Younger Precambrian Geology in Southern Arizona

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By ANDREW F. SHRIDE

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Stratigraphic, lithologic, and structural features of younger Precambrian rocks of southern Arizona as a basis for understanding their paleogeography and establishing their correlation

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YOUNGER PRECAMBRIAN GEOLOGY IN SOUTHERN ARIZONA

By Andrew F. Shride

ABSTRACT

Younger Precambrian rocks of southeastern Arizona consist of the Apache Group, the overlying Troy Quartzite, and large diabase bodies intrusive into and coextensive with the sedimentary formations. These rocks are widely exposed and little deformed in northern Gila County and in the southwest corner of Navajo County—areas that are structurally part of the Colorado Plateau. In the Basin and Range province farther south, outcrops are scattered and less extensive, and structural relations with younger formations are commonly obscure.

The Apache Group ranges in thickness from 1,250 to 1,600 feet, and includes, in ascending order, the Pioneer Shale, the Dripping Spring Quartzite, the Mescal Limestone, and unnamed basalt flows. The Scanlan and Barnes Conglomerate Members, formerly designated as separate formations, are considered basal conglomerate members of the Pioneer Shale and Dripping Spring Quartzite, respectively. The upper part of the Dripping Spring is divided into a massive arkose member and an upper thin-bedded member of siltstone and arkose. The Mescal Limestone comprises three members: a lower thin- to medium-bedded member of cherty dolomite or of calcitic limestone that is the metamorphic equivalent, a middle partly massive algal member of similar carbonate, and an upper argillite member.

The Troy Quartzite is locally at least 1,200 feet thick and is made up of three members. The lowest, a remnant only in northwestern Gila County, is well-sorted arkose characterized by very large scale crossbedding. The middle, or Chediski Sandstone Member, is poorly sorted, partly pebbly, sericitic or feldspathic sandstone, and is the only member that extends into the southern part of the region. The upper member is well-sorted evenly bedded clean quartzite.

All sedimentary formations are shallow shelf deposits; individual members reflect a variety of environments, each of which was notably uniform over as much as 15,000 square miles. Two volcanic episodes are apparent: (1) the Pioneer Shale is dominantly of water-laid rhyolitic(?) tuff; (2) basalt flows that locally underlie the argillite member of the Mescal and those that extensively overlie the formation are kindred and probably were not extruded at widely separate times. Evaporites and, incongruously, chert were deposited in the early sea of Mescal time; apparently during later carbonate deposition the salts were leached away, the result being a founder breccia of regional extent at the base of the formation. The uniform environments of subsequent dolomite deposition are shown especially in the intertidal deposits that make up the biostrome of the algal member. One and perhaps two aspects of the Troy suggest eolian activity. Even in areas of lapout, strata of the Pioneer and Dripping Spring generally do not vary greatly; the Troy, however, does exhibit changes where it laps onto older terrane. The recognition of these aspects of the Apache and Troy sequence provides new bases for speculation on interregional correlations of younger Precambrian rocks.

At least two major marine transgressions, sharply marked by the unconformities at the bases of the Pioneer and the Dripping Spring, occurred during Apache time; the Chediski Sandstone Member of the Troy was deposited during another transgression. The erosional unconformity between the Mescal and the Dripping Spring and the unconformity at the top of the algal member of the Mescal could also represent appreciable time intervals. The unconformities that separate the argillite member from the overlying and underlying basalt are apparently effects of brief subaerial exposures during recurrent broad warping which preceded uplift that terminated Apache sedimentation. Apache strata were variously eroded and locally stripped completely away before the Troy was deposited. Conceivably, unconformities also separate the members of the Troy.

Before and during the pre-Troy erosion, solution caverns were formed in the dolomite of the Mescal, and in a few areas a karst topography was formed. During this period the dolomite was extensively silicified and locally was converted to a massive rubble of chert. Hematite deposits of ore grade accumulated locally in the karst areas.

After lithification of the Troy, the upper Precambrian strata were intensely folded and faulted along narrow widely spaced north-trending belts. Between these belts strata remained virtually undeformed until they were extensively intruded and displaced by diabase sills. Many faults caused by diabase inflation have been visualized as post-Paleozoic in age, but all these rocks were deeply eroded and the structures truncated before Cambrian formations were deposited.

The diabase was emplaced largely as sills, some of which exceed 1,000 feet in thickness. In some areas their aggregate volume is as much as the strata they intrude. Sills occur throughout the Apache sequence but were intruded along certain stratigraphic horizons in preference to others; they are only locally conspicuous in the Troy. Commonly two or more were intruded along a given horizon, but in some areas multiple injections are rare. Dikes are insignificant, but some dikelike connections between concordant sheets at different horizons are regionally conspicuous. Because the sills are ubiquitous in the younger Precambrian terrane, thermal metamorphism related to them is widespread, and for the carbonate rocks it is a regional characteristic.

Olivine diabase is dominant; albite diabase is subordinate in these intrusions. Both are in part unusually coarse grained, a feature which sets those bodies apart from minor intrusions of dark-colored rock of later age. Gravitational settling of olivine is apparent only in the thicker sills. Extensive tabular bodies of diabase pegmatite occur just in the upper parts of thick sills at or near discordant parts of the intrusions. Quartzose differentiates, mainly granophyre and a highly quartzose aplite, occur in similar geometric habit but apparently are hybrid rocks formed by interaction between feldspathic xenoliths and late magmatic fluids.

The recognition of certain features in the basal formations of the Paleozoic is critical to unraveling the younger Precambrian history. In southernmost Arizona the Cambrian sequence consists of the Bolsa Quartzite and the overlying Abrigo Limestone, which is dominated by carbonate strata. Traced from south to north this sequence laps from a smooth surface on older Precambrian rocks, onto an irregular surface on the Apache Group, and ultimately onto the Troy Quartzite. In the same interval the Bolsa thins and in some areas laps out, and the Abrigo Formation becomes a sequence dominated by sandstone and mudstone. Because the Bolsa and Troy Quartzites seemed homotaxial equivalents and because in places fossiliferous sandstone—here recognized as Abrigo strata—seemed gradationally transitional into the underlying Troy, early workers correlated the Bolsa and Troy. In the northern part of the region, in a similar way, channel-fill sandstone at the base of the Martin Limestone (Devonian) has been included with the Troy. Criteria for recognizing the various sandstone and quartzite units as separate entities are given. Isotopic data indicate that a hiatus of at least 500 million years is represented in the pre-Bolsa unconformity.

INTRODUCTION

Younger Precambrian formations crop out in two parts of Arizona. In northern Arizona the Grand Canyon Series is exposed along the depths of the Grand Canyon in outcrops that aggregate less than 100 square miles in area. In southeastern Arizona outcrops of equivalent formations are much more widespread and occur mainly in a belt 20–60 miles wide that extends from the foot of the Mogollon Rim southward 170 miles to about the latitude of Tucson (fig. 1). Scattered outcrops also occur in an area 50–65 miles south of Phoenix.

The younger Precambrian rocks in southeastern Arizona consist of the Apache Group, the Troy Quartzite, and coextensive intrusions of diabase. Outcrop areas of these rocks are shown in figure 3. Figure 4, which covers the same area, shows the principal paleogeographic features that influence our understanding of the younger Precambrian terrane.

Ransome (1916, p. 133–135; 1923, p. 1) divided the State into three physiographic regions—the plateau region to the northeast, the desert region to the southwest, and the mountain region between the two (fig. 1). Apache and Troy outcrops are almost wholly in the mountain belt. It is commonly assumed, erroneously, that these geographic subdivisions are applicable in discussing structural disposition of these strata. If we apply the structural criteria commonly accepted for distinguishing the Colorado Plateau from the Basin and Range province, we find that the boundary between the geologic terranes is well within the mountain region and does not coincide with the Mogollon Rim as commonly assumed. Throughout this report, reference is made to this structural definition rather than to a physiographic delineation of provinces. In the report area the boundary is sharply defined and extends northwest through Gila County and is 40–50 miles south of the Mogollon Rim (fig. 3).

From this boundary northward to the Mogollon Rim, in the Colorado Plateau, the younger Precambrian formations are approximately horizontal and are laterally little interrupted by structure. The mesa-canyon topography of the plateau country, which has a relief of 2,500–4,500 feet, provides excellent exposures for the observation of stratigraphic detail of the sedimentary formations, the petrologic detail of the diabase, and the structural relations between formations. Conclusions drawn from this area can be more fully substantiated than those drawn from observations made elsewhere in the region of younger Precambrian outcrops.

In contrast, in the Basin and Range structural province most outcrops of younger Precambrian formations are in narrow mountain ranges, which are separated by broader deeply alluviated basins that are mainly floored on older Precambrian rocks. Huge mosaics of tilted fault blocks characterize many of the ranges. Thus, outcrops of the individual formations may be widely separated (fig. 3) and structural relations to younger and older formations obscured. Because the Troy Quartzite and formations of the Apache Group were first studied in and near the great copper-mining districts of Ray, Globe-Miami, and Superior, they were defined in this part of the region, and, until recently, conclusions derived in this area have dominated our understanding of younger Precambrian geology.

I first became aware that there were unreconciled descriptions and interpretations of the younger Precambrian terrane while examining an iron-ore deposit near the head of Canyon Creek during the winter of 1941–42. Prime stratigraphic details of the upper formations of the Apache Group and characteristics of the diabase intrusions, as well as the regional aspects of the metamorphism associated with them, were recognized incidental to the study of asbestos deposits in the Mescal Limestone between October 1942 and May 1944 and during the summer of 1946. These observations were confined mainly to the Sierra Ancha and to the canyon entrenched by the Salt River east of its junction with Canyon Creek. The basic data on the stratigraphy of the lower formations of the Apache Group and the Troy Quartzite and on regional structures of Precambrian age were collected

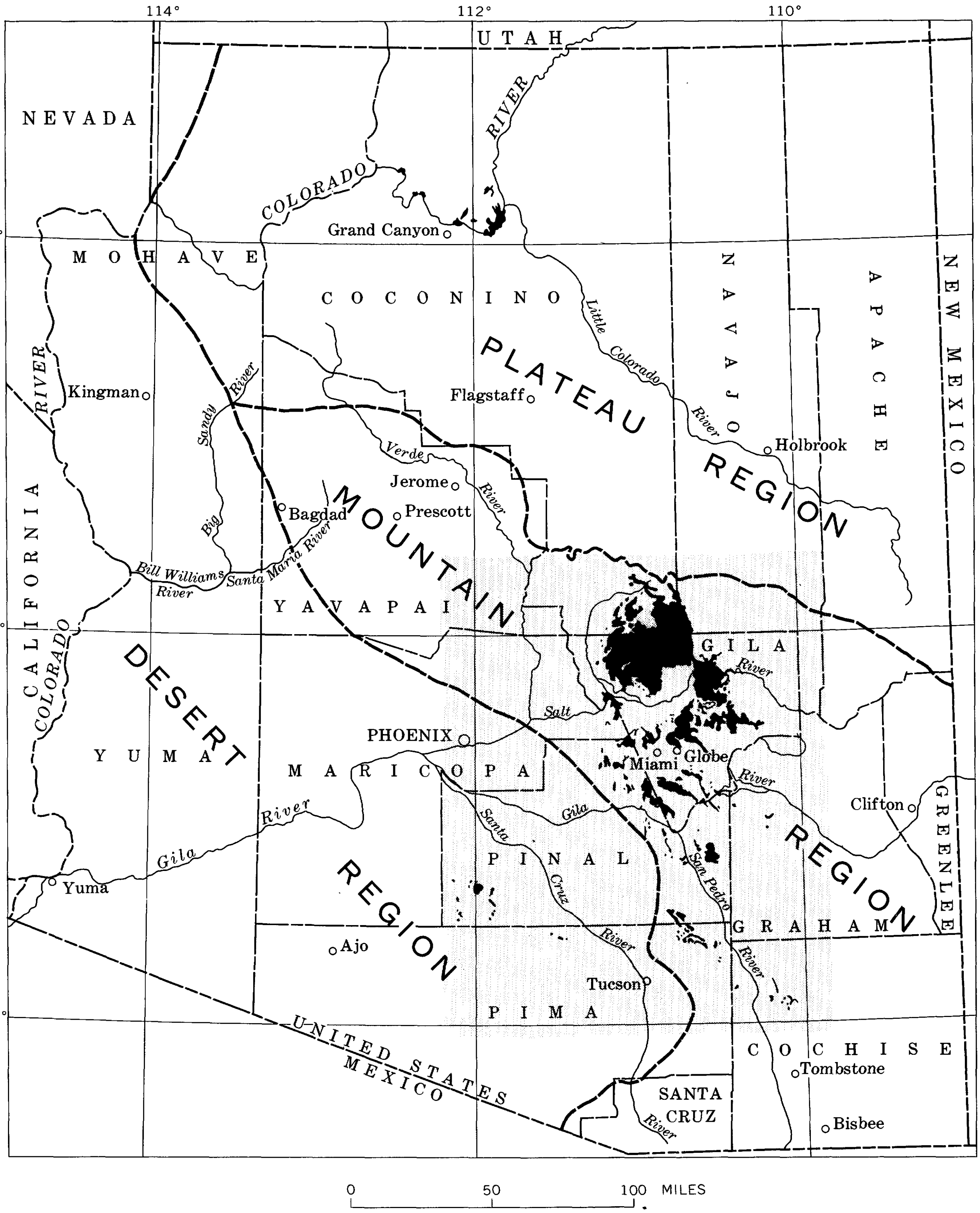

FIGURE 1.—Outcrops of younger Precambrian rocks in Arizona in relation to the three principal physiographic regions. Apache Group and associated formations (solid) crop out in shaded area; Grand Canyon Series (solid), in northern Coconino County. Area of figures 2, 3, and 4 shown by shading. Physiographic regions based on Ransome (1919, p. 27; 1923, fig. 2).

during detailed mapping of parts of the McFadden Peak quadrangle between July 1954 and June 1956. During the spring of 1957, a strip 10–25 miles wide, from north of the Salt River south almost to Globe along U.S. Highway 60, was mapped with C. T. Wrucke; this reconnaissance mapping aided in understanding details seen elsewhere. Premises thus evolved were tested, during June 1960, by the examination of Apache rocks as far south as the Little Dragoon Mountains in Cochise County, as far west as the Vekol Mountains in southwestern Pinal County, and as far east as Mount Turnbull in Graham County. At least a few observations of the younger Precambrian rocks have been made in almost every area of known outcrop. The large area of outcrop west of the 111th meridian and north of Theodore Roosevelt Lake has not been studied. Observations in this area might add to our knowledge of Pioneer and Dripping Spring strata, but probably would not provide significant data on the higher formations.

This report presents such reinterpretations of the regional geology of younger Precambrian rocks as can now be made, and provides background of stratigraphic and petrographic data so that these reinterpretations can be applied to deciphering the geologic settings of several mineral-bearing districts in southeastern Arizona. Rock descriptions in this report employ the "Rock-Color Chart" (Goddard and others, 1948) and the Wentworth grain-size scale (Wentworth, 1922). Bedding and splitting features of stratified rocks are described by the quantitative terms proposed by McKee and Weir (1953).

E. D. Wilson, geologist of the Arizona Bureau of Mines, has provided me with much information from his exceptional store of knowledge of the regional geology. In particular, Wilson guided me to areas in northern Gila County where relations at the base of the Apache section are especially well exposed and to three localities in the Mescal Mountains where Cambrian fossils had been collected. During the 1940's B. S. Butler, of the University of Arizona, offered much helpful comment. N. P. Peterson has advised me of many aspects of the geology of the Globe-Miami mineral belt. During 1954–56 H. C. Granger and R. B. Raup, of the U.S. Geological Survey, studied uranium deposits in the Dripping Spring Quartzite, and this work furnished many details on the stratigraphy. C. T. Wrucke examined many diabases petrographically that were collected during various studies and contributed careful observations and stimulating discussions. Other colleagues in the U.S. Geological Survey who contributed greatly to my understanding of the area include J. R. Cooper, S. C. Creasey, M. H. Krieger, C. R. Willden, D. W. Peterson, and A. R. Palmer.

OLDER PRECAMBRIAN BASEMENT

Rocks of the older Precambrian terrane that underlies the Apache Group are of interest here only as sources for the clastic materials of the Apache and Troy. Outcrops of the older formations, differentiated as to rock types recognized as detritus in the Troy and Apache, are shown in figure 2. Only the larger occurrences of the older Precambrian quartzites are shown, but additional thin units are fairly common in some of the pregranite formations north of the Salt River. Outside the region shown in figure 2, the outcrops of schistose rocks that have been studied are largely devoid of quartzite. Resistant rhyolites that have been recognized in detritus have been included with the nonresistant rocks. The proportions of rock types seen in present outcrops are very likely representative of those exposed in source areas just before Apache sedimentation began. The map indicates the dominance of granitic rocks and the sparsity of quartzites.

The Pinal Schist, dominantly a quartz-muscovite or quartz-muscovite-chlorite rock (Peterson, 1954; Gilluly, 1956, p. 10–11), underlies the Apache Group in many areas south of Globe. The Pinal consists mainly of metamorphosed sedimentary rocks, but also includes some intercalations of volcanic rocks (Ransome, 1919, p. 35–37; Gilluly, 1956, p. 11). Amphibolites and intrusive and extrusive rhyolitic units have been noted (Anderson, 1951, p. 1334–1335).

Slightly schistose and nonfoliated sedimentary and volcanic rocks, in part probably equivalent to the Pinal Schist, underlie the Apache Group north and west of the Sierra Ancha and crop out widely in the Mazatzal Mountains farther west (Wilson, 1939; Gastil, 1958). The sedimentary rocks are mainly shale, thin- to thick-bedded quartzite, and fine-grained to conglomeratic sedimentary units composed largely of volcanic debris. Many of the quartzites were thoroughly sheared or sheeted, then were firmly recemented, and are fine grained and highly vitreous. The relict sheeting effects as well as the dark shades of brown, red, or purple distinguish these rocks from later quartzite. The volcanic rocks are dominated by rhyolite flows, tuffs, and agglomerates, but basalt flows and agglomerates are also common. The southernmost large remnant of such rocks, within the area of Apache exposures, crops out along the south side of the Salt River southeast of the Sierra Ancha (fig. 2).

Much farther northwest, in the vicinity of Jerome, Prescott, and Bagdad, the Yavapai Series is made up of sedimentary and volcanic formations that are simi-

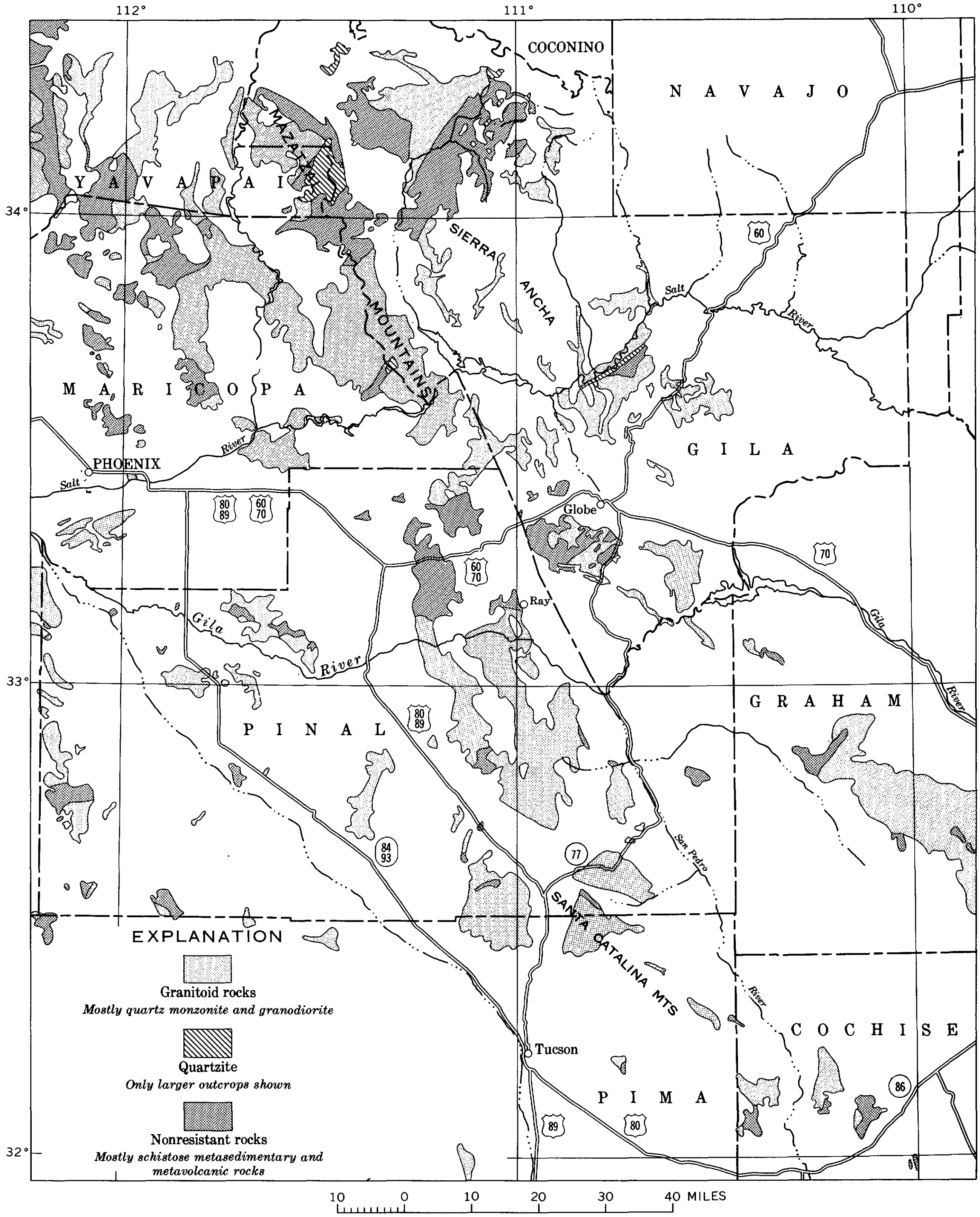

FIGURE 2.—Outcrops of older Precambrian terrane, showing types of rocks that were sources of clastics during late Precambrian time.

lar to those in the Mazatzal Mountains, except that quartzites and conglomerates are sparse (Anderson, 1951; Anderson and others, 1955, p. 7–12; Anderson and Creasey, 1958, p. 8–45). Small bodies of pyroxenite, gabbro, diorite, and rhyolite were intruded into all these rocks prior to invasion by granitic magmas.

All these rocks were strongly deformed, then invaded by granitic masses of batholith dimensions, and deeply eroded prior to deposition of the Apache Group. The granitoid masses in the area of Apache outcrop are dominantly of quartz monzonite or granodiorite (Peterson, 1938, p. 8–9; Peterson, 1954; Cooper and Silver, 1964).

Much of the quartz monzonite is a rather distinctive coarse-grained rock. In northern Gila County an abundance of large grayish-orange-pink to pinkish-gray microcline phenocrysts, which enclose many small grains of plagioclase and biotite, make the texture seem coarser on casual observation than it actually is. These poikilitic subhedral phenocrysts, ordinarily ½–1 inch long but sometimes as long as 2 inches, are conspicuously scattered through a coarse (1–8 mm) hypidiomorphic groundmass of pinkish potash feldspar, almost clear quartz, pale-greenish-yellow or yellowish-gray distinctly zoned chalky plagioclase, and thick books of black biotite. Sparse crystals of brownish-black sphene commonly can be noted megascopically. Apatite and minute grains of hematite(?) are accessories observed under the microscope. Orangish-pink aplite dikes seem to be the most common of minor associated facies. Quartz monzonite of this general description is termed Ruin Granite in the vicinity of Globe and is the dominant basement rock as far southeast as the Gila River; from Ray southward along the Tortilla Mountains a similar rock is said to make up most of the older terrane (Ransome, 1919, p. 37). The Oracle Granite of Peterson (1938), which crops out widely in and north of the Santa Catalina Mountains, is also similar.

Granitic detritus is abundant in the Pioneer, Dripping Spring, and Troy Formations of late Precambrian age. The coarser debris is readily recognized as petrographically similar to the Ruin Granite of the Globe-Miami area and the Oracle Granite of the Santa Catalina Mountains. Without doubt, these and similar monzonitic masses in adjacent regions furnished arkosic and quartzose detritus through most of late Precambrian time.

The Pinal Schist and lithologically equivalent formations—depicted mostly as nonresistant rocks in figure 2—seem to have furnished little detritus other than quartz and quartzite to the upper Precambrian formations. Some of the argillaceous materials of the several siltstone units in the Apache Group might have been derived from such sources. Reddish-brown jasper pebbles, reworked from the coarser fractions of the older quartzites, and pebbles from some of the denser rhyolites, are conspicuous but sparsely distributed in the conglomerates of the younger Precambrian. The bulk of the gravels in the conglomeratic units of the Apache Group and the Troy Quartzite are petrographically similar to the older quartzites. Granules and pebbles of quartz, which abundantly veins the older Precambrian rocks in some places, are locally abundant in the later conglomerates and are the principal coarse constituent of the Chediski Sandstone Member of the Troy.

During accumulation of most of the younger Precambrian sediments, the older Precambrian quartzites and resistant rhyolites—now known only within the area of figure 2—were buried by the basal units of the Apache Group and could not have been sources of the debris found in the succeeding formations. Therefore, a lithologic terrane similar to that shown in figure 2 must have been exposed somewhere outside the region of present Apache outcrops.

PRE-APACHE UNCONFORMITY

The mountains that were formed during the orogeny that terminated deposition in older Precambrian time in southern Arizona were leveled to a plain during the long episode of erosion that preceded Apache sedimentation. In degree of smoothness, this unconformity is comparable to the equivalent surface in the Grand Canyon, which has been interpreted as a peneplain formed by subaerial erosion and slightly modified by marine erosion (Sharp, 1940).

For most of the region the pre-Apache peneplain exhibits local relief of only a few feet at most. From the Sierra Ancha southward, variations in the thickness of the Pioneer Shale—the basal formation of the Apache Group—suggest that the surface may have had broad low hills rising slightly above the general plain or broad basins cut below it.

North of the Sierra Ancha the Pioneer Shale and at least the lower half of the Dripping Spring Quartzite lap out against a high on the basement surface (fig. 4). It seems more than a coincidence that this lapout occurs where sedimentary and volcanic rocks—generally the nonresistant basement formations—include small but numerous resistant units of quartzite and dense rhyolite. The line of lapout apparently trends east-northeast (fig. 4). It crudely parallels the northeast structural grain of the basement rocks and the boundary between these rocks and the less resistant granite—the prevalent basement rock as far south as Globe (fig. 2).

The northern belt of metasedimentary and metavolcanic rocks is the only area in which a prominent deviation from the pre-Apache peneplain has been observed. This is probably because no other extensive schistose zones include abundant ribs of resistant rock. Outcrops of the Apache formations are found overlying high parts of this northern belt, and undoubtedly Apache strata once extended much farther northwestward but were stripped away before the basal Paleozoic formations were deposited. Whether the area of pre-Apache lapout represents a large elongate monadnock rising above the general pre-Apache surface or a broad high of regional extent is not known.

The surficial granitic rocks disintegrated prior to deposition of the Apache Group. As remnant now, this zone of disintegration ranges from a few inches to a few tens of feet in thickness. In places the granitic debris, generally reddened by iron oxides, shows ill-developed bedding structures, but generally it represents residual material in virtually original position. Even where the basement rocks are not granitoid, the coarse fraction of the matrix of the overlying conglomerate consists largely of this disintegrated granite.

Where the unconformity truncates metasedimentary or metavolcanic rocks, the foliation or other planar structures may be somewhat obscured immediately adjacent to the erosion surface. In places a rubble of schistose fragments a few inches thick separates the basal conglomerate of the Apache Group from undisturbed schistose basement rock. Where such a regolith exhibits crude bedding, it commonly includes abundant angular to subrounded fragments of vein quartz, but is a unit lithologically different from the overlying conglomerate, which ordinarily consists mostly of debris from a distant source.

THICKNESS AND DISTRIBUTION OF YOUNGER PRECAMBRIAN ROCKS

Sections of individual formations of the Apache Group and of the Troy Quartzite are most fully represented in the Sierra Ancha. Farther south, owing to unconformities within and at the top of the sequence, sections generally are less complete.

Within the region between the Gila and Salt Rivers, the Troy Quartzite was everywhere thinned, and locally the Mescal Limestone and even the upper part of the Dripping Spring Quartzite were removed by pre-Paleozoic erosion. South of the Gila River the upper part of the Dripping Spring and all the overlying formations with their diabase intrusions were removed from large areas prior to Cambrian sedimentation; locally, however, most of the Mescal Limestone and moderately thick sections of the Troy Quartzite were preserved in blocks downfaulted prior to post-Troy erosion.

Thin to thick diabase sills are coextensive with the Troy Quartzite and the Apache Group, and they invaded the Apache in almost every known exposure; many sections of the Apache are displaced by sills at several horizons. Consequently, formations must often be pieced together from two or more partial sections found between sills. Actually, there are only a few places where the entire Apache section can be viewed from bottom to top, and these sections are not completely free of diabase sills.

Post-Paleozoic erosion of the younger Precambrian section was more extensive south of the Colorado Plateau than to the north. Also, these rocks to the south were much faulted during Tertiary time, and were partly covered by Cenozoic rocks. In the southern part of the region, therefore, thicknesses of the Troy Quartzite and the Apache Group are difficult to ascertain.

The thickness of the Troy Quartzite in any given area is in large part dependent on the degree of pre-Paleozoic faulting and the depth of pre-Paleozoic erosion. Therefore the only meaningful thicknesses of Troy, and those cited in table 1 and elsewhere throughout the report, are those remnant in exposures beneath a cover of Paleozoic rocks.

In the Colorado Plateau, where sections have been pieced together with considerable confidence at several localities, the Apache Group ranges in thickness from 1,250 to 1,600 feet. North of the Sierra Ancha, of course, these thicknesses are too great by roughly a factor of two because the Pioneer Shale and locally at least the lower half of the Dripping Spring Quartzite are lapped out. Throughout the Basin and Range area some sections of the Apache Group are as much as 1,600 feet thick, but other sections appear to be as little as 1,100 feet thick. The thinner sections are those reduced largely by (1) thinning of the Pioneer Shale, (2) erosion of the Mescal Limestone and the overlying basalt prior to deposition of the Troy Quartzite, or (3) a combination of these factors.

In the highest parts of the Sierra Ancha, the Troy is at least 1,200 feet thick and once was probably at least 800 feet thick over the entire region. Away from the Sierra Ancha as far south as the Gila River, remnants of the Troy generally range in thickness from 300 to 600 feet, but locally the formation is entirely eroded away. Farther south the formation is missing over large areas.

The younger Precambrian sedimentary sequences are thickest in the central Sierra Ancha, where the combined Troy and Apache total 2,650–2,800 feet. The thinnest sections are in the southern part of the region,

TABLE 1.—*Pre-Devonian formations of southeastern Arizona*

Age		Group, formation, and member	Thickness (feet)	Lithology and remarks
Paleozoic	Late and Middle(?) Devonian	Martin Limestone	30–400	Interbedded limestones, dolomites, shales, and sandstones. Carbonate strata are dominant. Northward from Globe, overlies Precambrian formations; unconformity at base becomes increasingly irregular; and basal sandstone, mostly less than 20 feet thick, locally thickens to more than 250 feet. The thick sandstones fill paleochannels as much as half a mile wide.
		Erosional unconformity		
Paleozoic	Late and Middle Cambrian	Abrigo Formation	0–840	South of Santa Catalina Mountains, mudstones dominate lower third; thin-bedded silty limestones, middle third; and sandy brown dolomites, upper third. Becomes increasingly sandy and dolomitic northward. North of Santa Catalina Mountains, thin-bedded and irregularly bedded mudstones and micaceous sandstones of lower part merge upward with sandstones and quartzites; sandy dolomites of upper 100–200 feet are the only conspicuous carbonate strata. Some northern occurrences rest directly on Precambrian rocks. Abrigo and Bolsa generally missing north of Globe owing to pre-Martin erosion.
Paleozoic	Middle Cambrian	Bolsa Quartzite	0–480	Thin- to thick-bedded gritty coarse-grained to well-sorted fine-grained cross-stratified quartzitic sandstone. Basal conglomerate typical. Gradational upward into Abrigo Formation. North of Dragoon Mountains, from older Precambrian rocks laps onto irregular surface on younger Precambrian formations. North of Santa Catalina Mountains, locally laps out completely, and north of Gila River, apparently is widely absent owing to lapout.
		Unconformity		
		Diabase		Mainly olivine diabase in sills, a few inches to 1,200 feet thick, and in narrow dikes. Within these intrusions are small differentiated bodies of diabase pegmatite, granophyre, and aplite. Sills are laterally extensive in all older formations.
		Intrusive contact		
		Quartzite member	0–500	Light-colored medium- to coarse-grained thin- to thick-bedded tangentially cross stratified quartzite, virtually free of pebbles and feldspar. Individual units well sorted.
				Friable to firmly cemented poorly sorted pebbly sandstone of well-rounded frosted quartz grains in clay and sericite matrix; basal conglomerate is conspicuous. Convolute lamination characterizes massive layers of

Younger Precambrian		Troy Quartzite	Chediski Sandstone Member	0–700	0–1,200		middle part, and medium-scale cross-stratification and thin conglomerate layers are typical of upper part. North of Salt River, is conspicuous owing to white color, massive outcrops, and odd erosion forms. Farther south, is less conspicuous, darker colored, irregularly bedded in part, more feldspathic, and locally quartzitic. Is basal member over most of region.
			Arkose member	0–450			Mostly fine-grained well-sorted arkose, conspicuously cross stratified on very large scale. Restricted to northwestern Gila County.
	Erosional unconformity						
	Apache Group	Basalt flows			0–375	1,250–1,600 (where all formations are present)	One to four flows of very hematitic amygdaloidal basalt; conspicuously porphyritic in many localities. Rests on algal member of Mescal south of latitude of Chrysotile.
		Erosional unconformity					
		Mescal Limestone	Argillite member	0–100	250–420		Flaggy- to massive-cropping reddish-orange to black siliceous argillite and subordinate mudstone; locally includes thin lenses of silicated limestone; typically a thin unit of chert conglomerate or breccia is at base. Member is best developed in McFadden Peak quadrangle and is absent south of Natanes Plateau.
			Erosional unconformity				
			Basalt flow (local)	0–110			Basalt flow like those above Mescal. Present only locally, in two small areas in McFadden Peak quadrangle and at Theodore Roosevelt Dam.
			Erosional unconformity				
			Algal member	40–130			Consists of two units, of about equal thickness: a lower thick-bedded stromatolitic biostrome and an upper thin-bedded and regularly bedded unit devoid of algal structures. Both are dolomite or silicated limestone. Generally chert is abundant only in upper beds of upper unit. Upper unit commonly thinned, owing to pre-Troy erosion.
			Lower member	150–270			Thin- to thick-bedded cherty dolomite originally; in some areas dolomite thoroughly silicified during formation of karst breccia; in most areas, has been metamorphosed to calcite limestone with abundant silicate minerals. In most of region, basal 15–100 feet is a coarse breccia of carbonate rock.

TABLE 1.—*Pre-Devonian formations of southeastern Arizona*—Continued

Age	Group	Formation	Member	Thickness (feet), member	Thickness (feet), formation	Thickness (feet), group	Lithology and remarks
Younger Precambrian—Continued	Apache Group—Con.	Unconformity				1,250–1,600 (where all formations are present)—Continued	
		Dripping Spring Quartzite	Siltstone member	200–370	550–700		Thin-parting feldspathic siltstone and subordinate quartzitic arkose. Siltstone is thinly laminated or cross laminated, and generally pyritic; mud cracks, scour-and-fill features, and stylolites are abundant.
			Arkose member	200–350			Thin- to thick-bedded massive-cropping firm arkose and feldspathic quartzite. Crossbedding characteristic but obscure. Member laps out in northern Gila County.
			Barnes Conglomerate Member	0–40			Mainly quartzite pebbles in arkosic matrix.
		Unconformity					
		Pioneer Shale			150–500		Mostly grayish-red tuffaceous siltstone or silty mudstone. Lower half contains units of arkose, which occurs in greatest volume in thick sections. Formation laps out in northwest Gila County and eastern Pinal County.
			Scanlan Conglomerate Member	0–30			Mostly quartzite or quartz pebbles in matrix derived from underlying rocks. Is absent or is merely a thin arkose with sparse pebbles through large areas. Locally at least 75 feet thick in northern lapout area.
	Nonconformity						
Older Precambrian	Granitoid rocks						Mostly coarse porphyritic quartz monzonite and granodiorite. Principal basement rock of region.
	Intrusive contact						
	Pinal Schist and equivalent metasedimentary and metavolcanic rocks						Small pendants of Pinal Schist in granitoid rocks in southern part of region. Poorly foliated metasedimentary and metavolcanic rocks, which include conspicuous quartzite units, increase in abundance in northern part.

where only Apache formations underlie the Paleozoic. In the southeastern part of the region, pre-Bolsa erosion was the principal cause of thinning and south of the Little Dragoon Mountains caused complete removal of the younger Precambrian rocks. If the diabase sills are included, the aggregate thickness of younger Precambrian layered rocks, in many areas north of the Gila River, are roughly twice those noted above, that is, as much as 5,000 feet.

The Apache and Troy rocks once extended over a 15,000-square-mile area (fig. 1) and undoubtedly once even had a much wider distribution. Individual stratigraphic units, whether formations, members, or even lesser units, are remarkably consistent in lithology and other features throughout this broad region and thus reflect a common site of deposition on a wide shallow shelf. The principal variations that do exist are due to postdepositional processes, which also affected large parts of the region.

Discussion of the origins of the younger Precambrian formations is largely deferred until all have been described, because they have much in common. Similarly, age and correlation are discussed in better perspective after description of relations to the diabase and to the Paleozoic formations.

APACHE GROUP

PIONEER SHALE

THICKNESS AND GENERAL CHARACTER

On the basis of the original descriptions, the Pioneer Shale has generally been supposed to range in thickness from 100 to 250 feet; actually the regional range is considerably greater. Near the west end of San Carlos Lake, the formation is thin or missing (fig. 3); and in exposures along the valley of the San Pedro River north of Mammoth, it is generally less than 30 feet thick and locally is absent. As far as is known the Pioneer does not exist east of this belt, and as already noted it definitely laps out north of the Sierra Ancha (fig. 4). In the canyon of Cherry Creek and along the southern front of the Sierra Ancha, the formation ranges in thickness from 250 to about 500 feet. Although as little as 150 feet of Pioneer Shale exists in many areas south of the Salt River—such as along the south escarpment of the Natanes Plateau and for several miles farther south and in places in the Ray quadrangle (Ransome, 1916, p. 136)—no pattern of thinning that would indicate other areas of lapout has yet been discerned. Indeed, as far south as the Apache Group extends, the Pioneer has maximum thicknesses of 300–500 feet. For example, Cooper and Silver (1964) measured a 306-foot section in the northwestern part of the Dragoon quadrangle, S. C. Creasey (written commun., Apr. 1961) measured thicknesses of 450–500 feet in the southwestern part of the Mammoth quadrangle, and Carpenter (1947) estimated a thickness of 400 feet in the Vekol Mountains.

The Pioneer Shale includes at its base the thin Scanlan Conglomerate Member; in many areas the remainder of the formation is mostly thin-bedded tuffaceous mudstone or siltstone. Where the Pioneer is more than 150 feet thick, mudstone ordinarily constitutes the upper one-half to two-thirds of the formation. The lower part is of fine- to medium-grained arkose or feldspathic sandstone intercalated with mudstone like that higher in the formation. These coarser beds may make up most or only a small part of the lower half of the Pioneer. In a few areas, as in the Dragoon quadrangle (J. R. Cooper, oral commun., 1958) and in a few places in the McFadden Peak quadrangle, similar arkose makes up as much as 50 feet of the uppermost part of the formation.

Outcrops of the Pioneer and encompassed diabase sills disintegrate readily and are commonly mantled in their own debris and the more resistant debris from overlying formations. Therefore it is frequently difficult to determine the proportions of mudstone, sandstone, and even diabase in a typical exposure. Along steep canyon walls the most firmly indurated siltstones crop out in thin ledges, and the sandstones form thin to thick ledges.

The contact between the Pioneer and the basal conglomerate of the overlying Dripping Spring Quartzite is everywhere sharp but in many places is also slightly undulatory. Rarely it is slightly discordant, or the conglomerate fills channels 1–2 feet deep in the top of the Pioneer. In some areas the upper 6 inches–3 feet of the Pioneer, normally grayish to blackish red, is bleached greenish gray or almost white. The bleached interval may represent a zone of regolithic material. No fragments of the Pioneer have been noted in the Dripping Spring. Only in the northernmost areas of outcrop does the lower part of the Dripping Spring have a purplish cast that might be attributed to fine material derived from the Pioneer. Because the upper part of the Pioneer is almost everywhere a monotonous poorly exposed sequence of mudstones lacking in distinctive marker beds, a regional angular unconformity would not be readily recognized. In general, the Dripping Spring and the Pioneer seem concordant. However, the few features that suggest post-Pioneer erosion and, in particular, the abrupt change in lithology indicate an unconformity of regional extent separating the formations.

LITHOLOGY AND STRATIGRAPHIC RELATIONS

SCANLAN CONGLOMERATE MEMBER

The Scanlan Conglomerate Member is merely a coarse-textured thin basal facies of the Pioneer. The

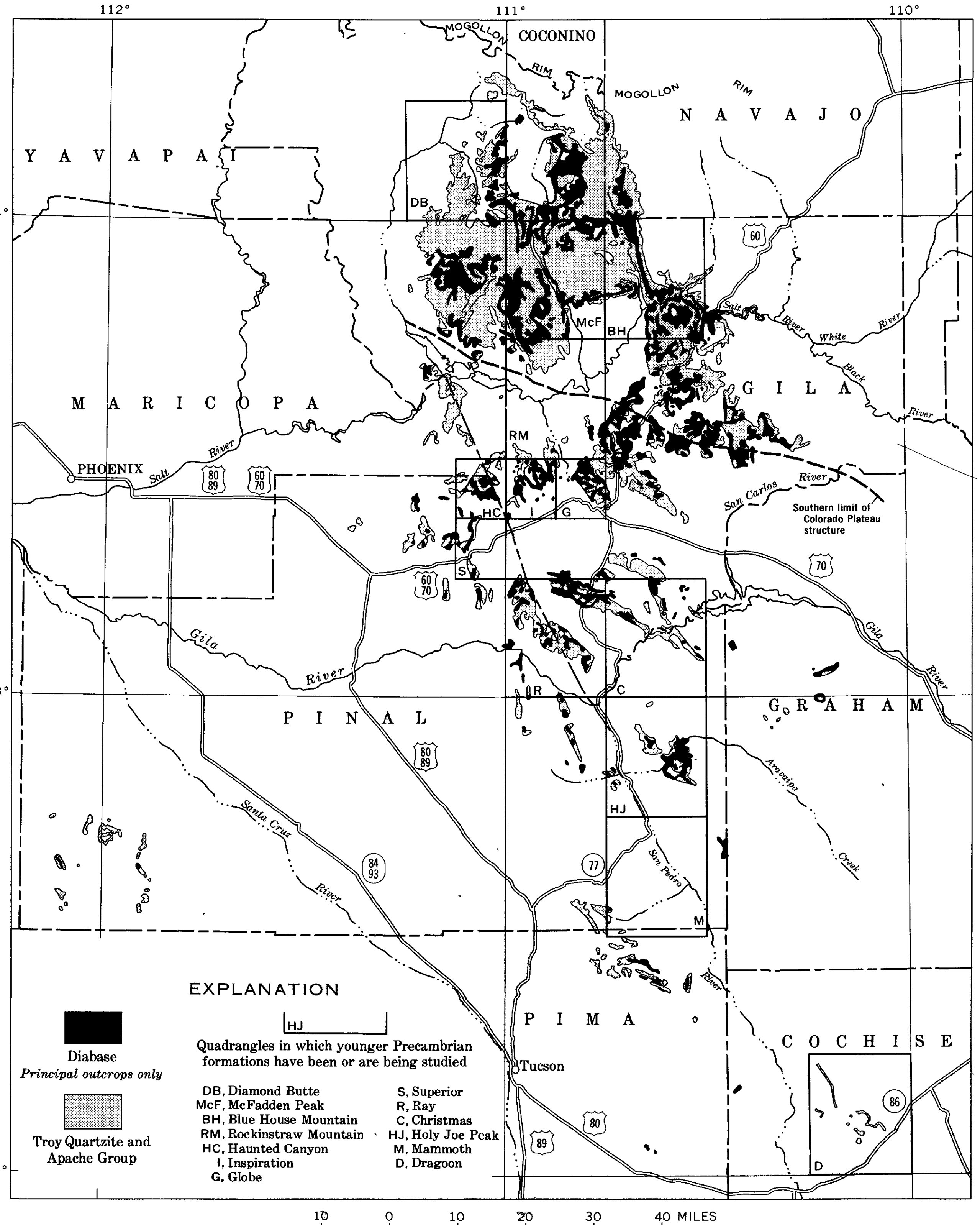

FIGURE 3.—Outcrops of younger Precambrian strata and coextensive diabase intrusions in southeastern Arizona. Modified from county geologic maps published by Arizona Bureau of Mines, 1958–60.

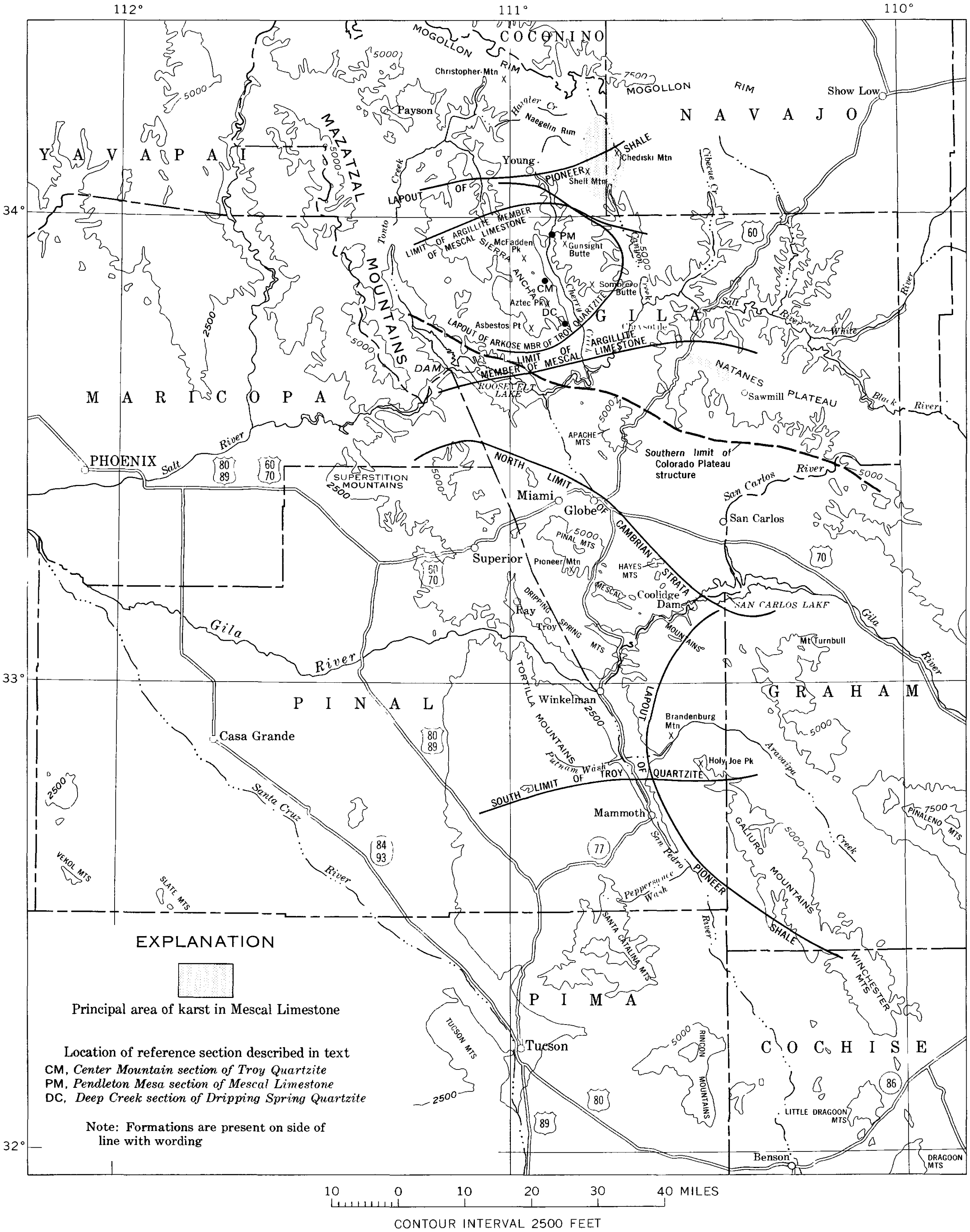

FIGURE 4.—Some paleogeographic features of the younger Precambrian formations in southeastern Arizona.

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conglomerate ordinarily crops out as narrow ledges or small cliffs and, except on dip slopes or in other unusual exposures, cannot be separately depicted even on large-scale maps. Therefore it should not be afforded formation status as in the past.

Typically the Scanlan Conglomerate Member consists of subangular to well-rounded pebbles or cobbles in a fine- to coarse-grained matrix and it is 1–8 feet thick. In many places the conglomerate is represented by only a few inches of granitic debris in which granules or small pebbles are sparsely distributed. Elsewhere, as at Roosevelt Dam, the conglomerate is as much as 30 feet thick, consists mainly of closely-packed cobbles, and comprises two or more beds. In a common variation the gravels of the Scanlan may include boulders as much as 1 foot in diameter and consist mainly of cobbles 4–6 inches in diameter; within half a mile along the outcrop, the gravels may be largely pebbles 1–2 inches in diameter. Similarly, along such a length of outcrop, the thickness of the conglomerate ranges from a few inches to—in a few places—20 feet; and in one locality the conglomerate consists of well-rounded gravels mostly of one lithology but in an adjacent area includes subangular gravels of the same or different rock types. In many localities pebbles and cobbles are disk shaped, and in some they are disposed in imbricate patterns. No consistent direction of imbrication has been noted. The regional variations in gravel sizes, composition, disposition, or thickness, which might suggest distance or direction from sources, seem to be no greater than the variations noted in relatively small areas.

Most of the gravels are of rock types foreign to the sites of deposition, but the matrix of the Scanlan Conglomerate Member reflects to some degree the composition of the underlying rocks. In most areas the pebbles and cobbles are mainly of light- to dark-gray and grayish-red quartzite and are closely packed in arkosic debris derived almost entirely from older Precambrian granite. White quartz pebbles are also common, and in some places white quartz constitutes most of the granules. Granules and small pebbles of reddish-brown jasper are sparse but conspicuous in many outcrops; the few pebbles of volcanic and dense schistose rocks, similar to those of the older Precambrian terrane in northwestern Gila County, are easily overlooked. Where the conglomerate overlies the more friable Pinal, it commonly includes abundant schist fragments, and the matrix may consist in large part of dark-colored shaly debris from the schist. Even here, however, angular grains of feldspar and quartz derived from granite are abundant in the matrix. Nearly everywhere the finer material of the matrix is hematitic, similar to the overlying mudstone or sandstone. In places, lenses of sandstone or tuffaceous mudstone are included in the conglomerate. Ordinarily the contact between conglomerate and overlying beds is sharp; however, transitional intervals are found which are a few inches to a few feet in thickness. Sporadic thin lenses of conglomerate or conglomeratic sandstone exist as far as 40 feet above the basal conglomerate.

TUFFACEOUS MUDSTONE

The so-called shale that dominates most Pioneer sections is in reality tuffaceous mudstone or tuffaceous siltstone that includes abundant grains of fine sand. The rock is minutely laminated or cross laminated (fig. 5) in beds ½–3 feet thick. Sparse small asymmetrical ripple marks occur in some sections. The rock is firmly indurated but closely jointed, both normal and parallel to bedding, so that it erodes to receding slopes. The mudstone is friable only in the sense that disk-shaped flakes, ¼–1½ inches across, occur in all stages from initial spalling on the outcrop to trains of chips that cover the slopes. The grayish-red mudstone is dotted with yellowish-gray to light-brown bleached spots, as much as 1 inch, but generally 0.1–0.2 inch, in diameter (fig. 5). Minute cubic grains of limonite, obviously derived from pyrite, can be seen in the centers of some of the smaller spots. In some localities, bluntly terminated lenticular bleached zones, as much as half an inch thick and several inches in diameter, lie parallel to either the bedding or, less commonly, the high-angle jointing. These bleached zones seemingly are most prevalent in the more quartzose beds.

The mudstone is composed of angular feldspar and quartz grains that are silt size to fine-sand size and of slightly larger irregularly lenticular devitrified glass shards all of which are set in a matrix of smaller shards and clay-size material (fig. 6). Quartz and sodium and potassium feldspar grains typically make up about a quarter of the rock. In some beds feldspar, mostly plagioclase, is dominant in this fraction. Minute flakes of muscovite spangle freshly fractured specimens of some units. Thin sections from specimens in which mica is not particularly noticeable to the unaided eye contain sparse to fairly abundant shreds of muscovite. The delicate lamination is defined by concentrations of platy grains—muscovite plates, shards, and tabular cleavage fragments of feldspar—all of which tend to be alined parallel to bedding. But the bedding is more obviously marked by the hematite that is thickly dusted throughout the rock and that opaques the finest grained parts of the matrix.

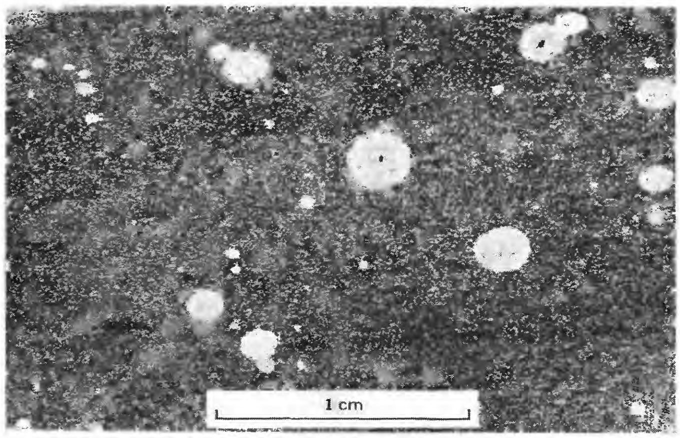


FIGURE 5.—Delicate cross-lamination in tuffaceous siltstone of the Pioneer Shale. Black specks in centers of bleached spots are limonite pseudomorphous after pyrite.

FIGURE 6.—Photomicrograph of tuffaceous siltstone of the Pioneer Shale. Wispy, irregular, or curved grains are relicts after glass shards; clear (white) nearly equidimensional grains are quartz; similar, but clouded, grains are feldspar; dark areas are heavily dusted with hematite. From same specimen as that of figure 5.

On the basis of X-ray studies, Gastil (1954) suggested that the devitrified shards consist largely of finely divided muscovite and very fine grained chalcedony. Some of the fine-grained matrix material appears to be optically similar to the shards, and probably much of it is of like mineralogic composition. Gastil also postulated that the glass was rhyolitic in composition. Robert L. Smith, who has studied tuffaceous rocks extensively, stated (oral commun., Aug. 1956) that the rodlike and bifurcated cross sections and other features of the shards suggest a composition no more mafic than andesite.

In some localities, hard tuffaceous beds, which look like very fine grained impure quartzite, are interlayered with the usual grayish-red mudstone. These beds generally are dusky red purple or blackish red. Planar lamination but no cross-lamination has been observed. Muscovite flakes are especially conspicuous in these beds and in the adjacent mudstones. Quartz and feldspar grains are slightly larger than those in the softer mudstone but still are mostly of fine-sand size. Quartz composes more of the silt-size and larger grains, clay-size materials make up less of the matrix, and there are fewer shards than in the usual mudstone. Beds of quartzitic mudstone have been observed throughout the region, but they seem to be more prevalent south of the Apache Mountains than north and are particularly noteworthy in the southernmost occurrences of the Pioneer.

ARKOSE

Arkose occurs mostly as thin- to thick-bedded cross-stratified layers intercalated with tuffaceous mudstone. Some sections include similarly bedded feldspathic sandstone in small amounts. These arkosic beds are grayish red; most are fine to medium grained, but some are coarse grained. Individual beds are fairly well sorted, but adjacent beds may differ in grain size. In a few areas the lower 40–90 feet of the formation is composed almost entirely of thick-bedded massive-cropping arkose that is medium grained and firmly cemented. Even in thin section these beds are not notably different from the flesh-colored sandstone of the arkose member of the Dripping Spring Quartzite. Isolated outcrops have been mistaken for Dripping Spring. In still fewer localities one or two beds of light-gray or pinkish-gray medium-grained feldspathic sandstone (more like parts of the Chediski Sandstone Member of the Troy than any other unit) also occur in this part of the formation.

Coarse-grained and very coarse grained arkose is usually confined to the basal 25–40 feet of sections that are at least 250 feet thick. This arkose is thick to very thick bedded and generally grayish red. Locally it contains thin beds of light-gray coarse feldspathic sandstone or of feldspathic conglomerate made up largely of quartz granules. In a few localities very coarse grained arkose that is mostly light colored, very poorly sorted, partly pebbly, poorly bedded, and obviously derived from granitoid rocks of the immediate vicinity constitutes the lower part of the Pioneer. A noteworthy outcrop, in which detritus of the basal 45–50 feet above the Scanlan Conglomerate Member is from the underlying Madera Diorite, is on the north flank of Pioneer Mountain, in the Ray quadrangle near the type locality. These various coarse strata are not similar to any other unit in the younger Precambrian sequence.

Generally, the abundant hematitic clay in the matrix of the arkose of the Pioneer imparts a reddish cast that distinguishes it from all other arkoses in the Apache Group. Furthermore, although individual beds of the Pioneer may be well sorted, probably no unit comprising several arkose beds is as well sorted as an equivalent thickness in the arkose member of the Dripping Spring. In sections of medium-grained arkose, where these features may not suffice for recognition, thin seams of brownish-black or blackish-red, dense, and in places exceedingly tough tuffaceous siltstone invariably separates some of the beds. Even in sequences where these seams are discontinuous and sparse, a few are as much as 8 inches thick.

LATERAL VARIATIONS

Significant lateral variations in the Pioneer Shale have been noted only in the northwestern part of the region and are restricted to the belt of lapout. There the Pioneer becomes appreciably coarser and particularly in its lower part varies greatly in distances of a mile or two. In the northwest corner of the McFadden Peak quadrangle, for example, the lowest 60–90 feet of the Pioneer is very coarse grained crossbedded feldspathic sandstone, which includes beds of granule conglomerate and many lenses of pebble conglomerate. The remainder of the 200-foot section is of medium- to coarse-grained gray arkose that, except for dusky-red weathered surfaces and a tendency to be more friable in parting, is similar to the overlying basal arkose of the Dripping Spring. To the northwest the lower half of the section is even more conglomeratic. In the nearest exposures to the southeast and northeast, 3–5 miles distant, Pioneer sections dominantly are tuffaceous mudstone, as are most sections farther south. This aberrant coarse facies of the Pioneer apparently marks areas where the pre-Pioneer surface not only was high but also had appreciable local relief. If mudstone is lacking or subordinate in a moderately thick section, the Pioneer thins abruptly nearby or is missing because of lapout.

This relation to local relief is even more strikingly exhibited in the Scanlan Conglomerate Member. Variations are seen in tracing the lower strata of the Apache north 6 miles from the northwest corner of the McFadden Peak quadrangle to Potato Butte, which is 4 miles west of Young. At the south end of this interval the Pioneer is about 200 feet thick, but it laps out about 3 miles to the north. Where the pre-Apache surface exhibits little local relief, the conglomerate is fairly typical; in places it is represented by a few inches to 3 feet of granitic debris which contains only scattered gravels of quartzite, and in nearby areas 3–8 feet of cobble conglomerate mark the base of the Pioneer. Northward from about the place where the Pioneer thins out completely, however, the basal conglomerate begins to thicken and become coarser; at Potato Butte—where possibly as much as 100 feet of the Dripping Spring Quartzite is missing because of lapout—the conglomerate is 85–110 feet thick. Here the conglomerate is mostly closely packed well-rounded cobbles 3–6 inches in diameter, but the basal 50 feet includes abundant larger cobbles and scattered boulders as much as 18 inches in diameter. This unusually thick conglomerate overlies an irregular surface cut on granite; local relief of as much as 35 feet exists along an outcrop length of 200 feet. Here, and in similar examples, the granite surface was swept clean of the arkosic debris usually seen below the conglomerate.

To the southwest 2–3 miles, where hills of the pre-Apache surface project up through about 200 feet of Pioneer and into the Dripping Spring, the Scanlan is as much as 75 feet thick, and the remainder of the basal two-thirds of the Pioneer is unusually coarse grained and ill sorted and contains many conglomerate beds. The beds are lenticular, and medium- to large-scale cross-stratification is conspicuous. Mudstone occurs only in the upper third of these sections. In contrast, east and north of Young, where the pre-Apache surface is high but virtually planar, the basal conglomerate generally is only 6–8 feet thick and in places is only a few inches thick. In this area the Pioneer is thin or missing; where present, in sections only 25–100 feet thick, mudstone beds make up at least one-third of the formation.

Some of the basal conglomerate that is continuous with the Scanlan but directly underlies the Dripping Spring Quartzite exhibits poor sorting and rounding. The most accessible exposure is 9½ miles due north of Young along the north wall of the canyon of Haigler Creek. (Specifically, this exposure is in the W½ sec. 6, T. 10 N., R. 14 E., at and immediately west of the hairpin turn of the Chamberlain Trail road, which there turns north to follow the crest of the ridge that lies east of Gordon Canyon.) At this locality the older Precambrian quartzite dips 55°–60° ESE., and the siltstone member of the Dripping Spring that here overlaps this quartzite dips about 15° E. The pre-Apache surface exposed in the roadcut along the canyon has relief of several feet. Above the unconformity is a breccia, 20 feet thick, composed of angular blocks of the older Precambrian quartzite in a matrix of medium- to coarse-grained sandstone. Some of the blocks, which are as much as 2 feet in diameter, traveled only a few feet and can be matched back into the irregularities on the surface from which they were plucked. A few hundred feet to the west, the breccia gives way to a conglomerate of poorly rounded boulders, some as much as 3 feet in diameter; at 600 feet west of the breccia exposure, this conglomerate is 40 feet thick. The matrix of this conglomerate and the beds several feet above the conglomerate are arkose, similar to the sandstone that ordinarily makes up the arkose member of the Dripping Spring. Though rare, such coarse breccia and conglomerate are not abnormal where basal units of the Apache Group lap onto very resistant older Precambrian rocks along a surface of some relief. Wilson (1939, p. 1151–1152) described other examples in this vicinity.

Where the Pioneer is thin along the San Pedro River valley (fig. 4), in contrast with the northwestern area of thinning, the basal part of the section shows no coarsening. The Scanlan Conglomerate Member is generally very thin or is represented only by scattered pebbles in the basal few inches of the formation, and the overlying strata are mostly the hard quartzitic mudstone. Such sections may have been deposited on a high of very slight local relief, or they may represent sections in which most of the Pioneer was eroded away prior to deposition of the Dripping Spring Quartzite.

The medium- to coarse-grained sandstone that occurs at the base of locally thick sections, away from the lapout area of northwest Gila County, probably has no regional paleogeographic significance. The differences in lithology apparently reflect local derivation of detritus.

DRIPPING SPRING QUARTZITE

In a discussion of Dripping Spring stratigraphy published since this report was prepared, Granger and Raup (1964) presented the three-member subdivision that is used here. They, too, regarded exposures in Bull Canyon—see reference section—as among the most representative of the formation. In presenting detailed descriptions of sections at several other localities in Gila County, they elaborated on some lithologic details that are given little consideration here. Thus, although the following may supplement, it does not take the place of their descriptions, which are easily the most definitive since the early definitions of Ransome (1915; 1916).

The Dripping Spring Quartzite, which was redefined to include at its base the Barnes Conglomerate Member (Granger and Raup, 1964), is generally about 600 feet thick, and it consists fundamentally of two elements of approximately equal thickness that contrast in composition, texture, bedding structures, and forms of outcrops. The lower half of the formation includes subordinate units of light-colored feldspathic quartzite but is mostly massive-cropping flesh-colored arkose. (In this report, a quartzose sandstone that includes 10–25 percent feldspar is termed feldspathic; one that contains more than 25 percent feldspar is described as an arkose.) The lower element, excluding the thin basal conglomerate, is herein termed the arkose member. Somewhat more than half of the upper element is dark-colored siltstone, which is interbedded with fine-grained arkose and feldspathic quartzite. In contrast with the arkose member, all strata are thin bedded, and outcrops are conspicuously flaggy or slabby; and overall they are considerably darker. The siltstone layers so strongly influence overall aspects of the upper element that it is designated the siltstone member. The middle and upper members of Granger and Raup (1964) are the arkose and siltstone members, respectively, of this discussion.

Although the Barnes has previously been designated a formation separating the Dripping Spring and Pioneer, mapping of the Apache Group has demonstrated repeatedly that separate delineation of the unit is impracticable, even at map scales as large as 1:12,000. The Barnes has obvious lithologic and genetic affinities with the Dripping Spring Quartzite. Put into proper perspective, this conglomerate is analogous to the basal sandstone of the Mescal Limestone, the basal conglomerate of the Troy Quartzite, and the conglomerate at the base of the Cambrian System. It merges with the Scanlan Conglomerate Member where the Pioneer laps out, and it is basal to the siltstone member where the Dripping Spring laps out; because of these features and especially because it is usually thin and locally absent, the Barnes is best visualized simply as the basal unit of the Dripping Spring, perhaps without even member status. In deference to its historical designation as a singular unit, however, and inasmuch as it does furnish conspicuous outcrops or float that serve to emphasize the unconformable contact between the Dripping Spring and Pioneer, it is termed the Barnes Conglomerate Member.

Gross aspects of the Dripping Spring Quartzite are apparent at the type locality (Ransome, 1903, p. 30–32) but, owing to faults, alteration, and cover, the piecing together of stratigraphic details at that locality is tenuous. The following section in Deep Creek canyon, measured 2½ miles west of Cherry Creek and near the south boundary of the McFadden Peak quadrangle, is representative of complete sequences and is considered the reference section.

Reference section of Dripping Spring Quartzite

[Measured on west wall of Deep Creek canyon, 3,000 ft north of junction of Deep Creek and Bull Canyon, in SW¼SW¼ sec. 18 (unsurveyed), T. 5 N., R. 15 E.]

	Thickness (feet)
Mescal Limestone:	
Dolomite breccia, in sandy dolomite matrix	20+
Arkose, pale-yellowish-brown; mostly medium grained but contains well-rounded very coarse quartz grains; feldspar content about 40 percent, dolomitic; firmly cemented; beds tabular, 6–24 in. thick, internally cross stratified mostly on large scale; unit crops out as massive locally vuggy ledge with poorly defined partings at ½–3-ft intervals; weathers moderate yellowish brown with lustrous black coating between grains. In adjacent areas, it is 4–9 ft thick and forms ledge that is prominent below slope-forming dolomite breccia; lower contact everywhere sharp	11
Unconformity.	
Dripping Spring Quartzite:	
Siltstone member:	
9. Arkose, pale-yellowish-brown to grayish-orange-pink, fine- to medium-grained, thin-bedded (2–36 in.); includes some beds of feldspathic quartzite; medium-scale planar cross-stratification conspicuous; weathers pale to dark yellowish brown, black surface coatings and grayish-red stains common; mud cracks and asymmetric ripple marks common. Lower half thicker bedded (1–3 ft), coarser grained, and less feldspathic than upper half and is cliff former; upper half crops out as ledges on slope; top 15 ft includes thin beds of siltstone like those of unit 8	99
8. Siltstone, dusky-yellow-green to very dark gray, very feldspathic; laminated, thinly laminated, and cross laminated; stylolites abundant, mud cracks common; undulant or irregular partings, 0.1–6 in. apart, characterize outcrops; unit weathers pale yellowish brown to grayish red; weathering obscures original texture; pyrite or minute yellow-green spots from which pyrite has been leached are abundant. Lowest 33 ft includes interbeds of arkose like those of unit 7 and crops out as small cliff; siltstone above exposed in a slope, which steepens so that uppermost 40 ft crops out as a cliff or steep ledgy slope	117
7. Arkose, pale- to dark-yellowish-brown; weathers slightly darker, and black coating common; very fine grained; low-angle medium-scale planar cross-lamination etches out poorly; outcrops part conspicuously at 2-in.–6-ft intervals; "pockmarks," ½–3 in. in diameter, on weathered surfaces are accentuated by black coating. Crops out as vertical face in recess below cliff	14
Dripping Spring Quartzite—Continued	
Siltstone member—Continued	
6. Siltstone, like that of unit 8, with subordinate thin beds and scour fills of arkose, like that of unit 7; sparsely micaceous, thinly laminated to thin bedded; many beds cross stratified; scour-and-fill channels, 3–24 in. wide, are abundant; slabby undulatory partings at 2–10-in. intervals prominently defined by shaly seams; shaly parting common in siltstone beds; stylolites abundant; weathered rock assumes porcelaneous texture. Slabs, weathered grayish red, thickly cover moderate slope, which steepens upward to almost vertical cliff locally	54
5. Arkose, like that of unit 7, with minor beds of siltstone like that of unit 8, except moderately micaceous. Basal 6 ft of dusky-red-weathering especially micaceous thin-bedded arkose grades upward into shaly siltstone that dominates interval 6–14 ft above base. Interval 14–23 ft up is hematite-flecked arkose in beds 6–30 in. thick. Upper 27 ft is arkose, in beds 2–8 in. thick, and subordinate thinner beds of siltstone; stylolites numerous; this part forms prominent hacky fractured ledge at top of slope-forming unit. Contact with arkose member sharp and planar	50
Thickness of siltstone member (units 5–9)	334
Arkose member:	
4. Quartzite, feldspathic, grayish-orange-pink to pale-yellowish-brown, medium-grained; weathers pale yellowish brown; crops out as massive rim-forming cliff; slightly etched face suggests large-scale low-angle thin cross-stratification; poorly defined partings at intervals of 2–15 ft (average 6 ft) suggest tabular beds; otherwise bedding structures obscure. Outcrops of basal 10 ft and uppermost 20 ft pock marked. Bottom 8 ft and top 12 ft less feldspathic. White quartz pebbles and irregular disks of white-weathering chert, as much as 1 in. in diameter, sparse in interval 20–30 ft above base; in adjacent areas, pebbles exist through thicker interval	65
3. Arkose, light-brown to pale-reddish-brown, fine-grained; in thinly laminated sets of beds 4–6 in. thick; breaks with quartzitic fracture; composition like that of unit 2, but appears more vitreous on fresh fracture; crops out everywhere as steep to vertical cliff that exhibits partings at intervals of 1 in.–15 ft, mostly at intervals greater than 6 ft; weathers pale red to pale reddish brown; on fresh break, weathered arkose exhibits minute patches of yellowish-orange clay(?). Cross-stratification, medium-scale and low-angle, apparent only in top 8 ft	81

	Thickness (feet)
Dripping Spring Quartzite—Continued	
Arkose member—Continued	
2. Arkose, moderate-reddish-orange, fine- to medium-grained, thin- to thick-bedded (1–4 ft); small- to medium-scale straight and concave planar cross-stratification etched out locally; concave crossbedding dominant in upper 20 ft, straight crossbedding dominant in lower 50 ft; asymmetric ripple marks sparse. Crops out on steep slope as bare rounded ledges with conspicuous partings at 51, 116, and 134 (top) ft above base of unit; in lower 50 ft of outcrop, obscure partings occur at intervals of 2 in.–4 ft (mostly latter); obscure partings of upper 80 ft at intervals of 1–8 ft (typically at 4–6 ft). Composed of angular to subrounded grains of clear quartz and reddish-orange potassium feldspar (30–40 percent of rock); scattered minute aggregates of hematite common; minute patches of dark-yellowish-orange limonite or limonitic clay characterize matrix of freshly broken weathered rock	134
Thickness of arkose member (units 2–4)	298
Barnes Conglomerate Member:	
1. Conglomerate, well-rounded pebbles and cobbles ½–8 in. in diameter (average 3 in.) in matrix of coarse-grained to very coarse grained light-brown to grayish-red arkose. Gravel mainly of fine-grained vitreous quartzite; pebbles of white quartz common; pebbles of reddish-brown jasper and andesite(?) scarce. Contains sparse lenses of arkose. Crops out as one massive ledge; to east across canyon, unit is 1–8 ft thick. Contact with Pioneer Shale sharp and slightly undulant	18
Thickness of Dripping Spring Quartzite	632
Unconformity.	
Pioneer Shale:	
Siltstone, grayish-red; has yellowish-gray to light-brown bleached spots; abundantly tuffaceous; forms steep slope; top 2 ft shaly and bleached light greenish gray	30+

Ordinarily a twofold subdivision, in which the arkose and conglomerate members are combined as one unit, proves most practical for general discussion and for detailed delineation of units as an aid in deciphering structurally complex areas.

Only small parts of the Dripping Spring can be described correctly as quartzite. Strata throughout the formation are so firmly cemented, however, that the rock fractures across component grains and many units form resistant outcrops comparable to those eroded from quartzites. In view of these aspects, rather than its having a highly quartzose composition, the formation was originally termed a quartzite, and this designation has unfortunately delayed recognition of the true makeup of the formation. Nonetheless the quartzite look is such a useful characteristic for distinguishing the Dripping Spring from adjacent formations that retention of the term "quartzite" in the formation name is deemed worthwhile.

The arkose member is 200–350 feet thick, the siltstone member is 200–370 feet thick, and the combined thickness of these subdivisions—in sequences normally underlain by the Pioneer and overlain by the Mescal—is apparently everywhere at least 550 feet. The Barnes Conglomerate Member adds negligibly to the thickness. Sections in excess of 700 feet have been reported, but those checked by me were suspect because of faulting or inflation by diabase intrusions.

From the latitude of Young south as far as the Gila River, the Dripping Spring is generally overlain by the Mescal Limestone, but in a few localities south of the Salt River, part of the siltstone member of the Dripping Spring has been stripped away during one of the intervals of erosion that preceded deposition of the Troy Quartzite, the Bolsa Quartzite, or the Martin Limestone.

South of the Gila River in the area presently drained by the San Pedro River, the arkose member of the Dripping Spring was locally exposed prior to deposition of the Troy, and in large areas the formation was again eroded before the Bolsa Quartzite was deposited. Fairly thick and presumably complete sections exist, farther west, in the Vekol Mountains and adjacent ranges 50–60 miles south of Phoenix (L. A. Heindl, oral commun., 1959).

The arkose member of the Dripping Spring laps out north of the Sierra Ancha, so that in a few localities the siltstone member rests directly on the older Precambrian terrane. Locally, as along the headwaters of Canyon Creek, the formation was thinned additionally by erosion that preceded deposition of the Martin Limestone.

BARNES CONGLOMERATE MEMBER

The Barnes Conglomerate Member, except for a different matrix and perhaps a tendency to be more persistent and uniform in texture, is similar to the Scanlan Conglomerate Member of the Pioneer. The Barnes generally consists of well-rounded granules, pebbles, and cobbles moderately well to firmly cemented in a matrix of arkose; locally the matrix is feldspathic sandstone. The gravels are mainly of light- to dark-gray and grayish-orange to grayish-red vitreous quartzite; granules and pebbles of white quartz are common; granules and small pebbles of jasper and pebbles of volcanic rock, largely rhyolite, are sparse. The gravels are mostly spheroidal or ellipsoidal, but in some localities many are discoidal. Imbrication is uncommon and shows no dominant direction. The matrix ranges from very fine grained to very coarse grained, and is ill sorted; in composition it resembles the arkose of the overlying beds, but it is generally much coarser.

The conglomerate ordinarily ranges in thickness from 5 to 30 feet. Locally, however, it is represented only by scattered granules and small pebbles along the bottom of the basal bed of the arkose member; at the other extreme it is as much as 40 feet thick. The thicker sections commonly consist of two or more beds of conglomerate, partly separated by lenses of pebble-free arkose. Variations in thickness of the conglomerate and variations in composition, size, shape, rounding, and sorting of the gravels seem to be no greater regionally than those that can be observed in a small area.

The transition zone from conglomerate to the overlying arkose is usually only a few inches thick, but transition zones of pebbly arkose several feet thick exist in some localities. A few feet or even a few tens of feet of the arkose immediately above the conglomerate is ordinarily slightly coarser grained, however, than the overlying beds.

ARKOSE MEMBER

The arkose member is so firmly cemented that details of bedding are obscure, and the member forms massive outcrops. The upper half of the member commonly forms a cliff, and the lower half, a series of steep ledges. In most areas three to five partings consistent in stratigraphic position are etched out strongly; other bedding separations are seen only on close examination. Only in a few areas, where the section was deeply weathered prior to the present cycle of canyon cutting, do all bedding structures etch out to refute the general impression that the arkose member is made up of very thick bedded units having little or no internal bedding structure. Usually the Barnes Member forms a separate prominent ledge, although in areas of extreme relief the two members may form one cliff.

The lower two-thirds to three-fourths of the member is composed of arkose, and the upper one-quarter to one-third is feldspathic quartzite. The arkose is pale brown to pale reddish brown or reddish orange, fine to medium grained, thinly laminated to thick bedded, cemented as firmly as a quartzite, and grayish orange pink to moderate brown when weathered. A unit of several beds of arkose may be uniformly fine grained or uniformly medium grained; little vertical gradation of sand sizes exists within a bed or even within a unit of several beds. The feldspathic quartzite is light gray to pale red and generally medium grained. Scattered small pebbles of light-gray chert and white quartz, not seen in every bed or outcrop, characterize this quartzite. Otherwise, in texture and bedding the quartzite is similar to the arkose. Outcrops of the quartzite tend to be pock marked, owing to the weathering out of less firmly cemented aggregates of sand.

In general the member consists entirely of tabular beds, most of which are internally cross stratified. Some thin beds are thinly laminated. Many of the beds are only 4–6 inches thick, but those that make up most of the member exceed 4 feet and in some areas are 12–20 feet thick. The crossbedding is of the straight or concave (tangential) type, of low to moderate angle, and of small to moderate scale (McKee and Weir, 1953).

Appreciable departures from the usual lithology of the arkose member have been noted only north of the Sierra Ancha—in or near the area where the Dripping Spring thins because of lapout. Bedding features etch out more conspicuously in those areas than farther south: cross-stratification of medium to large scale (5–40 ft between planes of truncation) is obvious in wedge-shaped sets of strata that constitute the thicker bedding units. Locally the dips of such cross-strata exceed 20°, but generally dips are less than 10°. Wedge-shaped bedding units have been noted only rarely farther south but may have been overlooked in the quartzitic outcrops. In some sections of the northern part of the region, the arkose is mostly medium to coarse grained, rather than fine to medium grained, but is still fairly well sorted.

The lack of local variation in the textural or bedding features in the northern area is well illustrated in exposures along Bryant Canyon, 7 miles southwest of Young. There, a moderately thick section of Pioneer and most of the arkose member of the Dripping Spring lap out against a small steep-sided hill of older Precambrian quartzite and volcanic rock. A zone of cemented rubble 4–12 feet thick, consisting of angular quartzite blocks, mantles this "fossil" hill. Thin layers of such blocks occur along certain bedding planes of the Dripping Spring for distances of a few tens to a few hundreds of feet from the hill, and sparse lone blocks are found even farther away. Otherwise, the member is uniformly of the medium- to coarse-grained, fairly well sorted arkose that is usual within a radius of several miles of this locality. Bedding structures immediately adjacent to the hill are no different from those in the same stratigraphic units a mile or two distant.

Near the hill the Barnes consists of one and in places two 6- to 10-inch layers of pebbles; locally there is no basal conglomerate. White quartz pebbles, unusual only in that they are mostly angular to subrounded, locally dominate the gravel. The quartzite and rhyolite pebbles, however, are well rounded. Although not representative of the average Barnes, these exposures illustrate that the Barnes in the area of lapout is not notably different from many outcrops farther south.

SILTSTONE MEMBER

The thin-parting upper member of the Dripping Spring Quartzite, in marked contrast to the topographic expression of the massive arkose member, forms slopes broken intermittently by ragged ledges or small cliffs. The boundary between the members is sharp and is commonly marked by a bench a few feet to a few tens of feet wide cut into the less resistant basal beds of the siltstone member.

The five units delineated separately in the reference section are easily recognized in most exposures. Three units are conspicuous either because of color or topographic expression. These are an uppermost cliff-forming arkose unit, which makes up the upper quarter to third of the member, and two fairly thick units of dark-colored siltstone, which make up much of the lower two-thirds of the member. The two siltstone units are separated by a relatively thin unit of arkose, and the lower siltstone is also underlain by, a thin arkose unit. Thin arkose beds, singly or in sets, are interlayered with the parts of the member that are mostly siltstone; similarly thin beds and seams of siltstone are numerous in parts that are mainly arkose. The siltstone causes the entire member to be fairly dark and to contrast strongly with the flesh-colored strata of the lower member.

The five units seemingly vary appreciably in thickness and in volume of arkose beds from one outcrop to another. These variations may be more apparent than real, however, because the boundaries between siltstone and arkose units are difficult to define consistently from one locality to another. This difficulty exists primarily because weathering obscures textural differences between siltstone and arkose. Secondarily, the rocks have been modified by metamorphism adjacent to the numerous diabase intrusions. As criteria for recognition of original lithology become better defined, future studies may prove the units to be more consistent in thickness than is now demonstrable.

LITHOLOGY

Strata that crop out and fracture like quartzites are mostly very fine grained or fine-grained arkoses; a few are medium or even coarse grained. On fresh fracture these rocks are grayish orange pink to dark yellowish brown. Silt- and clay-size material makes up 30 percent or more of the rock. Only in the coarsest grained sandstone that includes the least fine matrix can the feldspar content be accurately estimated. Such rocks are mostly arkose and contain from 30 to 50 percent clastic feldspar grains; a few beds are feldspathic quartzite with a feldspar content approaching 25 percent.

The siltstone is highly feldspathic and slightly to moderately micaceous. As noted by Granger and Raup (1959, p. 425, 439), the potassium content is abnormally high for a clastic sedimentary rock. Six specimens from widely separated outcrops of the lower siltstone (unit 6, reference section) contained 10.7–14.6 percent K_2O, the average of the six specimens being 12.3 percent.

The entire siltstone member is unusually radioactive, and in some areas the finer grained strata are notably uraniferous. During 1950–56 the member was intensively prospected, especially in northern Gila County, for the local concentrations of uraninite and secondary uranium minerals that approach ore grade. According to Granger and Raup (1959, p. 438–439), an appreciable part of the radioactivity is attributable to the high potassium content of the siltstone units.

Fine-grained pyrite is disseminated in the siltstone and is the main cause of the medium- to dark-gray color characteristic below the zone of weathering. Small amounts of carbonaceous material are also disseminated in the rock, and in some strata thin seams of fine-grained graphite have been noted (Granger and Raup, 1959, p. 442–443). Pyrite occurs in smaller amounts in the arkose, principally along joints, fractures, and stylolites.

WEATHERING MODIFICATIONS

Much of the siltstone in surface exposures is very dark, but leaching of pyrite causes it to weather yellowish gray to moderate brown, much like the arkose. Additionally, this process causes both the finer grained interbedded arkose and the siltstone to assume a coarsely porcelaneous texture. As a consequence the two rock types cannot be readily distinguished, and lateral correlation of individual beds or sets of beds is difficult. Weathering effects extend from a few feet to a few tens of feet below the outcrop, and in some areas the dark color of the siltstone can be seen only in deep excavations or mine workings. Adjacent to fractures, joints, and other partings, the bleached strata commonly are strongly impregnated with a grayish-red limonite stain; such stains are much more abundant in the siltstone units than in the arkose units.

In many areas, such as along the canyon of the Salt River, outcrops of the siltstone member are lightly coated or even crusted with reddish-orange to dark-reddish-brown limonite. This rusty coloration, coupled with the thin bedding, distinguishes these outcrops at a distance from all others in the younger Precambrian sequence. Because of slight differences in porosity, texture, composition, and exposure, the coatings produce a conspicuous surficial color banding, which parallels the gross bedding features.

Conspicuous color banding is characteristic of only the siltstone member. It is not an aid to recognition of the entire formation, as generally has been extrapolated from Ransome's (1903, p. 138) early, and apparently inadvertent, reference to color stripping in the arkose member. Only very locally, mostly in the southern part of the region, does a poorly comparable thin color stripping mark outcrops of the arkose member.

BEDDING AND SECONDARY STRUCTURES

The siltstone and arkose of the siltstone member occur in thin tabular beds and in irregularly pinching and swelling beds (fig. 7), which range in thickness from a fraction of an inch to 3 feet. The arkose is somewhat thicker bedded than the siltstone, and the lower 50 feet of the uppermost arkose unit is generally the thickest bedded. The beds are internally laminated and cross laminated (fig. 7). Flaggy to slabby partings characterize all outcrops. Partings along the internal laminae tend to be as conspicuous as those along planes between beds, causing some units—especially the siltstone—to appear thinner bedded than is actually true.

Mud cracks and ripple marks are abundant locally and are more prevalent in the siltstone than in the arkose.

Stylolites everywhere characterize the siltstone member and are especially abundant in the upper arkose unit. The stylolites ordinarily have an amplitude of less than a fourth of an inch, but amplitudes exceeding an inch have been noted. In unbleached siltstone many are filled with pyrite, and some include carbonaceous material. Those in bleached rock invariably are filled with limonite.

Large channels, filled with strata similar to those they were cut into, have been noted in a few localities north of the Salt River. Most range in depth from

FIGURE 7.—Typical lamination and irregular bedding in siltstone member of Dripping Spring Quartzite. Siltstone of unit 8, measured section, page 18. Photograph by C. T. Wrucke.

20 to 30 feet and in width from 200 to 250 feet, and they are generally restricted in stratigraphic position to the lower siltstone unit and to the arkose unit that separates the principal siltstone units. Granger and Raup (1959, p. 424) noted a channel, about 50 feet deep and 700 feet wide at the outcrop, that apparently was cut through the basal strata of the siltstone member and into the arkose member.

In the lower siltstone unit small subparallel cigarlike masses, of silt- to fine-sand-size arkose, are embedded in thinly laminated finer grained siltstone (fig. 8). These features, here interpreted as scour-and-fill phenomena, have been described briefly by Granger and Raup (1959, p. 437–438 and fig. 55) as "pseudochannels." They are unusual in trend and distribution and therefore warrant particular note as possible factors in interpreting the environment of sedimentation.

The scour-and-fill channels are 1 inch to as much as 8 feet in width, but most are from 6 inches to 2½ feet. The depth of many of the smaller channels is roughly the same as the width, but the depth of the larger channels is commonly one-third to three-fourths the width, and some of the wider channel fills are even lenticular or platelike in cross section rather

FIGURE 8.—Scour-and-fill and compaction features in the siltstone member of the Dripping Spring Quartzite. Canyon of Cherry Creek in NE¼ sec. 10, T. 7 N., R. 14 E. (unsurveyed). *A*, Cross-sectional view showing compaction of shaly-parting siltstone around cores of slightly coarser siltstone. *B*, Common longitudinal view in which cigarlike outlines of the cores and compaction of host siltstone are not apparent; head and pick and end of handle are against filled channels.

A

B

than crudely elliptical. Figure 8*A* illustrates a variety of the cross-sectional forms displayed by the fills and a typical range of sizes in a given outcrop. Longitudinal sections and cross sections commonly are not well exposed in the same outcrop, so that the ratio of length to diameter of the cigar-shaped fills is not well documented. At two localities, where fills ranged in diameter from 6 inches to 2½ feet, their lengths were determined to be from 20 to 30 times their widths.

The lower surfaces of the fills are mostly rounded and troughlike in cross section; the uuper surfaces arch gently upward or are almost flat. The fills are lighter colored and slightly coarser grained than the enclosing rock. The arkose fill is dense and may not show perceptible bedding structures, but where perceptible, fine laminae are undisturbed and horizontal.

The siltstone surrounding the cores shows marked compaction (fig. 8*A*). The laminated host strata are truncated and commonly warp abruptly downward where they terminate under and against fill material. The thin laminae are sometimes so intricately folded where they impinge against the fills that details of the minute folds are difficult to trace. Laminae of the siltstone host arch gently over the cores and are not truncated against them. The effects of compaction may be seen for several inches or even a few feet above and below the fills. Secondary shaly partings, developed in weathered siltstone, emphasize these effects.

In the fraction of an inch of siltstone that lies immediately over the sandstone fills, shrinkage cracks ordinarily are abundant. These cracks are somewhat irregular in outline, but all are alined approximately normal to the axes of the fills. Possibly these cracks also are the effects of compaction.

Where the scour-and-fill features are particularly numerous, the cigarlike cores impinge one on another. An earlier core was partly scoured away before the deposition of a second, and both may have been modified by the formation of a third. Invariably, such multiple fills are separated by thin seams of siltstone. In some beds, also, separate cores may be joined along the bedding by thin layers of sandstone. The lower surfaces of such connecting layers are commonly undulating in cross section.

In longitudinal sections the scour-and-fill structures are inconspicuous and may be overlooked (fig. 8*B*). Because joints ordinarily trend parallel to the fill, longitudinal sections are more commonly exposed than cross sections. Scour-and-fill structures exposed in a face that is neither normal nor parallel to their trend go unrecognized or are mistaken for undulant bedding, such as that depicted in figure 7. Relatively little of the undulant bedding seems related in origin.

A unique feature of the cigarlike fills is their near-parallel alinement. In a given outcrop the fills commonly show a variation in trend of only 10°–15°. In the McFadden Peak quadrangle, an area of about 250 square miles, they trend N. 5°–25° E., and at the latitude of Globe they trend from N. 15° W. to N. 10° E.

The scour-and-fill structures occur through stratigraphic intervals of 20–100 feet and may characterize the whole siltstone unit. They are especially abundant in outcrops of the northern half of the McFadden Peak quadrangle; southward to the latitude of Superior they are sporadic in occurrence. Their occurrence farther south remains to be determined. These structures have not been recognized in the upper siltstone unit.

FEATURES THAT DISTINGUISH PRINCIPAL UNITS

The lowest arkose unit of the siltstone member makes up the basal 20–50 feet of most sections. This unit is very fine grained, notably micaceous, and has a greater tendency to weather to reddish hues than has the overlying arkose. Individual beds are thin to very thin, and many are separated by seams of siltstone. The basal several feet of the unit commonly includes as much siltstone as sandstone. Consequently the upper part, having fewer siltstone layers, crops out as a ragged ledge or small cliff that overhangs the slope-forming lower part.

The sandstone unit that separates the two siltstone units, about one-third of the way up in the member, ranges in thickness from 10 to 35 feet and consists mostly of very fine grained to medium-grained arkose and feldspathic sandstone. Many thin beds of similar composition are interleaved in the basal 20–40 feet of the upper siltstone unit. These interbedded strata and the underlying arkose unit form one cliff in canyon exposures. In places other than canyons the arkose unit generally forms a ledge, and it may be the only exposure in the middle part of the member. Abundant pockmarks or pits, 1–4 inches across, mark outcrops of the coarser grained beds in this unit more commonly than they mark similar beds in other units.

In many areas the two siltstone units are mantled with flaggy and slabby rubble, but in much of northern Gila County both units crop out as smooth dark slopes—slightly convex upward—that are almost free of rock debris and vegetation. Apparently, where considerable sulfate is being formed by weathering of the highly pyritic siltstone, vegetation cannot flourish. Such barren slopes occur where the strata have been metamorphosed adjacent to a thick diabase sill. In many places in northern Gila County the separate units of the siltstone member are readily distinguished from distances of 1–3 miles because of the conspicuous parallel rock scars that mark the two siltstone units. Farther south, particularly at lower elevations where all outcrops support less vegetation, the siltstone outcrops are not so conspicuous, but they still are crudely marked by a thinner cover of shrubs and trees.

Thin-bedded fine-grained arkose dominates the upper 100–130 feet, or roughly the upper one-third (unit 9, reference section) of the member. In some places the unit is seemingly no more than 60 feet thick. Freshly broken rock ranges from grayish orange pink to light brownish gray or pale yellowish brown, and the rock weathers to rusty yellowish brown. Thus, in hand specimen or from a distance the uppermost unit appears lighter in color than the arkose units between and below the siltstone units. The lower 50 feet of the uppermost arkose unit includes many medium-grained beds, contains fewer siltstone beds, and is thicker bedded and less feldspathic than the upper part. In many localities the upper 5–30 feet includes numerous siltstone layers or is mostly siltstone. The lower half of the unit forms cliffs or a steep series of ledges; the upper half forms thin ledges on a moderate to steep slope. Where exposed near the bottom of a canyon, the entire unit is a cliff former.

PRE-MESCAL UNCONFORMITY

An unconformity, only recently recognized (Granger and Raup, 1964, p. 44), exists between the Mescal and the Dripping Spring. According to Ransome (1916, p. 138), upward in the Dripping Spring "the beds become thin, flaggy, and rusty, with a tendency to grade into the Mescal." On the basis of this early description, the boundary between the formations generally has been accepted as gradational. Actually the Dripping Spring is separated—everywhere north of the Gila River, at least—from stratigraphically higher limestone or dolomite by a distinctive poorly sorted sandstone unit a few feet thick. The contact between this basal sandstone of the Mescal and the underlying beds is sharp. In many places, but only for short distances, the sandstone overlies a breccia, 6–30 inches thick, of angular chips obviously derived from the uppermost beds of the Dripping Spring. This breccia has a matrix of dolomitic sandstone. In very few places, angular fragments from the Dripping Spring also occur in the sandstone unit or in the overlying carbonate rocks.

In places the thin-bedded siltstone at the top of the uppermost arkose unit of the Dripping Spring is thin or missing. Perhaps, therefore, appreciable erosion did occur after the Dripping Spring was lithified and before the Mescal was deposited. In some critical areas where the section seems thin, however, outcrops are not adequate to furnish unequivocal evidence of significant erosion. Furthermore, some apparent thinning may be attributable to variations in sedimentation. Pending better understanding of the stratigraphy of the upper member of the Dripping Spring, the unconformity probably should not be visualized as representing a profound hiatus—but the possibility must be admitted.

MESCAL LIMESTONE

The Mescal Limestone includes internal structures and lithologic and stratigraphic elements that previously have not been recognized as prime characteristics of the formation. The Mescal was described by Ransome (1919, p. 42–43) as "composed of thin beds that have a varied range of color but are persistently cherty. The siliceous segregations as a rule form irregular layers parallel with the bedding planes, and on weathered surfaces stand out in relief and give the limestone the rough, gnarled banding that is its most characteristic feature." He further described the formation as including limestones and dolomitic limestones with or without chert. Darton (1932) indicated that the formation includes "a large amount of algal material," and Wilson (1928, p. 30) noted that "the upper portion of most sections contains a massive, algal member that is an important horizon marker." In most descriptions, however, Ransome's descriptions have been paraphrased and the formation characterized as a hard thin-bedded cherty dolomitic limestone. Such descriptions do not typify complete sections and are wholly inadequate to identify certain ordinary facies of the Mescal.

The Mescal Limestone is divisible into three readily recognized members: a thin- to thick-bedded lower member, 150–270 feet thick; a stromatolite-bearing, in part massive, middle algal member, 40–130 feet thick; and an upper argillite member as much as 100 feet thick. The lower and algal members are dominantly carbonate and are present wherever the Mescal is found, though over much of the region a part of the middle member is missing owing to pre-Troy erosion. The argillite member was recognized first as a distinctive stratigraphic element overlying the Mescal by Hinds (1935, p. 32), but was not thereafter widely noted as a part of the Apache sequence because it is absent south of the latitude of Theodore Roosevelt Dam. The argillite and the two carbonate units have been enumerated as members of the Mescal (Bromfield and Shride, 1956, p. 622), but details of stratigraphy and regional relations of the individual members have not been delineated previously.

In different areas the carbonate (lower and algal) members have three distinctly different lithologies, two of which represent changes that took place after the strata were lithified. The original carbonate beds were cherty dolomites. During deep leaching, before and during deposition of the Troy, the dolomite was silicified completely in some areas and silicified at least to a small degree elsewhere throughout the region. The last and most widespread modification, also in Precambrian time, was the metamorphism adjacent to the diabase sills that converted the cherty dolomite almost wholly into silicate-bearing calcitic limestone. These metamorphic limestones generally lack chert and are so different from the unmetamorphosed dolomites that they are described separately (p. 39–41).

There are few extensive occurrences that have not been modified either by silicification or thermal metamorphism. However, cherty dolomites do occur throughout an area of about 30 square miles north and west of Sombrero Butte along the east side of Cherry Creek (McFadden Peak quadrangle). Much smaller areas exist at the south end of the Sierra Ancha (4 miles west of Cherry Creek along the south border of the McFadden Peak quadrangle), in the central part of the Sierra Ancha immediately north of McFadden Peak, along the upper reaches of Haigler Creek 10 miles north of Young, in the vicinity of Theodore Roosevelt Dam, and in a narrow belt that extends about 12 miles southeast from U.S. Highway 60 along the south rim of the Natanes Plateau. Farther south, areas of cherty dolomite are apparently smaller and even more sparse. Those known to me are on the southeast flank of the Apache Mountains, at widely scattered localities in the Mescal Mountains, and along the east flank of the Vekol Mountains. Ransome's descriptions and map of the Dripping Spring Mountains suggest a few scattered outcrops in that range. Most other southern exposures of the Mescal are partly or entirely metamorphic limestone.

The sequence of units and the lithologies shown in the following stratigraphic section on Pendleton Mesa, here proposed as the reference section for the Mescal, are representative of the unmetamorphosed sequence. The lower member is possibly the thickest seen to date; the middle member is thicker than most occurrences, although thicker sections do exist. The section is atypical only in that it was measured in one of the three small areas where a basalt flow separates the algal and argillite members.

Reference section of unmetamorphosed Mescal Limestone

[Measured on west slope of Pendleton Mesa, 0.6 mile north of Horse Tank Creek, in NW¼ SW¼ sec. 11, T. 7 N., R. 14 E. (unsurveyed)]

	Thickness (feet)
Troy Quartzite: Conglomerate overlain by pale-red pebbly arkose	20+
Unconformity.	
Mescal Limestone:	
Argillite member:	
12. Siliceous argillite, grayish-orange-pink to moderate-yellowish-brown; poorly exposed on gentle detritus-covered slope	44
Basalt flow: Basalt, hematitic and vesicular. Fossil soil 1–3 ft thick at top	52
Unconformity.	
Mescal Limestone—Continued	
Algal member:	
11. Dolomite, pale-red or yellowish-brown to medium-light-gray, finely crystalline, thin-bedded (½–2 ft); stromatolites of unit 10 disappear in basal 10 ft of this unit. Gray chert, in thin layers and lenses, increases in abundance upward; closely crowded hematite-flecked chert nodules occur locally along bedding in irregularly lenticular zones 2–4 ft thick. Several bedding partings marked by thin layers of grayish-orange argillite. Top-most bed very cherty; contact with basalt slightly irregular	28
10. Stromatolitic dolomite, pale red to grayish red at base grading to medium light gray and yellowish brown at top, finely crystalline, beds ½–6 ft thick, mostly about 4 ft. Unidentified stromatolites, of inverted cone form, mark basal 4 ft; the stromatolite *Collenia frequens* throughout rest of unit. Very irregular pinkish-gray to medium-gray chert lenses, ¼–3 in. across, sparse throughout but most common toward top. Very light gray claystone, which weathers grayish orange and is in lenticular layers 1–15 in. thick, caps beds at 4, 13, 50, 60, and 67 ft above base. Dolomite weathers brownish to yellowish gray, with extremely rough, locally fluted surfaces; bottom 60 feet crops out as cliff. Contact with unit 9 is sharp	77
Thickness of algal member (units 10–11)	105
Lower member:	
9. Dolomite with sparse chert: Dolomite, pale yellowish brown, weathers grayish brown with rough, fluted surfaces on which silty laminae etch out. Chert, light to medium gray, dense; weathers grayish orange; in thin layers and lenses (¼–4 in. thick). Thin bedded (8–24 in.), with slabby parting; crops out as thin ledges on partly covered slope. Contrasts with lower units in that it contains little chert	27
8. Dolomite interbedded sparsely wth cherty dolomite. Dolomite, pale yellowish brown, dense to finely crystalline, weathers pale yellowish brown to brown with silty surface. Chert, mostly light gray, weathers yellowish to brownish gray with brownish-black coating locally conspicuous. Beds 1–5 ft thick, slabby to massive parting, crop out as ledges on steep slope. Chert is mainly of the secondary variety, occurs as lenses that cap individual beds, and is sparse in lower 34 ft but becomes abundant in upper part of unit. Vuggy-weathering 3-ft zones, at 38, 49, and 58 ft above base of unit, consist of irregular nodules that contain fragmented early-formed chert. In places, 3-in.–2-ft-thick lenses of gritty sandstone border these zones or cut dikelike through chert and separating dolomite beds. Lenticular layers, as much as 8 in. thick, of reddish-orange argillite separate a few beds in upper half	68
7. Interbedded dolomite and cherty dolomite: Dolomite, grayish orange, yellowish brown and pale red brown, dense to very finely crystalline; in beds 2–4½ ft thick; includes clay and fine silt; weathers yellowish orange to yellowish gray and smooth. Chert, mostly medium gray, in coalescing layers ¼–1 in. thick that make up 10–80 percent of individual beds. The layers etch out conspicuously, have black coatings, and appear finely spongy to vesicular, reflecting poorly formed cubic molds after halite. Hopper-shaped molds as much as 1 in. across, occur in chert layer 13 ft below top of unit. Unit forms dark-colored cliff	47
6. Dolomite, grayish-red, dense; contains much clay and fine silt; weathers reddish to yellowish brown with smooth surface and inconspicuous splitting planes at 2–8-ft intervals; breaks conchoidally; bottom 6 ft forms ledge, top 13 ft smooth cliff with flat slope between. Top 6 in. includes small poorly rounded to angular pebbles of light-gray chert; interval 4–5½ ft above base includes similar granules and small pebbles of chert and sparse coarse grains of vitreous quartz	22
5. Dolomite, grayish-red, slightly silty; weathers to very rough surface mottled grayish orange to grayish pink. Single bed	3
4. Calcareous dolomite alternating with cherty dolomite: Dolomite, grayish red, very finely crystalline, some beds silty and cross laminated; weathers pale brown to pale red with silty surface; thin to thick bedded (1–5½ ft); five beds include sheaf-like impressions. Chert, brownish gray to white, in thin irregular layers that weather brownish black and etch out conspicuously; some layers include abundant small halite molds. Massive beds	

	Thickness (feet)
Mescal Limestone—Continued	
Lower member—Continued	
of top half grade from chert-free dolomite at bottom to very cherty dolomite at top of bed, and crop out as ledges on otherwise steep slope	29
3. Cherty dolomite: Dolomite, pale red to grayish red, thick bedded, massive parting, weathers brown to grayish brown; chert, brownish gray to black in very irregular vesicular layers, 1–4 in. thick, makes up 25–50 percent of each bed; chert juts out in very rough ledges that form steep slopes. Beds locally show slump and brecciation features, and many chert layers are seen as trains of angular fragments. Chert layers of basal 4 ft include abundant hopper-shaped molds after halite	44½
2. Dolomite breccia, of fragments of cherty dolomite like that of unit 3, in matrix of grayish-orange, silty and locally calcareous dolomite. Weathers pale to yellowish brown and to silty or gritty surface having rough blocky black-mottled projections; exposed on partly covered slope. Fragments range from minute chips to blocks 1×3×3 ft; large sizes most abundant in upper part. Bottom 3 ft includes abundant coarse sand grains as in unit 1; these diminish upward	23
1. Arkose, grayish-orange-pink to pale-brown, fine-grained; has dolomitic, clayey matrix; breaks with quartzitic fracture; includes coarse to very coarse grains of clear vitreous quartz. Single massive bed forms prominent ledge abundantly veined by limonite. Contact with Dripping Spring Quartzite planar	5½
Thickness of lower member (units 1–9)	269
Thickness of Mescal Limestone (units 1–12)	418
Unconformity.	
Dripping Spring Quartzite:	
Siltstone member:	
Arkose and siltstone, pale-red to pale-yellowish-brown, thinly cross laminated to thin-bedded (as much as 30 in.), very firmly cemented, slabby-parting (½–15 in.); stylolites abundant. Top 10 ft mostly siltstone	32+

LOWER MEMBER

LITHOLOGY AND STRATIGRAPHY

The lower member is mostly thin- to thick-bedded dolomite, in which chert is sparse to abundant. The dolomite is dense—almost lithographic—to finely crystalline; most beds include very little clastic material. The small amounts of fine sand, silt, and clay that are included cause minor bedding structures to etch out on weathered surfaces. The dolomite ranges from yellowish brown through pale red to grayish red. Brown hues, which are characteristic of sections that have had a minimum of postlithification solution and silicification, dominate in most sections. Red hues characterize strata deeply leached after lithification. The chert ranges from white through brownish black to black. The darker chert occurs mostly as irregular discontinuous laminae and thin layers that parallel bedding. These layers etch out conspicuously upon weathering and thereby produce the very rough gnarly banding supposedly typical of Mescal outcrops but actually absent from much of the formation. In most sections the dark cherts characterize only the lower one-half to two-thirds of the member. Upward in the lower member the chert is mostly light gray to medium and brownish gray and more massive. It commonly occurs in fairly thick irregular layers and as nodules or lenticular aggregates of nodules that only grossly follow bedding. The dark cherts formed nearly contemporaneously with the dolomite, whereas much of the light-colored chert formed later, contemporaneously with the development of karst topography in the Mescal. Although considerable lithologic modification accompanied the formation of the later chert, certain sequences of beds and bedding features are distinctive of individual units throughout the region; consequently the stratigraphic position of a given outcrop can be determined within narrow limits.

The basal sandstone of the Mescal (unit 1, reference section) probably averages 5–6 feet in thickness; in a few places it is only a few inches thick or is missing; in other places it is as much as 15 feet thick. Generally it crops out as a conspicuous massive ledge. Where the sandstone is especially thin, however, as along the canyon of the Salt River east of Canyón Creek, it may not be recognized readily because of a cover of debris. In places the unit is thinly crossbedded, in other places slump structures are common, and in still other areas the unit is massive and in effect structureless. Commonly the unit is a moderately well cemented arkose, made up at least in part of detritus from the Dripping Spring, but in places it is a quartzite virtually free of feldspar. The rock is slightly dolomitic or, if metamorphosed adjacent to diabase, calcitic. Reddish-brown limonite mottles and veins weathered surfaces. Abundant coarse to very coarse well-rounded grains of clear quartz and smoky opalescent chert, set in a matrix that is generally medium grained, characterize the sandstone. These distinctive grains, the carbonate cement, and poor sorting differentiate this sandstone from any in the Dripping Spring.

Immediately above and in sharp contact with the basal sandstone of the Mescal is a breccia of dolomite, which is generally 15–40 feet thick (unit 2, reference section). In and near the Sierra Ancha the breccia in a few places is as much as 100 feet thick. In the canyon of the Salt River, between the mouth of Canyon Creek and U.S. Highway 60, it is thin or missing, and in at least one locality in the Vekol Mountains there is no breccia. Everywhere else the breccia has been recognized. The overlying unit of cherty dolomite (unit 3, reference section) is relatively thin where the breccia is thick and vice versa. In the McFadden Peak quadrangle these two units make up roughly the lower fourth of the member.

The breccia is composed of fragments of cherty dolomite, ranging from small chips to blocks 5 × 10 × 20 feet, in a structureless matrix of pale-red or grayish-red to grayish-orange silty dolomite. The blocks may be sparsely distributed in the matrix or may be closely spaced in comparatively little matrix. Small fragments are most abundant in the basal part, where the matrix makes up a large proportion of the rock. Large blocks may be found anywhere in the breccia, but are most numerous near the top. No voids exist in the breccia. The matrix of the basal few feet includes abundant coarse grains of quartz and chert like those that characterize the basal sandstone, and in many places such grains are sparsely distributed upward through the matrix.

The breccia weathers to a smooth gentle slope, which collects much debris. Because exposures are poor and because weathering tends to obscure the outlines of the fragments and to emphasize the silty and sandy aspect of the matrix, in most exposures the breccia simulates a dolomitic mudstone. For this reason, the fine-grained upper strata of the Dripping Spring were formerly interpreted as grading into the Mescal.

The breccia is gradational upward into flat-bedded massive thin- to thick-bedded (1–5 ft) dolomite that is especially cherty (unit 3, reference section). Very irregular layers of brownish-gray to black chert, $\frac{1}{20}$–4 inches thick, make up 20–60 percent of each bed. In most places this unit crops out as a steep slope; in steep-walled canyons it forms cliffs.

The next overlying unit is 20–30 feet thick and consists of dolomite beds 1–6 feet thick that alternate with beds of cherty dolomite of comparable thicknesses (unit 4, reference section). Chert is concentrated only in the upper third or half of some beds; otherwise the cherty beds are similar to those of the underlying unit. Parting planes of a few of the chert-free beds are characterized by peculiar sheaflike markings. Locally, asymmetric ripple marks are abundant; ripple marks have been noted elsewhere in the lower member, but this is the only unit in which they are common. In all unmetamorphosed sections viewed, this unit is a striking marker because of the alternation of cherty and chert-free layers. However, this alternation is not everywhere an aid to recognition of stratigraphic units: in some areas of metamorphic limestone, the distribution of silicate minerals suggests that all beds of the interval were locally cherty. This unit occurs within the interval 50–85 feet above the base of the Mescal.

The fifth distinctive stratigraphic interval (units 6 and 7, reference section) crops out mostly as a cliff—the most prominent outcrop of the lower member—and makes up most of the middle third of the member. The basal 15–25 feet (unit 6, reference section) is dense virtually structureless clayey and silty dolomite, which may form ledges but as a rule is covered by debris from the cliff. This dolomite is free of chert except for detrital granules and small pebbles. The cliff-forming part (unit 7, reference section) is mostly thin bedded but does include beds as much as 5 feet thick. The beds are alternately very cherty dolomite (as much as 80 percent chert) and moderately cherty dolomite (10–25 percent chert). Where weathered, the dark chert etches out as wispy or coalescing pitted or spongy bands; most are less than an inch thick, and many are less than a quarter of an inch. Black coatings characterize the weathered surface of the chert and cause the cliff exposures to be darker than those of adjacent strata. This unit is generally the highest in which the early formed chert is prevalent.

Above the cliff-forming strata is a unit composed of 1–5-foot beds of dolomite, erratically variable in chert content (unit 8, reference section). This unit, 50–70 feet thick, forms prominent ledges on a slope. It differs from lower units in that the dolomite tends to be in light shades of brown rather than in dark shades of brown or red, and the chert is mostly light gray. Generally chert is sparse in the lower third to half of the unit. In some areas only a few of the higher beds are cherty; in others all the beds include chert and in places two or three thin layers of chert a few feet apart coalesce laterally to one zone as much as 10 feet thick. Chert occurs in some beds as thin wavy parallel layers, but it more commonly occurs as lenses and layers, 2–10 inches thick, that cap individual beds and project irregularly 1–8 inches down into the dolomite. Nodules, a fraction of an inch to 3 feet in largest dimension, are abundant within the beds in some of the most cherty sections. Lenses and dikelike masses of sandstone, which fill solution openings, are more common in this unit than in any other.

The uppermost unit, unit 9 of the reference section, is 20–30 feet thick and is similar to the one just described except that it is wholly thin bedded, parts conspicuously at intervals of a few inches, and forms gentler slopes. Layers and lenses of light-colored chert range from paper thin to 6 inches thick, but generally this unit contrasts with all lower units in containing very little chert.

Although the upper two units vary laterally in thickness and lithology, because of solution and late silicification, the sequence of beds within them is remarkably uniform from the Vekol Mountains to the Mogollon Rim, a distance of 135 miles.

The contact between the algal and lower members is sharp and readily recognized; the flat-bedded slabby dolomite below the contact contrasts strongly with the massive outcrops above in which algal structures are conspicuous.

STRUCTURAL AND TEXTURAL FEATURES

Abundant halite molds and small-scale sedimentary structures indicate that much of the dolomite was deposited in very shallow highly saline waters and that the dark cherts were formed early. Interrelations of such features show that the basal dolomite breccia and minor related breccias are not original sedimentation phenomena but are founder breccias formed before the upper part of the member was deposited. The origin of certain peculiar structures is not understood, but their description here may aid in future interpretations of the Mescal.

HALITE MOLDS IN CHERT

Many layers of the dark-gray to black chert include innumerable molds interpreted as representing halite crystals. In most cherts the molds are equidimensional pits, 1–5 millimeters across, filled with brown dolomite. These are sparse in some layers and constitute more than half the volume of others. On weathered outcrops, in which the fillings have been leached out, the highly pitted chert bands resemble vesicular basalt. Although most pits are rounded or have cubic outlines that are distorted and so are not diagnostic of halite, a few cubic forms can be found in most layers, and in some they are abundant. More diagnostic, however, are larger sharply outlined molds found in very few layers. These molds, which range from 5 to 30 mm in diameter, have the hopper-shaped form of skeletal halite crystals common in salt deposits (Shrock, 1948, p. 146–148; Dellwig, 1955, p. 89–95; Brooks, 1955).

Halite molds are particularly abundant in the cherts of the blocks and slabs in the founder breccia at the base of the member, and they characterize most chert layers in the cherty dolomite (unit 3, reference section) that overlies the breccia. Fewer of the chert layers in the unit of alternating beds of dolomite and cherty dolomite include the molds, and only sparse layers in the cliff-forming unit show them. They have not been preserved, if they existed, in the dolomite between the chert layers.

The association of halite and chert is an anomaly, and the molds in chert are almost unique. Hopper-shaped halite crystals apparently form in brine at the surface of hypersaline bodies of water and settle to the bottom when the surface of the water is disturbed. After settling, the fragile incipient cubes are subject to destruction by solution or modification by overgrowth (Dellwig, 1955, p. 88–92). Most of the large sharply defined hopper imprints are on the upper surface of chert layers, and apexes of the hollow pyramidal forms point down in the position of initial formation. Thus, these particular salt crystals were little disturbed by wave or current action, and little dissolved before leaving their imprints. Probably the halite crystals were impressed directly into silica gels, which accumulated on bottom muds before these sediments were diagenetically much modified.

During deposition of the lower 100 feet of the Mescal, circumstances favorable for the formation and preservation of halite crystals alternated innumerable times with those favorable for the deposition of silica-free carbonate muds. Possibly the latter were initially dolomite, because delicate laminae and crosslaminae outlined by clastic grains of carbonate are preserved. Anhedral molds of halite are ubiquitous in individual beds; it therefore seems unlikely that salt crystallized only during the desiccation of scattered pools, as has been reported where halite imprints are preserved as casts rather than molds (Brooks, 1955). Rather, the lower part of the Mescal probably accumulated in a broad uniformly shallow quiet body of water, such as a lagoon or other almost landlocked setting.

SHEAFLIKE IMPRESSIONS

Peculiar and distinctive markings characterize several chert-free beds in the unit of alternating dolomite and cherty dolomite (unit 4, reference section). On bedding planes these markings look like clusters of reeds or blades, which lie parallel to the bedding and radiate out from a center. They do not spread through 360° of arc, however, but subtend two opposite 20°–90° arcs, like a sheaf of straw gathered tightly at the middle. The long dimension of the markings mostly is from 2 to 8 inches but in places is as much as 2 feet. In a given bed and area the individual sheaves are fairly uniform in size, but in a different bed in the same area the markings may be uniformly of a different size. Everywhere they are so abundant that the sheaves overlap each other. In cross sections the features commonly are defined by discontinuous and slightly irregular films of silt arranged subparallel to the bedding. In a few outcrops, bedding edges show an anastomosing pattern of closely spaced silt films, parts of which diverge as much as 30° from the general plane of bedding.

These markings occur throughout beds or sets of beds, ranging in thickness from 1 to 6 feet, that are separated by beds comparable in thickness but free of such features. The markings are not obvious in the unmetamorphosed dolomite, but they weather out conspicuously in beds that have been converted to calcitic limestones. The metamorphic limestones tend to part readily, or even to be shaly, and if pulled apart they show these imprints on every plane of easy separation. The sheaflike markings commonly are confined to the lower 10–15 feet of the unit but have been seen throughout. Wherever this stratigraphic interval has been examined carefully north of Globe, the markings have been found. Farther south no search has been made for them.

These markings have not been explained. At first they were noted only on weathered outcrops of silicated limestone beds, and it was thought that they might be impressions of some silicate mineral. Recognition of the impressions on fresh parting planes, as well as in unaltered dolomite, made this idea untenable. The apparent occurrence of the sheaflike markings in certain beds throughout an area of at least 3,500 square miles suggests an environmental control. Some consideration has been given to the possibility that they represent imprints of ice crystals, which might have formed on exposed mudflats, or imprints of evaporite crystals now leached away, but no crystal forms like these markings have come to my attention. Ripple marks in adjacent and intervening beds indicate an environment in which agitation could cause mud-chip breccia (edgewise conglomerate), but the thinness of the layers and invariable sheaflike aggregation is unlike most such conglomerate. The uniformity in size of the sheaves in a given bed and their different but uniform size in another bed might be explainable if they are impressions of organic remains, probably plants. Such organisms might have been sensitive to environmental factors, such as water salinity or temperature, so that at a given place or time they developed only to certain sizes. The stromatolites of the algal member, undoubtedly of organic origin, exhibit analogous size variations that must reflect environment. Paleontologists apparently have not described organisms that would leave the sheaflike imprints.

MINOR SLUMP FEATURES

The regularly bedded dolomite (units 3 and 4, reference section), between the basal founder breccia and the cliff-forming unit, locally exhibits small slump features or slight brecciation. Bedding laminae may be slightly distorted or actually disrupted, and bedding details are blurred as though slumping occurred before the carbonate mud was completely lithified. The commonest manifestation of such disruption is seen in the dark chert layers, which appear to be pulled apart in the form of angular blocks; the narrow spaces between the blocks are filled with dolomite. In cross section the band of chert appears as a train of small blocks that are only slightly reoriented. Commonly these small-scale slump and breccia features are confined to one bed and are not found in the beds immediately above and below, though these adjacent beds may include similar features in an outcrop not far away. Where slump structures are widely spaced, they appear to be phenomena formed independently of similar structures in adjacent beds.

INTRAFORMATIONAL CONGLOMERATES

In a few areas the top of each disrupted bed is marked locally by a lense of conglomerate, generally only a few inches thick, that consists of subangular granules or small pebbles of chert. Many chert fragments show no abrasion and obviously were derived from the underlying bed. Sparse pebbles of dolomite are mostly well rounded and disk shaped. In a very few places slabs of cherty dolomite as much as 1 foot thick and 5 feet across make up a part of the intraformational debris. In such places the conglomerate may be as much as 6 feet thick, and many of the chert fragments are fairly well rounded. Delicate details of bedding are preserved in dolomite pebbles and in the reworked dolomite sand that makes up the matrix of the conglomerate. Thus, again, there is evidence that the beds initially, or almost initially, were composed of dolomite and that the dark cherts were formed early.

Lenses of conglomerate have been noted only in a few areas, all in northern Gila County; where they are especially numerous, however, local slump structures are also abundant through a stratigraphic interval of 30–50 feet, and therefore these features may be genetically related. Individual beds, immediately after or perhaps during lithification, apparently slumped to produce irregularities in the sea floor. Periodically, while certain areas were above wave or current base, parts of the last deposited bed or beds were truncated, and the eroded materials were redeposited in adjacent low areas as lenses of chert gravels with a clastic dolomite matrix. Undisturbed layers of conglomerate separate deformed layers of carbonate; again the indication is that slumping of the dolomites occurred bed by bed as the strata accumulated.

FOUNDER BRECCIA

The coarse breccia that makes up the basal part of the carbonate section is a stratigraphic entity throughout an area of at least 4,000 square miles. It represents the disruption of strata on a much larger scale than is apparent in the above described features. It is not a tectonic breccia, because there are no traces of related tectonic structures higher or lower in the Apache sequence.

The configuration of the breccia mass and the distribution of the coarse fragments within it indicate that the breccia is not of sedimentary origin. The base of the unit is planar; the top is irregular. Along the upper margins slabs of cherty dolomite, surrounded by typical matrix material, have moved only a few inches from their original position. Many slabs in the upper 15 feet of the unit have the orientations and positions relative to each other that might be expected if they had spalled from an overhanging cliff but had been arrested after falling only a few feet. Still lower in the breccia the slabs cannot be matched directly with the sites from which they spalled. Where the unit is thick, however, many of these blocks can be correlated—by pecularities in individual chert layers or by a sequence of chert layers or bedding features—with flat-lying beds that overlie a thin section of breccia in adjacent areas. The coarse breccia might reasonably be interpreted as sedimentary only if all these features were inverted in their order of stratigraphic succession.

All features of the breccia suggest the collapse of consolidated and almost consolidated cherty dolomite beds into an underlying mud—a mud so incompetent that it flowed readily to fill all voids. Such a soft base, mostly dolomite, that would persist while higher dolomite strata were lithified is difficult to conceive. The occurrence of halite suggests, however, an alternative process that could have caused foundering. The founder breccia, in typical 15–40-foot thicknesses, is largely confined to the interval in which halite was especially abundant. In this interval, also, slump features are conspicuous in the larger fragmental blocks of cherty dolomite; in adjacent areas where the breccia is thin the equivalent cherty dolomite is slightly folded. Intercalated with these dolomite beds are silty and clayey dolomitic layers that pinch and swell and are somewhat contorted—giving the appearance of having flowed. Thus these silty but otherwise aphanitic layers suggest dolomitic residuum left after the removal of gypsum from carbonate sections; no other indications of gypsum or anhydrite have been seen in the Mescal. It is here postulated that the basal few tens of feet of the section once included a much greater amount—and perhaps variety—of evaporite salts than is now apparent and that these salts were leached away, the residue being a carbonate mush into which higher strata collapsed. Locally, at least 100 feet of section foundered.

The founder breccia formed while the lower member was being deposited. It is not related in origin to the karst breccias that formed after the entire dolomite section was lithified, uplifted, and exposed to a still later leaching process. In a few places, where the founder breccia is about 100 feet thick, the broken strata are truncated and overlain locally by intraformational conglomerates. Minor slump features, noted in beds that overlie the founder breccia, occur in beds that apparently were in an environment of alternately high and low salinity, which would allow deposition and then leaching of evaporite salts. The minor slump structures and breccias probably were formed—as already noted—independently of those formed in earlier beds and distinctly later than the main founder breccia. Where the founder breccia is thick and minor slump features are numerous in higher beds, however, it cannot be demonstrated with confidence that the structures of the higher beds were formed wholly independently of the founder breccia. In any case, the relations of local unconformities to deformed beds indicate that all slumping and brecciation occurred before deposition of the cherty dolomite that makes up the cliff-forming unit (unit 7, reference section).

The mechanism of the leaching is an enigma. The distribution of halite-bearing chert layers and the decrease in halite molds in strata above the founder breccia suggest that lagoonal waters were alternately very saline and then only moderately saline; but overall, as higher carbonate strata were deposited, the salinity gradually decreased. Ultimately, late sea waters of lower salinity could have caused the solution of salts in the basal strata. But how did such waters gain access to these salts? Perhaps the hydration of anhydrite to gypsum caused initial deformation that allowed less saline waters to percolate downward. Once started, leaching of the salts could cause brecciation that would continue until all the salts were removed.

ALGAL MEMBER

The algal member, generally 50–100 feet thick, consists of two units: a lower thick-bedded cliff-forming unit characterized by stromatolites and an upper thin-bedded slope-forming unit. The algal structures grade out in the basal 10 feet of the upper unit. The stromatolites are not in mounded or reeflike masses (bioherms) but instead are in a regularly bedded blanket-like deposit (biostrome) (Cumings, 1932; Link, 1950). In most areas north of Globe the biostrome is 50–60 feet thick and makes up the lower one-half to two-thirds of the member. Farther south similar thicknesses exist, but in most places the biostrome is only 30–35 feet thick. Throughout the region, because of erosion that preceded deposition of the Troy, the upper unit is locally only a few feet thick or is missing. The contact with overlying rock units is everywhere sharp.

In recent years the biostrome of the Mescal has been widely recognized as a distinctive unit by those who prospect for and mine asbestos, and it is popularly termed the algal limestone. Generally, the upper unit has not been recognized because (1) in isolated outcrop it is quite similar to the upper parts of the lower member, (2) outcrops are commonly thin or not conspicuous, or (3) it is separated from the biostrome unit by a diabase sill. Because the biostrome dominates the middle member, the two units are here designated the algal member.

The member is composed of grayish-red to yellowish-brown dolomite. The brown dolomite weathers brownish to yellowish gray; the color of the red dolomite is little modified by weathering. From the Salt River south as far as Globe, most beds within the massive lower unit are more than 6 feet thick. In other areas, bedding planes are 6 inches–6 feet apart and typically are about 4 feet apart. The thinnest beds are near the top of the biostrome. The upper unit is thin bedded throughout.

Light-gray to gray chert, in many places tinted red by minute flecks of hematite, is generally sparse throughout most of the member. In each unit the chert content increases upward. In the biostrome, irregular lenses of chert, a fraction of an inch to a few inches thick, are typical. Toward the top of the upper unit, closely spaced chert nodules locally coalesce to form zones 1–4 feet thick, and the top few inches or feet is almost completely silicified.

A tabular chert zone, 8–30 inches thick, is a distinctive marker high in the upper unit because it displays conspicuous forms similar to cone-in-cone structures (Pettijohn, 1949, p. 155–156). Throughout several areas, each at least 20 square miles, this unusual zone is the effect of secondary silicification of one bed or a group of beds. It appears likely that one stratigraphic interval was especially susceptible to the formation of cone-in-cone structures throughout the region. In most of the area between the Salt River and the abandoned mining village of Chrysotile, this zone was a base that resisted erosion which preceded later sedimentation in Precambrian time. Elsewhere this chert zone is as much as 15 feet below the top of the member.

The algaloid structures are distinguished in cross section by laminae that have a wavelike pattern (fig. 9*B*) and are marked by chert or silt that etches out conspicuously when weathered. The structures appear in plan as flattened hemispheres (fig. 9*A*) or inverted saucerlike disks outlined by concentric rings. Exfoliated outcrops, such as the outcrop shown in figure 9*A*, are common only where the individual colonies are large. Outcrops showing concentric rings are more typical of smaller colonies. The structures are 2–30 inches in diameter, and the amplitude of the waves ranges from ¼ inch to 5 inches. Most structures are 5–10 inches across and 1–1½ inches in amplitude. In one bed in a given outcrop, the size is fairly uniform.

Such fossil forms preferably are termed stromatolites rather than algae. According to Rezak (1957, p. 129), "Fossil algae preserve recognizable organic microstructures that enable the examiner to determine their true biologic relationships." But all that generally remains of these algae in stromatolites are large headlike masses or the gross outlines of the original colonies. Johnson (1946, p. 1089, 1095), among others, has stated that each form-species, typified by a certain habit of growth and shape of colony, may have resulted from the life processes of several species or even several genera of algae that lived in constant association. Because binomial names applied to stromatolites have little or no biological significance, there is considerable difference of opinion as to the desirability of classifying stromatolites (Cloud, 1942, p. 364–366). Rezak (1957) has suggested the use of names of form-genera and form-species in a classification that is admittedly artificial but convenient in cataloging specimens with wide geographic and geologic ranges.

Rezak (oral commun., Apr. 1956) has identified the principal stromatolites in the Mescal Limestone as *Collenia frequens* Walcott. This form is also common in the younger Precambrian Belt Series of northwestern Montana (Rezak, 1957, p. 133, 136–140).

The basal bed, 4–6 feet thick, of the algal member contains a different stromatolite. In cross section the individual colonies are crudely conical. The apexes of the cones point downward, and the axes are inclined at 40°–80° to bedding surfaces. Laminae within each

A

B

FIGURE 9.—Structures typical of the stromatolite beds of the algal member of the Mescal Limestone. Stromatolites identified as *Collenia frequens* Walcott. From cliff above Regal mine, 3½ miles east of the junction of Canyon Creek and the Salt River. *A*, View of upper surface. *B*, Cross-sectional view.

cone are convex upward. The base of the cones is 2½–4 inches across. This stromatolite has not been identified.

The lack of variation in the biostrome of the Mescal warrants particular note. Throughout the region the laminated algaloid structures occur with lateral continuity comparable to that illustrated in figure 9, and they are interrupted vertically only by bedding planes that represent hiatuses in sedimentation. Sequences of stromatolite-bearing beds in the various exposures are remarkably similar—even at lateral intervals of as much as 10 miles. There are few exceptions to the 50–60-foot thickness typical through about 3,000 square miles in the northern part of the region. Similar stromatolites of other regions, studied in some detail as potential aids for stratigraphic correlation (Rezak, 1957), seemingly have not been as consistent. Elsewhere single stromatolite-bearing beds are generally discontinuous. Individual biostromes also differ considerably in thickness, in lateral makeup of bedding units, and in areal extent. Apparently biostromes of the makeup noted in a given exposure of the Mescal do not generally extend more than a few miles. Thus the continuity and uniformity that is a prime characteristic of the Mescal occurrence, bed by bed and overall, is very unusual.

ARGILLITE MEMBER

DISTRIBUTION AND STRATIGRAPHIC RELATIONS

Through much of northern Gila County an upper member, mostly argillite, unconformably overlies the algal member. The member generally is 50–80 feet and locally as much as 100 feet thick. It is separated from the overlying Troy Quartzite by flows of basalt. South of the latitude of Theodore Roosevelt Dam, this member apparently was eroded away before the overlying flows were laid down or before the Troy was deposited. (See fig. 4 for area of outcrop.) North of the 34th parallel, the Troy Quartzite generally rests on the algal member or on the lower member of the Mescal. In three small areas remnants of basalt flows, like those that overlie the Mescal, intervene between the algal and argillite members. South of the Salt River most sections are disrupted by diabase intrusions, and full sequences are difficult to piece together from widely separated outcrops.

The unconformity at the base of the argillite member represents more than one episode of erosion and probably represents an appreciable hiatus. Through most areas the argillite member seemingly is concordant with underlying strata; the only indication of unconformity is the silicified rock at the top of the algal member. Such concentrations of chert below contacts have been recognized as evidence of erosional unconformity (Leith, 1925). The chert zone continues beneath the basalts that locally separate the members. Angular truncation of the algal member occurs in a few places. Pebbles and angular fragments from the topmost silicified beds of the algal member are incorporated sparsely in the basal strata of the argillite member or in the bottom part of intervening basalt flows. Where flows occur, fossil soils cap the basalt and are included locally in the basal bed of the argillite member. The configuration and significance of the erosion surface at the top of the algal member is considered in more detail in the next section.

LITHOLOGY

The basal unit of the argillite member is made up of chert and generally is a few inches to 10 feet thick. Most commonly this unit is a thinly bedded breccia of small angular gray chert fragments in a matrix of pink or brown chert; in places it is a conglomerate of well-rounded chert pebbles in a matrix of silty chert, and in other places it is made up of thin beds of laminated chert. In a few small areas the conglomerate facies is as much as 40 feet thick. The breccia and conglomerates apparently are mostly intraformational—derived from the bedded chert, rather than from the underlying member. The chert beds form a prominent marker ledge. The unit is absent in few places north of the Salt River, but elsewhere is commonly absent.

The argillites that make up most of the member are yellowish brown, minutely laminated (but flaggy to massive splitting), siliceous, and very hard and dense. Subordinate medium-gray to black argillite beds are somewhat less siliceous and commonly have shaly partings. The original sediments apparently were entirely of clay-size material; however, because the upper and lower boundaries of the member were particularly favorable horizons for diabase intrusions, virtually everywhere the rocks are somewhat recrystallized, and their original mineralogy and texture is obscured. The argillite is abundantly spotted with aggregates, 0.2–3.0 mm in diameter, of finely divided mica or amphibole; smaller aggregates, even more abundant, are too fine grained for microscopic identification. Immediately adjacent to diabase intrusions certain argillite layers may be coarsely mottled with such aggregates.

East of Canyon Creek and south of the Salt River, the lowest 20–30 feet of the member tends to be shaly or crumbly greenish-gray to dark-gray mudstone, which is mostly concealed by debris from the overlying argillite. Calcareous or siliceous concretions, as much as 3 inches in diameter, are locally abundant in the mudstone.

In a few places in the Sierra Ancha, one or two units of thin-bedded lenticular limestone occur in the section. Included aggregates of silicate minerals indicate that this limestone was silty and moderately cherty dolomite before metamorphism. Isolated outcrops are readily mistaken for the metamorphic limestone that occurs below the member. A well-exposed lens, 2–6 feet thick, occurs 20 feet above the base of the member in the vicinity of Asbestos Point at the south end of the range. (See McFadden Peak quadrangle map for locations.) Occurrences near Workman Creek Falls 2 miles northwest of Aztec Peak and along the west wall of Cherry Creek canyon 3 miles southeast of McFadden Peak are unusual in that thin limestone beds dominate the upper 15–20 feet of the 60-foot sections. None of the limestone occurrences appears to be more than one-half square mile in area.

Radioactivity of the argillite member, determined in and near the Sierra Ancha by an airborne scintillometer, is higher than that of any other unit in the Apache Group except the upper member of the Dripping Spring Quartzite, which has about the same radioactivity (Magleby and Mead, 1955, p. 9). This characteristic led to considerable prospecting during 1954–55; however, only a few insignificant showings of uranium-bearing rock were discovered. One typical specimen of argillite, analyzed by semiquantitative spectrographic methods, contained more than 10 percent potassium (Neuerburg and Granger, 1960, p. 766).

If the one sample is representative of the member, the unusually high potassium content may account in large part for the radioactivity.

PRECAMBRIAN WEATHERING AND SILICIFICATION

Weathering, which occurred in several episodes after the algal member was lithified and before the Troy was deposited, caused extensive modifications of the dolomites of the Mescal. In at least two large areas(outlined approx. in fig. 4), and in several small areas, the algal and lower members of the Mescal were almost entirely converted to chert. In several areas of a square mile or less, all the algal member and the upper part of the lower member were similarly silicified. Silicification on such a scale apparently occurred through much less than 10 percent of the region. Individual beds of the two members were at least partially silicified, however, throughout the region. In certain areas, where solution was especially pervasive, a karst topography was formed on the Mescal, and in parts of this terrane the entire formation was converted to a massive collapse breccia. In certain of the karst areas, lateritic iron deposits accumulated.

The most extensive and complete areas of solution and silicification are those in which the argillite member and the basalt flows were removed by pre-Troy erosion. With few exceptions they are areas where the Chediski Member of the Troy, rather than the arkose member, rests directly on the algal member or lower member of the Mescal. Carbonate rocks displaying such relations to the Troy are not, however, everywhere strongly modified. The effects of solution and silicification became more pervasive with each successive episode of erosion before the Troy was deposited. Therefore, the degree of modification depends on whether or not the Mescal of a particular area was subjected to several periods of weathering.

SOLUTION FEATURES

The only solution-related features that are both wide-spread and conspicuous are massive bedlike layers of sandstone or quartzite, which occur locally in the carbonate section. These sandstone bodies are common in some areas but absent from most. They occur mainly in the upper third of the lower member of the Mescal but are not everywhere at the same stratigraphic horizons. The bodies are entirely devoid of bedding features, and all are apparently lenticular. Because they occur mostly as isolated bodies in the limestone terranes, where even their gross aspects are obscured by metamorphism and faulting, their origin was not understood at first. Recently the concordant layers and associated dikes have been observed in well-exposed outcrops of dolomite, where they have been recognized as solution cavities filled with sandstone. The evidence for fillings is especially clear in the central part of the McFadden Peak quadrangle. There, sinkholes and solution enlargements of joints are abundant in the algal member, and some sandstone-filled cavities have been traced downward almost to the bottom of the lower member. Even more numerous are the solution openings along bedding planes; the sandstone layers that fill these spaces pinch and swell but commonly are continuous for distances of several hundred yards from individual sinkholes. Some of these sandstone layers are as much as 15 feet thick for distances of at least 500 feet.

Where sinkholes are numerous and tend to coalesce, as near the headwaters of Canyon Creek, strata between sinkholes exhibit a variety of solution effects. Some beds are remnant only as layers of clayey and silty residues. Beds that were least affected by solution folded irregularly or fragmented into breccias to fill voids. In some places these beds drape and thin over great void masses of breccia or sandstone. Sandstone bodies in these settings tend to be very irregular in outline and are not of great lateral extent.

Solution of the dolomite began during the forming of the erosion surface that separates the algal and argillite members of the Mescal and continued while the argillite was being deposited. Sparse detrital chert fragments in the basal beds of the argillite member were derived from underlying beds that had already been somewhat thinned and silicified. In places the bedding of the basal 10–40 feet of the argillite member is undulatory and in cross section crudely approximates the configuration of underlying dolomite beds that subsided locally as a consequence of partial leaching. These undulations and angular relations between argillite beds indicate that subsidence occurred while some argillaceous sediments were still plastic and before later strata were deposited. In many sections several bedding planes well down in the carbonate section are conspicuously marked by thin seams of reddish-orange argillite. (See description of units 8 and 11, reference section.) These seams represent voids filled while the argillite member was being deposited. In a few sinkholes the matrix of the collapse breccia is similar argillite. The thinning of dolomite sections represented in all these features is minor compared to the thinning effected by still later leaching.

Collapse and probably additional solution continued during extrusion of the basalt flows that overlie the argillite member; such processes certainly continued during deposition of the two lower members of the Troy Quartzite. Angular blocks of argillite partly fill some sinkholes and in a few places are in a matrix of basalt that flowed into the cavities. A very few apophyses of basalt flowed downward and partly fill tabular voids in the carbonate rocks. A few cavities are filled with sandstone of the arkose (lower) member of the Troy. The prevalent filling is sandstone identical with that of the Chediski Member of the Troy. In some examples—those best exposed are 1–2 miles southwest of Gunsight Butte in the north-central part of the McFadden Peak quadrangle—the roofs of caverns collapsed, allowing plug-shaped masses of the two lower members of the Troy to drop and fill large sinks. Within the plugs of the Chediski Sandstone Member, there are almost no vestiges of bedding, and brecciation is not apparent. Therefore, the collapse must have occurred while these sediments were unconsolidated. No fillings equivalent to the quartzite (upper) member of the Troy have been recognized, and strata of this member apparently were not disrupted by collapse. Thus solution and silicification may have ceased before deposition of the quartzite member. Quartzitic lenses do exist, but these represent fillings of the earlier sandstone that were indurated adjacent to diabase intrusions.

LATE-FORMED CHERTS

Cherts that formed after the dolomites were lithified and first eroded differ in texture, form, and distribution from the dense dark chert that formed penecontemporaneously with the dolomites. These late cherts are mostly light in color, are flecked with hematite, and generally occur in thicker and more irregular masses than the older chert. In sections that are still mostly dolomite, the late cherts occur in greatest volume immediately adjacent to and out from the larger solution cavities. Elsewhere they seemingly occur only in dolomite that exhibits indications of considerable solution. In most of the region evidence of leaching is subtle and can be seen only on close inspection of the textures in the cherts themselves.

In sections that are mostly dolomite, the late-formed cherts are largely confined to the algal member and the upper third of the lower member, and certain beds obviously were more susceptible than others to leaching and silicification. Some beds contain only small amounts of secondary chert as discontinuous laminae; other beds contain numerous discrete ellipsoidal nodules, and in still others chert nodules coalesce to form layers or lenses that make up much of the bed. The nodules, or aggregates of nodules, are concentrated mostly in the upper parts of individual beds, and the upper surfaces are fairly smooth, whereas the lower surfaces of the chert masses may project very irregularly into the dolomite. On outcrops the unreplaced dolomite within the nodules and between closely spaced nodules has been partly leached, so that the nodules appear to be spongy and in a vuggy matrix.

A sequence of beds that includes much secondary chert may show striking lateral variations in thickness. In an ordinary section 20 feet thick, a sequence of sparsely cherty beds may be overlain and underlain by 1-foot zones of nodular chert. If traced a mile in one direction, the nodular zones of chert virtually pinch out, and the equivalent sequence may be 22–24 feet thick. In the opposite direction, however, the dolomite thins because of solution and the chert content may increase considerably; thus in a distance of a mile, the two chert zones may coalesce, and the equivalent stratigraphic interval—consisting almost entirely of chert and lacking in well-defined bedding—may be only 6–10 feet thick. In some places such modifications take place in a lateral distance of a few hundred feet. In these localities brecciation is obvious, and voids in and between secondary chert zones are commonly filled with sandstone.

Without doubt a considerable part of the secondary chert represents remnants of the first-formed chert mechanically concentrated by the removal of dolomite. Fragments of older chert cemented in the masses of secondary chert can be recognized in many places, although recognition can be difficult because much of the older chert was bleached during weathering. The spatial affinities with solution features suggests also that the silica which makes up the cementing chert was locally derived and locally reprecipitated. Completely silicified beds are made up of fragments of several generations of secondary chert. Apparently on initial solution only certain thin layers in the bed were silicified; with recurrent solution these collapsed, and additional layers were silicified; this process was repeated until the whole bed became chert. In a notable occurrence of such beds along the southeast flank of Shell Mountain, 8 miles east of Young, the sequence of thick and thin beds in the upper half of the lower member and in the basal 30 feet of the algal member is so perfectly mimicked that detailed stratigraphic correlations can readily be made. The sequence and most of the individual beds, however, are telescoped to only about one-third the thicknesses of equivalent intervals nearby where little silicification took place.

As a corollary to this interpretation, the spatial relations of different chert masses to other fillings in the solution openings indicate several stages of silicification. For example, argillite fillings along bedding planes may include fragments of chert, and, if the lithified argillite was then fragmented, it is cemented by similar chert. Also, dikelike fillings of Chediski

Sandstone Member cut across layers of secondary chert, but grains of this sandstone are included in other chert nodules or in zones that border the sandstone bodies. Spatial relations indicate that more than one episode of sandstone filling occurred.

KARST AND RELATED BRECCIAS

The fragmental cherts described above are but one aspect of the chert breccia formed concomitantly with the forming of karst surfaces on the Mescal. Most fragments in these breccias are less than an inch across; some relict beds include domino-shaped blocks a few inches across. The fragments and the cementing matrix may not be greatly different in texture or color. At the other extreme are the heterogeneous breccias that fill caverns whose roofs collapsed about the time the first Troy sediments were deposited. These breccias may include angular blocks of silicified Mescal that are several feet on a side, partly disintegrated boulders of basalt, and cobbles of older Precambrian quartzite from the basal conglomerate of the Troy; the matrix may be coarse sandstone or ferruginous siltstone.

In a few areas, which have mature karst features, outcrops are lithologically so different from the well-bedded carbonate rocks that they are hardly recognizable as parts of the Mescal. The algal member in particular was susceptible to thorough brecciation during the forming of solution cavities. The remnant unit is massive-outcropping coarse silica-cemented chert rubble, in which large blocks of dolomite can occasionally be found. The best examples of gradational relations between karst breccias and little modified carbonate terrane have been seen in the central and northwest parts of the McFadden Peak quadrangle. There a zone of rubble, containing recognizable but scrambled relicts from the algal member, may be as little as 20 feet thick; but half a mile away an equivalent dolomite section may be as much as 100 feet thick. Where the algal member is represented by a mixed rubble, the upper 20–30 feet of the lower member is similarly brecciated, and the breccias merge to obscure the usual sharp contact between members. Bedlike fillings of sandstone are more abundant in the dolomite section directly below such rubble zones than in other settings.

Completely silicified sections in which the entity of individual beds is preserved also grade laterally into extensive zones of heterogeneously mixed breccia. The mixed breccia is similar to the fillings seen in individual sinkholes and was probably formed—by the coalescence of sinks, which initially were widely spaced through the dolomite terrane. The vestigial strata were preserved in the areas between belts of active sink formation, where the beds were gradually thinned and recemented by chert in such a way that large-scale collapse, characteristic of the sinkholes, did not occur. Thus, locally within karst areas the stratigraphic sequence of whole units is still preserved—such as at Shell Mountain.

Where dolomite units were appreciably thinned by the mechanical erosion that preceded Troy sedimentation, layered karst breccias were formed intermediate in character relative to the breccias just described. In these breccias original gross bedding features can be followed in trains—as seen in cross section—of slightly reoriented blocks of chert, individually isolated in a matrix of hematitic siltstone or sandstone. Several parallel layers of blocks may make up the thicker layer that was originally a dolomite bed. In some relict beds the chert blocks are separated horizontally and vertically by little matrix; in other beds individual blocks and layers are separated by 2–4 inches of matrix material. Commonly, alternate relict beds are characterized by blocks of different size, separation or orientation. These distinctive breccias make up the upper 10–70 feet of the Mescal section that remained after erosion.

Delicate bedding structures, which are aids in the recognition of dolomitic equivalents, are preserved only in the individual chert blocks, where they are outlined by films of silt or pseudomorphed in porous laminae outlined by minute aggregates of specular hematite. In blocks relict from the stromatolite-bearing unit, the curvature of the stromatolites as seen in cross section is commonly greatly accentuated.

Chert blocks low in these breccias generally have reddish-brown rims; many blocks high in the breccias are stained reddish brown or grayish red throughout and could be termed jasper rather than chert. Near the bottom of these breccias, the chert blocks have sharp corners, are domino shaped, and may be more than a foot long. Upward in the sections their dimensions decrease, the edges are rounded, and generally the blocks are more widely separated. However, in many places, layers that represent the original beds remain little disturbed right up to the contact with the Troy. Therefore the rounding of the blocks and their wider separation must be solution phenomena and not the result of mechanical erosion. Apparently the higher cherts were more thoroughly impregnated with iron oxides during the solution process.

The largest area of outcrop in which the dolomite strata were thoroughly silicified and brecciated is in the Canyon Creek drainage basin north of the 34th parallel. This karst belt extends northwest as far as the head of Haigler Creek. Farther along Haigler Creek to the west, considerable thicknesses of the Mescal are locally silicified. Some parts of the main karst belt do include extensive remnants of dolomite, but most of these are interrupted by obvious solution features, such as local masses of sinkhole breccia. Dolomite is abundantly remnant only in the basal part of the section, where the older founder breccia lacked the bedding planes that were favored channelways for solution.

Another large karst area, only partly exposed in a belt of outcrops 1 to 3 miles wide, extends from U.S. Highway 60 southeast about 13 miles along the southern margin of the Natanes Plateau. In this area the Mescal was considerably eroded in pre-Troy time, and locally the Troy rests on the Dripping Spring Quartzite. Where the Mescal is very thin, even the founder breccia was converted to ferruginous karst breccia and was much silicified.

Just within the northeast corner of the Globe quadrangle and immediately to the east, in an area possibly not exceeding 6 square miles, a somewhat different aspect of the silicified facies exists. The Mescal consists almost entirely of chert, which occurs as thin layers and lenses separated by thin seams of silty material. The effects of repeated small-scale brecciation and cementation are obvious in many layers, but in wholly as many layers such effects are not apparent. Large-scale jumbling of breccia blocks is seen only very locally. Apparently the early-formed chert settled with a minimum of brecciation as the dolomite was removed; secondary chert cement exists in smaller amounts here than in most of the karst occurrences, and little extraneous sand, silt, or argillaceous material was introduced. This belt contains little hematite, whereas large parts of the two karst belts noted above contain abundant hematite. Chert sections representing the entire lower member and part of the algal member are 60–120 feet thick; 1–2 miles to the east, where solution phenomena are few, equivalent dolomite sections are 300 to almost 400 feet thick. In several other small areas where brecciation is inconspicuous, similar chert sections grade laterally into highly ferruginous obviously brecciated sequences.

HEMATITE CONCENTRATIONS

As a consequence of Precambrian weathering, hematite may be sparse or abundant in the upper part of the Mescal. Some concentrations are extensive and thick enough to be potential sources of iron. These have been prospected sporadically in the past (Burchard, 1931; Stewart, 1947), and since 1958 have been actively explored.

The larger iron deposits are in the upper parts of karst breccias in which layers of chert nodules mimic the bedding of original dolomites. Earthy but dense hematite and unctuous specular hematite, in massive layers, constitute most of the ore-grade material. Earthy hematite is dominant in the lower grade ores. Small botryoidal masses of hematite do not make up a great volume but are widely distributed. Micaceous specular hematite commonly lines small vugs and forms veins in the hematite and in hematitic rock that is interlensed with the ore. Hematite, chert, and detrital quartz grains generally make up 90–98 percent of material that assays more than 25 percent iron.

The best and largest known deposit (in 1961) is partly exposed along the walls of Canyon Creek just above its junction with Swamp Creek, a minor tributary 6 miles south of the Mogollon Rim. There, layers of hematite are intercalated with, and gradational into, highly hematitic layers of argillite, siltstone, silicified sandstone, and light-colored to dusky-red chert breccia in a zone that dips gently south and is 20–70 feet thick. The zone crops out almost continuously for a distance of 10,000 feet along the east wall of the canyon, is discontinuously exposed on the west wall, and is known from drilling to continue southward below stream level for at least 1,500 feet (A. P. Butler, Jr., written commun., 1959). Before erosion the zone undoubtedly was continuous beneath an area of at least 1 square mile. An interval 5–25 feet thick in the middle of the zone has the highest content of iron. Discontinuous crudely lenticular bodies 2–20 feet thick that contain more than 50 percent iron make up roughly two-thirds of this interval. An exceptional lens may be as much as 40 feet thick. Above the interval much of the zone is made up of thin layers of highly hematitic sandstone, and layered chert breccia dominates the part of the zone below the high-grade interval.

Lenses of hematite, too thin and small to constitute potential ore, occur locally in silicified strata 20–70 feet below the main iron-bearing zone. The basal few feet of the overlying Chediski Sandstone Member and thin lenses higher in the member also include local concentrations of hematite, some of which approach ore grade.

Several other iron deposits, petrographically similar to the one at Swamp Creek, occur in the belt of silicified Mescal that crops out along the upper reaches of Canyon Creek. Some are possibly more extensive than the deposit at Swamp Creek, but none has been found to contain as thick or continuous high-grade lenses.

Thinner zones of ferruginous rock occur at the top of the Mescal in some areas south of the 34th parallel, but widespread lenses of almost pure hematite are sparse. Instead, thin layers of chert or of slightly reoriented chert fragments occur in a hematitic matrix, which in many places consists of coarsely micaceous hematite. The karst belt along the south rim of the Natanes Plateau includes local concentrations of such material; some are several feet thick over several acres. This is the southernmost known locality of such concentrations. Commonly, thin lenses of hematite, no more than 6 inches thick, occur wherever an appreciable thickness of the Mescal is silicified, and in a few places, thin lenses even occur very sparsely in dolomite beds immediately below the silicified zone. Everywhere, regardless of the degree of silicification lower in the section, hematite flecks the chert that immediately underlies the unconformity at the top of the algal member.

In past speculations on origin, it has been erroneously concluded that the iron was deposited along the Troy-Mescal unconformity from hydrothermal solutions, which had a source in the diabase intrusions. However, well-rounded waterworn pebbles of cherty hematite are common in the basal beds of the Troy Quartzite and occur sporadically in conglomerate lenses tens of feet higher in the formation. These distinctive pebbles undoubtedly were derived from the Mescal, as was silt- and sand-sized hematitic material that locally forms lenses almost of ore grade in the basal Troy. Even without the evidence afforded by the pebbles, spatial relations between diabase bodies and the iron-bearing zones clearly indicate that the intrusions postdate the hematite deposits.

The hematite deposits are here regarded as fossil laterites, firmly cemented by silica released during laterization. The source rocks for the residual iron were either the dolomites—an unlikely possibility—or the basalt flows, which once everywhere overlay the Mescal.

The red ferruginous dolomite, which dominates the Mescal in certain areas, either contains considerable secondary chert or is vuggy or spongelike and obviously has been leached. Thick sections of red dolomite occur only near conspicuous solution features, such as sinkholes. Brown dolomite, which contains relatively little iron, is much more abundant and is characteristic of sections far removed from sinkholes. The spatial relations of the two types of dolomite suggest that the red dolomite is hematitic as a consequence of laterization and is not the source of great amounts of iron.

Basalt flows probably furnished most of the iron. Even the least altered basalt is dusted throughout with hematite, and specular hematite is abundant in vesicules and voids. Fossil soils on the basalt are almost indistinguishable from the hematitic argillite and siltstone that make up the matrix of many chert breccias or the basal parts of some iron-ore zones.

The iron-rich zones are here interpreted as layered karst breccia with a matrix of lateritic residues from basalt. These residues filtered down into the innumerable and in places large spaces left by the solution of dolomite. Some reworking and transport of these hematitic sediments occurred later, during the advance of the Troy sea, as shown by the makeup of narrow conglomeratic sandstone dikes in the tops of some iron zones. These dikes contain gravels mostly of foreign derivation in a matrix of locally derived hematite mud; a few also include well-rounded hematite pebbles. Some of the thicker layers of hematite and associated siltstone and sandstone in the upper parts of the hematite zones are discrete bodies with layering suggestive of sedimentation processes. However, some of these are between layered breccia units in which the chert fragments show little reorientation; these bodies must represent filled voids between beds, similar to the sandstone lenses seen in sections where relations are not obscured by hematite. Most tabular hematite bodies are probably fill features. In any event, silica was available as a cement during the initial accumulation of the iron, and the principal iron layers were lithified before encroachment of the sea in which the Chediski Sandstone Member of the Troy was laid down.

METAMORPHIC LIMESTONES

Adjacent to diabase intrusions, the dolomite of the lower and algal members was converted to calcitic limestone, and most of the chert was altered to silicate minerals. These effects of thermal metamorphism might be expected to be local, and in a few places where only one thin intrusion was emplaced, the effects are not widespread. Typically, however, several extensive diabase sills intrude the two members; consequently the metamorphic effects are regional in extent, and more than half of all Mescal exposures are silicated limestones. Indeed these limestones are so prevalent, and the included silicates are commonly so inconspicuous, that a sedimentary origin has been tacitly assumed. In the following paragraphs some of the physical rather than mineralogical effects are stressed, so that metamorphic and premetamorphic features of stratigraphic equivalents can be compared.

Mineralogic changes in the carbonate sections were brought about largely through a recombination of the sedimentary constituents. Silica and small amounts of alumina combined with the magnesia and lime of the dolomite to form silicates. The calcite remaining after this dedolomitization makes up the bulk of the reconstituted strata. With few exceptions these calcitic limestones are very fine grained and sugary. The silicates, with calcite, occur as mixed aggregates crudely pseudomorphic after masses of chert. They are mostly light colored, weather white, and do not etch out conspicuously, as does chert; therefore their presence in the metamorphic limestones may not be evident to the casual observer. Where zones of secondary chert were thick, the zones of silicate minerals include relicts of unreplaced chert; otherwise chert is largely absent from the metamorphic limestones. Highly siliceous equivalents of the dolomite were sparsely intruded by diabase, so that thermal metamorphic effects in these rocks are uncommon. The metamorphic minerals are described further under "Metamorphism associated with diabase."

The physical changes caused by metamorphism are striking. Dolomite that was silty or that contained chert in layers less than an inch thick was metamorphosed to shaly limestone. Unless viewed microscopically, silicate minerals seem to be lacking in these rocks. The partings simulate shaly bedding but are actually cleavages formed during metamorphism. At the other extreme, some of the thicker silt-free dolomite beds that were separated by massive irregular layers of secondary chert were literally welded together by silicate zones and formed layers more massive than their dolomite counterparts. In many areas, however, some silty material had been concentrated along solution planes between beds, and as the result of metamorphism the bordering carbonate rock became shaly; therefore such units appear thinner bedded than the original dolomites. Consequently, limestone and dolomite sections weather differently—units that as dolomites are ledge formers may as limestones form gentle slopes. The limestones are mostly gray to white, and from a distance shaly units are chalky in appearance.

Although mineralogic changes indicate the same degree of metamorphism throughout the Mescal, the various stratigraphic units differ in the physical modifications caused by metamorphism. The basal sandstone of the Mescal appears little changed—the matrix contains calcite rather than dolomite, and limonite stains are more common. The argilliceous and sandy matrix of the overlying founder breccia is altered to a greenish-gray massive rock. The dolomite of the breccia blocks was largely converted to calcite, but chert layers within blocks more than a foot thick are surprisingly little changed. The borders of the blocks grade into the silicated matrix. The remainder of the lower two-thirds of the lower member was converted to limestone with partings at intervals ranging from paper thin to 30 inches. The thicker pseudobeds are separated by a fraction of an inch to several inches of shaly limestone and are themselves very friable where moderately weathered.

These limestone layers—ledge and cliff makers as dolomite—form moderate to gentle slopes, are subject to slope creep, and commonly are covered by resistant debris from higher in the section. Thus in the metamorphic rocks it is difficult to distinguish between slumping that occurred when the founder breccia was formed and slumping that represents recent creep. Where the founder breccia was especially thin, as along the Salt River east of Canyon Creek, a few thick ledges of limestone, comparable with those in the upper third of the lower member, do crop out 40–50 feet above the base of the Mescal. Otherwise, resistant limestone beds are generally absent in the lower two-thirds of the lower member.

The upper third of the lower member (units 8 and 9, reference section) was modified to thin- to thick-parting units of white to medium-gray limestone separated by thin units of shaly limestone. Details of original bedding are preserved more faithfully here than in lower strata. In the thicker layers, metamorphic welding of beds is common. The resistant thick layers tend to form rounded ledges, accentuated by the etching out of the shaly seams. In general, metamorphic sequences of this interval crop out on steeper slopes than their dolomite equivalents.

Gray metamorphic limestone of the lower unit of the algal member retains original details of bedding and stromatolite structures, and the unit remains a massive cliff former. The upper unit of the algal member was converted to limestone much like that of the upper 50 feet of the lower member. Concentrations of secondary chert at the top of the algal member are represented by partially serpentinized remnants; otherwise relicts of chert are sparse.

It has not previously been recognized that most sections of the Mescal are entirely one lithologic type or the other—that is, entirely thin- to thick-bedded brown or red cherty dolomite or entirely silicate-bearing white-weathering shaly metamorphic limestone. Along the fringes of a metamorphic belt, pinkish-gray or brown dolomitic limestones—partially dedolomitized—do make up transitional zones, and cherty dolomites or cherty limestones are locally intercalated with silicate-bearing limestone, but such transition zones are narrow and readily overlooked. Descriptions that characterize the formation as consisting of dolomitic limestone interbedded with nonmagnesian limestone and descriptions that fail to correctly indicate that chert is almost ubiquitous probably were based on observations of metamorphosed sections in which the white silicates were not recognized. Certainly the stressing of thin-splitting characteristics, as though they were original, reflects such observations.

VARIATIONS IN THICKNESS

Six events or processes caused thinning of the Mescal before Paleozoic sedimentation was complete. These were (1) intraformational erosion before the argillite member was deposited, (2) pre-Troy erosion, (3) solution, (4) metamorphism, (5) pre-Bolsa erosion, and (6) pre-Martin erosion.

The thickness of the argillite member depends mainly on the amount of slumping due to solution of underlying strata and the amount of erosion prior to deposition of the Troy Quartzite. Over the large areas in which basalt flows cap the argillite member, thicknesses are fairly uniform except where basins were formed during deposition by slumping of the dolomite. In these basins, a square mile or less in area, argillaceous strata are abnormally thick. Beneath the basalt cover in the McFadden Peak quadrangle, the argillite member is typically 45–60 feet thick; locally, however, it is 100 feet thick. Toward the boundaries of the area of outcrop outlined in figure 4, the argillite member thins rapidly, and locally it is missing within that area, owing to pre-Troy erosion.

Thickness variations of the algal member mainly reflect four episodes of Precambrian erosion. The two erosional events that preceded deposition of the argillite member caused as much variation as later episodes. Along the north edge of the McFadden Peak quadrangle the upper flat-bedded unit is mostly missing, and the member is 50–60 feet thick; 3 miles to the south, thicknesses of as much as 105 feet have been measured. Along the canyon of the Salt River east of Canyon Creek few sections are less than 60 feet and most are 85–100 feet thick. One section of 130 feet, the thickest known, was measured near the Regal mine on the south rim of the canyon 5½ miles west of U.S. Highway 60. From the village of Chrysotile as far south as Globe, the upper strata of the member were generally truncated by erosion surfaces at the base of the basalt flows or at the base of the Troy; the upper unit is thin or missing, and the thickest remnants of the member are 50–70 feet thick.

In a few small areas north of Globe, such as along the south rim of the Natanes Plateau south of Sawmill, the Mescal Limestone is entirely eroded away, and the Troy rests directly on the Dripping Spring Quartzite. Further south, as at the southeast end of the Mescal Mountains and in the Holy Joe Peak quadrangle, the Mescal is missing over large areas owing to pre-Troy erosion.

Thinning on a large scale by solution is a local phenomenon, restricted as far as is known to the karst areas north of the latitude of Globe. In several of these areas, solution caused the combined thickness of the lower and algal members to be less than 100 feet. Throughout the region of outcrop, however, subtle thinning was caused by small amounts of solution along many bedding planes. This thinning, probably on the order of 10–50 feet, was largely obscured by later metamorphic effects.

Where metamorphosed and unmetamorphosed sections can be compared directly in immediately adjacent areas, the thickness of the lower and algal members seems to decrease toward the area of metamorphism. The cherty dolomite in the lower member of the reference section, for example, is 269 feet thick; but seemingly equivalent metamorphosed sections within a 1-mile radius northwest, north, and northeast of this section are about 220 feet thick. Also, in comparing widely separate sections apparently similar before metamorphism, no limestone section has been seen that is as thick as the equivalent dolomite section. Limestone or dolomite sections of the lower member in the Sierra Ancha and east to Canyon Creek range in thickness from 200 to 270 feet. Southeast, in the canyon of the Salt River and south as far as Chrysotile, the member ranges in thickness from 150 feet to slightly more than 200 feet. Only the thinnest of the western sections are reconstituted throughout to the same degree as the sections of the Salt River–Chrysotile area. Farther south comparable sections exceed 180 feet, and those that show the least metamorphism are 220–250 feet thick.

Because other processes contributed widely to thinning of the algal member, similar comparisons of metamorphic effects are not practicable for that member.

Although exact equivalence is difficult to determine, because of earlier solution thinning, seemingly comparable samples of the lower member suggest that metamorphosed sections are 15–20 percent thinner than cherty dolomite sections. A similar loss of volume on dedolomitization has recently been noted in a comparable metamorphic setting (Cooper, 1957, p. 582–588), and the probability of such shrinkage certainly should be considered for all the limestone of the Mescal.

The sequence of beds is the same wherever the Mescal is exposed, and partial sections similarly modified —whether by solution or metamorphism—are about equal in thickness; thus dolomite sections throughout the region were probably once uniform in thickness.

This thickness, if a few sections are representative, was about 400 feet for the lower and algal members combined. Because the original extent of the argillite member is not known, it cannot be included in a total that is applicable regionally.

Apache and Troy strata were differentially displaced by the intrusion of diabase sills, and later, during two episodes of erosion, the most elevated parts were stripped away (fig. 12). In places from Globe south to the Gila River, part of the Mescal remains under a cover of Bolsa Quartzite (Cambrian), but farther south, where most of the Mescal previously had been stripped away by pre-Troy erosion, the last remnants of large areas were removed by pre-Bolsa erosion. North of the latitude of Globe, pre-Bolsa erosion had little effect on the Mescal; but as far north as the Natanes Plateau, the Mescal was thinned or completely removed by an erosion cycle that predated the deposition of the Martin Limestone (Devonian). Farther north pre-Martin erosion breached the Troy cover locally, but only in areas of a few square miles.

To sum up, the Mescal is generally thickest (350–420 ft) in the northern part of the region, and southward progressively larger amounts of the Mescal are missing; even so the most complete local remnants in the south are only moderately thinner (250–300 ft) than those to the north. In the northern area, the Mescal was thinned locally to less than 100 feet during the development of karst features.

STRATIGRAPHIC NOMENCLATURE

The lower and algal members of the Mescal Limestone are distinctive units that logically can be grouped as a formation. However, in view of the lithologic differences and the erosional unconformity between the algal member and the overlying argillite unit, many geologists would be inclined to consider the argillite a separate formation. Such formational designation would be well within the principles stated by the American Commission on Stratigraphic Nomenclature (1961, p. 648–654). Indeed, in the only previous discussion of the argillite, formational status was proposed by Hinds (1935, p. 32), who said "At and near [Theodore] Roosevelt Dam, there are present locally above the vesicular basalt a maximum of 30 feet of chert and dark reddish and purplish siliceous shales which I include in the Apache group * * * and propose to call the Roosevelt member." Hinds here used the term "member" in a formational sense. The section described by him is incomplete, and because of deep weathering it is not typical of sections elsewhere; nor was it possible to determine that the principal basalt flows overlie the argillite.

The argillite member is preferably included as the uppermost member of the Mescal Limestone because (1) it is thin, (2) it crops out mainly in cliffs too narrow to depict except on large-scale maps, and (3) it almost certainly will be recognized and separately mapped only in an area of about 800 square miles in northern Gila County. Lithologic similarities between limestone of the argillite member and that of the underlying members indicate the recurrence of a similar environment of sedimentation and suggest a genetic, but not compelling, reason for including the argillite member in the Mescal. North and south of the area of outcrop shown in figure 4, the argillite member is virtually absent; west of Theodore Roosevelt Dam, where additional outcrops may yet be found, few remnants of the Apache Group exist; and east of the area the Apache Group is deeply buried by younger formations. In view of these facts, formational status would serve little practical purpose.

Although the Mescal originally was mostly dolomite, and although it now is locally a chert formation and includes an argillite member, limestone of metamorphic origin makes up most exposures. Therefore the name Mescal Limestone has been retained.

BASALT FLOWS

GENERAL FEATURES AND DISTRIBUTION

Basalt flows occur at two stratigraphic positions high in the Apache sequence. Throughout the region the principal occurrences separate the Troy Quartzite from the Mescal Limestone; flows petrographically similar but restricted in lateral extent locally separate the argillite and algal members of the Mescal. The two basalt sequences in their juxtaposition with the argillite member of the Mescal are best exposed in cliff faces in the southern part of the Sierra Ancha.

The basalt flows immediately beneath the Troy Quartzite generally aggregate 40–90 feet in thickness, and these apparently once extended in much greater thickness over the entire region. Variations in present thickness are the consequence of pre-Troy erosion. In the Sierra Ancha and east to Canyon Creek, this basalt ranges in thickness from 20 to 100 feet. North of the 34th parallel and also along the canyon of the Salt River east of Canyon Creek it is largely absent. In the vicinity of the village of Chrysotile, 5 miles south of the canyon, occurrences are slightly more than 100 feet thick. Southeast from Chrysotile along the Natanes Plateau, the basalt is generally absent; but in a small area 16–20 miles to the southeast, one flow at least 110 feet thick is locally remnant. Farther south the basalt is absent from large areas; where present, it directly overlies the algal member of the

Mescal. In the Apache Mountains the basalt is entirely missing; yet along the low chain of hills that extends 12 miles southeast from these peaks, it is commonly more than 250 feet thick, and in one place, 10 miles east of Globe, it is 375 feet thick. Along the mountain front 1–3 miles south of Superior, the basalt locally exceeds 200 feet in thickness. Elsewhere south of the Natanes Plateau most sections are 50–75 feet thick. South of the Gila River remnants are sparse but do exist as far as the Vekol Mountains.

Basalt is known to separate the argillite and algal members of the Mescal in only three localities: (1) along the southern front of the Sierra Ancha in an area of at least 9 square miles, (2) along Cherry Creek 6 miles northeast of McFadden Peak, over an area of about 1 square mile, and (3) in the vicinity of Theodore Roosevelt Dam. Basalt at this horizon is as much as 50 feet thick in the southern Sierra Ancha and at Theodore Roosevelt Dam and as much as 110 feet thick at the Cherry Creek locality.

The flows in the vicinity of Globe, Superior, and Ray, in the Mescal and Dripping Spring Mountains, and as far southwest as the Vekol Mountains could correlate with the basalt that underlies the argillite member of the Mescal in the Sierra Ancha, but this correlation cannot be made with any assurance. Obviously the upper (or principal) basalt overlies the argillite member where it thins along its southern extremities and could also be the basalt unit that directly overlies the algal member farther south.

The boundaries between flows are commonly obscure. In most sections only one flow is seen, but unusually thick sections contain as many as four flows and in some a fifth may be present. Individual flows are 50–125 feet thick.

The basalt formed as subaerial flows, which were very fluid. Individual flows poured out in uniform thickness over extensive areas. In many areas the flow breccias at the top of a flow were eroded away, and a planar erosion surface largely free of residual soils was formed before extrusion of the next flow. About half a mile south of Chrysotile two flows are separated by 10–12 feet of silty limestone, and between Cherry and Canyon Creeks a few inches of laminated argillite locally separates flows. These sedimentary layers and the planar surfaces suggest inundation of the region between outpourings of lavas. No feeder dikes for these lavas have been found.

PETROLOGY

The basalt is grayish red to blackish red or brown; a few outcrops weather pale brown to dark yellowish brown. Hematite is so abundantly disseminated through the rock that blackish red is the dominant color of fresh and weathered exposures. Most specimens show an intersertal or intergranular texture, defined by laths of plagioclase 0.1–0.3 mm long. Needle-like crystals of apatite are abundant in some of the basalt. The rock has been altered to such a degree that identification of the original plagioclase has not been possible. Under the microscope the alteration minerals seen are albite, calcite, serpentine, quartz, and chlorite; all are dusted with hematite and so aggregated that the original texture is not as apparent as in a hand specimen. In some altered specimens, magnetite, epidote, and an amphibole (hornblende (?)) are common.

Large tabular phenocrysts of plagioclase are sparse to abundant in most flows. Some vertical sections of a given flow are porphyritic throughout; only parts of other sections are porphyritic. The vesicular parts of some flows include the greatest number of phenocrysts. They typically range in thickness from 0.5 to 2 mm and in diameter from 5 to 20 mm. Plates as much as 40 mm across have been seen, and in a few places phenocrysts make up as much as 40 percent of the rock. In many outcrops the phenocrysts weather almost white and are very striking. In especially hematitic basalts they are moderate red to dusky red and are inconspicuous. Regardless of the size or abundance of the phenocrysts, the groundmass varies little in texture or grain size. Except that no porphyritic varieties have yet been observed in it, the basalt below the argillite member resembles that above the member.

Vesicles and amygdules are especially abundant in the tops and bottoms of flows; ropy flow breccia is conspicuous in places. Partly filled elongate vertical vesicles (pipe vesicles) occur locally and abundantly in the bottoms of flows. Uncommonly the basal breccia of the lowest flow of the upper sequence includes fragments derived from the argillite member of the Mescal. Variolitic basalt, a few inches to a few feet thick, makes up the basal part of the lowest flow in a few localities. Voids and vesicles are filled by calcite, quartz, specular hematite, and an unidentified grayish-green material. In a few localities blue-green copper minerals coat weathered joint and fracture surfaces.

CRITERIA FOR DISTINGUISHING BASALT FROM DIABASE

Typically the basalt is readily recognized as conspicuous blackish ledges or cliffs just below the Troy Quartzite. In places, however, the distinctive outcrops are obscured by talus; and in other areas they do not exist, because the basalt that was altered adjacent to diabase intrusions disintegrated on weathering. The basalt has been mistaken for diabase where it is not exposed in cliffs.

The two rock types can usually be differentiated by noting the following characteristics: (1) The plagioclase phenocrysts of the basalt have no counterparts in the diabase. Some coarse facies of the diabase contain plagioclase crystals even longer than the phenocrysts of the basalt; however, they are lath shaped and intergrown with equally large grains of other minerals, whereas the phenocrysts of the basalt are tabular and set in a fine-grained groundmass. (2) Amygdules of quartz, quartz-hematite, or hematite are characteristic of the basalt. (3) The basalt decomposes to a fine powdery dark-reddish-brown soil, which contains abundant residual amygdules even in areas where amygdules are not obvious in adjacent outcrops. The soil typically derived from diabase is granular, contains abundant fragments of plagioclase laths in a yellowish-brown clayey matrix, and tends to be adobelike. Surfaces of some diabase soils are powdery, but the typical granular detritus can be found by digging a few inches. (4) The finer grained diabase commonly breaks down into a rubble of small ellipsoidal pebbles which represent the pyroxene host crystals in the ophitic-textured rock. The small plagioclase laths in these pebbles tend to weather chalky white and are conspicuous. In contrast, basalt soils include only scattered pebbles and fragments of irregular form, and the feldspar laths, if distinguishable, almost invariably weather reddish orange to yellowish brown. The basalt pebbles where freshly broken do not show the flashing cleavage surfaces of pyroxene characteristic of the diabase pebbles.

PRE-TROY UNCONFORMITY

The angular relations of the unconformity that separates the Troy Quartzite from the Apache Group are apparent only on a regional scale. Apache strata were broadly warped, then eroded before Troy deposition, as shown diagrammatically in figure 11. Except for channels that occur in a few areas and that locally exceed 100 feet in depth, the pre-Troy surface was virtually planar. In the northwestern part of Gila County, this surface was downwarped to form a structural basin a few hundred feet lower than the general surface. Basal strata of the Troy lap out against the margins of this basin, but elsewhere in the northern part of the region they are practically concordant with underlying strata. In the southern part of the region, the surface truncates the underlying strata; therefore, at localities several miles apart, the Troy rests on different formations of the Apache Group.

TROY QUARTZITE

DEFINITION AND SUBDIVISION

On redefinition of the Troy Quartzite as Precambrian rather than Cambrian and on delineation of three distinctive members, two of which are regional in extent, it becomes obvious that the Troy encompasses a greater thickness and variety of strata than is usually conceived. Nonetheless, there can be no doubt that Ransome (1915, p. 384–385; 1916, p. 139–141, 154) included all these strata in his earliest descriptions of the Troy. In the Ray quadrangle, where it was originally defined as separate from the Dripping Spring and was first mapped (Ransome, 1919, p. 44–45; 1923), the Troy consists mainly of strata that are here assigned to the Chediski Sandstone Member. The first section Ransome described, 9½ miles southeast of the type locality on Troy Mountain in the Ray quadrangle, is 362 feet thick. From reconnaissance outside of the quadrangle, Ransome realized that the Troy occurs in much more extensive outcrops to the north and noted that it is especially thick in the Sierra Ancha. He therefore described, in generalities, a 900-foot section on Baker Mountain, which is 47 miles north of the type locality. The Baker Mountain section is 2½ miles southwest of the Center Mountain section designated as a reference section in this report. The two sections are similar, but the Baker Mountain section is not so well exposed and lacks, owing to pre-Devonian erosion, the upper 200 feet or more of the sequence remnant in the Center Mountain section. Ransome also recorded, apparently with some misgivings, a third section at Theodore Roosevelt Dam. That section is only 160 feet thick and is partly or perhaps wholly of Devonian sandstone. It is now apparent that none of the southern sequences of the Troy are so complete as several remnants that are in the high part of the Sierra Ancha, within a few miles of the reference section described here.

From his most detailed observations, Ransome considered 400-foot thicknesses representative of the Troy in the Ray quadrangle and preferred not to give credence to sections seemingly much thicker at and near the type locality. Local abrupt changes in thickness are readily reconciled, however, when one recognizes that the Troy was differentially uplifted along faults, then variously eroded before deposition of Cambrian strata and, in the northern part of the region, again deeply eroded before Devonian strata were deposited. As a consequence adjacent Troy sections separated by a Precambrian fault and overlain by an unfaulted capping of Paleozoic strata may vary several hundred feet in thickness. In the northern part of the region, most sections are more than 400 feet thick, and a few are at least 1,200 feet thick; in much of the southern part, the Troy is absent, but where remnant it ranges in thickness from 300 to 700 feet.

In a few places small amounts of rusty-weathering thin-bedded sandstone, now recognized as Cambrian, were included in the top of the unit first designated as Troy Quartzite. The 362-foot section described by Ransome includes 50 feet of Cambrian sandstone. North of the Ray quadrangle, except in a few small areas between Globe and Superior, Cambrian strata are entirely missing from outcrops designated as Troy by Ransome and later workers. Therefore, published maps that show the Troy in the northern areas require only slight modification. However, quartzitic strata called Troy in the area between the latitudes of Globe and Tucson must now be reassessed. Some of the outcrops are Troy; many others are entirely the Cambrian Bolsa Quartzite or are wholly sandstone facies of the overlying Abrigo Formation.

In the Sierra Ancha the Troy can be readily subdivided into three members: the lowest member is a reddish well-sorted arkose, the middle member is a white sericitic and pebbly sandstone, and the uppermost is a gray clean quartzite. The following section, measured along the east face of Center Mountain in the highest part of the range, is one of the most complete yet found.

Reference section of Troy Quartzite

[Measured along east face of Center Mountain; arkose member described largely from exposures in NE¼ sec. 16, T. 6 N., R. 14 E.; Chediski Sandstone Member measured in NW¼SW¼ sec. 15 and NE¼SE¼ sec. 16, T. 6 N., R. 14 E.; quartzite member descriptions from SE¼ sec. 16, NE¼ sec. 21, and NW¼ sec. 22, T. 6 N., R. 14 E.]

	Thickness (feet)
Troy Quartzite:	
Quartzite member:	
Top eroded. Probably one of the thickest remnants of the member.	
11. Quartzite, dominantly medium light gray but many beds pale red to grayish-red purple; mostly medium-grained with sparse content of coarse grains; a few beds coarse grained; in tabular beds 3–12 in. (mostly 4–8 in.) thick, most cross stratified on small to medium scale; flaggy to slabby parting is everywhere conspicuous. Lower part crops out in steep slope partly covered with talus; slope merges upward into cliffs as much as 150 ft high	250
10. Quartzite, very light gray to pale-red-purple, medium-grained; in tabular beds 4 in.–3 ft (mostly 1–2 ft) thick; more vitreous than unit 11, and internal stratification obscure; on weathering, outcrops stain slightly to pale or moderate reddish brown; joints may be thinly coated by limonite. Forms cliff with narrow to broad bench at top; upper 80 ft not as resistant as lower part	185
Troy Quartzite—Continued	
Quartzite member—Continued	
9. Quartzite, pale-red to grayish-red-purple, coarse-grained; voids between rounded grains imperfectly filled by quartz overgrowths. Otherwise like unit 10, and crops out as steep ledgy slope or cliff continuous with that of unit 10	60
Thickness of quartzite member (units 9–11) remnant	495
Chediski Sandstone Member:	
8. Sandstone, very light gray to pinkish-gray; mostly coarse grained, some beds very coarse grained; sand-grain and matrix characteristics as in unit 6; tabular beds 3–8 ft thick cross stratified on medium scale. Well-rounded pebbles, mostly less than 1 in. in diameter, of white to pink quartz and very sparse pebbles of reddish-brown jasper are scattered through a few beds and concentrated in upper 1–2 in. of many beds. Pebbles most abundant in upper 30 ft, and topmost bed conspicuously conglomeratic. Lower 44 ft friable; upper 30 ft tends to be quartzitic. Crops out as massive rounded ledges, in which bedding structures are obscure	74
7. Sandstone, light-gray to pinkish-gray; not pebbly; otherwise similar to unit 6	80
6. Sandstone, pinkish-gray to grayish-red-purple; progressively darker and more mottled by reduction spots upward, fine- to coarse-grained (mostly medium-grained), poorly sorted, friable; of well-rounded minutely pitted quartz grains in matrix of sericite and clay. Convolute lamination conspicuous throughout massive layers, mostly 10–30 ft thick; layers display little continuity of bedding. A few thin (2 ft maximum) tabular beds, with medium- to large-scale tangential cross-stratification, separate these layers. Unit contains scant pebbles like those in unit 5. Forms gentle to steep slope, locally surmounted by hoodoos	67
5. Conglomeratic sandstone, light-brownish-gray, poorly to firmly cemented; bedding obscure but dominantly horizontal; mostly coarse grained, but part very coarse grained; includes lenses of granule-size gravels. Sand grains well rounded, minutely pitted, mostly of quartz; very sparse granules of orange-pink feldspar conspicuous. Pebbles mostly of white to moderate-red quartz; about 20 percent are of pale-yellowish-brown to grayish-red quartzite; a few of feldspathic quartzite, reddish-brown	

	Thickness (feet)
Troy Quartzite—Continued	
Chediski Sandstone Member—Continued	
jasper, or rhyolite. Most pebbles well rounded and less than ¾ in. in diameter; maximum diameter, 3 in.; some of larger pebbles are ventifacts. Crops out as massive ledges or small cliff	39
Thickness of Chediski Sandstone Member (units 5–8)	260
Arkose member:	
4. Arkose, light-brown to pale-brown or pale-red (pale red dominant), medium-grained, firmly cemented; in tabular beds 2–8 ft thick that exhibit large-scale cross-stratification; some beds wedge shaped; unit forms cliffs. Cross-stratification truncated at sharp slightly irregular contact with unit 5	110
3. Arkose, pale-brown to grayish-red, fine-grained, firmly cemented; cross stratified on very large scale; planes separating tabular bedding sets are mostly tens of feet apart; slabby parting; crops out only on narrow hoodoo-studded spurs, talus covered elsewhere	165
2. Arkose, like unit 1 except in tabular beds 4 in.–3 ft thick. Some beds separated by silty seams that preserve mud cracks and, uncommonly, ripple marks. Crops out as receding cliff of massive ledges	60
1. Arkose, pale-brown, fine-grained, very firmly cemented; mostly in wedge-shaped bedding sets (1½–8 ft thick at thickest part) cross stratified on medium to large scale. Lower 15 ft in cross-stratified tabular sets 4–18 in. thick. Basal 6–18 in. includes abundant fragments from underlying basalt. Forms ragged cliff	108
Thickness of arkose member (units 1–4)	443
Thickness of Troy Quartzite	1, 198
Unconformity.	
Basalt flow: Basalt, blackish-red, fine-grained, porphyritic, amygdaloidal; crops out as steep slope or cliff	40+

As indicated in the reference section, the three members have distinctive lithologies and bedding structures. These distinctions are conspicuous north of the Salt River, but they are subdued in most areas to the south owing in part to lithologic changes and in part to the incompleteness of sections. The arkose member is restricted to the pre-Troy basin of northwestern Gila County; the Chediski Sandstone Member is represented throughout the region; the quartzite member thins drastically southward and probably is missing from all areas south of the Mescal Mountains.

ARKOSE MEMBER

DISTRIBUTION AND THICKNESS

The arkose member is exposed in an area of only a few hundred square miles, and outcrops are numerous only within the McFadden Peak quadrangle. In the highest parts of the Sierra Ancha, within 2–4 miles of the Center Mountain section, the member is 400–450 feet thick; to the north, east, and south it thins against successively higher parts of the pre-Troy surface; within 8–15 miles it laps out completely, and the Chediski Sandstone Member becomes the basal subdivision of the Troy (fig. 11). The arkose member can be traced into areas in which it is less than 100 feet thick, but it is now missing—owing to Cenozoic erosion—from areas where it thinned further and ultimately pinched out. Nevertheless, the extrapolated limit of the member shown in figure 4 is narrowly restricted except along the southern segment, where the first remnants of the Chediski that rest directly on Apache strata are as much as 10 miles south of the southernmost remnants of the arkose member. Judged from the pattern of thickening, the arkose member might be as thick or even thicker in the western Sierra Ancha than in the high parts of the range; but west of the 111° meridian, erosional remnants of the Troy are so few, thin, and widely scattered that this has not been confirmed.

STRATIGRAPHY AND LITHOLOGY

Where the member is thin, or where it rests mainly on the argillite member of the Mescal, a conglomerate or conglomeratic sandstone 5–20 feet thick lies either at the base of the arkose or is 10–20 feet above the base. The pebbles and cobbles mostly are well-rounded quartzite and rhyolite from the older Precambrian terrane and are in a matrix similar to the overlying arkose. Where thickest, the member generally rests on the upper basalt of the Apache Group, and the basal part is not conglomeratic. In a few areas a thick residual soil mantled the basalt; this debris was incorporated in the basal few feet of the member and made it dark, earthy or friable, and from a distance difficult to distinguish from the basalt.

The arkose above the conglomerate is fine to medium grained, firmly cemented, and distinguished from other Precambrian sandstone by striking bedding features. This arkose is mainly pale red to grayish red and secondarily light to pale brown on both fresh and weathered surfaces. Individual beds are uniformly sorted. Most are fine grained, tabular, and more than 10 feet thick; beds as much as 40 feet thick are common. In areas of low relief, the arkose erodes to gentle smooth slopes and, in areas of high relief, to cliffs or to steep slopes studded with rounded to very angular ledges. Regardless of outcrop form, large-scale or very large scale cross-stratification within the tabular beds is conspicuous. Sets of cross-strata are a few tens to several hundreds of feet in length. All examined by me dip eastward; most dip 18°–20°, but dips as low as 7° and as high as 22° have been noted.

Such bedding features characterize all except the lowest parts of most sections. Where the member is more than 400 feet thick, the lower 80–175 feet may be made up of tabular beds 3 inches–10 feet thick, rather than tens of feet thick; these beds are cross stratified on a medium to small scale. In some basal sections wedge-shaped beds are abundant. Where the member is less than 250 feet thick, wedge-shaped beds and small-scale cross-strata are seen only in the lowest few feet of the section. In a few exposures along the canyon of Cherry Creek, particularly in the north half of the McFadden Peak quadrangle, the lowest arkose is notably different. In that area, channels, generally broad and shallow but in places narrow, steep-walled, and deep, were eroded into the basalt and in places several tens of feet below it. In most channels the arkose is poorly sorted, sparsely pebbly, and devoid of well-defined bedding structures. Locally these conglomeratic arkoses are 100 feet thick.

This arkose member has been mistaken for the arkose member of the Dripping Spring Quartzite; in hand specimen the two are somewhat alike, but in outcrop they are entirely different. Actually, specimens of the Dripping Spring arkose differ in being orange pink or reddish orange, in being somewhat less uniform in texture, and in breaking with a quartzitic fracture. Owing to erosional etching or well-defined partings along the inclined planes, the large-scale cross-stratification of the Troy arkose is everywhere conspicuous, whereas that of the Dripping Spring arkose is small to medium scale and obscure. The basal conglomerate differs from the Barnes Conglomerate Member mainly in that it contains some fragments of Mescal Limestone and basalt.

CHEDISKI SANDSTONE MEMBER

A white sericitic sandstone, which forms rounded slopes surmounted by knobs and hoodoos bizarrely sculptured by erosion, was recognized by Burchard (1931, p. 54, 56–57) as a distinctive unit that separates the principal iron deposits of the Mescal and quartzites of the Troy. Without additional definition, he termed this unit the Chediski White Sandstone Member of the Troy Quartzite. The type locality was designated as Chediski Mountain, a high flat-topped ridge on the west side of Canyon Creek, 23 miles north of the Salt River. The member forms an almost horizontal characteristically conspicuous outcrop around the east end of this ridge. (The member derives its name from this outcrop: "chediski" as used locally by the Apache means long white rock.) South of the Salt River the member is locally quartzitic and is neither so distinctively white nor so spectacular in its erosional forms; it therefore has not been distinguished previously as a separate member. Less obvious features do set it apart everywhere, however, and under the designation Chediski Sandstone Member it is here recognized as the most widespread subdivision of the Troy. It is the basal member, as Burchard specified, except where it is underlain by the arkose member in northwestern Gila County.

THICKNESS

In a general way the Chediski Sandstone Member thickens from north to south. North of Young it thins by lapping out; and in its northernmost exposures along and near the Naegelin Rim, it is only 75–100 feet thick. Between the latitude of Young and that of Globe, it is about 250 feet thick. Deviations reflect irregularities in the pre-Troy surface: where the surface truncates uparched Apache strata along the south rim of the Natanes Plateau northwest of Sawmill, for example, the member is less than 200 feet thick. South of Globe the member commonly is directly overlain by Paleozoic strata, and many sections are obviously incomplete; nevertheless, thicknesses of 250–300 feet are common, and several occurrences are much thicker. Sections 2–4 miles east of Ray are at least 500 feet thick, and in the Holy Joe Peak quadrangle some sections may exceed 700 feet.

LITHOLOGY AND STRATIGRAPHY

The features described below are typical of outcrops north of the Salt River; the less distinctive southern facies are described later.

Three subdivisions are characteristic of the Chediski Sandstone Member. Although variable in thickness they are easily recognized by differences in bedding features. More subtle are the lithologic differences, which reflect composition and manner of occurrence of the coarsest constituents. The upper unit crops out in well-defined ledges and cliffs in which cross-stratification is obvious, and it is consistently characterized by thin layers or lenses of small pebble conglomerate. The middle unit is largely free of granules and pebbles and is characterized throughout by convolute lamination. In areas of high relief this is the part of the member that eroded to fluted cliffs and hoodoos; in areas of low relief rounded massive outcrops are typical. The lower unit is also massive but differs from higher strata in being sparsely and erratically pebbly and devoid of well-defined bedding structures. Generally there is a ledge-forming conglomerate layer at the base of the lower unit, and the rest of the unit forms a gentle concave slope partly hidden in debris. Where the Chediski exceeds 200 feet in thickness, the crossbedded unit makes up the upper third of the member, and the lower unit is 25–60 feet thick. Everywhere in northern occurrences the middle unit is the thickest of the three. In sections appreciably thinner than 200 feet, it commonly comprises all except the uppermost and lowermost 10–15 feet of the member.

Distinctive colors also cause the member to contrast with overlying and underlying strata. In hand specimens the sandstone is mostly pinkish gray; viewed from a distance on a bright day, outcrops of this hue have so white a glare that the casual observer tends to overlook lesser details essential to regional recognition of the member. Commonly the lowest 10–80 feet of the member—that is, part, or all, of the lower unit and sometimes part of the middle unit—is grayish red to grayish reddish purple. Parts of the red sandstone are mottled or streaked by reduction spots. In some sections the mottled parts are light brownish gray to light greenish gray, but in most they are almost white. The coarser grained strata or the boundaries between layers of different grain size are mottled in preference to the rest of the rock. In several areas the lower unit is largely devoid of red-purple coloration, and the lower part of the middle unit retains that color. Thus an irregular but conspicuous band of color 20–40 feet thick may roughly mark the boundary between the two units.

The member consists largely of very friable to moderately well cemented very poorly sorted porous sericitic sandstone. Subrounded to well-rounded pitted and frosted clear quartz grains, ranging in size from fine sand to granule, make up most of the rock. The matrix consists of sericite and a whitish clay; the sericite is usually the more abundant. Cleavage fragments of orange-pink microcline or orthoclase are very sparse; clay aggregates that have the outlines of such cleavage fragments occur somewhat more abundantly. The feldspar grains are generally only a fraction of the size of the quartz grains, but in the coarsest grained beds where feldspar grains are the most numerous some may be granule size. Although much of the rock appears devoid of feldspar, except under the microscope, chemically the sericitic sandstone is probably equivalent to a moderately feldspathic sandstone.

In general as the grain size of the sand increases, the clay or sericitic matrix decreases, and the sandstone is more firmly cemented by quartz. The crossbedded upper unit is generally the only part of the member that is firmly cemented; but commonly even the lower part of this unit is very friable. In some sections a bed or two of quartzite can be found near the top of the unit. In places the voids in a coarse-grained lens in the middle unit are completely filled with quartz. Where the matrix is sparse, clusters of minute needle-like quartz crystals—visible only under a microscope—form meshworks between grains; these needles are not attached to the clastic quartz grains. Indeed, except in the minor occurrences of quartzite and in the firmly cemented sandstones just noted, individual sand grains do not exhibit overgrowths of quartz.

Lithologic details, considered in stratigraphic sequence, provide additional criteria for delineation of the three units and are especially useful in regional correlation.

Where the Chediski is the lowest member of the Troy, the basal 2–40 feet is pebble to cobble conglomerate or conglomeratic sandstone. Generally the conglomerate is less than 20 feet thick and is a sharply defined ledge-forming layer that does not grade into the overlying pebbly sandstone. The gravels of these conglomerates are mostly quartzite and quartz derived from the older Precambrian rocks; pebbles of basalt and chert from the Mescal and angular fragments from the argillite member of the Mescal are sparse. Locally the gravels are almost monolithic, especially where the conglomerate overlies a terrane of thoroughly silicified hematitic Mescal. Where the Chediski overlies the arkose member, the basal few feet may be conglomerate or erratically pebbly sandstone, or may be practically free of pebbles. Detritus from Apache strata is virtually absent from these occurrences. The contact between the two members is sharp, and though small channels and truncated cross-strata in the arkose member obviously indicate erosion, debris from the lower member has not been recognized in the Chediski.

The sandstone of that part of the lower unit immediately above the basal conglomerate is generally sparsely pebbly, but lenses of granule, pebble, or in a few places even cobble conglomerate do occur in this interval. Unit 5 of the Center Mountain section, which happens to be an occurrence lacking the basal conglomerate, is typical of many such intervals. The vague bedding features are defined almost entirely by thin discontinuous lenses of very coarse sandstone or conglomerate, which suggest an irregular horizontal stratification. Lenticular beds of cross-stratified sandstone do occur sparsely. The pebbles and cobbles mostly are fine-grained vitreous quartzite. In some sections quartz pebbles predominate; part of these differ from quartz pebbles higher in the section in being shades of red or brown, rather than white. Some gravels, and especially the largest, show polished facets and the forms characteristic of windworn pebbles or ventifacts (Bryan, 1931, p. 29–36). In some sections these ventifacts make up 5 percent of the pebbles more than 1 inch in diameter; in most they do not approach this abundance.

Because soft-rock deformation caused the mixing of layers of different grain size and obliterated textural characteristics, it is difficult to be specific about the degree of sorting represented in the middle unit. Perhaps sands of certain layers were once moderately well sized. As a unit it is not well sorted, but overall it is much better sorted than the adjacent units.

Apparently thin to thick beds, individually dominated by either fine or medium sand, made up most of the massive sets of strata as each was deposited. Many thin lenses or tabular layers of coarse sand, not as well sorted as the finer sands, were intercalated in these sets in most localities. There are occurrences, however, in which such layers are few. The upper half to three-fourths of the middle unit is free of gravels, but the lower part contains pebbles and cobbles distributed as erratics without apparent relation to original bedding features. In one outcrop small pebbles are scattered at intervals of 2–5 feet but nearby, in outcrops of equivalent beds, only cobbles 15–50 feet apart are seen. The sandstone of this unit differs considerably from that above and below in lacking layers or lenses of very coarse sand and granule conglomerate and in including relict masses suggestive of layers that were once moderately well sorted.

Internal structures of the forms collectively termed convolute lamination or convolute bedding (Ten Haaf, 1956) mark all massive tabular layers (sets(?) of beds), which make up most of the middle unit. The layers of contorted sandstone, mostly more than 10 feet thick, are defined by inconspicuous horizontal planes; a few are separated by cross-stratified beds, 1–2 feet thick, that are tabular and sharply defined top and bottom. Convolute lamination is absent only from the well-defined thin beds. The convolute lamination is obscure except where outlined by layers of contrasting grain size. In a small proportion of outcrops, the structures are everywhere conspicuous, because textural differences in adjacent layers are accentuated by differential cementation or by color mottling if the sandstones are red. The most abundant structural forms are contorted wispy laminae or swirl-form mixtures of sands that occur as though separate layers had been gently stirred together. The coarse sand layers that survived plastic deformation as crude lenses are more striking, because they project prominently from outcrops and are the most continuous features. Many such lenses are saucer- or bowl-shaped in cross section; others are intricately thickened and thinned and contorted. Where such marker layers were displaced no more than a foot or two, saucer-shaped lenses thinned to form sharp narrow "anticlines" where their "rims" meet, and these may be alined laterally to give a scalloped effect in cross section. Less common are isoclinally folded laminae that occupy the full thickness of tabular layers, some as much as 8 feet thick. Folds of this and the scalloped type persist uninterrupted as far as the enclosing layer can be traced—in some places, several hundred feet.

The convoluted structures were formed mainly while the individual massive layers were being deposited. Truncation of the internal convolutions at the planar surfaces shows that a sand layer was deposited, deformed, and then eroded before the next similarly marked layer was laid down. If the internal corrugations had formed as a consequence of submarine sliding on a depositional slope, irregular thinning or thickening of the layers or the piling up of disrupted strata within wedge-shaped masses might be expected. No features suggestive of this or any other sort of extensive lateral slump have been noted. The separating planes are continuous for considerable distances, the indication being that after a set of beds crumpled, there generally was no additional deformation. Locally and at widely separate places, the thin cross-stratified beds and separating planes are obliterated by slumping. Thus, in one outcrop, layers of contorted sandstone may be separated by planes 10–20 feet apart, whereas in an adjacent outcrop of the same interval, well-defined planes are as much as 50 feet apart. Such relations are uncommon, but they do suggest composite causes for some apparent slump structures. Perhaps gross slump effects were superimposed on smaller scale convolutions that were entirely the effects of earlier intrastratal crumpling and stretching of very plastic sand layers.

Rounded outcrops, in which several layers differ little in texture and in which the slump structures are inconspicuous, may be additionally marked by peculiar patterns of V-shaped furrows. These furrows or etched-out surficial joints in some places penetrate the outcrop to a depth of 4 inches. The furrows generally are in two sets with about equal spacing; together the sets define a reticulate pattern, so that the outcrop simulates a rock wall built of crudely squared but tightly matching sandstone blocks. One outcrop may have furrows 6 inches–1 foot apart; another may have furrows 2–3 feet apart. In a few outcrops the blocks have five or more sides, rather than four.

The boundary between the middle unit of contorted sandstone and the lower unit is poorly defined; that between the contorted sandstone and the upper unit is sharp and apparently represents a surface of intraformational erosion.

The upper unit of the Chediski Sandstone Member is readily recognized in most localities by tangential cross-stratification that etches out in bold relief. The beds are grossly tabular, but if traced far enough many prove to be wedge shaped or lenticular. In some areas the beds are 3–10 feet, in others 10–20 feet thick. Dips of cross-strata exceed 10° and commonly approach or slightly exceed 20°. No dominant direction of dip has yet been noted.

Where crossbedding is not conspicuous in the upper unit, as in unit 8 of the Center Mountain section, the composition and disposition of the small gravels serve to distinguish it from all other Troy strata. Granules are possibly as voluminous as pebbles. Most pebbles are less than three-quarters of an inch in diameter; some are as much as 2 inches across. The gravels are well rounded and are mostly of white or pink quartz, with a few being reddish-brown jasper. The beds are coarse to very coarse grained, and perhaps half of them are sparsely pebbly throughout. Gravels become more abundant and coarser upward both in individual beds and in the section. In the top 25–35 feet of the upper unit, the top 2–6 inches of almost every bed is conglomerate. The uppermost 1–6 feet of the member is everywhere a conglomerate or a conglomeratic sandstone. The positions of the pebbles and their increase in size and number upward indicate that the finer constituents in the top parts of each bed were winnowed away—leaving behind only the coarser fractions of what was originally a much thicker layer—and that such winnowing was increasingly more effective and prevalent as successive beds were deposited.

LATERAL VARIATIONS

North of the Salt River, lateral variations are mainly in thickness; lithologic details do not differ enough to hinder immediate recognition of the Chediski Sandstone Member or of its three units. South of the river, lithologic variations, which are mainly manifestations of slight changes in the makeup of matrix materials and in the degree of cementation, overshadow and obscure the stratigraphic entities so apparent farther north. Thus, detailed regional correlations have not been attempted previously. A quartzitic facies, generally considered the facies typical of the entire Troy, actually is not important regionally. The lateral variations in occurrence and composition of the gravel-size constituents and variations in bedding structures are the significant modifications to be interpreted in making regional correlations. The tripartite subdivision of the member is not as sharply delineated but is still valid.

Compositionally the sandstone of the southern part of the region is akin to that of the northern part. The matrix is dominated by sericite and clay, but in a general way these constituents decrease in amount southward, especially in the upper half of the member. In some localities coarse muscovite, which is uncommon in the northern facies, fills interstices between grains and is abundant in a few beds. Orange-pink feldspar is more abundant in the southern facies, and in the coarse grained beds compound fragments of feldspar and quartz, obviously derived from granitoid rocks similar to those that underlie the Apache Group, are locally conspicuous.

As the content of clay decreases southward, the sandstone becomes more firmly cemented by quartz, and the Chediski becomes almost entirely quartzite over areas of a few square miles and partly quartzitic over even broader areas. Nevertheless, if emphasis had not been placed on occurrences in the vicinity of the abandoned mining camp of Troy, near Ransome's type locality, and in and near the northeastern part of the Globe quadrangle, where Ransome apparently first noted extensive outcrops of the Troy, the Troy might have been designated a sandstone. The northernmost exposures in which the whole member is quartzitic extend from Cibecue Creek east along Salt River canyon and southeast across the western part of the Natanes Plateau. There the three units of the member are obscure and large parts of the section are crossbedded massive red quartzite comparable with that previously emphasized as typical of most of the formation (Ransome, 1916, p. 139–141 and pl. 27*A*). Large parts of the member are vitreously quartzitic and except for feldspar and pebble contents could be mistaken for the overlying quartzite member. Bedding structures distinctive of the Chediski, such as convolute lamination, are almost wholly obliterated by silicification. In perhaps 50 percent of all exposures farther south, the distinctions between members are similarly obscure because the crossbedded upper unit of the Chediski is cemented almost as firmly as the overlying member. The two lower units, however, are generally sandstone, and parts of most sections are loosely cemented and friable.

The southern sandstone and quartzite are mostly yellowish gray or light brownish gray to medium dark-gray or pale red, and in these drab hues they do not contrast with the adjacent stratigraphic units as do the white-hued outcrops of northern Gila County. Certain medium- to dark-gray beds do, however, contrast with interlayered beds that are brown or red hued, and where they occur these gray beds are distinctive of the Chediski Member. The gray sandstone owes its color to abundant minute aggregates of black hematite that partly fill voids between quartz grains. Similar gray sandstone does not occur in the northern area, although the red-hued basal sandstone of that area does owe its color to hematite. In the widespread exposures in the range of hills southeast of the Apache Mountains, the basal 50–75 feet of the member is grayish red or grayish red purple, a darker red than generally seen farther north. Such hues characterize as much as 120 feet of the member in a very few localities, but in most southern areas dark sandstone is subordinate. In only a few localities, such as the Hayes Mountains, are white outcrops as conspicuous as in the northern area.

In the northern part of the region, about the same degree of poor sorting, characterizes stratigraphic intervals 30–100 feet thick, whereas in the southern part, adjacent beds contrast because of differences in sorting. Sands of coarse-grain to granule size, in particular, tend to be segregated as lenses and beds in the southern sections.

This contrast in sorting is emphasized because gravels like those found only in the basal conglomerate north of the Salt River, as well as the ubiquitous quartz and quartzite gravels, are seen throughout. Angular to well-rounded fragments derived from resistant strata of the Apache Group are found through the lower half of the member. Lenses of gravels may also include sparse fragments of granite and Pinal Schist, which are absent from the northern sandstone. In places, such as near the mouth of Putnam Wash and farther east in the Holy Joe Peak quadrangle, many beds of the lowest 50–70 feet of the Chediski contain abundant angular fragments, as much as 1½ inches across, of orange-pink feldspar derived from the pre-Apache granite and of reddish-orange siltstone from the Dripping Spring Quartzite. Typical polished ventifacts are extremely sparse; however, some faceted pebbles have rounded edges and were possibly windworn, then modified by subaqueous abrasion. In northern parts of the region the upper half of the contorted sandstone unit is virtually pebble free; in southern areas pebbles are fewest in the upper half, and throughout the unit they are mainly confined to layers or lenses that separate pebble-free beds. In the northern outcrops nearly all the pebbles are quartz in the cross-stratified upper unit; southward, quartz is dominant, but small pebbles of other composition are also common.

Southward from the Natanes Plateau individual beds become increasing well defined. The lower 100–150 feet of the member is mainly of tabular beds ½–4 feet thick. Some beds are tangentially cross stratified on small to medium scale; others are horizontally laminated. The tops of many were channeled before deposition of the overlying bed, and lenses of conglomerate or conglomeratic sandstone, ½–3½ feet thick, may fill the channels. South of Globe a large proportion of the undulant partings between the beds are marked by gravels. Where lenses of conglomerate are absent, granule-size gravels thickly coat some bedding surfaces, small pebbles sparsely litter other surfaces, and in extreme examples cobbles as much as 8 inches in diameter are scattered very sparsely along the bedding planes. Some beds wedge out laterally, and the gravel layers coalesce. Sharply defined layers and lenses of gravels, laterally continuous though anastomosing as viewed in cross section, do not characterize northern exposures of the member.

In many areas south of Globe, the individual layers of convoluted sandstone, 8–15 feet thick, are sharply separated by gravel layers or by thin sets of undeformed beds. The middle unit does not crop out as massively, and if convolute lamination is inconspicuous, as in quartzitic sections, the casual observer may include beds of this unit with those of the lower unit and assume that the contorted sandstone interval is very thin or missing. Actually, in southern areas the stratigraphic interval marked by convolute lamination does vary considerably in thickness from 50 to at least 125 feet.

Bedding features of the cross-stratified upper unit are not notably different from those of equivalent strata in the northern part of the region, but because this unit is commonly quartzitic, they are not as conspicuous. Perhaps fewer beds are capped by concentrations of winnowed pebbles, but such pebble layers are still common enough to characterize the unit.

QUARTZITE MEMBER

The upper member is a fairly clean well-bedded quartzite, which lacks the distinctive bedding features that characterize the two lower members. The section measured at Center Mountain is one of the thickest in northern Gila County and is typical.

STRATIGRAPHY

The quartzite member can be subdivided into three units, although the differences are slight. Coarse-grained quartzites in which individual grains appear angular because the interstitial voids are only partly filled by overgrowths characterize the lowest unit; a few beds are very coarse grained. This subdivision (unit 9, reference section) is 50–80 feet thick and grades upward into the middle subdivision (unit 10, reference section), which is 150–200 feet thick and is medium grained and highly vitreous. Both units are made up of tabular beds, which range in thickness from 3 inches to 6 feet but are mostly 1–2 feet. Tangential cross-stratification on small to medium scale apparently is characteristic, but in most outcrops internal bedding details and partings between beds are obscure. The uppermost subdivision (unit 11, reference section) is thinly bedded, bedding structures etch out distinctly, and flaggy to slabby partings are everywhere present. It differs more subtly in not being as uniformly sorted as the lower units; some beds are coarse grained, and a scattering of coarse grains is common in the medium-grained beds. The beds are tabular and range in thickness from 2 to 30 inches; most are less than 12 inches thick. Some are horizontally stratified, most are cross stratified. In some outcrops the more conspicuous splitting surfaces follow cross-strata rather than planes between beds and give a false impression of undulatory bedding.

LITHOLOGY

The quartzite member differs from the Chediski and arkose members by being remarkably free of detrital grains other than clear quartz. Thin sections show well-rounded quartz grains locked together by overgrowths that, except in the lowest beds, fill the voids between original grains. A few beds include random grains of orange-pink feldspar. Granules of quartz are extremely sparse. South of Globe, feldspar and quartz granules, the quartz at places in thin lenses that include a few small pebbles, occur in the lowest quartzite unit; otherwise such constituents are practically nonexistent.

The quartzites are mostly grayish pink to pale red, although light gray or yellowish gray dominates in some localities and beds of dark grayish red or grayish red purple are conspicuous in other places. The red is due to a thin dust of hematite(?) on the original detrital grains and to minute aggregates of hematite along boundaries between overgrowths. The coarse-grained lowest unit is generally grayish red or grayish red purple, but in some localities it is very light colored; light-gray or yellowish-gray quartzite seems to be characteristic in the southern part of the region.

Weathering accentuates the red. Where the quartzites are deeply weathered, detached blocks may be stained throughout. Such extremes of discoloration particularly characterize the slabby upper unit of the member where it is exposed on benches or mesa tops. Detached slabs are stained moderate brown to very dusky red throughout, exhibit knobby or pitted surfaces, and lie in a thin red soil of earthy sand.

Generally the grayish quartzite is unmodified by weathering, but in some localities a few beds are pitted where partially cemented aggregates of sand weathered free. The pockmarks, commonly not more than an inch across, may show thin limonite stains, coatings of black oxides, or lichen growths and are very conspicuous against a background of light-colored quartzite. In different localities pock-marked layers occur in different parts of the member, so that they have little use as stratigraphic markers.

DISTRIBUTION AND TOPOGRAPHIC EXPRESSION

Thick remnants of the quartzite member are common north of the Salt River but not south, and in a general way the proportion that is remnant below Paleozoic formations can be judged from the topography. The two lower units commonly form a steep ledge-studded slope or a single cliff. The contact between the middle unit and the slabby upper unit is marked by a broad topographic bench. Because the upper 30–40 feet of the middle unit is somewhat less resistant than the underlying quartzite, in many areas there is a steplike transition zone of ledges between the top of the cliff and the bench. The lower part of the upper unit forms a steep slope that merges downward into the bench and upward into ragged cliffs. The lower 25 feet is largely covered by flaggy and slabby talus from higher beds. Where the upper unit is 150–200 feet thick, as in the higher parts of the Sierra Ancha, the quartzite member is readily recognized at a distance of 2–5 miles by its distinctive profile of two prominent cliffs separated by a steep slope and bench. Elsewhere north of Globe the slabby unit is thin or absent, and only the lower cliff typifies the member. Farther south, Paleozoic formations commonly rest directly on the Chediski Member; if thin remnants of the lower unit of the quartzite member do intervene, they do not crop out distinctively from the Chediski.

DISTINCTIVE FEATURES OF MEMBERS

In most areas north of Globe, the boundary between the Chediski and the quartzite member is a sharp easily recognized contact between sandstone and quartzite. Where strata below this contact are quartzites, as in many of the southern occurrences, the contact may seem rather vague. Although the large-scale cross-stratification of the upper beds of the Chediski does distinguish them from the overlying quartzites, in sections of vitreous quartzite or sections much shattered by faulting, bedding structures are not easily discerned; even convolute structures of much lower strata may be overlooked. Differences in sorting and composition do serve, nonetheless, to distinguish the members. The upper part of the Chediski, even where it is a vitreous quartzite, includes some clay or sericite between grains or includes appreciable feldspar, whereas the lower unit of the overlying member is free of such constituents, except for sparse feldspar in a very few beds. The upper beds of the Chediski are very poorly sorted, and most beds are pebbly. Comparatively, the coarse-grained lower unit of the quartzite member is much better sorted. A conspicuous conglomerate or conglomeratic quartzite or sandstone is the uppermost bed of the Chediski; no comparable bed occurs in the overlying member. Where the conglomerate zone locally is absent, in the southern part of the region, one of the lower, thinner, and less persistent zones might be mistaken for it. Even so, the highest fairly persistent pebbly layer in a sequence of quartzite probably can be taken as the contact between members, without being in error more than 20–30 feet stratigraphically.

DIABASE AND RELATED ROCKS

FORM AND DIMENSIONS OF INTRUSIONS

Through much of the region, diabase sills and dikes were emplaced in the Apache Group almost to the exclusion of other formations, but in places extensive bodies also intruded the older Precambrian formations and the Troy Quartzite. Paleozoic formations rest unconformably on the diabase. Figure 3 serves well to show that the larger intrusions of diabase are singularly coextensive with the younger Precambrian sedimentary rocks. Small intrusions and even sills hundreds of feet thick, but narrow in outcrop, are omitted; so the map does not show how virtually every sedimentary outcrop is somewhere interrupted by diabase bodies. The bodies that are shown separate from Apache and Troy exposures were mostly intruded into older Precambrian formations only slightly below the Apache; some are exposed in windows eroded through younger formations. None show structural or other features that would cause them to be categorized as different in age, and all show petrographic details peculiar to the intrusions in the Apache rocks.

The sills, which are virtually horizontal in the Colorado Plateau and which generally dip less than 25° in the Basin and Range region, tend to erode into broad concave slopes, whereas the sedimentary host rocks tend to form cliffs. Therefore, outcrops of diabase are conspicuous in most landscapes where younger Precambrian formations are exposed. Indeed they may dwarf outcrops of the cliff-forming rocks, as in areas where the aggregate thickness of diabase sills is as great as the combined thickness of Apache and Troy strata.

Sills predominate in volume, and range in thickness from a few inches to 1,200 feet. Thicker sills have been reported, but these are almost certainly sheets that comprise two or more separately injected sills. The sills are remarkably persistent. Some only a foot thick extend laterally more than 1 mile, and several a few hundred feet thick extend at least 20 miles.

Dikes are numerous but insignificant in volume. Perhaps some are as much as 100 feet wide, but most range in width from a few inches to a few feet. A few dikes 10–30 feet wide can be traced several miles.

Though sills—by definition—are tabular concordant intrusive bodies, all the Arizona sills are locally discordant, and this is especially apparent in the principal diabase masses (fig. 14). Concordant sheets at different horizons are connected by discordant tabular bodies, which are as much as 1,000 feet wide and can be traced for several miles. These connecting elements generally do not extend above the upper contact of the higher concordant body or below the bottom of the lower body, and they should be visualized as the discordant parts of a stepped sill rather than as dikes. (See feature C, fig. 14.) Many sills change horizon by high-angle abrupt steps up or down across the host strata; less obvious is the gently undulant contact that transects the sedimentary layers of the host formations at a very low angle. Most discordant parts of sills are not simple tabular bodies but are stairlike masses bounded in one place by high-angle discordant contacts and in another by low-angle or concordant contacts. Depending on the attitudes of the high-angle "riser" and of the low-angle "tread," some steps are consistent in height for miles along the strike, whereas the height of others varies considerably in a short distance. Lest reiteration of discordant features be misleading, it should be emphasized that only a small proportion of the diabase of each sill is encompassed between discordant contacts.

Because a cover of younger formations and post-Paleozoic deformation obscured intrusive relations in mining districts of the Basin and Range structural province, some misconceptions exist concerning the forms of the diabase bodies. There has been a tendency to visualize the sills as masses that bulge and pinch irregularly and have bulbous or irregularly ragged terminations. Such outlines have also been considered typical of discordant sections of sills. The structural features seen in extensive cliff and canyon exposures in the Colorado Plateau and in the better exposures farther south wholly refute such interpretations. Actually, the diabase bodies are so regular in outline that they can be projected in cross sections with almost as much confidence as can the layered sedimentary rocks.

Most contacts of sills, including those that bound the stepped discordances, are virtually planar and parallel; they circumscribe masses of diabase that are distinctly tabular. (See fig. 14.) Even the terminations tend to be planar, because many sills end abruptly with chilled contacts against high-angle faults. Such terminations have been noted for a few sills that are as much as 800 feet thick and are common for sills less than 200 feet thick. High-angle connections between concordant parts of sills were also emplaced along faults. Most discordant and concordant boundaries join abruptly in an angle, rather than along a rounded or irregular surface.

Because the separate concordant parts of a given intrusion emplaced at horizons hundreds of feet apart are not exposed in some areas and are entirely eroded in other areas, the extent and the shape in ground plan of individual thick sills is largely indeterminate. Thick sills do not seem to thin significantly by wedging. The parts at different horizons may differ in thickness, and sills may thin by steps, as shown between points B and C of figure 14; but each concordant element of a sill is very uniform in thickness through areas of as much as a few tens of square miles. Some thin sills that are stepped apophyses from still thicker sills do wedge out, but the wedging is gradual. Near the edges of these apophyses, the contacts tend to become undulant, and concomitantly the sills thin rapidly to less than 100 feet, and many then terminate abruptly in a narrow discontinuous dike that marks a high-angle fault. From the habit of such minor elements and the planar terminations of some thick elements, it seems likely that large parts of sill perimeters are marked by abrupt terminations. Such apparent or actual terminations are linear in plan, and some do not deviate for several miles, except for small steplike irregularities.

Large tabular plates of sedimentary rock, common in many areas and seemingly isolated in the diabase, provide insight into the mode of intrusion. In the exceptional exposures of the Salt River canyon, the relations of representative plates of Mescal Limestone and Dripping Spring Quartzite can be deciphered unequivocally. These plates are 5–300 feet thick, and some are at least 2 miles long and ½ mile wide. They are separated from roof or floor of the main host by layers of diabase a few feet to several hundred feet thick. Few of the larger plates can be demonstrated to be entirely separate, but many a few hundred yards across are definitely isolated in diabase. Some plates wedge out, whereas others end abruptly. The bedding within most is parallel to the strata above and below the diabase. Where a plate is terminated discordantly, the truncated beds match perfectly in thickness and sequence the truncated beds on the opposite wall of the discordant mass (fig. 10). The complete lack of even small-scale assimilation by the diabase can be demonstrated in many examples. In some canyon walls only one plate is seen; in other localities several plates are separated by layers of diabase.

The diabase masses over and under the individual plates were injected at different times. As shown diagrammatically in figure 10, a simple sill first intruded the section. A second intrusion followed the same general path but locally deviated and pried away a plate or plates of sedimentary rock, which therefore appear suspended in the composite layer. The sequence of intrusions is generally recognized only from the interrelations of chilled borders; in places an early sill is obviously crosscut or displaced by a later sill. In places tabular segments of an intrusion intermediate in age occur between sills in the same manner as plates of sedimentary rock. Individual intrusions invariably completely crystallized before the next one was injected. Excellent exposures of multiple intrusions along a single irregular plane and of chilled contacts between sills are readily inspected along the part of U.S. Highway 60 that traverses the north wall of the Salt River canyon. Actually, it is uncommon that one chilled border will be welded tight to the border of an adjacent sill. Instead a thin seam of hornfelsed sedimentary rock, easily overlooked, separates adjacent sills. Where the diabases are altered and chill zones are obscure, the recognition of these seams can aid in deciphering the geometry of the sills.

The general lack of warping, or of wedging phenomena, even in the thinner plates of incompetent strata and the lack, along discordant contacts, of breccias or folds that would suggest intrusive drag indicate that the magma was introduced passively. Certain noteworthy small-scale faults and folds, considered in the section on structure, did originate with the inflationary process. These in no way invalidate the above conclusion; indeed, the occurrence of delicate tenuous tongues of diabase along some of the structures enhances it.

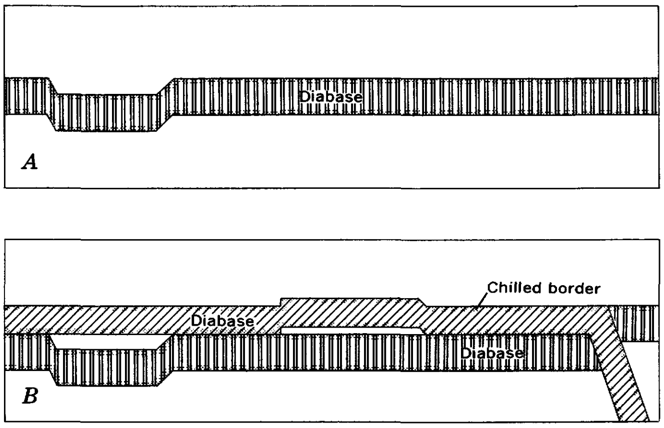


FIGURE 10.—Relations of multiple injections of diabase to plates of host rock isolated between intrusions. *A*, First locally discordant diabase sill. *B*, Second sill intruded generally along same plane as first but locally discordant, so that large plates of strata remain between sills.

The large plates can in no way be confused with true xenoliths. Furthermore, the average sill can be traced for miles without xenoliths being seen. In the few localities where xenoliths do occur, they are confined to the margins of intrusions, and most of them are measureable in inches. The largest that has come to my attention is 4 feet thick and 15 feet long. The large plates are so different and so consistent in habit that they can be considered everywhere as proof of multiple intrusions.

Recognition that the intrusions are typically multiple has practical aspects for geologists concerned with the geometry of sills in the mining areas. Certain diabase bodies and related features in the Magma mine at Superior, for example, have long been an enigma (Short and others, 1943, p. 37; Sell, 1960). Of two sills exposed by the mine workings, the upper one is said to be 2,000 feet thick, and the two are reported to have a combined thickness of more than 3,000 feet. Described as xenoliths within the 2,000-foot sill "are numerous isolated bodies of quartzite, which range from a few feet to 300 in thickness. * * * one of them extends * * * for a distance of 3,000 feet." It is stressed that these xenoliths "have the same dip and strike as the main mass of quartzite from which they are separated" and, that, aside from other considerations, "this attitude is difficult to understand if the blocks sank into a molten magma." Descriptions of the extent and attitude of the xenoliths, considered together with the known relations of such sedimentary blocks elsewhere and the complete absence of sizeable xenoliths within any single diabase intrusion, provide the basis for an explanation. Almost certainly the larger bodies of quartzite exposed in the supposedly simple thick sill of the Magma mine are blocks or plates that locally separate two or more sills intruded at different times. The lack of any particularly thick sill in nearby surface exposures, which must represent the same complex of diabase intrusions seen underground, tends to support this interpretation. Because formations of the area are intricately broken by Tertiary faults, many of the smaller so-called xenoliths could be fault segments offset from the principal inclusions between separate sills.

If this 3,000-foot sequence of diabase is a composite unit, it is the thickest one known in Arizona. In review of this sequence and of other thick sheets, however, no single intrusion thicker than the 1,200-foot sill beneath Aztec Peak in the Sierra Ancha has been unequivocally confirmed.

In one area the overall habit of the diabase intrusions may be considerably different from that in another. The extremes of habit are fairly well represented by intrusions of the Sierra Ancha and by intrusions in and near the Salt River canyon: (1) Multiple injections are characteristic of the canyon but are almost nonexistent in the Sierra Ancha, (2) the ratio of diabase to sedimentary rock is about 1:1 in the canyon, or perhaps 2–4 times that of the Sierra Ancha, (3) the representative sill of the Sierra Ancha is probably much thicker than the individual sill of the Salt River area but is not greatly different from the composite sheets of the latter. Apparently corollary, (4) in the Salt River canyon the formations especially susceptible to diabase intrusion were inflated much more intricately and at more horizons, (5) although a few steps of several hundred feet have been observed, most sills do not deviate more than a few tens of feet from a given stratigraphic horizon in outcrop intervals of 5–10 miles, and (6) the many dikelike bodies that connect concordant parts of sills are mostly thin and of no great extent, whereas several connecting bodies in the Sierra Ancha are more than 500 feet wide and 2–10 miles long and traverse 500–1,500 feet of section. Comparable contrasts in habit have been noted farther south in the region.

STRATIGRAPHIC DISTRIBUTION

Through some understanding of the characteristics of strata most commonly split by sills, supplemented by knowledge of the horizons preferentially inflated —perhaps because of some local factor—in immediately adjacent areas, moderate success is possible in predicting where a discordant step in a sill will feed into a concordant segment. Particular application is afforded in predicting sites of unexposed asbestos deposits, since they are contact metamorphic phenomena adjacent to diabase. Such details could be used more widely to interpret displacements of the sedimentary rocks.

Sills are commonest and most consistent along horizons that separate layers of markedly different competencies. The extensive sills in the Mescal, for example, are at or near the contact between the massive algal member and the thinner bedded less competent lower member. Intrusions that are not at the contact between competent and incompetent strata are mostly within a few feet of the contact and in the incompetent unit. Sills in the Dripping Spring split the siltstone units of the siltstone member—within a few feet of the more competent quartzitic arkose units—or are along a horizon not far above the contact between the siltstone member and the massive arkose member. As a corollary, sills also are common along all unconformities in the Apache Group except the one between the Mescal and the Dripping Spring. The stratigraphic units that were particularly susceptible to intrusion are also those that contain the thickest intrusions.

Massive units of the younger Precambrian sequence are in few places hosts for concordant sills. These units are the arkose member of the Dripping Spring, the founder breccia of the lower part of the Mescal Limestone, the karst breccia that makes up the Mescal in some areas, and the arkose member and Chediski Sandstone Member of the Troy. Dikes and discordant parts of sills do cut across these units but are not so numerous as in the units that are hosts for concordant intrusions.

In most areas, the Mescal Limestone was intruded along more horizons and contains a greater amount of diabase than do other formations. The sills may be thick or thin. The Pioneer Shale was intruded by one or two sills, which are thin but persistent for long distances. Locally the siltstone member of the Dripping Spring is host to sills, but these are not so extensive as in the Mescal or Pioneer; over many areas of several tens of square miles, no intrusions occur in the formation. Through much of the Sierra Ancha southeast of McFadden Peak, one sill, 500–1,000 feet thick, splits the siltstone member. This is atypical; elsewhere in northern Gila County one or two sills, which are less than 200 feet thick, inflate the member; south of the Apache Mountains, sills that are 200–500 feet thick are common.

Sills are almost ubiquitous along or adjacent to the unconformity between the Troy Quartzite and the Apache Group; their configuration depends on the character of the rocks beneath the unconformity. Where the Troy directly overlay the massive unit of the algal member of the Mescal or was separated from it only by basalt, one thick sill was typically intruded; where the argillite member intervened and the thin-bedded upper unit of the algal member was remnant, two to five thin sills intruded the upper part of the Mescal.

Because even discordant intrusions are very sparse in the Chediski Sandstone Member, which makes up most of all Troy exposures, there has been argument as to the existence of diabase in the Troy. Actually, sills as much as 200 feet thick are common in the quartzite member, and the thick sill at Aztec Peak happens to be in the upper slabby-splitting unit of the member. Sills would be seen more widely in the Troy if this unit were extensively remnant.

Sheets of diabase 100–400 feet thick are common, and in some areas widespread, in the upper part of the older Precambrian terrane. In the granites these sheets are fairly regular in thickness and were emplaced along joints that had formed parallel to the pre-Apache surface. In the layered and schistose rocks of the older Precambrian, the diabase intrusions tend toward the same habit but are more irregular in outline. At depths of more than 500 feet below the base of the Apache Group, extensive diabase intrusions are practically nonexistent. Therefore, where broad exposures of granitoid or schistose formations do not include diabase, it seems reasonable to postulate that the Apache plus a considerable thickness of the older formations has been eroded and that remnants of the Apache Group are not likely to be found in the vicinity.

PETROLOGY

The normal diabase is a fine- to coarse-grained medium- to dark-gray holocrystalline rock of ophitic or subophitic texture. The aspect that distinguishes this rock from other dark-gray intrusive rocks of southeastern Arizona and causes it to be notably different from most diabases is its coarse-grained facies. Though definitely subordinate in volume to the medium-grained rock, the coarse facies makes up conspicuous parts of most bodies and, in places, whole sills. These coarse elements might well be termed gabbro if they were not prevailingly subophitic in texture. Feldspathic differentiates or the so-called red-rock types—diabase pegmatite, aplite, and granophyre—make up only a minute part of the diabase bodies but are conspicuous because of their light color and distinctive textures.

The diabase weathers light olive gray to moderate yellowish brown and can be distinguished at a distance by the characteristic greenish-gray or olive-gray hue of barren slopes. The larger poikilitic grains of pyroxene characteristically protrude on weathered surfaces, which therefore appear knobby or warty. Spheroidal weathering to rounded boulders is common. The rock disintegrates to a light-brown or yellowish-brown granular soil that commonly includes abundant kernels that also represent the ophitic aggregates.

Much of the diabase disintegrates rapidly. Some newly blasted outcrops, first observed as tough and very firm rock faces, crumbled to sand during periods ranging from 3 to 10 years. The tough knobby outcrops that disintegrate very slowly are indicative of diabase which has had a minimum of deuteric alteration. In many localities the only resistant parts of the outcrop are conspicuous light or dark ribs, mostly less than a quarter of an inch in width, which project as much as 2 inches above the weathered surface of the diabase. The light, commonly pink ribs are concentrations of albite along joints, or minute fractures, in the diabase; the dark ribs are concentrations of hornblende, biotite, and chlorite. These ribs mark diabase that has been much altered deuterically.

Very fine grained chilled borders are characteristic of the normal diabase. These selvages are in few places more than 3 feet thick, and most are 6–18 inches. The effects of chilling on grain size are commonly reflected through an additional 15–30 feet of sill borders. Very dense borders, in which only scattered microlites of plagioclase and magnetite can be recognized, are restricted to intrusions less than 2 feet thick.

Most of the normal diabase is composed largely of labradorite and clinopyroxene; olivine may be absent, but it generally occurs in sufficient amount that the rock can be characterized as olivine diabase. A texturally similar but subordinate facies consists mainly of albite and either pyroxene or amphibole. All variations between olivine diabase and albite diabase have been seen in one sill.

DIABASE

Thin sections of fresh olivine diabase show that labradorite (An_{55}–An_{70}) composes 45–65 percent of the rock; clinopyroxene, 10–25 percent; olivine, 5–20 percent; biotite and hornblende, each 1–5 percent; and ilmenite-magnetite, apatite, and sparse sphene, a total of 2–8 percent. Subhedral plagioclase laths range in length from 1 to 4 mm. Commonly the outer third of the crystals is normally zoned. Grains of colorless or slightly brown or pink augite, apparently the only pyroxene in much of the rock, range in diameter from a fraction of a millimeter to 40 mm and are commonly 5–20 mm. A small part of the clinopyroxene in some thin sections is clear pigeonite ($2V < 30°$). A little hypersthene has been observed in some specimens. Olivine occurs as small rounded grains or clusters of rounded grains. A composition of about $Fo_{60}Fa_{40}$ is indicated from refractive indices of a very few specimens by the method of Deer and Wager (1939, p. 21). Strongly pleochroic yellowish-brown to reddish-brown biotite is widely distributed, but in few places abundant. Pyrite is sparsely distributed through most of the normal diabase, and in some specimens chalcopyrite can be identified.

Much of the olivine diabase shows late magmatic or deuteric alteration. Olivine is replaced by chlorite or by serpentine that includes the usual fine granules of magnetite. Olivine was also converted to talc, iddingsite, or bowlingite. The pyroxenes were replaced mostly by green hornblende and chloritic minerals; locally stilpnomelane or tremolite are abundant. The hornblende appears dark green or black in hand specimens. The reddish-brown biotite, seemingly an early mineral in the fresh rock, at least in part is a replacement of both pyroxene and olivine in the altered rock. Green biotite is abundant in diabase so thoroughly altered that the primary mafic minerals are difficult to identify. Epidote occurs only in a few localities, where it replaces both mafic minerals and plagioclase. Cores of labradorite laths commonly were converted to fine-grained aggregates of sericite, clay minerals, and chlorite and in places were converted to prehnite. Rims became turbid; in some specimens these turbid rims are mostly albite. Ilmenite-magnetite is especially conspicuous in the coarser facies because it is there partly altered to leucoxene.

Albite diabase consists essentially of albite, augite, and hornblende. The dominant mafic mineral may be augite only moderately altered to hornblende or may be a fibrous gray-green hornblende. The fibrous hornblende is plumose in habit and strongly pleochroic from yellowish green to blue green, and thus is very different from the darker hornblende which occurs more sparsely in the olivine diabase. Olivine is absent. The reddish-brown variety is the dominant biotite, but the green variety is more abundant than in the olivine diabase.

In most of the albite diabase, the albite is grayish pink to orange pink. The turbid laths or turbid rims of the grayish-pink mineral are, in a small proportion of occurrences, microperthite. Commonly, alteration products completely obscure the twinning of the albite. Prehnite is a more abundant alteration product of the plagioclase in this rock than in the olivine diabase. In some intrusions the albite is mostly white to gray; where the diagnostic pink color is lacking, albite can be surmised in hand specimen from the clouded rims and from the vague serrate boundaries which are characteristic of the albite laths.

Albite and quartz, as graphic intergrowths on a microscopic scale, occupy interstices between plagioclase laths in some of the diabase. This micropegmatite has been observed in small amounts in little-altered olivine diabase but occurs more abundantly in the albite diabase. Textures appear normal in hand specimen, but micropegmatite can be surmised in the field where exposures are lighter in color than most of the diabase. Ilmenite-magnetite occurs in extraordinary amounts with micropegmatite in some of the albite diabase, however, and causes these rocks to be almost black.

Much of the rock in which hornblende is conspicuous and several of the separate sills of albite diabase are coarser grained than the average diabase. Some is very coarse grained and includes labradorite or albite laths, 5–25 mm in length, and augite crystals, which may be slightly to largely altered to gray-green hornblende and which are as much as 50 mm in length. In some of this rock the augite tends to fracture as curved blades; many of these augite grains are twinned. Where the hornblende is obvious it is megascopically fibrous. Biotite tends to be more abundant in such rocks than in the average olivine diabase. Apatite needles are larger, seemingly more abundant. These needles and skeletal crystals of ilmenite-magnetite are readily seen by the unaided eye. Where the diabase is coarser, the ophitic texture may not be apparent in outcrops but is readily seen under the microscope. It is generally true, however, that ophitic texture is not so well formed in the coarser albite diabase and that it is lacking in some layers in which the plagioclase content is uncommonly high.

Gravity differentiation is subtly apparent in some sills of the Sierra Ancha that are more than 500 feet thick. Certain layers are more feldspathic and others are more olivine rich than the average diabase. Olivine, however, is not largely restricted to a zone low in the sill, as in the classic Palisade sill of New Jersey (Walker, 1940, p. 1067), but is dispersed through three-fourths of the body. The olivine concentrations nowhere appear high enough to account for the low silica content (43–49 percent) that seems to be general in these diabases. Olivine occurs in greatest amounts through an interval 200–300 feet thick, which is slightly below the middle of a sill. The olivine of certain layers is coarser than that in the rest of the interval. These layers and zones are not sharply defined but are gradational into the rock above and below. Nevertheless, the zone in some places forms a resistant bench that can be mistaken for a separate intrusion.

FELDSPATHIC DIFFERENTIATES

Outcrops of the feldspathic differentiates contrast strongly with outcrops of diabase because they have different textures, orange to pink colors, and general staining by limonite. Furthermore, the transition between the differentiates and the diabase of normal texture is rather abrupt. Commonly it is a zone not more than a few feet wide; in many places the transition to the larger masses of granophyre or pegmatite is only a few inches wide, and some dikes of quartz-free aplite have knife-edge boundaries with diabase.

DIABASE PEGMATITE

Diabase pegmatite is the most voluminous of the feldspathic differentiates. It occurs in two habits. As lenses or crudely lenticular dikes, a few inches to a few feet in width, or less commonly as irregular masses, a few inches to several feet in diameter, it may be found in any part of a sill, and such masses occur in many sills throughout the region. As roughly tabular masses, a few feet to tens of feet thick and tens to several hundreds or even a few thousands of feet across, the pegmatite roughly parallels and is not far below the upper boundary of a sill. The large tabular masses have been seen only in a few areas in the Sierra Ancha.

The pegmatites consist of irregular or blotchy aggregates of very coarse grained clusters of light and dark minerals. Plagioclase laths commonly are 3–6 mm wide and 1½–4 centimeters long, and some as long as 9 cm have been seen. Augite grains are 1–5 cm across and may be as long as 10 cm. Skeletal plates of ilmenite-magnetite tend to be concentrated with augite crystals. The turbid light-gray to grayish-orange-pink plagioclase is apparently all albite. Sphene is a conspicuous accessory in some occurrences. Pyrite occurs only sporadically.

The pegmatite ranges from a dark rock which contains abundant augite and ilmenite-magnetite, through a pinkish-gray rock which contains dark-greenish-gray fibrous hornblende but little other essential mafic mineral, to a yellowish- or pinkish-gray rock which is 95 percent plagioclase. A particular mass may include this entire range or be nearly uniform in mineralogy. The darkest type apparently approximates the mineral composition of the diabase that includes no olivine and no hornblende except as an alteration of the pyroxene. The variety that is probably the most plentiful is composed almost entirely of albite and hornblende.

GRANOPHYRE

Granophyre occurs mostly as irregular tabular bodies which are a few feet to a few tens of feet thick and a few tens to a few hundreds of feet in horizontal dimensions. These bodies closely parallel the upper boundaries of sills, and most are contiguous to a discordant contact. Most outcrops are pervasively decomposed owing to leaching of ubiquituous pyrite. Some of the tabular bodies of pegmatite include small masses of granophyre.

Equidimensional grains of orange-pink feldspar, intergrown graphically with quartz, give the granophyre a predominantly orange hue. Euhedral or subhedral quartz grains can be observed with the aid of a hand lens. Under the microscope, the feldspar is so clouded with hematite and limonite that identification is difficult; a large part is albite, but potash feldspar may also be present. The feldspar grains range in diameter from 1 to 10 mm. Some occurrences of granophyre are very rich in quartz and almost lacking in mafic minerals; others include 10–25 percent of a rustic-weathering fibrous gray-green amphibole and abundant thin plates of ilmenite-magnetite. Transitional zones between granophyre and diabase contain augite.

APLITE

The diabase aplites comprise two varieties, one quartz free and one quartz bearing. The two varieties are texturally similar but differ considerably in mode of occurrence.

The quartz-free aplite occurs mainly as veins and subordinately as dikes, which are insignificant in volume but are nonetheless conspicuous because they are equigranular and pink against the dark diabase that is everywhere their host. The veins are a fraction of an inch to 6 inches in width and are characterized by gradational borders which are from one-tenth of an inch to a few inches in width. The less common dikes are sharp walled, and few are more than 12 inches wide. The quartz-free aplite bodies occur sparsely in most sills and are not restricted to any particular part of a sill; they are most abundant in the diabase adjacent to a discordant contact. Where the veins are especially numerous, the host diabase commonly is albitic.

Quartz-free aplite is composed mainly of pink feldspar, which constitutes 80–95 percent of the rock. The anhedral equant feldspar grains typically range in diameter from 0.05 mm to 2 mm. In thin section, the feldspar is clouded with claylike alteration products. Most of the feldspar is untwinned albite, but potash feldspar may also exist. Typically the aplite is mottled with gray-green hornblende, and commonly with epidote, both interstitial to albite. Sparse small miarolitic cavities are lined with hairlike needles of hornblende. Sphene is sparse but ubiquitous. In the borders between aplite veins and diabase, the aplitic texture appears to be superimposed on the intergranular texture of the diabase, the result being a salt-and-pepper effect.

The quartzose aplites occur mostly as irregular crudely lenticular bodies, a few inches to several feet thick, that invariably are adjacent to a feldspathic host rock. Most bodies lie parallel to contacts between diabase and the siltstone member of the Dripping Spring Quartzite, and these may have outcrop lengths of as much as several hundred feet. Similar smaller lenses are along contacts between the argillite member of the Mescal and the diabase or are along contacts between basalt flows and diabase. Some quartz-aplite layers form a vague-boundaried transitional zone between hornfelsed host rock and other feldspathic differentiates. To a lesser extent, the aplite occurs as dikes, a few inches to 5 feet wide, that either are in diabase sills but close to their borders or are in the host rock and adjacent to a sill. All quartzose aplite occurrences noted to date have been in or near the Sierra Ancha.

Quartz micrographically intergrown with pink to gray feldspar makes up a few percent to more than 50 percent of this rock. In many specimens, minute rounded grains of diopside are scattered through the rock. Biotite and chlorite have been seen in some thin sections. Pyrite, absent from much of the quartz-free aplite, is minutely disseminated in patches in parts of virtually every body of quartz aplite. It occurs throughout some bodies and causes their outcrops to be conspicuously stained or even crusted with limonite.

MODE OF FORMATION

The spatial relations of the feldspathic differentiates indicate that they do not represent later and independent intrusions but that they can be visualized as products that evolved in two ways from volatile-rich residues fractionated from nearby parts of the host intrusion. Once accumulated, the residues either (1) crystallized to form the quartz-free differentiates or (2) reacted with certain wallrock materials to form the quartz-bearing differentiates.

Most of the small randomly oriented pegmatite bodies are narrow and elongate and were obviously emplaced along joints and faults which opened after the diabase had crystallized enough to sustain fractures. That they could not have been effects of assimilation is indicated by the form of these pegmatite bodies, of the narrow dikes and veins of quartz-free aplite, and of the albite veinlets that are small-scale manifestations of the aplites, as well as by their distribution through sills. Apparently volatile-rich materials migrated locally to fill the fracture openings and to crystallize in and near them.

The positions of concentrations of quartz-free differentiates suggest the paths of migration and a reason for large accumulations of the late fluids. The small bodies are especially numerous, and somewhat more persistent, in the diabase adjacent to hanging walls of some steep-dipping discordant bodies and thus mark the sites favored as channelways. Extensive bodies of pegmatite exist only in the upper parts—though not necessarily at the very margin—of thick sills. The thickest parts of these layers occur where the discordant contact of a tabular feeder body joins the concordant roof of the sill. The layers have their maximum lateral dimension along this intersection and their minimum dimension—exceptionally as much as a mile—away from it. As the high point and terminus of the channelway, this is the logical site, and empirically the only one, for large amounts of residual materials to accumulate.

Quartzose differentiates only occur contiguous to tops of sills, they commonly border flat-dipping discordancies, and the thicker and more extensive bodies invariably contain feldspathic xenoliths. The only extensive masses are in sills that intrude the siltstone member of the Dripping Spring. Lesser bodies do exist along walls of sills intruded into the argillite member of the Mescal and into the basalt flows. These associations set them apart somewhat from the tabular masses of diabase pegmatite; nevertheless their details provide cause to speculate on a metasomatic origin for large masses of all the differentiates.

Details are best seen in the tabular masses of quartzose aplite that separate siltstone of the Dripping Spring from diabase or one of the other differentiates. In a few of these masses the relict bedding in xenoliths seems without exception to parallel the bedding of the adjacent siltstone. Apophyses from aplite masses commonly extend as narrow dikes into the host rock. From many such dikes thin tongues of similar aplite extend a few inches to several feet along the bedding of the hornfelsed siltstone. The tongues merge gradationally into the siltstone, and within the aplite, minute details of the delicately laminated strata are preserved. Furthermore, where the parent dike crosses strata that include several aplite tongues, the dike may be wider than elsewhere. These features suggest that the quartzose aplite is of metasomatic origin.

In some occurrences, certain dikes of quartz aplite appear definitely intrusive. These dikes cut cleanly across and dilate coarser differentiates, normal diabase, the host sedimentary rocks, and even vague-boundaried xenoliths dispersed in the main tabular mass of aplite. Where these dikes do vary slightly in width, the variations do not seem to reflect the rock type traversed, as is common in dikelike bodies that are replacement phenomena. Furthermore, though most of the xenoliths in the main layer of aplite have vague boundaries, appreciable numbers have sharp boundaries, and in some occurrences all are randomly oriented.

An elongate pod-shaped mass of differentiates exposed on the north wall of the canyon of Reynolds Creek 2½ miles south-southeast of McFadden Peak (in the SW¼ sec. 7, T. 6 N., R. 14 E., McFadden Peak quadrangle) shows relations to host rocks particularly well. Here an extraordinarily thick layered mass, exposed continuously for a mile along the strike of a locally discordant step in the sill contact, separates normal diabase from overlying sedimentary formations. The upper part of the differentiate body consists of quartzose aplite as much as 60 feet thick. In places the top of the aplite layer is against limestone 15 feet above the Mescal–Dripping Spring contact; in other places it is against siltstone or arkose 35–40 feet feet below this contact. Everywhere the contact between aplite and hornfelsed sedimentary rock is vague. Dikes of the aplite occur in both the overlying sedimentary strata and the underlying granite. Downward the aplite layer merges, through a transition zone a few inches to a few feet thick, with a light-gray uniformly coarse grained rock termed hornblende granite in the field. This granite, generally 30–50 feet thick but locally about 80 feet thick, merges abruptly downward with a coarse-grained gabbroic rock. This gabbroic rock grades downward through an interval of 5–20 feet into ordinary medium-grained diabase. The body of layered differentiates is locally as much as 150 feet thick where its contact with the host rock is markedly discordant. As the contact becomes concordant downdip, the body pinches out, and normal diabase is in contact with the sedimentary formations.

Xenoliths apparently settled differentially into this mass, perhaps partly as a function of size (Lovering, 1938) and partly as a function of impedance offered by different strata of the crystallizing magma. Small angular fragments of siltstone and arkose, heterogeneously oriented, occur in great numbers above the granite. Many are so thoroughly reconstituted that they are difficult to distinguish from the aplitic matrix. Two small rounded blocks of fine-grained diabase, perhaps remnants of a chilled selvage of normal diabase, were seen a few feet below the top of the aplite. Xenoliths are sparse through most of the granitic layer, but they increase in number downward and can be seen every few feet along the basal 10 feet of the layer. Those 20 feet or more above the bottom of the granite range in largest dimension from 6 inches to 2 feet and are randomly oriented, whereas those in the basal 20 feet of the granitic layer are larger, as much as 15 feet long and 1–4 feet thick, and many are oriented parallel to the bottom of the granite layer. Some of these large xenoliths are bordered by a reaction zone 1–6 inches thick that is texturally like some of the highly feldspathic diabase pegmatites. Many, however, have fairly sharp contacts with the granitoid host. Xenoliths are uncommon in the underlying gabbroic facies; the few seen were small and had narrow pegmatitic rims.

About a mile to the southeast, in a similar setting and probably along the same contact, a layer of granophyre 10–30 feet thick is separated from the siltstone by only a few feet of quartzose aplite. The layers of granophyre and aplite are ill defined; the aplite includes moderate numbers of xenoliths, but the granophyre includes only a few. This occurrence of granophyre and aplite is more representative of the composite bodies of differentiate generally seen.

Seemingly the extensive pegmatite bodies are the differentiates that formed where xenoliths did not contaminate the accumulating residuum. Some bodies, as much as 20 feet thick, border the upper contact of a sill. Sporadic patches of granophyre that might represent assimilated xenoliths occur in these bodies, and some pegmatite layers merge upward or laterally with bodies of granophyre that are in turn against the host rock. In a few places one or two 10–20-foot pegmatite layers also occur separately within the interval 60–90 feet below the top of a sill. These layers are commonly uniform in thickness along several hundred feet of outcrop; but in places they pinch and swell, and where one layer thins the adjacent layer may thicken. Some layers have subparallel apophyses, separated from the main mass by thin wedges of diabase. Small irregular lenses of pegmatite may be abundant in the diabase between the principal pegmatite layers. The geometry of all except the border layer would seem to preclude an origin that involved reaction with xenoliths. Only the border layer contains granophyre, and peculiar coincidences of xenolith distribution would be required to account for the interrelations between the other layers and their apophyses.

The quartzose differentiates occur not only in the part of a thick intrusion where residual fluids accumulated in volume, but also where xenoliths dropping into a mush of crystallizing material triggered the formation of rock different from the pegmatite. The only effective contaminants were those with at least a modest feldspar content. Xenoliths of quartzite or limestone lack any features that would suggest assimilation and nowhere have caused similar effects. Possibly a considerable volume of contaminating materials was required for this reaction; the quartzose differentiates generally occur where diabase magma was insinuated along preexisting shatter zones that could furnish great volumes of material. Further, some hint exists in the Reynolds Creek examples that xenoliths in different sizes, numbers, and surface areas in different parts of the residual magma may have had a considerable effect on the type of rock that finally crystallized. The quartzose aplite in part formed in place by the soaking of feldspathic rocks with magmatic fluids. But in part it must also represent a melt modified by reaction and injected as discrete dikes into a variety of earlier formed rocks.

METAMORPHISM ASSOCIATED WITH DIABASE

Except for thermal metamorphic effects induced at the time of intrusion of the diabase, the Apache and Troy strata generally are metamorphosed no more than the Paleozoic formations that overlie them. However, because diabase intrusions are virtually ubiquitous in Apache strata, the effects of thermal metamorphism associated with the diabase intrusions are of regional scale. The heat was probably transferred mainly by solutions emanating from the diabase, but carbon dioxide, released as a result of the dissociation of dolomite, must have been an important transporting agent in the Mescal Limestone. Intrusion of diabase caused small-scale fracturing, particularly in carbonate rocks, and this fracturing was an additional factor that caused wider distribution of metamorphic products in the carbonate rocks than in other rock types. The mineralogic changes were brought about largely by reconstitution of the original rock-forming minerals. Some constituents were added from the diabase, but whether or not the additive process had an appreciable role in the metamorphism is uncertain.

Where the metamorphic effects are inconspicuous, as in many weathered outcrops, the original lithology may be misinterpreted—as generally has happened with the lower and algal members of the Mescal. Except in isolated small outcrops, however, the metamorphic effects are not such as to cause misidentification of a formation.

Quartzose rocks and the coarser grained more massive feldspathic rocks were the least susceptible to alteration. Quartzitic parts of the Troy in few places exhibit effects of metamorphism; immediately adjacent to a diabase intrusion the quartzite member locally becomes more dense and vitreous. Locally, the porous sandstone of the Chediski Member was converted to quartzite through widths of at least 100 feet. Some wedges of the arkose member of the Troy and of the lower member of the Dripping Spring Quartzite, which are engulfed in diabase, were recrystallized, through thicknesses of as much as 100 feet, to a hornfels in which discrete grains of quartz appear to be embedded in a dense matrix of potash feldspar. The original bedding structures were largely obliterated, and in isolated outcrops the hornfels of the Troy arkose is difficult to distinguish from that of the Dripping Spring. In other settings the two arkoses were little altered.

Generally the pre-Apache rocks are little modified adjacent to diabase intrusions. Weathered outcrops of quartz monzonite adjacent to diabase tend to be more abundantly stained by limonite than those that are not intruded, but textural modifications are generally lacking. Thoroughly sheared quartz monzonite or the granitic regolith that underlies the Pioneer Shale may be more strongly metamorphosed. One modification consists of abundant egg-shaped relicts of partly absorbed microcline phenocrysts in a fine-grained olive-gray or dark-greenish-gray matrix of feldspar, quartz, dark micas, and other dark minerals. From a distance this rock would seem to be an "orbicular" diabase. In other occurrences, the phenocrysts have been completely destroyed, and their places are marked by elongate blebs of bluish-gray opalescent quartz in a similar matrix; the quartz blebs are as much as 1 cm in diameter and in places have coalesced as vermicular aggregates. In some of the less altered regolithic material, broken fragments of feldspar remain angular, but the quartz grains are rounded. Certain regoliths, in which only granules of quartz and feldspar remained after the fines had been winnowed, were reconstituted to an orange-pink rock in which abundant coarse blebs of quartz lie in an almost textureless matrix of feldspar.

Because strata of the siltstone member of the Dripping Spring were intruded more extensively than the more massive sandstone, they exhibit metamorphic effects more widely. The siltstone units were more susceptible than the arkose units of the member. Strata a few feet to a few tens of feet from diabase are better indurated, extremely tough, and where weathered are more abundantly stained by limonite; but to the unaided eye most of the hornfels otherwise is little different from the original siltstone or arkose. Hornfels coarse enough to contrast texturally occurs in only a few areas adjacent to some of the thicker sills. Where a thick intrusion discordantly invaded strata previously shattered by faulting, zones of hornfels as much as 100 feet thick exist, and the coarse hornfels may be conspicuous in a zone 5–30 feet thick. The coarser hornfels is pinkish gray to medium gray and occurs as thin layers or lenses that crudely parallel bedding or as veinlike zones or irregular masses that follow joints or fractures across the bedding. All are gradational into the dark fine-grained hornfels. Some of the more extensive examples have been considered on pages 59–60, in discussion of the quartzose aplites of metasomatic origin.

As seen under the microscope, the potassium feldspar of the hornfels is coarsened in texture, and in places is replaced by albite. In the coarsest hornfels, quartz and feldspar commonly exist as micropegmatitic intergrowths, and the mafic minerals are clear anhedral grains of pyroxene and an amphibole. Biotite may be common. Locally, sphene is an abundant accessory. In freshly broken hornfels thickly disseminated patches of fine-grained pyrite may be apparent megascopically.

The tuffaceous mudstone and siltstone of the Pioneer Shale generally are metamorphosed for a few feet adjacent to thin sills, and the entire formation may be recrystallized to some degree where it is intruded by thick sills. Outlines of the microscopic shards are obliterated wherever there has been the slightest reconstitution of the rock. The dusky red strata were metamorphosed to hard dense grayish-orange to grayish-green hornfels, in which bedding structures are largely obliterated. The hornfels is mottled throughout with gray patches, mostly less than 1 mm in diameter, of finely divided dark minerals. In partially recrystallized strata the hornfels retains a pink or red tinge.

The claystone and siltstone strata that originally made up the upper member of the Mescal were converted to argillite virtually everywhere. The sparse dolomite beds that were interlayered in a few localities were invariably converted to silicate-bearing limestones. Adjacent to thick sills the argillite is a reddish orange to almost black very dense novaculitic rock, which contains abundant conspicuous metacrysts or knots of metamorphic minerals. In some beds the knots are elliptical, are as much as 3 mm in diameter, and are mostly aggregates of fine-grained dark mica. Other beds are irregularly mottled with light-brown to dark-yellowish-brown aggregates, which consist of a very fine grained mineral or minerals and which do not exhibit cleavage, or are thoroughly flecked with lathlike metacrysts, 1–10 mm long, of greenish-gray amphibole. Parting planes of the rock containing the amphiboles resemble surfaces covered with miniature bird tracks.

The basalt flows adjacent to some sills were thoroughly albitized and locally veined by epidote. Where recrystallized the basalt texturally resembles the diabase aplite but is mottled greenish gray to reddish orange. The gray patches are mostly chlorite, biotite, and fibrous amphiboles(?), whereas the orange rock consists mainly of very clouded feldspar, probably mostly albite. In a few localities both argillite and basalt were converted to an orange-pink rock which cannot be distinguished in hand specimen from the diabase aplites. This rock forms irregular masses, a few inches to 30 feet across, which gradationally interfinger with less altered rock along bedding in the argillite and along planes of fracture in the basalt.

The carbonate members of the Mescal are the units most widely and pervasively metamorphosed by diabase. The original strata of dolomite and cherty dolomite were dedolomitized and converted to calcitic limestone that contains concentrations of magnesian silicate minerals. Where a concordant sill at least 50 feet thick intruded the carbonate section, the entire section was generally dedolomitized. Where wide dikelike parts of diabase sills cut across the carbonate strata, the dedolomitized zones extend as much as 2,000 feet along the bedding. Metamorphic halos a few tens to a few hundred feet wide border dikes that are not more than 25 feet wide. The thin-bedded lower member of the Mescal tends to be more completely converted to limestone than the massive algal member.

The cherty dolomite yielded a variety of products that largely reflect the original composition. It was most commonly reconstituted to mixtures of tremolite, diopside, talc, serpentine, and calcite in a host of fine-grained sugary-textured calcite limestone. Diopside, tremolite, and talc were formed first, and they occur as irregular layers, lenses, and nodules that mimic the form of the chert concentrations. Where chert was moderate in amount, it was entirely replaced, and relicts are nowhere abundant, except in strata that once contained extraordinarily large concentrations of secondary chert. Dolomite virtually free of chert was altered to limestone in which the sparse silicate concentrations commonly are almost entirely serpentine. Serpentine is by far the most abundant silicate in the limestone and tends to obscure the interrelations of the earlier formed minerals and chert. It is also the most widely distributed of the silicate minerals and is found in parts of the dedolomitized zones most distant from diabase intrusions, where the other silicates apparently are absent.

Serpentine may be in paper-thin wispy discontinuous layers, in tabular zones locally as much as 8 feet thick, in microscopic blebs, or in elliptical nodules 3 feet in maximum diameter. Most of the serpentine replaced the early silicates or chert, and grossly outlines the original chert zones or chert nodules. Considerable amounts of serpentine, however, replaced the carbonate rock along points and faults, and some of the last-formed serpentine is in sharp-walled veins that cut across earlier concentrations of all minerals.

In hand specimens, masses dominantly of diopside are light gray to greenish gray. Diopside occurs as minute anhedral grains or aggregates of such grains. Tremolite occurs as very fine needles or radiating aggregates of needles. Aggregates of both minerals are tough and difficult to break. Small amounts of fine-grained talc are associated with diopside and tremolite; its paragenetic relations are not clear, owing to envelopment in the ubiquitous serpentine. All the silicates weather white and are much less conspicuous than the chert, which etches out boldly where remnant.

The dense serpentine is soft, exhibits a waxy or greasy luster on freshly exposed surfaces, and has a wide range of colors in hues of yellow, green, brown, and black. Microscopic inclusions of other silicates, calcite, and magnetite commonly cause variations in luster, color, and hardness. The late-formed vein serpentine is mostly grayish orange pink and in a very few places pale green. The dense serpentines are fibrous, but the fibers generally are too small to be resolved with the optical microscope. Selected specimens from the various occurrences studied by optical microscope, differential thermal analysis, electron microscope, and X-ray diffraction have all been identified as chrysotile (E. J. Young, written commun., 1959). Many serpentine layers contain sharp-walled veins of megascopically fibrous chrysotile. Locally these occur in volume and constitute the asbestos deposits that have been prospected or mined at more than 150 sites in Gila County.

The silicate minerals just described were formed from dolomite that contained very little impurities other than chert; the metamorphic products derived from clayey or silty dolomites have been given little attention. The silty or sandy dolomite matrix of the basal breccia of the Mescal has been converted to a dark-greenish-gray soft rock, apparently a very fine grained mixture of serpentine, chlorite(?), and calcite. In some areas much argilliceous material was concentrated near the top of the algal member during pre-Troy weathering. As a result of metamorphism, this clayey and siliceous dolomite yielded large amounts of gray-green micaceous minerals, which occur in thick hexagonal books as much as 5 mm in diameter. These metacrysts are in a matrix of calcite, serpentine, and other silicate minerals too fine grained for microscopic identification. They consist of at least two mineral species, one apparently a chlorite, the other unidentified. Locally, beds of apparently similar original lithology are now composed of greenish-gray tremolite as randomly oriented blades which are as much as 1/4 inch wide and 3 inches long. In places the tremolite occurs as irregular veins of woodlike cross fibers. Small amounts of still other silicates occur sparsely; an example is dark-yellowish-brown garnet, which has been noted in only three localities.

Magnetite is present in many localities but is not as prevalent a contact-metamorphic mineral as in many other areas where carbonate rocks are intruded by diabase. Locally magnetite is sparsely to abundantly disseminated through 20–100-foot thicknesses of limestone; a few such occurrences have an areal extent of several acres. The woody tremolite and coarse micaceous minerals occur in the same or adjacent beds. This magnetite was probably reconstituted from dolomites that were highly hematitic and argillaceous as a consequence of weathering. In many places, massive magnetite replaced limestone, which was much fractured within a few feet of a discordant diabase intrusion. Some of this magnetite is disseminated in the limestone, but most of it forms discrete narrow veins or small pods; a few veins are 1/2–3 feet thick and several tens to a few hundred feet long. These occurrences may represent iron expelled from the diabase magma or iron picked up from the dolomite and concentrated along shear zones by solutions emanating from the diabase.

LATE PRECAMBRIAN DEFORMATION

The Apache Group and the Troy Quartzite were folded and faulted prior to intrusion of the Precambrian diabase. However, the effects of this deformation have been obscured by the intrusion of diabase and, in the southern part of the region, by deformation of post-Paleozoic age, so that this early period of deformation has not previously been recognized. Details of prediabase structures have been studied mainly in the Colorado Plateau.

PREDIABASE DEFORMATION

The earliest deformation that affected the younger Precambrian rocks were the episodes of gentle warping that elevated the Mescal for subaerial erosion before the basalt flows were laid down and again before the Troy was deposited. Downwarping in northwestern Gila County formed the basin in which the arkose member of the Troy was deposited (fig. 11).

After deposition and lithification of the Troy, but before diabase intrusion, younger Precambrian strata were strongly folded and faulted along widely spaced narrow belts; between these belts the formations re-

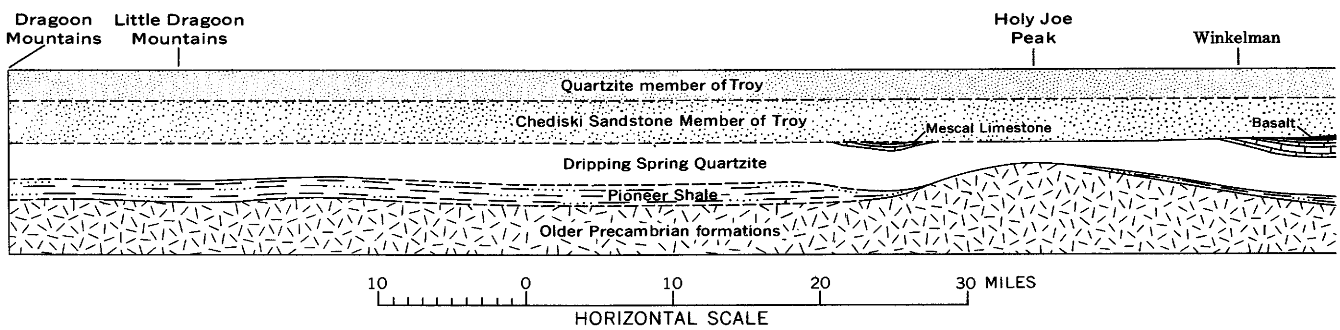


FIGURE 11.—North-south geologic section after the Troy Quartzite was deposited. Vertical Contacts dashed where there are no

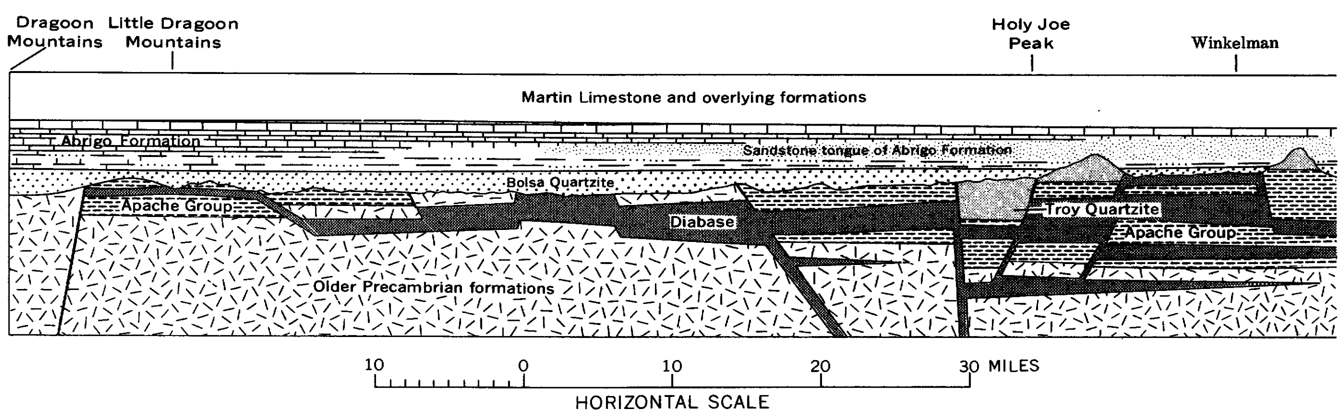


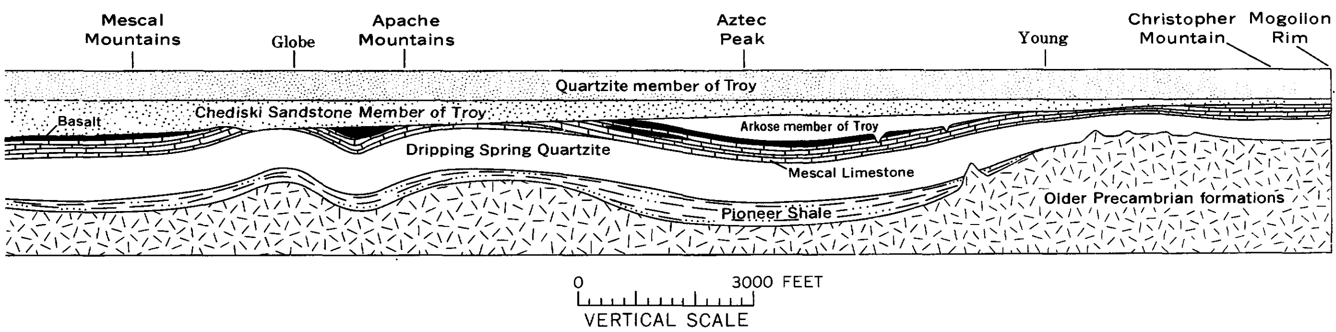
FIGURE 12.—North-south geologic section at the end of Devonian time. Structures are diagrammatic, but the positions of the

mained horizontal. Some of these belts can be followed 20–40 miles before they disappear beneath a cover of younger rocks. The belts commonly have a sinuous trace, but all have a general north or north-northwest trend. Because they were loci for deep erosion during late Cenozoic time, some are readily seen along major north-south tributaries to the Salt River, such as Cherry Creek and Canyon Creek.

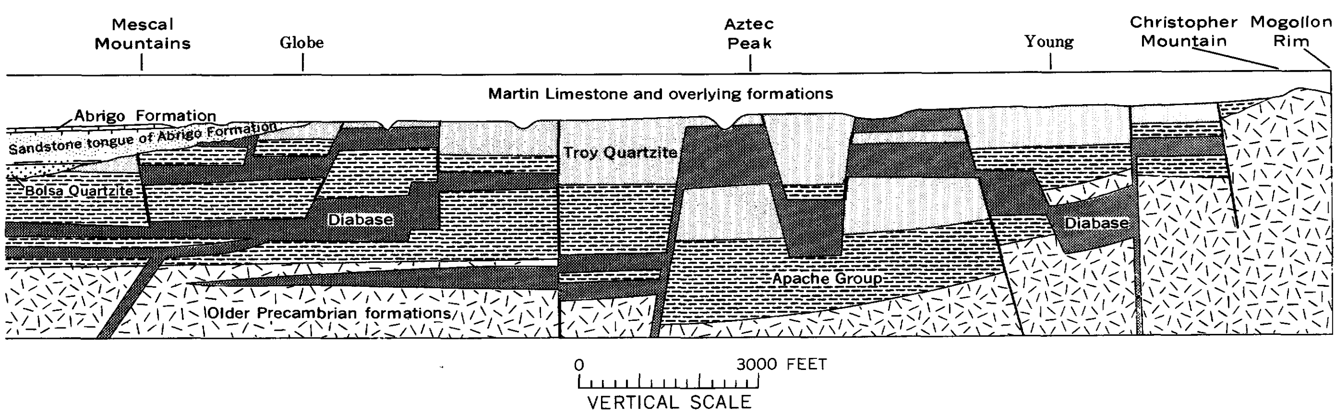
The canyon of Cherry Creek in its southerly course through the McFadden Peak quadrangle exposes a greater variety of structural details than is generally found along one belt. A single nearly vertical fault, or in places a zone of steep subparallel faults 1,000–5,000 feet wide, has been traced along the west wall of the canyon from a locality 2 miles south of the quadrangle to a locality 2 miles north of it; the belt probably extends much farther to the north and south. In the southern 10 miles of this exposure, most of the many faults in a zone about a mile wide strike N. 5°–10° W. In the central part of the quadrangle, for a distance of about 8 miles, the zone of faults is narrow or is represented by one fault that strikes N. 30° W. Along the northernmost 5 miles, a single fault, intruded by a wide dikelike apophysis from a diabase sill, strikes N. 10° E. Thus, along a 23-mile length, major segments of this fault zone differ in strike by as much as 40°. A fault of greater throw than the rest generally defines the west edge of the zone; the lesser faults that strike obliquely to the local trend of the belt terminate against this principal fault.

Along or immediately east of Cherry Creek proper, the formations were folded into a westward-dipping monocline about a mile wide that persists the length of the structural belt. From the horizontal strata east of the belt, there is a gradual transition westward to dips of 3°–10° W., which are characteristic of most of the monocline. Near Cherry Creek the dip increases rapidly, and where the monocline terminates against the principal fault west of Cherry Creek, dips of 20°–30° are common; in a few places near the fault, beds are almost vertical. Along parts of the canyon, some displacement downward to the west is due to faulting as well as monoclinal folding, and thus in places the Cherry Creek structural belt resembles half of a graben. The faults are intricately intruded by diabase dikes or dikelike discordant parts of sills along most of the length of the belt, and the monocline generally terminates against an intrusion.

Immediately west of the principal fault of the structural belt, sedimentary formations generally dip steep-



exaggeration overemphasizes pre-Troy folding and angularity between Troy and Apache strata.
remnants to confirm interpretation.



Troy Quartzite and the Apache Group relative to the pre-Bolsa and pre-Martin unconformities are probably representative.

ly west. For considerable distances along the fault, the formations are vertical or nearly vertical, and in places the strata within a few hundred feet of the fault are overturned to the east, in a few places dipping less than 50°. This intense folding affects only a narrow zone; 500–2,000 feet west of the principal fault, the formations are virtually horizontal and continue so westward into the Sierra Ancha. The Troy Quartzite, the Apache Group, and the underlying granite are involved in this fold. Because some of the formations are very competent and because the stratigraphic thickness (about 3,000 ft exposed) was too great to be accommodated in such a tight fold, the folded section is thinned by shearing. In places incompetent parts of formations are missing, and in a few places competent parts are duplicated along a shear zone roughly parallel to the steep fault that bounds the fold on the east.

Displacements caused by faulting and folding within the Cherry Creek structural belt are noteworthy, but corresponding horizontal strata on both sides of the canyon outside the deformed belt are—if one ignores local displacements caused by later diabase inflation—at about the same elevations. Within the belt, stratigraphic displacement locally exceeds 1,500 feet; the Troy Quartzite, for example, is locally in juxtaposition with pre-Apache granite.

Including all elements, the Cherry Creek fold-fault belt ranges in width from 2 to 4 miles. Other structural belts are similar in their gross details. Few, however, show such intense folding, and most are 1–2 miles wide.

In general, the belts are several miles apart; between them the formations, though locally displaced by high-angle faults of later date, are practically horizontal. Canyon Creek, which is 10–12 miles east of Cherry Creek, also follows a north-northwest fault zone downstream from a locality 23 miles north of its junction with Salt River; south of the Salt River this belt can be traced an additional 19 miles before it is covered by Cenozoic gravels. West of Cherry Creek 4–10 miles, another narrow belt of north-northeast trend has been traced from the southwest corner of the McFadden Peak quadrangle northward to within 4 miles of Young—a distance of 20 miles. A fault is continuous the length of this belt, except perhaps across the east flank of McFadden Peak where an east-dipping monocline defines the structure. Along much of its length north of McFadden Peak, the fault is dilated by a diabase dike(?), which is 300–1,000 feet wide. Narrow zones of sharply folded strata border the dike. South of McFadden Peak the position of the fault is marked by the discordant part of a thick diabase sill. The belts just described are the principal ones in the McFadden Peak and Blue House Mountain quadrangles. Other north- or north-northwest-trending structural belts of considerable length apparently exist east and west of these quadrangles.

The narrow but persistent zones of intensely folded strata that border some structural belts are not everywhere readily identified as drag phenomena adjacent to major faults. These zones are so intimately associated with dikelike bodies of diabase that the folds might be interpreted to be the result of forceful injection of diabase magma. As shown earlier, the diabase magma was a very fluid material that insinuated its way along splitting planes or fractures, rather than a viscous mass that punched its way along planes of weakness and aggressively folded the wallrocks. Therefore, except for minor superimposed details, it is wholly unlikely that any part of the fold structures can be explained in this way. This conclusion is reinforced by spatial relations of diabase in the belts. In some localities diabase intrusions do not crop out in juxtaposition with the fold, and the complete lack of thermal metamorphism, especially in rocks particularily susceptible to reconstitution, indicates that diabase must be remote if present at all. In some places the fold adjacent to a fault is discordantly split by the sill; such a split indicates that the fold predated the intrusion. (See horizon *y*, fig. 13*B*.)

Along much of a belt, the displacement now apparent in the bordering fault does not reflect the movement that caused the sharp folds. Where drag shown by a fold is opposite that which is expected from apparent displacement, the sense of the original movement was reversed by later faulting caused by diabase inflation (fig. 13*B*). In other places, inflation caused additional displacement in the same direction as the original throw (fig. 13*C*). Whatever the sense of later movements the folds generated in the early episode of faulting were not notably modified.

The dikes along the north-trending structural belts may represent the principal feeder conduits for the sills of the region. These are the only thick dikelike bodies that can be inferred to extend from far down in older Precambrian rocks up through entire sections of younger Precambrian strata. Elsewhere, most thick tabular crosscutting masses merely connect concordant parts on one sill.

The areal distribution of sills that join with the possible feeder dikes is particularly noteworthy, because they occur in patterns that may aid in the recognition of prediabase structural belts farther south. Thick or extensive sills on one side of a structural belt—perhaps the west side—commonly terminate abruptly against the principal fault or merge with the dike that occu-

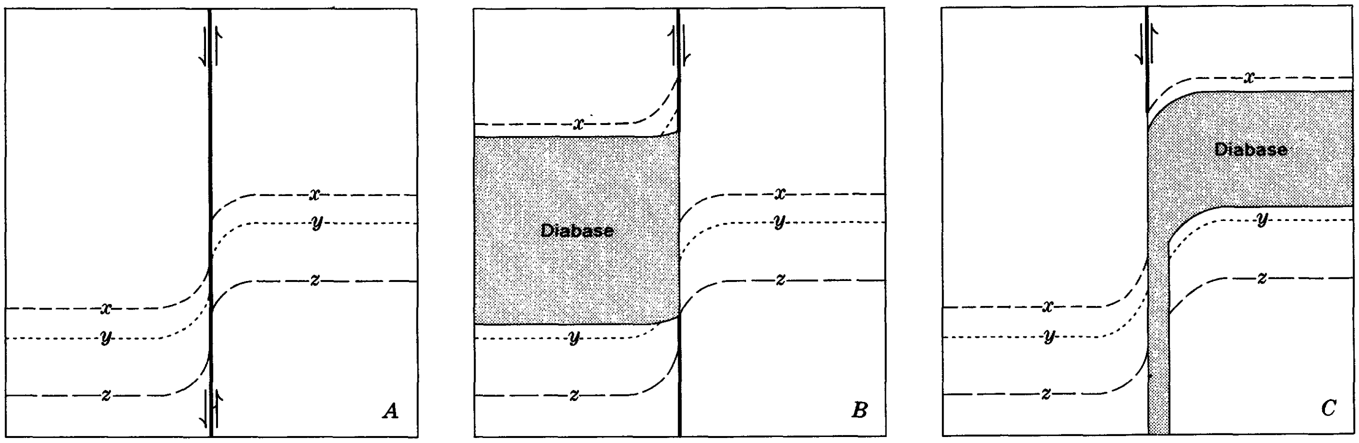


FIGURE 13.—Effects of sill inflation on displacement along a preexisting fault. Arrows indicate direction of relative displacement. *A*, Prediabase fault bordered by sharply folded strata. Lettered horizons are those shown in *B* and *C*. *B*, Sill terminating at fault of *A*; displacement along the upper part of the fault has been reversed by inflation. Note that fold is cut discordantly by sill. *C*, Sill terminating against fault occupied by dike; displacement along upper part of the fault is increased.

pies the fault, and then may be insignificant or missing on the opposite side of the fault. Elsewhere along the belt, thick sills may inflate strata on the east side of the fault but be absent on the west side. Finally, sills on one side of a structural belt commonly were intruded along horizons that were different from those on the opposite side. Perhaps such spatial relations of sills can be used as clues to the existence of such belts. Whether prediabase structures are also dominantly of northerly trend in the southern part of the region is not now known. It is of interest that the Cherry Creek belt, if projected south, could extend into the northeast corner of the Globe quadrangle, where the fault and intrusion patterns mapped by Peterson (1954) are similar to those along the east wall of the canyon of Cherry Creek.

DEFORMATION ASSOCIATED WITH DIABASE INTRUSION

High-angle normal or reverse faults, which displace younger Precambrian strata but not Paleozoic formations, are common throughout the region. Most of these faults owe their dislocation entirely to diabase inflation. Some interrelations between faults and diabase are shown in figure 14, which is idealized from clear-cut examples in the Sierra Ancha and the Salt River canyon.

For most of these faults the stratigraphic throw closely approximates the thickness of the inflating sills. Some faults obviously terminate either at the top of a sill (see loc. A, fig. 14) or the bottom (loc. D, fig. 14). The throw of such faults does equal the thickness of the sill. (Compare intervals x and x',

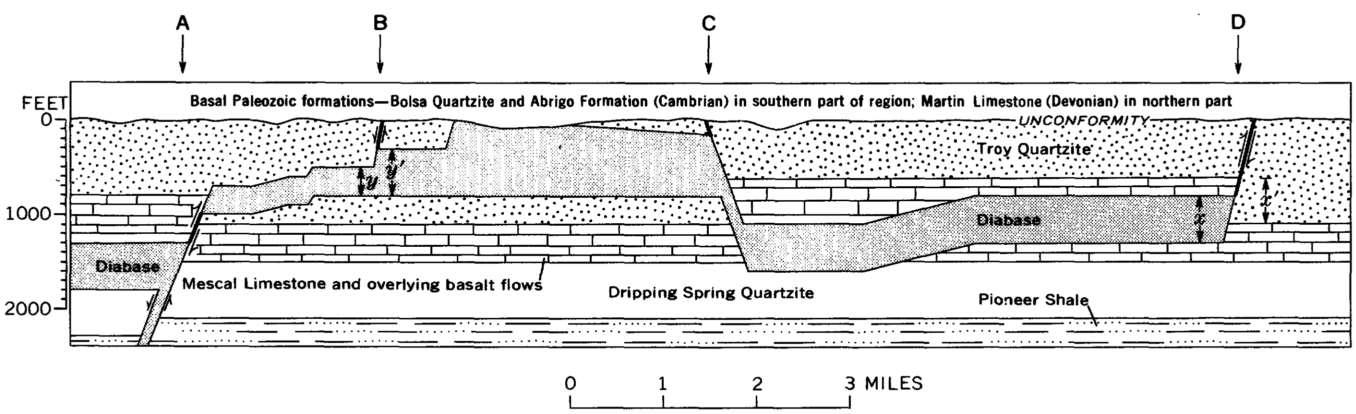


FIGURE 14.—Idealized geologic section showing relations of faults to diabase intrusions and to pre-Paleozoic unconformities. Note vertical exaggeration. Diagram is greatly simplified; typically two or more sills inflate the Apache sequence, and several minor steps exist along sill boundaries. Letters A–D indicate features discussed in text.

fig. 14.) Faults of this type occur where sills step discordantly from one horizon to another but may not be recognized if they are occupied by a thick dikelike body of diabase (loc. C, fig. 14). The footwall contact of a sill may be concordant, but a discordant step and fault may offset concordant parts of the hanging wall (loc. B, fig. 14). The throw of the fault in such an example is not the thickness of either adjacent part of the sill but the height of the step (the difference between y and y', fig. 14). As a consequence of inflation at different horizons, a fault may be displaced in one sense along its upper part and in the opposite sense along a lower part. (See loc. A, fig. 14.) Such reversals in throw can also be observed at the junction along the strike of two sill-related faults of opposite throw. Where more than one sill inflated a sedimentary section or where successive sills intruded along one horizon displaced one another, the relations between intrusions and faults can be more complex.

In the Colorado Plateau the trends of diabase-related features were influenced by two conjugate systems of joints. The joints of one conjugate system strike N. 10°–25° E. and N. 70°–85° W.; those of the other system strike N. 5°–30° W. and N. 45°–60° E. The principal inflationary faults, dikes, and discordant parts of sills of a given area generally parallel one of the joint sets. Subordinate faults and intrusions mostly parallel the complementary set of that system, but may instead follow one of the other sets. As seen in plan, discordant intrusions and faults of the dominant trend deviate locally along joints of the secondary trend. The exceptions to these generalizations occur in and near the belts of prediabase structure; there the diabase-related features follow one or the other of the north-trending sets of joints.

Sill-related faults of various trends differ in lateral persistence and displacement. Those of northerly trend are the most numerous, and most of the individual faults of appreciable throw and length fall in this group. But those of west-northwest and northeast strike are conspicuous because they cut across the prevailing structural and topographic grain of the area. A few of west-northwest trend are extensive or show considerable displacement, and are particularly noteworthy because they are entirely inflationary in origin. An outstanding example is a fault of N. 70° W. strike that has been traced westward from the northwestern corner of the Blue House Mountain quadrangle to a point on Cherry Creek 2 miles north of the McFadden Peak quadrangle; the fault probably continues farther west. Throughout its known length of 11 miles, it is followed by the discordant part of a diabase sill. Another example, only about 3 miles long but of greater throw, extends from a point about half a mile south of McFadden Peak east to Cherry Creek canyon. This fault strikes about N. 85° W., is downthrown on the north about 1,000 feet, and terminates a diabase sill in exactly the manner shown at locality *D* of figure 14. Eastward, this fault ends abruptly against the principal fault of the north-trending Cherry Creek structural belt; south of McFadden Peak, the throw of the fault decreases abruptly at its intersection with a northeast-trending discordancy, and not far to the west, the fault probably dies out. Faults of N. 30°–60° E. strike are numerous, but most are short and of small throw. They perhaps seem more prevalent than comparisons warrant because discordant bodies of diabase were commonly intruded along them and in certain areas many asbestos occurrences were opened adjacent to these intrusions.

In the southern part of the region, where later deformation has obscured them, the Precambrian structural features presumably are similar and equally as abundant. The prediabase faults are probably deep seated, but those of inflationary origin are literally without "roots". The latter would not likely be reopened, and they do not extend to such depths that they would likely tap sources of metal-bearing solutions. Thus most of the younger Precambrian faults were probably not significant in the localization of metal deposits, and in the metal-mining districts they should be distinguished from later faults that were effective as channelways for ore-bearing solutions.

Throughout the region small-scale monoclines, synclines, anticlines and domes locally occur along or near discordant parts of intrusions. Many folds are so subtle that they are recognized only by detailed mapping on mine scales. Small bedding-plane faults, thrust faults, and near-vertical strike-slip faults occur directly adjacent to discordant contacts and are especially abundant in the vicinity of the more strongly defined small folds. These faults and folds represent mild structural adjustments that occurred during and immediately following the emplacement of diabase, as is apparent from their geometric relations to successive intrusions of diabase. They occur mainly in the least competent units of the stratigraphic sequence—the lower member of the Mescal Limestone, the lower two-thirds of the siltstone member of the Dripping Spring Quartzite, and the tuffaceous siltstones and mudstones of the Pioneer Shale. Those in the Mescal are particularly noteworthy as the prime structural factors in the localization of the chrysotile-asbestos deposits.

RELATIONS OF PALEOZOIC FORMATIONS TO PRECAMBRIAN ROCKS

The base of the Paleozoic sequence in southeastern Arizona is defined everywhere by an erosional unconformity that truncates diabase or diabase-related structures. There are two of these erosion surfaces, and each—in different areas—separates the basal Paleozoic strata from older Precambrian rocks. Therefore a variety of relations must exist between the Paleozoic and the younger Precambrian rocks. South of a line that extends from the vicinity of San Carlos Lake northwest almost to Theodore Roosevelt Lake (fig. 4), the younger Precambrian formations are commonly overlain by the Bolsa and Abrigo Formations of Cambrian age. As drawn, this line represents the present northern limits of Cambrian outcrops, and it approximates their northern limits in Early Devonian time. A few miles to the north, the Martin Limestone of Devonian age is everywhere the basal formation of the Paleozoic sequence.

Figure 12 is an idealized reconstruction of the north-south geologic section that must have existed at the end of Devonian time. The significance of certain features depicted in this figure has not been appreciated until recently. The pre-Troy and pre-Bolsa unconformities have been assumed to be the same surfaces; furthermore it has not been generally recognized that sandstones of the Abrigo Formation can rest directly on Apache strata, as do the Bolsa and Troy. Consequently the Bolsa Quartzite, certain sandstone facies of the Abrigo Formation, and in a few areas the locally thick basal sandstone of the Martin Limestone have been correlated erroneously with the Troy Quartzite. In the following section, emphasis is placed on the structural relations of the various basal units of Paleozoic sandstone to the Precambrian rocks, and on criteria for distinguishing between formations previously presumed to be lithologically similar.

ROCKS ADJACENT TO THE PRE-BOLSA UNCONFORMITY

As originally defined in the Bisbee area by Ransome (1904, p. 28–30) and as described for several localities northward to the Little Dragoon Mountains by Gilluly (1956, p. 14–15, 24) and by Cooper and Silver (1964, p. 45–47), the Bolsa Quartzite is a white to grayish-red rusty-weathering quartzite which ranges in thickness from 390 to 480 feet. This quartzite occurs in regular internally cross stratified beds that range in thickness from a few inches to 10 feet but perhaps average 2–4 feet (Gilluly, 1956, p. 15). A basal conglomerate a few inches to a few feet thick is characteristic. The overlying 100–200 feet of the formation is made up of medium- to coarse-grained beds that have a considerable content of white quartz granules and, in places, of small quartz pebbles. These are the eye-catching "gritty" strata frequently stressed, out of proportion, as definitive of the whole formation. Actually this unit is transitional upward into much better sorted medium-grained strata and ultimately into fine-grained and very fine grained beds that make up the upper third of the formation. The Bolsa is transitional upward, by an increase in sandy and micaceous mudstone layers, into the Abrigo Limestone. Gilluly described this transition zone as generally about 50 feet thick and noted that at least in one locality it is almost 100 feet.

Ransome (1904, p. 30) originally considered the Bolsa Quartzite to be Middle Cambrian in age because of its comformable relation with the overlying Abrigo, in which Middle Cambrian fossils had been found. No diagnostic fossils have yet been found in the Bolsa, but in recognition of the transitional relations between the two formations, the original designation is generally accepted.

Between Bisbee and the Little Dragoon Mountains, the Abrigo Limestone ranges in thickness from 700 to 844 feet. Green-tinted claystones, siltstones, and sandy mudstones, and sparse units of limestone ribboned by mudstone, make up the lower third. Limestone ribboned by thin layers of similar mudstone dominate the middle third. These strata give way to yellowish- or reddish-brown dolomite, in which quartz sand becomes coarser and increasingly conspicuous toward the top of the formation. The Abrigo is distinguished by thin bedding, beds marked by edgewise conglomerates, numerous units that exhibit irregular partings, and irregularly anastomosing layers of silty and sandy material that separate lenticular layers of sandy carbonate. Worm trails and fucoidal markings are abundant. Beds of massive limestone form a few ledges, and beds of quartzite, sandstone, or sandy shale occur sporadically throughout the sequence. Northward from the type locality near Bisbee, the amount of sand increases (A. R. Palmer, in Gilluly, 1956, p. 20).

Fossils of both Middle and Late Cambrian age have long been recognized in the Abrigo (Stoyanow, 1936, p. 466–481). The list of recognized forms has recently been expanded from collections made by Gilluly and associates, and the implications of the several faunal zones have been concisely summarized by A. R. Palmer (in Gilluly, 1956, p. 20–24).

Sequences of the sort seen in the principal outcrops of western Cochise County, visualized by many as typical of the Cambrian elsewhere in southern Arizona, are clearly separated from and lithologically different from the underlying rocks. For the most part they rest on a planar surface cut on older Precambrian rocks; in the Dragoon quadrangle they rest on the Pioneer Shale, on Dripping Spring Quartzite, and on diabase sills which intrude into these formations. Although they exhibit many features in common, the western Cochise County sequences are hardly typical of the sequences farther north, which are more arenaceous and locally seem transitional into underlying strata.

The first of the aberrant sequences demonstrated in some detail to be temporally equivalent to the type Abrigo was described by Stoyanow (1936, p. 476–477), from outcrops in the northern part of the Santa Catalina Mountains, near the head of Peppersauce Wash, in the southwest corner of the Mammoth quadrangle. This section, 750 feet thick, includes considerably more sandstone and quartzite than sections farther south. The lower 400 feet, separately designated the Santa Catalina Formation by Stoyanow, consists largely of rusty-weathering thin-bedded and irregularly bedded micaceous mudstones and quartzitic sandstones, in which are intercalated thin layers of dolomite. This unit is capped by a conspicuous ledge-forming unit of clean quartzite, which Stoyanow termed the Southern Belle Quartzite. The Southern Belle is 25–30 feet thick along the upper canyon of Peppersauce Wash but occurs only locally elsewhere in the northern Santa Catalina Mountains (S. C. Creasey, oral commun., 1960). Overlying the Southern Belle is almost 300 feet of thin-bedded sandy limestone and intercalated thin sandstone beds, and between this unit and the Martin Limestone is 20–25 feet of sandstone and quartzite.

Stoyanow (1936, p. 465–481) recognized several faunal zones during his study of the Cambrian strata in the Santa Catalina Mountains and elsewhere in southeastern Arizona, and consequently proposed that Ransome's terminology should everywhere be revised. He restricted the designation Abrigo Formation to the limestone at the top of the section and to paleontologically equivalent sections farther south. The overlying sandstone-quartzite unit he designated as a separate formation, the Peppersauce Canyon Sandstone. Other formational designations were also proposed for southern limestone equivalents of the Peppersauce Canyon and Santa Catalina. The proposed subdivisions make up an apparently unbroken sedimentary sequence and, for reasons enumerated by Palmer (in Gilluly, 1956, p. 24), cannot be consistently defined over broad areas. Therefore Ransome's original usage, which applied the term Abrigo to all Cambrian strata between the Bolsa Quartzite and the overlying Martin Limestone, is preferred.

The units of the Peppersauce Wash locality undoubtedly grade southward with a decrease of clastic constituents into dominantly carbonate beds. As an example, the Santa Catalina Formation and Southern Belle Quartzite of Stoyanow can be readily correlated with the lowest of the three members of the Abrigo Formation recognized by Cooper and Silver (1964, p. 47–50). In the northwestern part of the Dragoon quadrangle, 35–40 miles southeast of Peppersauce Wash, this lowest member has similar bedding characteristics and contains mudstones and sandstones like those of Stoyanow's Santa Catalina Formation, but includes considerably more brown-weathering dolomite. It is capped by a thin unit of quartzite, which pinches out to the south. On the basis of distinctive faunal zones, Stoyanow defined the Santa Catalina and Southern Belle units as Middle Cambrian, and the overlying strata as Late Cambrian. In the Dragoon quadrangle the boundary between the lower and middle members of the Abrigo is shown by fossils to be the boundary between the Middle and Upper Cambrian (A. R. Palmer, oral commun., Nov. 1960).

Stoyanow's work in the northern Santa Catalina Mountains provides the background for recognizing the more northern Abrigo facies, which include an even greater content of coarser clastics. Briefly stated, a thin- to medium-bedded quartzitic sandstone unit in the relative position of the Southern Belle Quartzite of Stoyanow thickens northward, at the expense of thin-bedded finer grained strata like those of Stoyanow's Santa Catalina unit. At Zapata Mountain, 1½ miles northwest of Holy Joe Peak, for example, regularly bedded typical Bolsa Quartzite, 160 feet thick, is overlain by 170 feet of beds which resemble the Santa Catalina, and this 170 feet in turn is overlain by 110 feet of cliff-forming quartzitic sandstone beds (M. H. Krieger, written commun., 1960). Except that *Scolithus* conspicuously marks beds in the upper and lower part of this cliff-forming unit, it might easily be mistaken for the Bolsa Quartzite of the same area. Above the unit, similar quartzitic beds form ledges in an otherwise friable sandstone that is transitional, through an interval of about 100 feet, into the brown dolomite and dolomitic limestone unit that makes up the upper 150 feet or more of the Abrigo at this and other northern localities. The Peppersauce Canyon Sandstone of Stoyanow is not a recognizable element of this or any other northern section. Generally several dolomitic sandstone beds occur in the carbonate unit, and many of the dolomite beds contain an abundance of coarse quartz sand, which etches out to outline the cross-stratification distinctive of this unit. Many beds are conspicuously glauconitic. In these and other aspects, this is the only stratigraphic unit in the northern part of the region that has a direct lithologic counterpart in the type section at Bisbee and in the thickest sections of western Cochise County.

In areas north of the 32d parallel, where it rests on Troy and Apache strata, the Cambrian sequence was deposited on an irregular erosion surface; furthermore the Bolsa is thinner there than to the south, and in places it has lapped out. As much as 300 feet of local relief is recorded for the northeastern part of the Little Dragoon Mountains. There the surface is generally plateaulike, reflecting the stripping of flat-lying Dripping Spring strata, but it is surmounted by hills as much as 100 feet high and is cut by valleys 150–200 feet deep. In this area the Bolsa ranges in thickness from 14 to 335 feet (Cooper and Silver, 1964, p. 43–46). The influence of resistant formations on the pre-Bolsa surface is readily seen in the area between Holy Joe Peak and Brandenburg Mountain, in the east-central part of the Holy Joe Peak quadrangle. The lowest part of the surface is on diabase, and there the Bolsa is as much as 150 feet thick. At Brandenburg Mountain the Bolsa laps out, and at least 350 feet of the Abrigo is absent where the formations pass from diabase onto higher surfaces formed on the Dripping Spring and Troy Quartzites. In several exposures farther north in the quadrangle, the Bolsa is less than 100 feet thick; and through much of the Dripping Spring and Mescal Mountains, sandy strata of the Abrigo directly overlie the younger Precambrian formations.

Though the Bolsa Quartzite is notably thinner in northern localities than in areas farther south, it is not significantly different in lithology. In areas of obvious local relief, the basal conglomerate contains boulders and varies locally in thickness and composition. Even in sections less than 100 feet thick, the Bolsa exhibits the usual gradation from gritty strata above the basal conglomerate to finer grained beds at the top. It is not as well cemented as in many southern outcrops, and in many places sandstone would be a better designation for it than quartzite.

Where the Bolsa Quartzite thins drastically the mudstone that intervenes between it and the 100–200-foot sandstone tongue of the middle part of the Abrigo Formation varies locally in texture and composition. In places fine-grained micaceous sandstone dominates the interval; elsewhere clayey mudstone alternates with sandstone. Dolomite is easily overlooked but exists as a cement or as nodules or thin lenses in sandy mudstone of the basal 20–100 feet of the Abrigo. Regardless of composition, these strata retain the character of the fine-grained claystone and mudstone of southern sections by being thinly and irregularly bedded, drab in color, and rusty weathering. The relatively clean resistant sandstone of the Bolsa generally gives way rather abruptly to the dirty friable sandstone of the Abrigo, so that the contact between formations is readily identified in most areas.

Fossils diagnostic of key faunal zones in southern sections bracket the northern sandstone tongue of the Abrigo above and below, and confirm correlations based on the recognition of comparable bedding features in lateral facies that are different in composition. Trilobites have been noted in the brown-weathering sandy mudstone of the lower part of the Abrigo at widely scattered localities from the Santa Catalina Mountains north into the Hayes Mountains, mostly as fragments that permit only the generalization that the rocks are Cambrian in age (A. R. Palmer, oral commun., Nov. 1960). In collections of trilobites made by C. R. Willden at Poverty Flat, 5 miles southwest of Coolidge Dam in the southeastern part of the Mescal Mountains, however, an association of bolaspidellid and marjumiid forms generally characteristic of the upper Middle Cambrian has been recognized (A. R. Palmer, written commun. to C. R. Willden, Aug. 5, 1960). These fossils and compatible occurrences confirm the earlier suggestion by Darton (1925, p. 255) that "brown to greenish-gray shales and soft earthy sandstone" of this and other northern localities should be correlated with the Abrigo Limestone. Ransome (1919, p. 44) considered such "layers of fine-grained unevenly colored brown, pink, and green quartzite, an inch or two thick, separated by films of olive-gray shale whose cleavage faces are ridged and knotted with numerous worm casts" to be the "most characteristic material" of the upper few feet of the Troy in the Ray quadrangle. The sandstone of the outcrops which Ransome discussed is seemingly in gradational contact with the underlying Chediski Sandstone Member of the Troy. It is more to the point here, therefore, to recognize that the northern occurrences of Middle Cambrian fossils show that the supposed transitional strata at the top of the Troy are correlative with the Middle Cambrian mudstone member that is at the base of the Abrigo in the Dragoon quadrangle and farther south. Specimens of *Billingsella*, a brachiopod of Late Cambrian (Franconia) age that characterizes the uppermost parts of the Abrigo Limestone in southern sections, have been found in most northern occurrences of the glauconitic sandy dolomite unit at the top of the Abrigo.

In a recent review of the data of earlier workers, Lochman-Balk (1956, p. 538–549) rightly concluded that the northern sandy facies that is here considered

Abrigo "should be kept distinct" from the underlying Precambrian sandstone of the Troy, but she also proposed that the northern facies should be interpreted as lithogenic equivalents of the Bolsa. In making this interpretation she did not have advantage of the following facts: (1) The Bolsa-Abrigo transition is everywhere fairly abrupt, and northern occurrences of the contact are strictly comparable with the contact at the type locality. (2) Sandy elements at the top and bottom of northern Abrigo sequences really are not greatly different in lithology from units of equivalent age in the type section. (3) Where the Abrigo is most fully remnant in northern outcrops, it is 600–650 feet thick—or almost as thick as the classic southern sections. (4) Further, the stratigraphic interval between the Middle Cambrian and Upper Cambrian faunal zones, cited in the preceding paragraph, is about the same in northern sequences as in southern sequences. Therefore, the Bolsa-Abrigo contact is not notably time-transgressive from south to north.

Where nondiagnostic fossils occur in abundance, they can be considered lithologic aspects that distinguish the Abrigo Formation from the Bolsa Quartzite. Throughout the region of Abrigo outcrops north of the Santa Catalina Mountains, primitive phosphatic brachiopods of Cambrian aspect are locally abundant. These brachiopods are particularly numerous in sandstone in the transition interval at the very base of the Abrigo, in similar friable sandstone within and at the top of the mudstone unit, and in the transitional sandstone above the cliff-forming sandstone tongue of the Abrigo. None have been seen in sandstone certainly identified as Bolsa. Rod-shaped fucoids—of the size of cigars—parallel the bedding and, with *Scolithus*, are locally very abundant and conspicuous in the coarser and cleaner sandstone of the Abrigo. *Scolithus* has been reported in a very few places in the Bolsa; apparently it occurs only sparsely and would be easily overlooked even in careful search of that formation. Narrow worm(?) or trilobite(?) trails and other smaller scale markings of possible organic origin characterize bedding surfaces throughout the lower two-thirds of the northern Abrigo sections. These markings have not been seen in the Bolsa.

Recognition of such features is especially helpful in identifying certain of the more northern outcrops of the thick sandstone tongue of the Abrigo. In these occurrences the tongue is made of light-gray almost white, fine- to medium-grained sandstone that does not necessarily weather rusty to make dark outcrops. The thin bedding may be obscure, and the unit commonly forms massive ledges or cliffs, parts of which have gnarly surfaces easily taken for convolute structures when seen from a distance. Thus, misidentification with the Chediski Sandstone Member is possible where outcrops are light colored, and with Bolsa Quartzite where they are dark, especially where the unit rests directly on Precambrian rocks. On close inspection the gnarled surfaces prove to be sites of abundant *Scolithus* or fucoids. Where these fossil features were not made obvious by differential silicification, they can generally be found in a brief search.

The reasons for previous correlations of Bolsa and Troy and for inclusion of certain occurrences of the Abrigo with the Troy are now apparent. The increase in sandy strata, particularly quartzite, and the decrease in carbonate content northward from the least clastic occurrences in western Cochise County, which are generally considered representative of the Abrigo, cause incomplete sequences of the Abrigo to appear grossly similar to parts of the Troy and Bolsa Quartzites. In many places the Bolsa or sandstone units of the Abrigo rest on the Troy in seeming concordance, and the contact between formations is obscure. In other areas the Troy is entirely missing, and the combined Bolsa and Abrigo sections, which occupy the same relative stratigraphic position, are approximately the thickness of the Troy noted in other southern occurrences. In places, however, the Bolsa Quartzite or the Abrigo Formation rests in angular or sharp erosional unconformity on the Troy Quartzite, on formations of the Apache Group, and on diabase. In these places, criteria for distinguishing sandstone and quartzite of Precambrian age from those of Cambrian age can be unequivocally defined. These diagnostic features are summarized in table 2, which also highlights some features not mentioned in the preceding generalized descriptions. Details of relative positions of quartzites and sandstones in the sequences, differences in textures of the sandstones, and the common occurrence of fossil features in sandstones of the Abrigo are particularly noteworthy. By the use of such criteria, for example, the so-called Troy Quartzite of the northern Santa Catalina Mountains (Ransome, 1916, p. 144; Stoyanow, 1936, p. 473–477) clearly should be redesignated Bolsa Quartzite.

Where other criteria for identifying the pre-Bolsa unconformity are not readily applicable, the recognition of regolithic debris from the diabase can be helpful. Beneath Cambrian strata the diabase is decomposed through thicknesses generally ranging from 2 to 25 feet. In this interval the diabase is platy or shaly parting and generally is grayish red. Very little of the remaining regolith was reworked and transported. In a few places the decomposed diabase is greenish gray; in such outcrops, in particular, the original ophitic texture may be relict up to the contact with the Cambrian strata. Although diabase soil incorporated in the basal part of the Bolsa or the Abrigo commonly causes several feet of strata to be dusky red, fragments of diabase in the basal conglomerate are almost nonexistent. In a few localities the diabase debris below the contact is bedded. In an unusual example 2 miles southeast of Superior, grayish-green diabase debris, mixed with a considerable amount of well-rounded quartz sand, is conspicuously cross stratified through a vertical interval of 25–30 feet. Here the zone of platy-parting diabase underlies the reworked material. In some areas Cambrian sandstone is separated from older strata by a few inches to a few feet of transported reddish-weathering diabase debris.

PRE-MIDDLE DEVONIAN UNCONFORMITY

The Martin Limestone generally overlies Cambrian formations in the southern part of the region and younger Precambrian formations in the northern part. The hiatus between the Cambrian and Devonian is represented in the southeasternmost Arizona by a disconformity, along which even minor channeling is not apparent (Gilluly, 1956, p. 25–26). As the disconformity is traced northward from Cochise County, however, erosional relief gradually becomes marked, and in parts of northern Gila County relief of a few hundred feet is apparent. The characteristic form of the erosion surface is seen along the Salt River canyon, where the unconformity can be traced many miles without interruption. Along the canyon, the pre-Devonian surface is a planar surface incised at wide intervals by steep-walled channels of considerable extent. Generally hills do not rise above this plateaulike surface, but in the area between Canyon Creek and Payson there are hills of moderate relief. The 400-foot carbonate section that makes up most of the Martin practically laps out against some of these hills of Precambrian rock.

In most places south of the Pinal Mountains, Cambrian strata probably intervene between the Martin and the Precambrian formations, but at least locally, such as near the abandoned settlement of Troy in the Dripping Spring Mountains, the Martin rests on Troy Quartzite or on the Apache Group.

The Martin Limestone generally overlies the Troy Quartzite in the area north of the line that defines the limit of Cambrian outcrops (fig. 4). In some areas of as much as a few square miles, however, the Martin Limestone rests on the Mescal Limestone or on the Dripping Spring Quartzite, or on diabase sills intruded into these formations. Northwest from Canyon Creek along the foot of the Mogollon Rim, where the Apache Group was already thin as a consequence of lapout on a pre-Apache high, the pre-Martin unconformity progressively truncates strata lower in the Apache Group. Ultimately at the west end of Christopher Mountain, northwest of Young, the Martin rests on older Precambrian formations (fig. 12). Farther west the Apache and Troy strata were everywhere removed during the pre-Martin cycle or an earlier erosion cycle.

The pre-Martin unconformity truncates the younger Precambrian formations and structures in the same manner as the pre-Bolsa surface. Angular discordances of as much as 30° exist, but in most areas the Martin appears to be concordant with the older strata. Except that it is generally thinner, the regolith zone on diabase below the Martin is comparable to that below the Cambrian formations. Basal strata of Devonian age contain even less regolithic debris than do their Cambrian counterparts.

The Martin sequence that surmounts the usual plateaulike surface is mostly limestone and dolomite but does include beds of shale and sandstone; regionally the formation is somewhat variable in lithology. (See Huddle and Dobrovolny, 1952, p. 73–76, 83–85; Gilluly, 1956, p. 26–29.) In the northern part of the region a dolomitic sandstone, a few inches to 20 feet thick, occurs everywhere at the base of this sequence. As thus delineated the Martin has generally been considered Late Devonian in age, but recently Teichert and Schopf (1958, p. 213–215) suggested that plant remains, collected a few feet above the base of the formation, indicate that the lower part of the Martin in northern Gila County is probably not younger than Middle Devonian and do "not rule out the possibility of assignment to the Early Devonian."

Of particular concern here is the correlation of sandstone that fills the widely separated channels below the general level of these dated carbonate strata. Previously, the sandstone has been considered part of the Troy. The channels range in depth from a few feet to at least 300 feet and in width from a few hundred feet to as much as half a mile. At least two exceed 5 miles in length. The channels everywhere transect the latest rocks and structures of the younger Precambrian terrane and merge with the higher pre-Martin surface; therefore the sandstone fillings are almost certainly Paleozoic rather than Precambrian in age. Because the carbonate section has been stripped from some of the prime examples and the contacts of other sandstone bodies with overlying strata are not definitive, it can be speculated that some and perhaps most

TABLE 2.—*Criteria for distinguishing clastic strata of the Bolsa Quartzite and Abrigo Formation from those of the Troy Quartzite*

[These descriptions apply only to the part of the region where Bolsa and Abrigo Formations overlie Troy Quartzite]

	Strata of younger Precambrian age		Strata of Middle and Late Cambrian age	
	Quartzite member of Troy	Chediski Sandstone Member of Troy	Bolsa Quartzite	Sandstones of Abrigo Formation
Position in sequence	At top of Troy. If remnant, is underlain by Chediski Member.	Basal part of Troy; generally makes up most if not all of Troy where Cambrian strata are overlying.	At base of Cambrian sequence; transitional by intercalations of sandy mudstone into Abrigo. Contact with Precambrian rocks generally marked by basal conglomerate.	Sandstone tongue in middle part of Abrigo, separated from Bolsa by fissile sandstones and sandy mudstones. Either unit may lap directly onto Precambrian rocks, and then commonly has—but in places lacks—basal conglomerate.
Lithology	Well-sorted clean quartzite, practically free of feldspar and pebbles. Basal unit coarse grained; rest mostly medium grained.	Poorly sorted pebbly sandstone of well-rounded frosted quartz grains in sericite-clay matrix; weakly cemented to quartzitic. Coarse feldspar grains may be conspicuous. Locally includes coarse muscovite.	Quartzite or firmly cemented sandstone; basal part feldspathic and gritty, with sparse pebbly beds; grades up into medium-grained and ultimately into fine-grained moderately well sorted strata.	Tongue of fine- to medium-grained moderately cemented to quartzitic sandstone. Mudstone is clayey to very sandy; commonly alternates with shaly-parting fine-grained micaceous sandstone. Dolomite sparse; occurs as cement, in thin seams, or local small pods or lenses.
Bedding and related features	Tabular beds thin to thick; cross-stratification may not be conspicuous.	Irregularly bedded; tops of many beds channeled and marked by scattered pebbles or by thin layers or lenses of conglomerate. Convolute lamination in middle part of section. Uppermost beds cross stratified on large scale. Where quartzitic, bedding structures are obscure.	Tabular beds, ½–4 ft thick; well defined in outcrop by partings that etch out much more conspicuously than those of Troy. Internal small- to medium-scale cross-stratification is common.	Sandstone tongue, thin-bedded; bedding tends to be tabular but may be obscure and seem massive where fucoids are abundant. Undulant thin to very thin bedding characteristic of mudstones.
Color	Light-colored. Generally pinkish gray to light gray; may be light brownish gray to medium gray.	Mostly yellowish gray or light brownish gray, but locally includes medium-dark-gray beds. Basal strata locally pale to grayish red. Limonite stains largely restricted to fracture faces in quartzite outcrops of mineralized areas.	Very light gray to grayish red; weathered surfaces generally stained reddish brown. Minimal rusty weathering in vitreous quartzites, which may display conspicuous color banding. Color banding more common in Bolsa but also occurs in other quartzites.	Sandstone tongue pinkish gray to yellowish gray; in some areas, especially where quartzitic, is rusty weathering and forms pale-yellowish-brown to reddish-brown outcrops. Mudstones and shaly sandstones yellowish gray to dark greenish gray, characteristically rusty weathering.

Fossils	None.	None.	*Scolithus* sparse and rarely recognized; apparently can ignore for practical purposes. Transitional mudstones immediately above Bolsa commonly marked by phosphatic linguloid brachiopods.	*Scolithus* and large fucoids sparse to abundant in sandstone tongue, especially in parts transitional into overlying and underlying strata. Phosphatic brachiopods almost characterize very sandy mudstones. Small worm(?) casts and trilobite(?) trails characterize partings in finer grained beds. Trilobite fragments sporadic.
Remarks	Generally more vitreous and massive cropping than Cambrian quartzites. Bedding comparable to basal part of Bolsa, but strata are not gritty.	Where quartzitic, is distinguished by massive outcrops, poor sorting, pebbles, and sparse coarse grains of orange-pink feldspar.	Commonly cliff former. Gritty part lacks undulant bedding planes and lenticular layers of pebbles that characterize lower coarse-grained strata of Chediski.	Mudstones and fine-grained micaceous shaly sandstone beds unknown in Troy and Bolsa. Light-colored sandstones lack pebbles that occur in Chediski.

of these fillings are the outlying remnants of Cambrian sandstone that once extended over the area.

The various channel fills differ in that medium-grained sandstone dominates some and coarse-grained or very coarse grained sandstone dominates others. These differences apparently reflect variations in amounts of locally derived sands. Where a channel cut through the especially friable parts of the Chediski Sandstone Member, for example, the fill may be compositionally very like the Chediski; features seen along the margins may be the only aspects readily of use in distinguishing host from fill. Some fills are bottomed by as much as 40 feet of cobble or boulder conglomerate, well rounded in one occurrence and angular in another; other fills lack a conspicuous basal conglomerate.

The fills are similar in that all contain many beds of very poorly sorted sandstone and in that scattered small pebbles and lenses of granule conglomerate are typical. In some places the sandstone ranges in color from almost white to dusky red, but generally it is yellowish brown to reddish brown. In perhaps four out of five occurrences, certain layers in the upper parts of the fills are conspicuously marked by abundant hematite-cemented concretions. The sandstone is generally very friable. Most is in tabular beds, a few inches to 10 feet thick, in which small- to large-scale cross-stratification etches out conspicuously. Shaly-parting thin layers of sandy siltstone locally separate sandstone beds. A few beds are highly feldspathic, and a few beds are cemented by dolomite or calcite.

This sandstone is wholly different from any sandstone of the Troy, Bolsa, or Abrigo Formations. Features common to most fills suggest that all the sandstone is of one age. These features in the more southern occurrences stand in obvious contrast to the diagnostic features of nearby outcrops of Cambrian strata. It is therefore unlikely that the fills can be correlated with the Bolsa or Abrigo. In general these sandstone beds merge upward into the sandstone that is basal to the Martin, and in a few places sandstone high along the margins of channels seemingly grades horizontally into limestone within the channels. Thus these are almost certainly local basal units of the Martin.

From Christopher Mountain west to the vicinity of Payson, an arkosic sandstone about 50 feet thick separates carbonate strata of the Martin from older Precambrian rocks. Certainly this sandstone, usually speculated to be Cambrian in age, does postdate the Troy Quartzite. Whether it in part or in entirety can be correlated with the channel sandstone remains to be determined.

AGE OF DIABASE AND ASSOCIATED ROCKS

In the southern part of the region, the Troy Quartzite, the Apache Group, and the diabase are separated from Middle Cambrian strata by an unconformity so profound that their assignment to the Precambrian is quite reasonable. In the northern part of the region, though similar age designations can be made by analogy (Shride, 1958), stratigraphic relations directly prove only that the diabase is pre-Devonian; radiometric data that prove Precambrian age have recently become available for this area.

Because there has been much question concerning whether the diabase intrusions are all of the same age, a brief review of the critical relations is worthwhile. The age of the extensive intrusions has been variously inferred to be: (1) Precambrian, (2) Precambrian and Cambrian, (3) post-Cambrian but pre-Devonian, (4) post-Pennsylvanian and probably early Mesozoic or late Paleozoic, or (5) Late Cretaceous or early Tertiary.

A Late Cretaceous or early Tertiary age, based on inferred relations in the structurally complex mineral-bearing belts in and near southern Gila County, where exposures of contacts commonly are poor, is the one most widely accepted for the diabase. In these belts large masses of Paleozoic strata have been interpreted as masses that foundered into some of the larger sills. Furthermore, large diabase bodies that underlie Paleozoic strata locally show steplike contacts with the Paleozoic formations, and faults alined with these steps suggest offsets attributable to intrusions by diabase. These discordant contacts are either obvious fault contacts or are so poorly exposed that their nature cannot be directly determined. Nowhere are diabase sills both overlain and underlain by Paleozoic formations; that is, none of the supposedly engulfed blocks contain sills. In contrast, wherever Apache or Troy strata adjoin or are engulfed by large masses of diabase, they are intruded by sills.

The Martin Limestone is the Paleozoic formation most commonly supposed to be intruded by diabase. The Martin consists partly of cherty dolomite beds, which are comparable to the cherty dolomite of the Mescal and should be converted to silicate-bearing limestone if intruded by diabase. They have proved quite amenable to such metamorphism adjacent to quartz monzonite of Tertiary(?) age (Cooper, 1957, p. 582–587). These rocks, even where exposed within a few inches of diabase, are completely lacking in contact-metamorphic minerals. Where fully exposed contacts are found in adjacent areas, they prove that the Martin rests in sedimentary contact on the diabase body that supposedly intrudes it. I have not found a single example of a chilled selvage in diabase against Paleozoic rocks, although selvages are readily seen along contacts of the diabase body with Apache rocks in the same area. All the supposedly engulfed blocks of Paleozoic strata are blocks that have been faulted against diabase rather than torn loose from their original position by diabase intrusion.

Dark dikes of diorite(?) and andesite basalt that occur very sparsely in Paleozoic rocks have been equated erroneously with the diabase. Ransome (1919, p. 53, 56) clearly stated that small dikes of diabase cut into Paleozoic formations only in a few places and that in assigning a post-Paleozoic age to the diabase he "supposed" these dikes "to represent parts of the same magma that solidified in the larger masses." Darton (1925, p. 254) confirmed Ransome's observations for at least one locality, but he took exception to Ransome's correlation of these dikes with the large diabase sills and suggested that the small dikes were feeders for some of the Cenozoic basalt flows of the region. I have made a particular effort to confirm the reported examples and to find others, and have yet to find in a Paleozoic host rock a dike that compositionally or texturally resembles the diabase.

It is now very clear that the diabases are petrographically somewhat distinctive, that all large masses are kindred, and that none postdate any of the Paleozoic rocks. Abundant stratigraphic evidence shows that they antedate the Paleozoic formations throughout the region.

The Precambrian age of the diabases in the northern part of the region has recently been confirmed by various isotope determinations. In 1955 H. C. Granger and others of the U.S. Geological Survey collected specimens of uraninite and galena for isotopic analyses from four localities in the Sierra Ancha. The uraninite and galena occur mainly as discontinuous veinlets in the Dripping Spring Quartzite, but some of the sampled veinlets transect quartzose aplite dikes which are associated with the principal diabase sill of the Sierra Ancha. This sill transects the highest parts of the Troy Quartzite and is overlain by Devonian strata. On the basis of Pb^{207}/Pb^{206} ratios, determined from both galena and uraninite, L. R. Stieff (written commun. to H. C. Granger, Jan. 26, 1956; Neuerburg and Granger, 1960, p. 775–776) estimated the age to be 1,100 m.y. (million years). Determinations by Silver (1960; oral commun., Nov. 1960) of Pb^{207}/Pb^{206} ratios in uraninitite collected from Workman Creek (3½ miles northwest of Aztec Peak, McFadden Peak quadrangle) and in zircons collected from the granitic differentiate on Reynolds Creek give a minimum age of 1,075±50 m.y.; Silver extrapolated a probable age of at least 1,200 m.y. From potassium-argon ratios determined for biotite in very slightly altered diabase of the same sill, an age of 1,140±40 m.y. has been determined (P. E. Damon, written commun., Dec. 1961; Damon and others, 1962). Thus, throughout southern Arizona the large sills of diabase considerably predate the Cambrian, and the Troy and older formations may be considerably older than 1,200 m.y.

A maximum age for the sedimentary formations can be stated only relative to granitoid rocks that unconformably underlie the Apache Group. In the Dragoon quadrangle an age of 1,660 m.y. has been estimated for the underlying Johnny Lyon Granodiorite (Silver and Deutsch, 1961), and a minimum age of 1,500 m.y. has been determined for Ruin Granite at the south end of the Sierra Ancha (Damon and others, 1962).

GEOLOGIC HISTORY

The following summary of younger Precambrian events emphasizes some features that persist laterally and some that vary. Where possible, the emphasis is on features indicative of an environment or process, in the hope that recogition of like or related geologic phenomena may aid in correlating the Apache and Troy strata with those of younger Precambrian age in surrounding regions.

Throughout the outcrop area of the Apache Group, most of the individual members of the Precambrian formations show remarkably little lateral variation in composition, texture, and bedding features or other internal structures. The bedding structures of the Chediski Member of the Troy Quartzite do change considerably from north to south, but no marked variation in the overall lithology is apparent in the same interval. Lateral differences in thicknesses are largely the effects of erosion before a succeeding unit was deposited. Certain units, the Pioneer Shale and the arkose member of the Troy for example, vary in thickness because of the configuration of the basin of accumulation. Some lateral lithologic changes, such as from dolomite to ferruginous chert or from cherty dolomite to silicated limestone, as observed in the Mescal, are postdepositional but are related to a significant tectonic event. Other variations known to exist seem of little paleogeographic significance.

Prior to deposition of the Apache formations, the older Precambrian rocks were deeply eroded to a surface of generally low relief which was largely on granitoid rocks. In northwestern Gila County, this surface stood above the rest of the region and exhibited noteworthy local relief. The surface was a peneplain, additionally smoothed and swept clean of most of its residual debris by marine abrasion.

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As the first sea of Apache time encroached across the region, wave and current action must have been vigorous. In places the Scanlan Member of the Pioneer is composed of subrounded or angular gravels, almost entirely of local derivation, but the bulk of the Scanlan contains well-rounded pebbles and cobbles of highly resistant quartzite that were deposited several tens of miles from any quartzite source. A moderately thick blanket deposit of arkose might be expected to have accumulated on a granitic terrane. Arkose does make up the lower part of the Pioneer in many areas, but even the cleanest beds are separated by thin layers of tuffaceous siltstone; in many areas fine-grained tuffaceous strata immediately overlie the basal conglomerate and make up most of the formation. Small-scale lamination and cross lamination in these beds, and lack of abrasion of the fragile glass shards, suggest that currents were incompetent to transport great amounts of arkosic debris during much of Pioneer time. Voluminous and extensive falls of ash, from a source as yet unrecognized, must have literally flooded out the arkosic strand deposits of an encroaching sea that was placid during most of Pioneer time. The arkoses found in places at the top of the Pioneer may be remnants of coarser sands laid down during regression of the sea.

An erosional episode must have intervened between the deposition of the Pioneer and the deposition of the Dripping Spring. Except that the contact between the Pioneer Shale and the overlying Barnes Member is sharp and the conglomerate does fill a few shallow channels, there is no direct indication of erosion. The conglomerate and overlying arkose are entirely lacking in tuffaceous material, and thus must mark a break in sedimentation and herald new environments at the source of sediments and at sites of deposition. The thin Barnes Conglomerate Member is not readily explained except as a deposit of a transgressive sea. Local sources for the typical gravels were buried by the Pioneer. Thus the well-rounded and in part very coarse gravel of the Barnes must have been transported long distances during a second episode of vigorous marine erosion at the start of Dripping Spring time. The source terrane, perhaps not now exposed to any appreciable degree, was dominantly granitoid but contained a considerable amount of quartzite of the sort that ribs the older terrane in northwestern Gila County.

The composition and bedding features of the arkose member of the Dripping Spring Quartzite suggest vigorous mechanical erosion of an elevated granitic terrane and rapid transport into a shallow sea. Rather uniform grain size through thick sets of cross-stratified tabular beds implies frequently repeated winnowing by tidal or wave action that was competent to sort and spread the sand widely in thick sheets. Toward the top of the member, feldspar content decreases, the grain size increases slightly, and scattered small pebbles are included in the beds. Perhaps these features hint the termination of this period of shallow marine sedimentation.

The significance of the rather abrupt change from well-sorted sand of the arkose member to ill-sorted mud of the siltstone member is not understood. The environment of the source area at the beginning of late Dripping Spring time may have been modified abruptly, or there may be a hiatus in deposition for which the evidence is not clear.

The thin and irregular bedding of the siltstone member, the many scour-and-fill structures, the few large-scale channels, and the abundant ripple marks and mud cracks collectively indicate deposition in very shallow water. No feasible explanation for the consistent northerly alinement of the scour-and-fill features can be suggested. Subaerial exposure and reworking of the muds occurred repeatedly for some parts of the section and provided opportunity for oxidation. The effects of aeration must have been largely nullified by rapid burial; otherwise the sulfur now represented in the ubiquitous pyrite and the carbonaceous matter would not have been preserved. The muds probably accumulated on broad tidal flats of a shallow sea that sporadically advanced onto a granitic terrane of low relief.

The Dripping Spring sediments were indurated and eroded before the Mescal Limestone was deposited. The thin basal sandstone of the Mescal Limestone was reworked in large part from the upper part of the Dripping Spring Quartzite, but some coarse sand of distinctive type was derived from another source. Similar coarse sand also was supplied during early stages of carbonate sedimentation in the Mescal sea, but otherwise little clastic debris was contributed.

During the early stages of carbonate deposition, circulation in the Mescal sea must have been restricted to permit the high salinity necessary for the precipitation of halite. Probably other evaporites were deposited. Such an environment would also favor the deposition of dolomite or the penecontemporaneous conversion of calcareous mud to dolomite. The formation of chert in such sediments is an enigma. The sea floor was so shallow that wave action was repeatedly effective in reworking mud and partly lithified cherty carbonate strata. After about a third of the lower member had accumulated, the sea waters freshened, and perhaps the freshening was the cause of leaching of the evaporite section. This leaching caused the peculiar founder breccia that makes up the lower part of the Mescal. The sea remained sufficiently saline thereafter, however, to favor the continued formation of dolomite. An almost invariable sequence of clastic-free beds suggests that conditions for sedimentation were uniform and sea-floor abrasion less vigorous during deposition of the upper third of the lower member.

During the deposition of the algal member of the Mescal, the sea floor also must have been remarkably stable and uniform in depth; otherwise the stromatolites would probably occur in bioherms rather than in biostromes (Link, 1950). This environment was all the more remarkable in that the stromatolites are probably intertidal accumulations and in that these tidal flats were persistently hospitable to algal growth for some time over an area of at least 15,000 square miles.

The carbonate members of the Mescal were elevated above sea level, at least briefly, and eroded before additional rocks were laid down. At this time solution cavities began to form in the dolomite. Basalt was extruded onto this erosional surface and in turn was mostly eroded away. At least the northern part of the region then subsided, and very fine grained siliceous mud and subordinate carbonate mud of the argillite member of the Mescal accumulated in a quiet body of water. Following another episode of erosion, several thin basalt flows were extruded.

During the period of instability that followed the deposition of the carbonate members of the Mescal, the Apache strata were broadly warped; consequently the extensive subaerial planation that preceded deposition of the Troy Quartzite locally exposed formations low in the Apache Group, and southeast of the Mescal Mountains all the Apache was removed. The enlargement of solution cavities in the Mescal had continued with each episode of erosion, and solution became particularly effective during the pre-Troy episode. Few, if any, of the dolomite sections completely escaped effects of solution and related silicification. In some areas a karst topography was formed and part of or all the dolomite section was converted to a massive rubble of chert. Concomitantly, hematite, probably a result of laterization of the basalt flows, was concentrated in certain karst areas. Solution and collapse of the dolomite continued during deposition of the Troy.

In the northwestern part of the present area of outcrops, a structural basin a few hundred feet deep existed during early stages of Troy sedimentation, and arkosic sand was literally poured into this basin. The prevalent eastward dip of crossbeds indicates transport from the west, but because the arkose that is visible today in the local basin probably represents only a small part of a much larger deposit, these attitudes may not be significant. Good sorting, fine grain, and very large scale steep-dipping cross-stratification suggest eolian deposition, but these deposits might as likely be a local phase of a rapidly growing delta.

The sharp contact between the arkose member and the Chediski Member of the Troy may reflect a significant erosional hiatus or may merely indicate a minor break in sedimentation. Through most of the region the base of the Chediski Member is marked by conglomerate comparable to the Barnes Conglomerate Member and, like the Barnes, was necessarily derived from a distant source and was spread widely by a vigorously transgressing sea. The Chediski Member represents an episode in sedimentation quite different from that of the arkose member and therefore was perhaps separated from it by an appreciable length of time.

The sands of the Chediski Member are coarser or otherwise entirely different from any sands deposited previously in the younger Precambrian sequence, and no part of the member, except small amounts of gravel, could have been derived from these older strata. The sands were derived, however, from granitic terranes similar to those that furnished the arkoses that are lower in the sequence. Angular pebbles and cobbles of Pinal Schist, Ruin Granite, Dripping Spring Quartzite, and Mescal Limestone, which in places are in abundance in sections south of Globe, indicate vigorous erosion of Apache strata in nearby areas during Chediski time.

Certain features suggest an eolian origin for sand in the Chediski Member, but other features oppose this interpretation. Typical ventifacts at the base of the member do confirm an eolian stage in the formation of the sand; the excellent rounding and characteristic pitting of quartz grains throughout the member and the moderate to high-angle cross stratification in the upper third of the member also suggest eolian deposition. Modern dune sand tends to be fine grained and well sorted (Pettijohn, 1949, p. 232–236), and in this respect the coarse fractions of the upper and lower units of the Chediski are definitely atypical. Perhaps the convoluted strata are sorted adequately enough to be of eolian deposition, but they are more reasonably explained as individual beds hydroplastically deformed in place as each was being deposited (Ten Haaf, 1956), rather than as dune sands that slumped when saturated by water of an encroaching sea. In the southern part of the region, the Chediski

Member lacks virtually all features suggestive of eolian origin, except for the degree of rounding and the abundant pitting of the grains. Irregular bedding, channeling, and lenses of conglomerate between beds might suggest fluviatile deposits. However, many of the conglomerate lenses seem to be lag gravel concentrated by the winnowing of pebbly sands. This gravel tends to form broad sheets rather than local channel fills, and thus suggests concentration on a sea bottom.

All features considered, the Chediski Sandstone Member is possibly best interpreted as a deposit laid down in a sea which transgressed the area from south to north or from southwest to northeast and which was so abundantly supplied with coarse quartz-feldspar sand that only crude sorting could be accomplished. Great volumes of sand and subordinate amounts of quartz gravel, previously rounded, mechanically etched, and partially sorted by wind, must have been available for deposition into this sea. The sea bottom must have been a site of considerable agitation for formation of blanket deposits as extensive as those of the Chediski Member.

The quartzite member of the Troy calls for a different paleogeographic interpretation. The quartzite member may well have had a source of clastics in common with the Chediski Member; if so, these clastics were subjected to considerable transport and sorting, and accumulated in a marine environment as thin layers of well-rounded well-sorted quartz sand, virtually free of other minerals or pebbles. Its uniformly sharp contact with the Chediski Member suggests a hiatus rather than an abrupt change in environment.

Undoubtedly the Troy was once deeply buried by additional Precambrian strata; otherwise the diabase now seen in the uppermost parts of the thickest sections of Troy could not have been emplaced as thick sills. We do not have a sound basis for estimating the additional thicknesses of younger Precambrian strata once in southern Arizona. The diabase in the top of the Troy is petrographically identical to that emplaced at least 3,000 feet lower and might indicate an equivalent minimum cover on the Troy. If the Apache Group and Troy are correlative with the Unkar Group of the Grand Canyon, as Darton (1925, p. 36) surmised, additional thicknesses of at least 10,000 feet may have been deposited and removed in Precambrian time.

All strata of the Troy Quartzite and the Apache Group were deposited on a shallow stable continental shelf. Formations and even members of the sequence are individually very different from each other; the differences suggest abrupt changes in the environment of deposition, in the mode of transport, or in the environment at the source, but not in the kind of materials. Some of the abrupt changes are marked by unconformities, and additional unconformities may yet be recognized. The sequence is analogous to the sequence of Paleozoic formations in the same area, and I believe, even without consideration of the missing strata, that it may well represent as long a time interval.

After lithification of the Troy, the younger Precambrian strata were first locally faulted and folded, then extensively intruded and displaced by diabase intrusions, and finally uplifted regionally and deeply eroded, all in Precambrian time. Despite this long history of deformation, most of the younger Precambrian strata were virtually flat lying when Cambrian seas encroached across the region. Northeast and southeast of the area of present outcrops, these formations were completely removed prior to Cambrian sedimentation, and elsewhere they were further eroded before Devonian sedimentation. The remnants were then deeply covered by Paleozoic formations and everywhere remained almost horizontal at least into Mesozoic time. During Tertiary time the younger Precambrian formations were extensively faulted and tilted throughout the Basin and Range region. Only since this orogeny have the Troy Quartzite and Apache Group been again widely exposed to erosion. Farther north, in the Colorado Plateau, these strata have remained almost horizontal to the present time.

INTERREGION CORRELATIONS

Our knowledge of younger Precambrian rocks elsewhere in the Southwestern United States is still so scant that interregion correlation is largely impracticable; but past speculations can be improved. There is now reason to postulate that the sequence of environmental features and unconformities recognized in the Apache and Troy section may be partly duplicated in the Grand Canyon Series. A search for similar indicators of sequence in other shelf deposits of the Southwest may provide basis for additional correlations. Aside from being a potential source of material for isotope dating, diabase should be viewed with the thought that the singular petrography of the southern Arizona intrusions may be common to the diabase of each of the widely separate terranes; further, the diabase may mark a tectonic event common to all.

Darton (1925, p. 36; 1932) noted that formations of the Apache Group have some similarity to the younger Precambrian rocks of the Franklin Mountains and Van Horn areas of western Texas, and was particularly impressed by their similarity to parts of the Grand Canyon Series of northern Arizona. The Grand Canyon Series has been divided into two groups of formations: the Unkar, or lower group, about 7,000 feet thick, and the Chuar, or upper group, about 5,000 feet thick (Walcott, 1894, p. 508–516; Van Gundy, 1946, p. 1902). Although Darton suggested that the basalt flows of the Apache Group might have counterparts in the basalt flows found in the upper part of the Unkar Group, he speculated in particular that the Troy Quartzite and parts of the Apache Group might correlate with the lower part of the Unkar. Stoyanow (1936, p. 473–474), who preferred a provisional correlation with the Chuar Group, and Hinds (1935, p. 30; see 1937 supp., p. 1994–2000, of Stoyanow, 1936) have taken exception to Darton's suggestions.

Formations of the Apache Group crop out through a north-south interval of 160 miles, and features distinctive of the individual units are remarkably consistent throughout this interval. The southernmost exposure of the Grand Canyon Series is only about 135 miles distant from the northernmost outcrop of the Apache Group (fig. 1); therefore, some of the distinctive characteristics of the Apache Group and Troy Quartzite might well be duplicated in the Grand Canyon Series.

The descriptions of the Grand Canyon Series by Noble (1910; 1914, p. 37–60, 80–83) and Walcott (1894, p. 508–518), supplemented by the general statements of Hinds (1935; 1936) and Van Gundy (1934; 1951), do suggest that comparable features can be found. Near the bottom of the Bass Limestone, which is 200–330 feet thick and is the lowest formation of the Unkar Group if the underlying thin Hotauta Conglomerate is excepted, Noble noted distinctive chert layers "dotted with small cubic depressions" that "strongly resemble salt hoppers." Perhaps these layers are counterparts of vesicular chert layers in the lower part of the Mescal that contain halite molds. Massive irregular chert nodules, jasper, intercalated red or purple shale and thin arkosic sandstone, undulatory banded cherty limestone, and their interrelations and metamorphic equivalents in higher parts of the Bass Limestone certainly suggest the solution phenomena and related secondary features common in the upper parts of the Mescal dolomite and limestone sections. Beds of stromatolitic limestone are noted and illustrated for the Bass Limestone, but their relative stratigraphic position has not been stated. It may be significant that where sections of the Bass are thin and solution(?) features suggest erosion of the top of the formation, no mention of algal structures has been made. According to E. D. McKee (oral commun., Aug. 1959) the algal structures in the Bass Limestone are of *Collenia* aspect like those of the Mescal. To the best of his knowledge, the stromatolites in the thin limestone units of the Chuar Group are all "large biscuit-forms," entirely different from those found in the Mescal. Indeed, the scant descriptions of the limestone of the Chuar hint that they have few features in common.

The Hakatai Shale, which overlies the Bass Limestone and ranges in thickness from 500 to 580 feet, is collectively characterized as comprising argillaceous shale, dense, hard, and varicolored jasper, blue slate, and fine-grained pink or red sandstone, quartzite, or mudstone—a description that could fit the argillite member of the Mescal very well. Neuerburg and Granger (1960, p. 766) have presented semiquantitative spectrographic analyses of one specimen of the Hakatai Shale and one specimen from the argillite member of the Mescal. Both specimens were unusual in their high content of potassium—about 10 percent in the Hakatai Shale and appreciably more than 10 percent in the Mescal argillite (H. C. Granger, oral commun., Oct. 1960). The two specimens show remarkable agreement in trace-element content, and for no element except sodium (Mescal argillite, 0.7 percent; Hakatai Shale, 3 percent) were the analyses notably different. The two specimens, chosen only as seemingly representative of their respective units, are virtually indistinguishable megascopically. However, they are very different from siltstone of the upper member of the Dripping Spring, which also contains unusually large amounts of potassium. If such data are representative, singular chemical characteristics may be an aid in correlating younger Precambrian formations of different parts of the Southwest.

In certain aspects the Shinumo Quartzite, which overlies the Hakatai Shale and which is about 1,500 feet thick, resembles the Troy Quartzite. Although Noble (1914, p. 51–53) did note lenses of conglomerate, he characterized the sandstone and quartzite of the Shinumo as composed of well-rounded well-sorted fine grains[1] of quartz. One of Noble's statements implies that the lowest 400 feet of the Shinumo section is arkosic. Higher parts of the section are at least in part massive and are white, purple, or red. Walcott (1894, p. 511) and Noble made special comment on "curiously twisted and gnarled" layers of white or purple sandstone which are suggestive of the convoluted sandstone of the Chediski Sandstone Member. None of the descriptions indicates that quartzite like

[1] From the context of Noble's descriptions, his fine grained designation may be equivalent in part to medium-grained or even coarse-grained designations of the Wentworth scale.

that of the upper member of the Troy constitutes a distinct separate unit of the Unkar section. However, E. D. McKee (oral commun., Aug. 1959) confirmed that arkose and quartzite like those of the lower and upper members of the Troy do occur and stated that white, purple and mottled sandstone texturally similar to the Chediski Sandstone Member is conspicuous in the Shinumo Quartzite. The large-scale cross-stratification so conspicuous in the arkose member of the Troy was not noted by McKee, but he did see convolute structures very like those in the Chediski.

The Dox Sandstone, which overlies the Shinumo Quartzite and is 2,300–3,000 feet thick, is apparently highly micaceous, is commonly friable, and has shaly partings; except for its basal beds, it seemingly has no counterpart in the southern Arizona sections. The basalts of the Apache Group are thin compared to the 800–1,000 feet of basalt flows noted near the top of the Unkar Group. The differences in thickness can be overlooked because the southern basalts were greatly thinned by pre-Troy erosion; if the Troy and Shinumo Quartzites are correlative, however, the basalt sequences can hardly be stratigraphic equivalents. The sedimentary formations, especially those of the Chuar Group, that overlie the Unkar basalts mostly are thin-bedded nonresistant strata unlike any of the Apache formations.

Correlatives of the Dripping Spring Quartzite and Pioneer Shale have not been suggested in this résumé, but it may be reasonably surmised that, owing to lapout, they were not deposited at the base of the Grand Canyon sequence. Past descriptions have not hinted of unconformities that should be recognized in the Unkar Group if the ideas presented have merit. It would seem that collection of a modest amount of additional information might, despite the differences, permit firm correlation of the Apache and Troy rocks with those of the lowest 1,500 feet of the Unkar Group.

Not far west of the Nevada-California boundary and astraddle the Inyo County–San Bernardino County line in California, a younger Precambrian sequence of formations known as the Pahrump Series (Hewett, 1940) crops out. The belt of Pahrump outcrops, according to Noble (1941, p. 949), does not exceed 25 miles in width and extends northwest from the Kingston and Shadow Mountains 75 miles into the ranges that border the southern part of Death Valley. The Pahrump belt is 200–250 miles west of the westernmost outcrop of the Grand Canyon Series and about 350 miles northwest of the Apache outcrop area.

The Pahrump Series, 5,500–7,000 feet thick where described in some detail (Hewett, 1956, p. 25–28; Wright, 1952, p. 7–15), has been subdivided into three formations—in ascending order the Crystal Spring Formation, Beck Spring Dolomite, and Kingston Peak Formation. The Crystal Spring Formation, 1,600–2,300 feet thick, includes units of feldspathic quartzite and siltstone, red siltstone or shale, quartzite, and dolomite that in places is associated with massive chert units. The Beck Spring Dolomite, 1,000–1,300 feet thick, is largely a monotonous sequence of massive-bedded dolomite but includes subordinate beds of quartzite and shale. Most of the Kingston Peak Formation, 1,000–2,000 feet thick, is conglomerate or conglomeratic quartzite, but the upper and lower parts are shaly-parting sandstone.

Noble (1934, p. 174), who has worked extensively in both the Grand Canyon and Death Valley regions, has commented without giving details that parts of the Pahrump Series are strikingly similar to the Grand Canyon Series. Apparently the Crystal Spring Formation, in particular, includes units comparable with those in the Apache Group. In May 1952, I viewed briefly some of these strata, in the vicinity of Tecopa and the Saratoga Hills, under the guidance of L. A. Wright of the California Division of Mines and Geology. The Saratoga Hills area, 35 miles northwest of Baker, Calif., is the locality of the most detailed stratigraphic descriptions of the series yet published (Wright, 1952). A thick feldspathic quartzite unit makes up the lowest part of the Crystal Spring Formation. Overlying this quartzite, and midway in the section, is a purple shale or siltstone unit, which appears in outcrop remarkably like the siltstone of the Pioneer Shale. Overlying this shale in the Saratoga Hills is a brown-weathering dark fine-grained quartzite unit which somewhat resembles the siltstone member of the Dripping Spring Quartzite. The dolomite and chert that make up much of the upper part of the Crystal Spring Formation have many features, including stromatolite beds, in common with the Mescal Limestone. It would be presumptuous, however, to suggest any correlation of the Apache Group and Pahrump Series at this time. As Hewett (1956, p. 26–28) has pointed out, lateral variations of the Pahrump are known, and problems exist in correlating units of the series from place to place within the Death Valley region.

In the Franklin Mountains, immediately north of El Paso, Tex., a Precambrian sequence that is structurally little disturbed was recognized by Richardson (1909). The uppermost formation of Precambrian age is extrusive rhyolite, as much as 1,400 feet thick, which unconformably underlies the Bliss Sandstone of Cambrian age. The rhyolite in turn unconformably overlies a sequence dominated by beds of quartzite and sandstone, 2,600 feet thick, which Richardson termed the Lanoria Quartzite. These formations, probably owing to the lack of deformation, have been provisionally regarded as occupying the same place in the Precambrian sequence as the Apache Group and the Grand Canyon Series (Darton, 1932). Recently, Harbour (1960) recognized that the Lanoria Quartzite unconformably overlies an indurated rubble of basaltic debris, which he termed the Mundy Breccia. The Mundy Breccia, as much as 250 feet thick, in turn rests on the channeled top of a limestone formation, which Harbour named the Castner Limestone. Excluding diabase sills from the type section Harbour measured, the Castner Limestone is exposed through a thickness of almost 800 feet; its base is intruded by granite, and the section is therefore not complete. The stromatolite form-species *Collenia frequens* characterizes a 4-foot bed near the bottom of the Castner section, and in other aspects much of the formation resembles the Mescal Limestone. The Mundy Breccia might, of course, have a counterpart in the basalt flows of the Apache Group. The Lanoria Quartzite seemingly is not particularly like the Troy Quartzite.

Strata of younger Precambrian aspect also are exposed near Van Horn, Tex., about 100 miles southeast of El Paso. These strata were tentatively correlated with those of the Franklin Mountains by King (in King and Flawn, 1953, p. 125–131) after a careful consideration of the pitfalls in equating the formations. King also considered the problems of correlating the Precambrian strata of west Texas with those of Arizona. Harbour (1960) took exception to King's correlations of formations between the Franklin Mountains and Van Horn areas.

The information bearing on the problems of correlation of the west Texas formations is given in full in the cited publications. Without going into detail, I would suggest that if limestone formations were not common to both the Texas and Arizona sections, there would be little inclination to equate the sections. If structural and plutonic events were given more emphasis in comparisons, one might be inclined rather to correlate the west Texas strata with those underlying the Apache Group. Until additional bases for correlation are available, little weight should be given to lithologic similarities of the Precambrian formations in the two regions.

Diabase intrusions of Precambian age are common in all the Precambrian terranes mentioned in the preceding discussion. However, if the relatively thin Apache and Troy terrane is excepted, the intrusions are restricted stratigraphically to only certain parts of each sequence. In the Grand Canyon, diabase occurs in great volume only in the Bass Limestone and the Hakatai Shale, although thin sills do exist in the Shinumo Quartzite and the Dox Sandstone (Noble, 1914, p. 55 and pl. 1). Diabase has not been reported in the Chuar strata, which by analogy should be favorable host types for diabase inflation. The inconspicuous unconformity that marks the boundary between the Unkar and Chuar Groups (Walcott, 1895, p. 325), or one of the similar unconformities stratigraphically not far distant from that contact (Van Gundy, 1951, p. 954–957), may therefore have more significance than is now appreciated. In the Death Valley region, diabase intrusions have been recognized only in the Crystal Spring Formation. According to Wright (1952, p. 15) the Beck Spring Dolomite's lack of diabase intrusions and the Kingston Peak Formation's abundant diabase detritus, almost certainly derived from the known sills, strongly suggest that "the intrusion had taken place before the basal Beck Spring dolomite beds were deposited." In the Van Horn area of Texas, greenstone bodies that possibly represent diabase sills seemingly are restricted to the Allamoore Formation of the Precambrian sequence (King and Flawn, 1953, p. 78–79). In the Franklin Mountains, several thin diabase sills intruded the Castner Limestone; these may predate the Lanoria Quartzite and definitely are older than the Precambian granite that intruded the Castner Limestone (Harbour, 1960). If diabase intrusions have any significance for correlation, the suggestion might be made that the Apache and Troy strata correspond to the oldest of the younger Precambrian formations in the Southwest.

Feldspathic differentiates, which might provide zircons suitable for isotope dating, have been noted in the thickest sill in the Unkar strata. Such differentiates, if they exist in the diabase of the Death Valley region, have not come to my attention. Descriptions indicate that some of the thicker diabase bodies of that region are multiple thin sills in which granitoid differentiates would not be expected. Sills of the Franklin Mountains are also too thin. Greenstone of the Van Horn area is not a likely source of suitable material. Owing to pervasive alteration, biotite suitable for potassium-argon dating would be difficult to find in any of the occurrences, except possibly those of the Grand Canyon. There are hints in the descriptions that diabases of the canyon are petrographically kindred to those of southern Arizona. For lack of data, this speculation cannot be extended to the other terranes.

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