMORPHISM OF THE BASALTIC ROCKS

General Sequence of Metamorphism

The general paragenesis or succession of mineral assemblages that have developed in rock of basaltic composition can be summarized, with metamorphic grade increasing downward, as follows:

chlorite-albite-epidote-muscovite-carbonate
(with chlorite after hornblende?)
(chlorite formed from biotite?)
↓
chlorite-albite-epidote-biotite
↓
chlorite-oligoclase-epidote-biotite
↓
oligoclase-epidote-biotite-hornblende
↓
hornblende-oligoclase-(epidote)
↓
hornblende-andesine-(epidote)

Minerals in each assemblage are listed in order of decreasing abundance.

Greenschists

The lowest grade metamorphic rock of basaltic composition in Las Tablas quadrangle is chlorite-albite-epidote-sericite-carbonate schist a rock that can be included in the muscovite-chlorite subfacies of the greenschist facies of Turner (1948, p. 96-98). Grains of partly to wholly chloritized biotites are common in much of this schist, and suggest that a retrograde change from the biotite-chlorite subfacies of Turner (1948, p. 94-95) has occurred. The partial nature of the pseudomorph (or incomplete replacement of biotite by chlorite) implies that equilibrium was not established under the later imposed conditions of the muscovite-chlorite subfacies.

The pseudomorphs of chlorite, which commonly contain minor epidote, magnetite, sericite, and albite, that have formed from amphibole(?) indicate that the greenschist was at one time an amphibole-bearing rock of higher rank. Such an amphibole may have formed from original pyroxene or olivine phenocrysts during the regional metamorphism, followed by a retrogressive alteration to chlorite.

An albite-muscovite-quartz-biotite-epidote schist, which is exposed one-half mile north of Moppin ranch, contains 10 percent of biotite and thus can be classed in the biotite-chlorite subfacies. This rock forms a zone about one-half mile wide. The biotite is associated with plagioclase of albitic composition.

In the metamorphism of metabasalt the albite appears to react with epidote to form oligoclase before chlorite and epidote react to form hornblende. The chlorite-oligoclase-epidote-biotite schist along the north fork of Duran Creek implies such a change, because its feldspar is oligoclase, and not albite as in all of the greenschist to the west, and because its chlorite, epidote, and biotite have not reacted to produce hornblende. This rock forms a zone about one mile broad; too few thinsections were studied, however, to delineate the zone accurately. Thus, the albite-epidote amphibolite facies of Turner (1948, p. 88-92) is not present between the greenschists and the oligoclase amphibolites.

Amphibolites

The first appearance of hornblende in the progressive metamorphism of the metabasalt is in oligoclase-epidote-biotite-hornblende schist, as in the amphibolite at the west end of Tusas Mountain. The boundary between the amphibolite and the greenschists to the west already has been discussed in connection with the lithology of the Moppin metavolcanic series. Similar amphibolite was found in Vallecitos Canyon, La Jara Canyon, and Canada del Oso. The general grain size and content of hornblende increase eastward, along with a rather abrupt decrease in content of chlorite and biotite. Biotite is wholly transformed to hornblende in the progressive metamorphism of this rock before the chlorite is completely transformed in a similar reaction. Rocks containing chlorite and hornblende, and which are of higher metamorphic grade than any of the biotite-bearing amphibolites, are present east of Canada del Oso and south of Kiawa Lake.

The amphibolites along Cow Creek are fine-grained hornblende-oligoclase rocks, in which the earlier-formed chlorite, epidote, and biotite evidently have been wholly converted to hornblende.

The grain size of the plagioclase in the amphibolite is mostly between 0.12 and 0.25 mm, which is only slightly larger than that in the greenschists. The grains of hornblende, however, range from 0.1 to more than 10 mm long, and reach their maximum length in the rocks in the vicinity of American and Spring Creeks.

The composition of the hornblende is essentially the same in all of the amphibolites, as indicated by constancy of optical properties. There may be slight changes in the soda and alumina contents of the hornblende from Aveta Creek, where the associated plagioclase is oligoclase, to Kiawa Canyon, where the plagioclase is andesine. The zoning of the plagioclase, in which the rims of the grains are more calcic than the cores, implies growth of such grains during a gradual rise in metamorphic intensity, perhaps as a result of increasing temperature only. The initially albitic cores apparently grew as epidote was resorbed, becoming more calcic and aluminous. The zoning suggests growth of individual grains by accretion, rather than by diffusion of atoms to sites within the lattices, because the latter process probably would have yielded homogeneous crystals.

All the amphibolites can be grouped in Turner's staurolite-kyanite subfacies of the amphibolite facies (1948, p. 81-85).

Stability Relationships

The sequence of mineral assemblages given above for the progressive metamorphism of the metabasalts is believed to have developed in response to increasing temperature. The sequence can be divided into two general reactions:

albite + epidote--> oligoclase -->andesine
chlorite + biotite + epidote --> hornblende.

A reaction of less significance is:

Mg-rich chlorite + muscovite --> Al-rich chlorite + biotite,

which involves a mutual interchange of aluminum and magnesium between the chlorites and micas. However, the extent to which this reaction takes place is not known.

The metabasaltic rocks of Las Tablas quadrangle are generally similar to those of other areas (Turner and Verhoogen, 1951, p. 446-472), except that, with increasing metamorphic grade, oligoclase is first developed in the greenschist, rather than in the amphibolite. One possible explanation of this process is that the water pressure during metamorphism was abnormally high. If this were the case, the albite, an anhydrous mineral, probably would follow its normal trend with increasing temperature, and react with epidote to form oligoclase, a reaction that is largely independent of water pressure. The chlorite and biotite of the greenschist are hydrous minerals, and would be stable to higher temperatures than in most similar metamorphic rocks because of the higher water pressure. The high water pressure would increase the stability fields of the chlorite and biotite to higher temperature ranges.

Sequence of Crystal Structures

The mafic minerals that have been developed in the metabasalt show a definite sequence with increasing metamorphic grade of crystal structures. The sequence is chlorite —> biotite —> hornblende. It involves a change of structural types from one with alternating micalike and brucitelike layers to the double-sheet mica structure and thence to the double-chain amphibole structure (Bragg, 1937). The tendency is to give, with increasing temperature, structures with simpler linkage of the silicon-oxygen tetrahedra, as stated above in the quotation from Buerger (1948, p. 116-117).

The silicate groups, such as the double chains in hornblende, are bonded together more strongly by metallic cations in the higher-temperature forms. The two types of sheets in the chlorite structure are not bonded to each other by atoms lying between; the double sheets in mica are bonded together by potassium atoms; and the double chains in hornblende are bonded together by calcium, magnesium, and bivalent iron atoms (Bragg, 1937). The more strongly the silicate units are bonded together, the more the structure resists disintegration by thermal agitation (Buerger, 1948, p. 117).

The transition, with increasing temperature, from albite to andesine is typical of metamorphic rocks (Ramberg, 1952). This relationship also has been explained by Buerger (1948, p. 117). Aluminum and calcium in calcic feldspar substitute for sodium and silicon in albite. Substitution of trivalent aluminum for tetravalent silicon decreases the bonding strength within the linked silicon-oxygen and aluminum-oxygen tetrahedra, and allows for stronger bonds between framework units.

Diffusion

The mineralogical changes that took place during the transformation of basalt --> greenschist --> amphibolite must have involved diffusion of atoms over distances of several millimeters. Units of space as large as 20 cu mm were initially occupied by plagioclase and mafic minerals of the basalt, later by chlorite, albite, and epidote of the green-schist, and lastly by single grains of hornblende or by aggregates of hornblende and plagioclase. Thus, there were marked changes in the distribution of atoms in such units of space during the metamorphism.

Progressive metamorphism of the greenschist to amphibolite involved formation of minerals with decreasing contents of water. There must have been loss of water from the rock during metamorphism, probably by migration along grain boundaries, as the water or hydroxyl group is too large to pass readily through most silicate lattices. Thus, the intergranular spaces and surface phases of crystals were saturated with water while the rock was metamorphosed, a condition that facilitated diffusion of atoms, nucleation, and grain growth.

Nucleation and Growth

The average grain size of the plagioclase is similar to that of the mafic minerals in the greenschists and may be due to similar rates of nucleation and growth. In the amphibolites, however, the grains of plagioclase are about 0.12 to 0.25 mm in average diameter, whereas the hornblende grains range from 0.1 to 10 mm in length and are about 5-10 times larger in volume than the plagiodase grains. This difference may be due to growth about fewer nuclei of hornblende than of plagioclase. With a given amount of constituents available for the growth of hornblende, the grain size would vary inversely with the number of stable nuclei produced or nourished.

There is no petrographic evidence that scraps of one crystal structure served as seeds for nucleation of different crystal structures, except in the pseudomorphism of biotite and chlorite. Such processes may have occurred, as they involve breakage of fewer bonds than does complete disintegration of one crystal and growth of another, and hence they require less energy and would constitute a more probable mechanism.

Biotite shows a platy habit and hornblende a prismatic habit in the metabasalt. Ramberg (1952, p. 132-133) has explained this habit of biotite. Ionic forces from one Si_40_{10} sheet-unit to another are much weaker than the forces parallel to the sheets. Hence, atoms added to the crystal are taken preferentially at the edges of the sheets, or prism faces. If a new sheet is initiated, however, it will grow rapidly as atoms are attached to the edges, or prism faces, of the sheets. The explanation of the habit of hornblende is similar. Atoms are attached preferentially to the ends of the double chains, rather than to the sides of the chains, or prism faces, because of stronger unsatisfied bonds at the ends of the chains. The hornblende crystals, therefore, grow much faster parallel to their c-axes, or the long dimensions of the chains, than normal to their c-axes.

The marked parallelism of the c-axes of the hornblende grains in much of the amphibolite may be due to rotation of grains by shear, or to preferential growth of nuclei that were initially oriented with their c-axes parallel to the lineation. Available petrographic evidence does not favor one hypothesis over the other.

METAMORPHISM OF PELITIC SCHISTS

The muscovite-oligoclase-staurolite-kyanite-magnetite schist north of Aveta Creek is a metamorphosed mudstone. It belongs, of course, in Turner's staurolite-kyanite subfacies of the amphibolite facies. The staurolite and kyanite in this schist are oriented randomly. Crystals of neither mineral show any rotational effects.

The thin layer of quartz-muscovite-biotite-plagioclase-almandite schist exposed in the Spring Creek Canyon road cut contains more silica and potash than the staurolite-kyanite rock discussed above. The rotated crystals of garnet were developed before the folding ceased; they may well be older than the unrotated crystals of staurolite and kyanite in the schist to the north.

GENESIS OF KYANITE IN THE ORTEGA QUARTZITE AND KIAWA MOUNTAIN FORMATION

Nearly all the kyanite in the Ortega quartzite, Kiawa Mountain quartzite, and Jawbone conglomerate is believed to have been derived from alumina and silica that were present in the sediments as originally deposited. Kyanite, as mentioned above in the section on stratigraphy, occurs as grains disseminated along bedding planes in nonhematitic quartzite, with hematite in original dark laminae, and as rosettes with quartz in veins.

The kyanite in the nonhematitic quartzite probably formed in situ from original kaolin or bauxite. There is no reason, beyond the presence of alumina in the kyanite, to postulate that alumina migrated into the quartzite and became uniformly distributed through it. Formation from kaolin would liberate silica and water:

$$\text{kaolin} = \text{kyanite} + \text{quartz} + \text{water}$$
$$Al_2Si_2O_5(OH)_4 = Al_2SiO_5 + SiO_2 + 2H_2O$$

Similarly, the kyanite in the hematite-kyanite laminae is thought to have been formed from bauxite and quartz. The association of kyanite, commonly as grains too small to be seen with the naked eye, with the hematite in the dark laminae is constant throughout all of the vitreous quartzite and conglomerate. This association suggests original deposition of both hematite and bauxite.

The quartz-kyanite veins also are present only in the vitreous quartzose metasedimentary rocks. They cut across the bedding of the sedimentary rock, and therefore are of secondary origin. The small size but rather uniform and wide areal distribution of these veins can be accounted for if they represent redeposition of indigenous alumina and silica. Expulsion of water, whether connate or the product of metamorphic reactions, may have been responsible for movement and deposition of the quartz and kyanite. The rosettes of kyanite have not been sheared, and hence the veins must have been formed during a late stage of the folding, or after the folding had ceased.

Kyanite contains aluminum in both tetrahedral and octahedral co-ordination (Bragg, 1937, p. 165). In the other aluminosilicates, andalusite and sillimanite, the aluminum is in four- and five-coordination. These two minerals are less dense than kyanite, and hence may be less stable than kyanite under high hydrostatic pressure. High temperature, in contrast, would favor formation of the more open-packed minerals, andalusite and sillimanite. Experimental investigation of the stability relationships of these minerals is needed.

METAMORPHISM OF THE BURNED MOUNTAIN METARHYOLITE

The only major changes that appear to have taken place in Burned Mountain metarhyolite during the regional metamorphism were the transformation of sanidine (or orthoclase) to microcline and perthite, granulation and change of shape of bipyramidal quartz phenocrysts, and rotation of all of the original phenocrysts. Neither the groundmass nor the phenocrysts appear to have increased significantly in grain size during the metamorphism.

It is interesting to note that some of the metarhyolite contains relict phenocrysts of perthite, as in La Jara Canyon, Canada del Oso, and immediately north of Poso Lake, and some contains microcline and plagioclase — but no perthite. The crystals of perthite probably are pseudomorphs after original phenocrysts of the high-temperature form of alkali feldspar. These phenocrysts probably crystallized between 800°C and 900°C (see phase diagram of Bowen and Tuttle, 1950, p. 497), if a water pressure of about 1,000 kgm/cm^2 is assumed. The potassium and sodium atoms were randomly distributed in alkali feldspar that formed at such a high temperature. Subsequent chilling quenched these dynamically disordered crystals into a state of static disorder (Buerger, 1949, p. 110). The disordered crystals did not unmix at the lower temperature into the ordered, stable phases microcline and lowalbite, however, because the diffusion rates at this temperature are too low to allow the migration necessary for exsolution to take place.

The temperature of the metarhyolite was raised during metamorphism, however, to a level at which diffusion was sufficient to effect separation of the potassium and sodium atoms into separate crystals and to allow ordering of the silicon and aluminum atoms. In the nonperthitic metarhyolite the same process of unmixing must have occurred, and subsequently the albite migrated out of the perthite grains to form separate grains in the relict groundmass (Tuttle, 1952, p. 121).

RELATIONSHIP BETWEEN FOLDING AND REGIONAL METAMORPHISM

The folding and the regional metamorphism of the Precambrian rocks of Las Tablas quadrangle were broadly synchronous. This is suggested by the well-developed lineation of hornblende grains in much of the amphibolite. If these grains had formed prior to folding, they would have been fractured during the folding; if they had formed after the folding, their c-axes probably would have been oriented more randomly, as in many contact-metamorphic rocks.

There is a general lack of fractured, granulated rocks in this area, a lack which would seem to preclude pretectonic metamorphism. The almost ubiquitous grains of lamellae-bearing quartz and the rotated crystals of garnet in the schist on the north side of Spring Creek Canyon suggest some deformation after recrystallization. The lamellae in the quartz, however, may have been formed during recrystallization. The crystals of garnet in the schist may have grown to their present dimensions before the metamorphism and folding ceased.

RELATIONSHIP BETWEEN REGIONAL METAMORPHISM AND PLUTONIC

Rocks

The boundaries between the various mineral assemblages of the metabasalt are not parallel to their contacts with the Maquinita granodiorite and the Tusas Mountain pluton of the Tres Piedras granite, but these assemblage-boundaries make moderate to high angles with the contacts of the metabasalt and plutonic rocks. Greenschist is present adjacent to the west end of the granodiorite pluton that lies south of Cow Creek, and amphibolite and staurolite-kyanite schist lie near the eastern limit of exposure of the same pluton.

There is no apparent contact metamorphism of the Moppin meta-volcanic series by either the Maquinita granodiorite or the Tres Piedras granite. The temperatures of the Moppin rocks may well have been close to the maximum temperatures that prevailed during their metamorphism when the Maquinita plutons were injected. If this were so, and if there were steep gradients of temperature across the margins of the plutons, the greenschist at the contacts would have been heated only slightly by the igneous rock, perhaps only 50°C or 100°C.

Tres Piedras granite is in contact with amphibolite and quartz-feldspar rocks, all of which probably would be physically stable, or nearly so, at contacts with a cooling pluton of granite.

The metabasalt decreases in metamorphic grade southwestward from the Tres Piedras granite exposed in Tusas Canyon and along the lower Rio Tusas. There may be a correlation between metamorphic grade and the position of this pluton, but too much of the Precambrian terrane is masked by Tertiary rocks to permit any meaningful determination.

On La Jarita Mesa and on the south side of Kiawa Mountain, the metamorphic grade of the amphibolite increases toward the La Jarita pegmatites.

PEGMATITIC-HYDROTHERMAL METAMORPHISM

GENERAL FEATURES

Two general rock types have been affected by an aureole of metamorphism that surrounds the area of pegmatites on La Jarita Mesa. The most abundant type includes the quartz-feldspar rocks — quartzite, quartzose conglomerate, feldspathic quartzite, and metarhyolite — and the other, much less abundant type is amphibolite. The quartz-feldspar rocks have been partially replaced by muscovite, and to much lesser extents by biotite, garnet, gnd feldspar. As mentioned above, muscovite is present in amounts ranging from a few percent to about 15 percent in the quartzite, conglomerate, and metarhyolite that are grouped together as the Petaca schist. In some contact zones around individual pegmatite bodies the schist contains as much as 40 percent of muscovite. The amphibolite has been partly to wholly replaced by chlorite, biotite, muscovite, quartz, and garnet.

METASOMATISM OF THE QUARTZ-FELDSPAR ROCKS

Wall rock alteration of quartzite at contacts with masses of pegmatite has been discussed by Jahns (1946, p. 52), who recognized four general zones of alteration.

A fine-grained contact zone of muscovite-rich rock, zoned, consists mostly of randomly oriented, 1- to 10-mm flakes of pale-green or silvery muscovite, with interstitial quartz and feldspar. Part of this rock originally was quartzite; it is the most highly altered quartzite in the entire wall rock. This zone ranges in thickness from a fraction of an inch to a foot or more.

Zone c consists of coarse-grained quartz-muscovite-albite-oligoclase-microcline schist, which in general is roughly foliated and highly crenulated. Muscovite is present as irregular sheets and sheaves that show a fair to marked schistosity parallel to that in the closely adjacent less micaceous quartzite, and have been folded into crinkles that define an excellent lineation. Quartz and feldspar occur as irregular sheets and knots, whose longest dimensions are parallel to the axes of the muscovite crinkles. The grain size of the quartz, muscovite, and feldspar is variable, and most grains are 2-10 mm in maximum dimension. Quartz characteristically shows sutured boundaries, and both the albite-oligoclase and microcline, which in general occur with quartz in porphyroblastic knots, are irregular in shape. Hematite is common as an accessory constituent of this schist.

The thickness of zone c varies from one pegmatite body to another; in some it is absent or only a few inches thick, and in others it is 10 or 15 feet thick. Marked variations in thickness characterize many contacts, and generally the thickest part occurs where the schistosity is about normal to the pegmatite-wall rock contact.

Quartz-muscovite schist forms zone b and is gradational into zone c. The muscovite content of this rock generally amounts to 10-30 percent, and decreases in a direction away from the pegmatite. The pale-green to colorless flakes of mica occur mostly as disseminations in granulose mosaics of quartz, and in part as mica-rich partings. This rock is much finer grained than that of zone c, and contains flakes and foils of muscovite 0.25-1 mm in diameter and anhedra of quartz 0.25 mm in average size. The muscovite is very well foliated, and gives an excellent schistosity to the rock. Microcline and albite-oligoclase constitute as much as 10 percent of the rock.

Zone b ranges in thickness from a few feet to several hundred feet, as measured horizontally, and has a marked tendency along concordant pegmatites to be thicker along the schistosity than normal to it.

Zone a is the slightly micaceous quartzite already described in connection with the lithology of the Petaca schist.

There are two major differences between the normal Burned Mountain metarhyolite and that included in the Petaca schist. The latter type contains muscovite in amounts ranging from a few percent to more than 25 percent; and the color of this rock is very light flesh to light gray or light greenish gray. All gradations from slightly muscovitic to heavily muscovitized metarhyolite are present, and they correspond to zones a through c of the muscovitized quartzite.

The only major difference between the quartzite and the metarhyolite in zones b and c is the content of feldspar. The metarhyolite appears to retain much of its original potash feldspar and plagioclase, whereas the quartzite originally had little or none and has received some alteration in zone c, but not in zone b.

The relict microcline and quartz are preserved in the metarhyolite in all of zone a and most of zone b. Muscovite is present as uniformly disseminated 0.25- to 0.5-mm flakes and as thin partings of pale-green flakes 0.5-2 mm in diameter. The rock is given an excellent schistosity by the muscovite flakes, which are commonly parallel to the axial planes of minor folds. Garnet and biotite are locally present in small amounts in the muscovitized metarhyolite; these minerals probably are of meta-somatic origin.

The muscovitic metarhyolite contains about 5 percent more $Si0_2$, 2 percent less Na_20, and slightly less Fe_20_3, FeO, MgO, CaO, and TiO_2 than the nonmuscovitic metarhyolite from Canada del Oso (table 8). The compositional variations of the metarhyolite in Las Tablas quadrangle are not known; hence the differences between the two chemical analyses cannot be discussed accurately in terms of gains and losses of material by metasomatism. The writer believes, however, that the differences are largely the result of metasomatism. The 10 percent of muscovite in the altered metarhyolite appears to have formed at the expense of preexisting potash feldspar in the groundmass, rather than by addition of K_20, $A1_20_3$, and H_20.

TABLE 11. CHEMICAL ANALYSIS OF TYPICAL MUSCOVITIC METARHYOLITE
(In percent)

SiO_2	80.54	Na_2O	1.47
TiO_2	0.14	K_2O	4.52
Al_2O_3	10.25	P_2O_5	0.00
Fe_2O_3	1.58	H_2O+	0.49
FeO	0.21	H_2O-	0.00
MnO	0.02	CO_2	0.00
MgO	0.07	Total	99.64
CaO	0.35		

• From northeast of Poso Spring, NE¼SW¼ sec. 23, T. 27 N., R. 8 E. Analyst: H. B. Wiik, Helsinki, Finland.

METASOMATISM OF AMPHIBOLITE

Lithology

The amphibolite layer in the amphibolite member of the Kiawa Mountain formation contains knots of chlorite, as described above in the section on lithology of that member. The percentage of these knots in the rock generally increases to the southeast, toward the pegmatites of La Jarita Mesa. There is a marked increase, from several percent to about 20 percent, of chlorite in the amphibolite from Kiawa Canyon and the south rim of Tusas Canyon to the Kiawa mine.

Biotite-epidote-quartz-plagioclase schist is interlayered with micaceous quartzite and metarhyolite near Poso Spring (NE¼ NE¼ sec. 27, T. 27 N., R. 8 E.), as shown on Plate 3. This is the only known occurrence of schist of this unusual composition in Las Tablas quadrangle. The rock is believed to be an altered amphibolite.

This schist contains 40-50 percent biotite, which gives a marked planar structure to the rock. The biotite is pleochroic, from pale olive to dark greenish brown. It forms tabular grains, 0.5 mm in average diameter. Biotite grains commonly enclose smaller epidote grains. Equant to elongate grains of colorless epidote, 0.12 mm in average diameter, are both disseminated throughout the rock and grouped in clusters. Epidote is present in amounts of 25-30 percent. Anhedral quartz grains, 0.12 mm in average diameter, occur as mosaic aggregates of lenticular and irregular shapes. Quartz forms 15-20 percent of the rock. The remaining 10 percent of the rock comprises oligoclase-andesine, magnetite-ilmenite, and muscovite. A modal analysis of this rock is given below (table 12).

Variants of the biotite-epidote-quartz-plagioclase schist that contain as much as 50 percent of muscovite are present along parts of the contacts between the schist and the enclosing micaceous quartz-feldspar rocks. These variants are similar to the muscovite-biotite-garnet skarn rock described in the following paragraphs.

A layer of muscovite-biotite-garnet skarn rock, 5 feet thick, is present in the Petaca schist about 400 yards east-southeast of Poso Spring. This rock has a decussate texture. Muscovite, present in amounts from 60 to 70 percent, occurs in tabular to chunky subhedral crystals, some of which have ragged edges. The grains are 0.1-5 mm in diameter, averaging about 1 mm. The birefringence of the muscovite is about 0.040, and Z A 001 = 2°. Biotite, which forms 20-25 percent of the rock, is similar in habit and grain size to the muscovite. The biotite is pleochroic, with X = yellow, Y = brownish green and Z = greenish brown; the birefringence is 0.033; and Z A 001 = 11/2-2 degrees. Colorless, equant crystals of garnet, as much as 5 mm in diameter, form 7-10 percent of the rock. Quartz, magnetite-ilmenite, and apatite form the remainder of the skarn. A volumetric mode of the muscovite-biotite-garnet skarn is given below (table 12).

Sequence of Alteration

The general sequence of alteration of the amphibolite can be summarized as follows:

unaltered hornblende-plagioclase amphibolite

↓

hornblende-plagioclase amphibolite with knots of chlorite

↓

biotite-epidote-quartz-plagioclase schist

↓

muscovite-biotite-garnet skarn rock.

Compositional Changes

Volumetric modes of typical hornblende-plagioclase amphibolite with knots of chlorite, of hornblende-plagioclase amphibolite rich in knots of chlorite, of biotite-epidote-quartz-plagioclase schist, and of muscovite-biotite-garnet skarn rock are given in Table 12.

The chemical compositions of typical unaltered amphibolite, of the biotite-epidote-quartz-plagioclase schist as calculated from the modal analysis, and of muscovite-biotite-garnet skarn rock are given in Table 13.

The general chemical changes in the alteration of hornblende-plagioclase amphibolite to muscovite-biotite-garnet skarn, assuming constant volume, have been shown above (table 13). The net losses and gains of constituents in this process are given in Table 14.

Potash, alumina, water, and phosphorus pentoxide were added to the amphibolite; silica, lime, magnesia, ferric oxide, and soda were subtracted. The assumption as to constant volume cannot be evaluated. Other elements, such as the halides and lithium, may be present in the skarn, but were not chemically determined.

The mineralogical reactions in the transformation of amphibolite into the skarn, in approximate order of their occurrence, can be represented as follows:

$$\text{hornblende} + \text{plagioclase} + H_2O \rightarrow \text{chlorite} + \text{quartz} + CaO + Na_2O$$

$$\text{hornblende} + \text{chlorite} + \text{plagioclase} + K_2O \rightarrow \text{epidote} + \text{biotite} + \text{quartz} + Na_2O$$

$$\text{biotite (in part)} + Al_2O_3 \rightarrow \text{muscovite} + MgO$$

$$\text{epidote} \rightarrow \text{garnet} + Al_2O_3 + H_2O$$

TABLE 12. VOLUMETRIC MODES OF METASOMATIZED AMPHIBOLITE
(In percent)

	(1)*	(2)	(3)	(4)
Hornblende	55	63	—	—
Plagioclase	27	13	6	—
Quartz	5	7	18	3
Epidote	tr	—	28	—
Chlorite	6	12	—	—
Biotite	—	—	45	24
Muscovite	—	—	tr	62
Garnet	—	—	—	8
Magnetite-ilmenite	6	4	2	2
Apatite	1	1	1	1

* Specimens arranged, from left to right, in the order of increasing alteration:
1. 36-D-60. From SE¼SE¼ sec. 14, T. 27 N., R. 8 E.
2. 36-D-56. From NW¼SW¼ sec. 11, T. 27 N., R. 8 E.
3. 36-D-65. From NE¼NE¼ sec. 27, T. 27 N., R. 8 E.
4. 36-D-18. From NW¼NW¼ sec. 26, T. 27 N., R. 8 E.

TABLE 13. CHEMICAL ANALYSES OF FRESH AND ALTERED AMPHIBOLITE

(In percent)

	(1)*	(2)	(3)	(4)
SiO_2	52.53	48.78	51.8	45.01
TiO_2	1.41	2.28	1.5?	1.54
Al_2O_3	14.22	14.82	16.4	18.46
Fe_2O_3	4.33	6.15	4.0	3.32
FeO	7.72	8.52	6.3	7.38
MnO	0.25	0.28	†	0.21
MgO	5.49	5.02	5.1	3.65
CaO	9.00	8.67	7.6	4.85
Na_2O	2.05	3.46	0.4?	1.47
K_2O	0.56	0.25	5.3	8.43
P_2O_5	0.45	0.64	†	0.77
H_2O+	1.58	0.79	2.6	4.26
H_2O-	0.00	0.00		0.04
CO_2	0.00	0.00	—	0.00
Totals	99.59	99.67	100.00	99.66

* Specimens arranged, from left to right, in the order of increasing alteration:

1. 36-D-2. From SW¼SE¼ sec. 9, T. 27 N., R. 8 E. Analyst: H. B. Wiik, Helsinki, Finland.
2. 36-D-60. From SE¼SE¼ sec. 14, T. 27 N., R. 8 E.
3. 36-D-65. From NE¼NE¼ sec. 27, T. 27 N., R. 8 E. Calculated from modal analysis.
4. 36-D-18. From NW¼NW¼ sec. 26, T. 27 N., R. 8 E. Analyst: H. B. Wiik, Helsinki, Finland.

† Not calculated.

TABLE 14. NET LOSSES AND GAINS OF CONSTITUENTS IN METASOMATISM OF TYPICAL AMPHIBOLITE TO MUSCOVITE-BIOTITE-GARNET SKARN ROCK

(In percent)

	Loss	Gain
SiO_2	7.5	—
TiO_2*	—	—
Al_2O_3	—	4.2
Fe_2O_3*	1.0	—
MnO*	—	—
MgO	1.8	—
CaO	4.1	—
Na_2O	0.6	—
K_2O	—	7.9
P_2O_5	—	0.3
H_2O	—	2.7
Totals	15.0	15.1

* No essential change.

Undoubtedly the actual reactions were much more complicated, and their exact determination would require very detailed studies.

Altered amphibolite overlies much of the large Harding pegmatite in Taos County, New Mexico. Part of this amphibolite has been metasomatized to a rock that is almost wholly muscovite. Such a rock represents the end stage of alteration of amphibolite by pegmatitic magma, or by hydrous fluids in equilibrium with pegmatitic magma.

Relations of Altered Amphibolite to Exposed Pegmatites

The only amphibolite layer that is immediately adjacent to a pegmatite body is one at the west end of the Kiawa mine (jahns, 1946, pl. 6). The actual contact is not exposed. There are no pegmatite bodies of 10 feet or more horizontal length exposed within several hundred yards of the highly altered amphibolite layers near Poso Spring.

GENERAL DISCUSSION OF PEGMATITIC-HYDROTHERMAL METAMORPHISM

Source of Introduced Material

The direct association of the altered quartz-feldspar rocks and amphibolite with the La Jarita pegmatites implies that fluids emanating from the pegmatites were the cause of the alteration. The general subject of pegmatitic magmas and separation of a gas phase from them has been discussed by Bowen (1933), and the discussion that follows is based largely on his presentation.

Granitic pegmatite magmas contain relatively large amounts of hyperfusible constituents, especially the volatiles H_2O, CO_2, $C1_2$, and **F2.** The silicate melt can be considered as a partially polymerized mixture of silicon-oxygen tetrahedra, which form a crude network. The network contains interstitial metallic cations, such as Na, K, Mg, etc., and the hyperfusibles. Thermal oscillation is high in such a system, and bonds between polymers, cations, and hyperfusibles are continually formed, broken, and reformed. The hyperfusible constituents form limited numbers of bonds with the silicate portion of the magma.

As the magma crystallizes, the amount of silicate liquid decreases, and the concentration of hyperfusibles in that liquid increases (assuming that the precipitated crystals contain less of the hyperfusibles than did the wholly liquid magma). As solid phases continue to separate, the ratio of hyperfusibles to silicate liquid increases, and ultimately the solubility of the hyperfusibles in the liquid may be exceeded. Boiling follows, during which a vapor phase of high pressure forms in the silicate liquid. The vapor consists largely of H_2O and HC1, with lesser amounts of alkali halides, $SiC1_4$, CO_2, various metallic cations, halogens, and other of the more volatile components of the magma (Bowen, 1933, p. 119). The vapor is decidedly acid in character as it leaves the magma chamber. The vapor migrates into the wall rock, down gradients of vapor pressure and temperature. In open fractures in the wall rock the vapor will cool to a liquid. Bowen (1933, p. 123) has proposed

> . . . that a mass of intrusive rock and its envelope, containing interstitially and in fractures a pegmatitic liquid that is experiencing "second boiling," will constitute itself a fractional-distillation column. Throughout this column there occurs a liquid through which gas bubbles slowly rise and suffer selective (fractional) condensation. Close to the boiling source the liquid will be continually enriched in the hyperfusible constituents that are relatively nonvolatile; at points more and more remote the liquid will become increasingly richer in those hyperfusible constituents that are more volatile — water, various halogen compounds including acids, and the other substances already suggested as prominent in the vapor phase.

The vapor produced in the pegmatite liquid contains sodium, as evinced by albitization of crystals in the outer zones of the pegmatite bodies. Potassium forms compounds that are more volatile than those of sodium (Bowen, 1933, p. 123), and so the vapor phase that separates from the magma not only contains less sodium than potassium, but the sodium that is present is largely used up in the formation of albite within preexisting zones of the pegmatite body.

Migration of Material

In the pegmatitic-hydrothermal aureole that encloses the La Jarita pegmatites, the hydrothermal solutions must have migrated along grain boundaries, as there is no evidence that larger channeiways existed. The surfaces of grains and intergranular openings were probably saturated with water. The K, Al, and P atoms that were added to the Petaca schist and amphibolite, and other atoms that also migrated but were not captured by crystallizing minerals, probably moved as chloride- or oxide-groups of low charge and rather large size. The K and Al may well have had higher mobilities in this water-saturated environment than in a drier rock, such as the metarhyolite during regional metamorphism.

In the wall rock that surrounds the La Jarita pegmatites, the replaced minerals include quartz, oligoclase-andesine, hornblende, chlorite, epidote, and biotite. The chlorite, epidote, and biotite were stable during early stages of the alteration, but, as the amounts of potash and alumina in the rock increased, they became unstable and were replaced by muscovite and garnet. The chronological sequence of mineral changes in the amphibolite implies addition of water first, then potash, and lastly alumina. Thus the most volatile constituent of the vapor phase, water, probably migrated fastest; potassium, which may have migrated as KC1, was intermediate in velocity of migration; and aluminum, which also may have migrated as a chloride, was lowest in velocity of migration.

Nucleation and Grain Growth

Nucleation of crystals in the altered rock probably was facilitated by the relatively high speeds of migration. Enlargement of intergranular pores by solution may have increased the rate of nucleation, by allowing many atoms to congregate in a small space, polymerize, and form nuclei.

RELATION OF PEGMATITIC-HYDROTHERMAL METAMORPHISM TO REGIONAL METAMORPHISM

The pegmatitic-hydrothermal metamorphism probably occurred later than the regional metamorphism. The best evidence for this is the chlorite pseudomorphs that were formed from hornblende in amphibolite, and the knots of chlorite that crosscut the lineation of the amphibolite. The muscovite-biotite-garnet skarn has a decussate texture; if this rock were formed during folding, and hence during regional metamorphism, a schistosity would have been developed.

Most of the plates of metasomatic muscovite in the Petaca schist lie parallel to the bedding on the flanks of folds, and parallel to the axial planes at the noses of folds. This feature may be due to replacement while the rock was being sheared, or to replacement following folding along planes of weakness parallel to the axial planes that were formed during the folding.

FORMATION OF THE BODIES OF QUARTZ-KYANITE ROCK AT LA JARITA MESA

Six bodies of quartz-kyanite rock lie in muscovite quartzite and metarhyolite of the Petaca schist at Big Rock, about 150 yards south of Poso Spring, and in an area about five-eighths of a mile northwest of Poso Spring. The bodies are oval in plan, with maximum horizontal dimensions from a few tens of feet to several hundred feet. Lenses of quartzkyanite rock, which contain as much as 80 percent kyanite as single grains and rosettes, lie scattered in slightly to moderately kyanitic quartz rock. In each of the deposits the quartz-kyanite rock is enclosed in a shell of silvery coarse-grained muscovite-quartz schist.

These deposits were studied in detail by Corey (1953), who believes (p. 2) that they are metamorphosed pelitic silt lenses, and that within the larger quartz-kyanite bodies

> Lenticular or irregular masses of interlacing coarse kyanite-quartz and rosette kyanite-quartz rocks within the kyanite schist formed in the absence of stress toward the end of the metamorphic period. Metamorphic differentiation or deposition from hydrothermal solutions contaminated by dissolved kyanitic material, are believed to have formed these bodies. Still later in pre-Cambrian time, hydrothermal solutions with assimilated kyanitic material formed quartz-kyanite veins in joints in the kyanite schist lenses.

The writer does not believe that these bodies of quartz-kyanite are metamorphosed pelitic lenses, because: 1) Their shapes are very different from those of intensely folded layers in adjacent rocks, such as those shown on Plate 3; 2) metarhyolite can be traced into one of the bodies of quartz-kyanite; 3) lenses of pelitic rock are not found elsewhere in the Ortega and Kiawa Mountain formations; 4) other minerals typically found in metamorphosed pelitic rocks, such as garnet, staurolite, and biotite, are not present or are very scarce in the quartz-kyanite bodies; and 5) if the bodies of quartz-kyanite had formed prior to crystallization of the La Jarita pegmatites, they undoubtedly would have been muscovitized.

A hydrothermal origin seems probable for these masses of quartz-kyanite and muscovite-quartz schist. They may have been formed by alumina-silica metasomatism similar to that which formed the veins of quartz-kyanite in the Ortega quartzite and quartzite members of the Kiawa Mountain formation, but on a much larger scale. In such a case, alumina and silica would migrate from kyanitic quartzite and would replace micaceous quartzite and metarhyolite. The formation of quartz-kyanite rock in micaceous quartzite and metarhyolite would involve replacement of muscovite, microline, and plagioclase, as well as quartz. The potash, alumina, silica, and water derived from this replacement would be deposited at the margins of the quartz-kyanite bodies, where conditions for their deposition would be favorable. As the bodies of quartz-kyanite grew outward from the source conduit of the hydrothermal solutions, the rim of muscovite-quartz schist would migrate outward, its inner margin becoming unstable and dissolving with concomitant redeposition at its outer margin. The time relationships of such a process are uncertain. The small veins of quartz and kyanite in the Ortega and Kiawa Mountain formations probably were formed during a late stage in the regional metamorphism, yet the large bodies of quartz and kyanite would have to be of similar age or younger than the pegmatitic-hydrothermal metamorphism.

The quartz-kyanite bodies may have been formed from hydrothermal solutions that were expelled from pegmatite magmas. In such a case, the solutions would have an unusually high Al:K ratio, in order to develop kyanite rather than muscovite. It is doubtful whether a solution of this composition would be formed by a pegmatitic magma.

Another possibility is that these bodies are pegmatites, high in alumina and silica, and low in alkalis. The major difficulty with this hypothesis is in accounting for generation of such a magma. Assimilation of kyanitic quartzite might produce a magma of this composition. However, a mechanism to extract the soda, potash, and lime from a granitic magma is necessary, unless dilution by kyanitic quartzite were very great, which is not likely. Reaction with a mafic rock, prior to emplacement and crystallization as a quartz-kyanite body, also might produce a drop in alkalis and lime.

If these quartz-kyanite masses have formed from contaminated granitic pegmatites, one might expect to see bodies of intermediate composition, but these are not found.

The quartz-kyanite bodies have not been so intensely folded as the enclosing rocks; hence the folding occurred prior to formation of these bodies. The lack of muscovitic alteration of these bodies also suggests that they were formed after, or possibly contemporaneously with, the pegmatitic-hydrothermal alteration of the adjacent rocks.

Geologic History

The geologic history of Las Tablas quadrangle can be summarized as follows:

1. Deposition of quartz sand in Precambrian time, probably at depths from a few feet to several hundred feet.
2. Deposition of the Big Rock conglomerate as a beach gravel during a temporary drop in sea level.
3. Extrusion of the Moppin volcanic rocks, with some contemporary deposition of quartz sand, followed by deposition of the Jawbone conglomerate, extrusion of the basalt, and deposition of the upper quartzite member of the Kiawa Mountain formation.
4. Intense deformation, with development of northwest trending and west plunging folds, accompanied by metamorphism of the rocks to low and moderate rank.
5. Intrusion of Maquinita granodiorite just before cessation of folding but after metamorphism. Emplacement of the Tres Piedras granite and the La Jarita pegmatites after folding. Pegmatitic-hydrothermal metamorphism in quartzite and metarhyolite surrounding the pegmatites, with formation of the Petaca schist. Big Rock quartz-kyanite bodies formed.
6. Deep erosion after the orogeny, continuing intermittently from Precambrian to Miocene time. Possible sedimentation and vulcanism in this interval, with later loss of any such rocks by erosion.
7. Deposition in Miocene to Pliocene time, of terrestrial sandstone, conglomerate, and tuff, with contemporaneous vulcanism forming welded tuff, flows ranging in composition from olivine basalt to rhyolite, and minor intrusive andesite breccia.
8. Tilting to the northeast, block faulting, mainly along Tusas and Vallecitos Valleys, with slight tilting to the southeast. Elevation of the Jawbone Mountain-La Jarita Mesa highland relative to the Tertiary rocks to the southwest, and depression of the highland relative to Tertiary rocks of the Taos Plateau to the northeast.
9. Slight erosion and deposition of alluvium in Quaternary and Recent time.

References

Atwood, W. W., and Mather, K. F. (1932) *Physiography and Quaternary geology of the San Juan Mountains, Colorado,* U. S. Geol. Survey Prof. Paper 166.

Billings, M. P. (1928) *The chemistry, optics, and genesis of the hastingsite group of amphiboles,* Am. Mineralogist, v. 18, 287-296.

Bowen, N. L. (1933) *The broader story of magmatic differentiation, briefly told,* in Finch, J. W., et al. (ed.), Ore deposits of the Western States, Am. Inst. Min. and Met. Eng., New York, 106-128.

———, and Tuttle, O. F. (1950) *The system $NaAlSi_3O_8$–$KAlSi_3O_8$–H_2O,* Jour. Geology v. 58, 489-511.

Bragg, W. L. (1937) *Atomic structure of minerals,* Ithaca, N. Y., Cornell Univ. Press.

Buerger, M. J. (1948) *The role of temperature in mineralogy,* Am. Mineralogist, v. 33, 101-121.

——— (1951) *Crystallographic aspects of phase tarnsformations,* in Smoluchowski, R., et al. (ed.), Phase transformations in solids, New York, John Wiley & Sons, Inc., 183-209.

Butler, A. P. (1946) *Tertiary and Quaternary geology of the Tusas-Tres Piedras area, New Mexico,* Ph. D. dissertation, Harvard University.

Corey, A. F. (1953) (abs) *Kyanite deposits of the Petaca district, Rio Arriba County, New Mexico,* Ph. D. dissertation, University of Michigan.

Cross, W., and Larsen, E. S. (1935) *A brief review of the geology of the San Juan region of southwestern Colorado,* U. S. Geol. Survey Bull. 843.

Dapples, E. C. (1947) *Sandstone types and their associated depositional environments,* Jour. Sed. Petrology, v. 17, 91-100.

Daly, R. A. (1936) *Average chemical analyses of certain types of igneous plutonic rocks,* in Birch, F., et al. (ed.), Handbook of physical constants, Geol. Soc. Am. Spec. Paper 36, 2.

Eskola, P. (1934) *A note on diffusion and reactions in solids,* Comm. géol. Finlande Bull. 104, 150-154.

Fairbairn, H. W. (1949) *Structural petrology of deformed rocks,* Cambridge (England), Addison-Wesley Press.

Goldschmidt, V. M. (1911) *Die Kontaktmetamorphose im Kristianiagebiet,* Kristiania Vidensk. Skr., I. Math.-Naturv. Kl. 11.

Grim, R. E. (1953) *Clay mineralogy,* New York, McGraw-Hill Book Co.

Holmes, J. A. (1899) *Mica deposits in the United States,* U. S. Geol. Survey, 20th Ann. Rpt., 706-707.

Jahns, R. H. (1946) *Mica deposits of the Petaca district, Rio Arriba County, New Mexico,* N. Mex. School of Mines, State Bur. Mines and Mineral Res. Bull. 25.

Johannsen, A. (1932) *A descriptive petrography of the igneous rocks,* v. 2, Chicago, Univ. of Chicago Press.

Just, E. (1937) *Geology and economic features of the pegmatites of Taos and Rio Arriba Counties, New Mexico,* N. Mex. School of Mines, State Bur. Mines and Mineral Res. Bull. 13.

Kay, M. (1951) *North American geosynclines,* Geol. Soc. Am. Mem. 48.

Krumbein, W. C., and Sloss, L. L. (1951) *Stratigraphy and sedimentation,* San Francisco, W. H. Freeman & Co.

Krynine, P. D. (1941) (abs) *Paleogeographic and tectonic significance of sedimentary quartzites,* Geol. Soc. Am. Bull. v. 52, 1915-1916.

——— (1943) *Diastrophism and the evolution of sedimentary rocks,* Am. Assoc. Petrol. Geol., Dist. Lect. Series (mimeo).

Lindgren, W., Graton, L. C., and Gordon, C. H. (1910) *The ore deposits of New Mexico,* U. S. Geol. Survey Prof. Paper 68.

Mansfield, G. R., and Ross, C. S. (1935) *Welded rhyolitic tuffs in southeastern Idaho,* Am. Geophys. Union Trans., 16th Ann. Meeting, 308-321.

McKee, E. D. (1945) *Cambrian history of the Grand Canyon region,* Carnegie Inst. Wash. Pub. 563.

Niggli, Paul (1954) *Rocks and mineral deposits,* San Francisco, W. H. Freeman & Co.

Pettijohn, F. J. (1949) *Sedimentary rocks,* New York, Harper & Brothers.

Ramberg, Hans (1952) *The origin of metamorphic and metasomatic rocks,* Chicago, Univ. of Chicago Press.

Seitz, F. (1951) *Fundamental aspects of diffusion in solids,* in Smoluchowski, R., et al. (ed.), Phase transformations in solids, New York, John Wiley & Sons, Inc., 77-145.

Smith, H. T. U. (1938) *Tertiary geology of the Abiquiu quadrangle, New Mexico,* Jour. Geology, v. 46, 933-965.

Smoluchowski, R. (1951) *Nucleation theory,* in Smoluchowski, R., et al. (ed.), Phase transformations in solids, New York, John Wiley & Sons, Inc., 149-180.

——— (1952) *Movement and diffusion phenomena in grain boundaries,* in Shockley, W. (ed.), Imperfections in nearly perfect crystals, New York, John Wiley & Sons, Inc., 451-475.

Turner, F. J. (1948) *Mineralogical and structural evolution of the metamorphic rocks,* Geol. Soc. Am. Mem. 30.

———, and Verhoogen, J. (1951) *Igneous and metamorphic petrology,* New York, McGraw-Hill Book Co.

Tuttle, O. F. (1952) *On the contrasting mineralogy of extrusive and plutonic salic rocks,* Jour. Geology, v. 60, 107-124.

Verhoogen, J. (1952) *Ionic diffusion and electrical conductivity in quartz,* Am. Mineralogist, v. 37, 637-655.

Walker, F., and Poldervaart, A. (1949) *Karroo dolerites of the Union of South Africa,* Geol. Soc. Am. Bull. 60, 591-705.

Williams, H. (1942) *The geology of Crater Lake National Park, Oregon,* Carnegie Inst. Wash. Pub. 540.

Winchell, A. N., and Winchell, H. (1951) *Elements of optical mineralogy,* pt. 2, New York, John Wiley & Sons, Inc.

Wiseman, J. D. H. (1934) *The central and south-west highland epidiorites: a study in progressive metamorphism,* Geol. Soc. London Quart. Jour. v. 90, 354-416.

Yoder, H. S., Jr. (1952) *The MgO-Al_2O_3-SiO_2-H_2O system and the related metamorphic facies,* Am. Jour. Sci., Bowen volume, pt. 2, 569-627.

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Numbers in **boldface** indicate main references.