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Some aspects of younger Precambrian geology in southern Arizona

by

Andrew F. Shride

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3. Some aspects of younger Precambrian geology in southern Arizona, by A. F. Shride. 387 p., 16 figs., 5 tables.

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Some aspects of younger Precambrian geology

in southern Arizona

by Andrew F. Shride

Abstract

The younger Precambrian rocks of southern Arizona crop out only in the southeastern quarter of the state. They are widely exposed and relatively little deformed in northern Gila County and the southwestern corner of Navajo County, areas that are structurally a part of the Colorado Plateau. Farther south, where Basin and Range structure prevails, outcrops are less extensive and are scattered; and -- largely as a consequence of Late Mesozoic and Cenozoic deformation -- their structural relations ~~of~~ with younger formations are comparatively obscure. Nevertheless, in the southern part of the region of outcrop, these Precambrian rocks are well exposed for study in the Apache, Hayes, Mescal, and Dripping Spring Mountains, near Roosevelt Dam and where Aravaipa Creek breaches the west flank of the Galiuro Mountains.

A principal purpose of this paper is to report the structural interrelations of the younger Precambrian formations, as a base for a better overall interpretation of the structural geology of southeast Arizona.

The younger Precambrian rocks of southern Arizona comprise the Apache group of layered formations, the overlying Troy quartzite, and large diabase bodies intrusive into and coextensive with the sedimentary formations. The Apache group, 1,230 to 1,600 feet thick, is herein redefined to include, in ascending order: the Pioneer shale, the Dripping Spring quartzite, the Mescal limestone and an unnamed formation of basalt flows. The Scanlan and Barnes conglomerates, formerly designated as separate formations, are considered merely basal conglomerates of the Pioneer shale and Dripping Spring quartzite, respectively. Although arkoses constitute a part of the Pioneer formation, throughout the region tuffaceous siltstones and mudstones compose the bulk of the formation. The Dripping Spring quartzite is divisible into two distinctive members: a lower massive-cropping, firmly indurated arkose member; and an upper highly feldspathic, thin and irregularly bedded, slabby parting member. An unconformity is now known to separate the Dripping Spring and Mescal formations, and necessarily the boundary between these formations is redefined. The Mescal limestone is comprised of three members: a lower thin- to medium-bedded member of cherty dolomite or metamorphic limestone; a middle member of like composition, but rendered conspicuous by the massive-cropping stromatolitic biostrome that everywhere makes up the lower one-half or more of the member; and an upper member of thinly-laminated argillite, which now crops out only in the northern half of Gila County. An erosional unconformity and, very locally, small remnants of basalt flows separate the middle and upper members of the Mescal. Like basalt flows unconformably overlie the Mescal throughout the region; generally the basalt formation is thin, but in some areas a sequence of flows almost 400 feet exists -- a thickness much greater than previously recognized.

The Troy quartzite is not accurately or simply described as a cross-bedded pebbly quartzite formation, as generally conceived. Rather it is comprised of three lithologically distinctive members. The lowest member, of which remnants exist only in northwestern Gila County, is a medium-grained well-sorted arkose conspicuously characterized by very large-scale cross-stratification. The middle or Chediski sandstone member, which is the basal member through most of the region, is largely a very poorly sorted, light-colored, pebbly sericitic or feldspathic sandstone. In the northern part of the region a large part of the member is massive-cropping and is characterized by pre-consolidation slump structures; in the southern part the member is less massively bedded, but is typified by irregular bedding and by thin layers and lenses of conglomerate. Locally the Chediski member is a quartzite, and recognition of its distinctive bedding features and poor sorting is required to differentiate it from the upper quartzite member. The latter is an evenly bedded, well-sorted clean quartzite, virtually free of feldspar and pebbles. Where the three members are most fully remnant, the Troy is at least 1,200 feet thick, or almost three times the thicknesses usually cited for the formation.

The thin-bedded shaly-parting fossiliferous sandstones and quartzites found at some places in seeming continuous sequence with the Troy formation, and previously held to characterize the upper part of the formation are strata equivalent to the Bolsa quartzite (Middle Cambrian) and a northern clastic facies of the Abrigo limestone (Middle and Upper Cambrian). These sandstones and quartzites are actually separated from the Troy quartzite by an unconformity that represents a hiatus of about 500 million years, so the Troy must now be considered Precambrian rather than Cambrian in age.

Formations of the Apache group were deposited in shallow shelf seas, and are products of at least three and possibly as many as five separate episodes of marine invasion. The arkosic sediments available to the first Apache sea were overshadowed by voluminous contributions of fine-grained felsic ash, widely distributed from a source unknown to form the bulk of the Pioneer shale. During a second transgression of the sea, marked only subtly as an erosion surface but defined strongly by a basal conglomerate (Barnes conglomerate) typical of transgression, a 300-foot sequence of fairly well sorted arkoses was first deposited. Then finer feldspathic muds accumulated rapidly, probably on extensive tidal flats. After lithification and some erosion of these strata, the carbonate members of the Mescal accumulated in a very shallow sea in which circulation was somewhat poor, as indicated by the deposition of evaporites and primary or early-formed cherty dolomites. The environment of the sea floor must have been remarkably uniform throughout a broad area, because a uniform sequence of deposits was formed everywhere in the region of present outcrops. The region was uplifted, solution cavities began to develop in the cherty dolomites, and continued to form while basalt flows poured out over their eroded upper surfaces and were in turn eroded. Then very fine-grained siliceous muds were deposited in a quiet body of water and additional basalt flows accumulated on these.

During this period of instability the Apache group was broadly and differentially warped. And before the first Troy sea encroached across the region there was a long period of erosion, in which karst topography and karst breccia formed in areas where the dolomites were exposed, and some outcrops were partially to almost completely silicified. Some secondary chertification was affected everywhere in the region. In a few karst areas lateritic debris, probably formed by decomposition of the basalts, accumulated as siliceous hematite deposits. Generally the formations below the Mescal were not affected by this erosion, but in a few areas a considerable part of the Apache section was eroded away.

The three members of the Troy quartzite are products of three distinctly different environments. The well-sorted arkosic material of the lowest member probably deposited rapidly as foreset beds of a delta, but accumulated only in the deepest parts of the Troy basin. The Chediski member probably represents great volumes of eolian sand rapidly redistributed in a transgressing sea which was not competent, although a site of violent agitation, to affect good sorting. Material of probably similar sources was available during the third phase of Troy sedimentation, but it was well sorted and laid down in relatively thin beds of clean quartz sand, and cemented as a quartzite. Still additional Precambrian strata, not now remnant in southern Arizona, must have been deposited in considerable thicknesses on the Troy quartzite.

After lithification of the Troy, at least in the northern part of the region, the late Precambrian sequence was faulted and folded along narrow northerly trending belts. Between these belts, which are widely spaced, the strata remained virtually horizontal. Thereafter, possibly with the narrow structural belts as principal zones of access for the magma, the Apache formations and Troy quartzite were extensively inflated and displaced by diabase intrusions.

The greatest volume of diabase was emplaced as sills, which are from a few inches to more than 1,000 feet thick and in some areas constitute as much as half of the thickness of younger Precambrian formations. The sills are largely concordant, but different parts of a given sill were intruded along different stratigraphic horizons. The concordant portions of the intrusions at different horizons are connected by tabular discordant masses, which variously transect the host rocks at low to high angles and commonly have step-like boundaries. The tabular masses joining sills are commonly thick, but dikes that do not connect sills are narrow and relatively few. In some areas multiple sill intrusions along a given horizon are characteristic; in other areas multiple injections are rare. Most of the many pre-Paleozoic faults that offset Apache and Troy strata are direct effects of diabase inflation.

Although the sills are coextensive with all the younger Precambrian strata, they are particularly voluminous in the least competent stratigraphic units and are largely absent in the most competent units. As a consequence of this habit the cherty dolomites of the Mescal were so widely metamorphosed to silicate-bearing limestones that calcitic strata are ordinarily considered representative of the formation. Adjacent to the intrusions, the other formations were metamorphosed in lesser degree.

Gravity stratification of the olivine-bearing diabases is subtly developed in some of the thicker sills, but is not noteworthy in most intrusions. The intrusions include a variety of highly feldspathic differentiates; diabase pegmatite, granophyre, and a variety of aplitic rocks are the most common. The differentiates comprise only a very small part of the diabase and are in greatest amount in certain sills. The largest masses occur along or near the hanging walls of discordant portions of thick sills. Quartzose differentiates apparently formed in volume only where xenoliths of feldspathic host rocks reacted with the diabase magma. Otherwise assimilation was virtually a nonexistent process.

One particularly extensive granitoid mass has furnished zircons suitable for dating studies. The lead isotope ratios determined for these zircons confirm ratios derived in 1956 from galena and uraninite veinlets, which establish the minimum age of the diabase as 1.1 billion years.

The later Precambrian terrane was uplifted anddeeply eroded before the transgression of Cambrian seas, which probably had shorelines not far north of central Gila County, or at the most accumulated only thin deposits of clastic sediments in the northern part of the region. In the southernmost part of Arizona the Cambrian section comprises a basal quartzite formation (Bolsa quartzite) overlain by the Abrigo formation, which is predominantly of carbonate strata. Farther north the Abrigo formation is dominated by clastics in part much like those of the Bolsa quartzite. In gross aspects these clastics and those of the Bolsa resemble the quartzites of the Troy formation, and in places have been identified with the Troy. Adequate criteria for separating these formations have now been recognized.

After another long episode of non-deposition and erosion, which resulted in Cambrian deposits of the northern part of the region being stripped away, the Martin limestone of Devonian age was deposited. The Martin is mostly of dolomite, limestone and shale, but in the northern part of the region channels in the pre-Martin surface were sites of local accumulations of sandstone, in places at least 250 feet thick. These sandstones also have been mistakenly included with the Troy; again, these rocks are readily distinguished.

Through this long history, spanning at least three-quarters of a billion years after the Troy was deposited, and apparently until the Laramide orogeny of Late Cretaceous or early Tertiary time the younger Precambrian formations remained virtually horizontal. During the Tertiary the region was uplifted; the southern part was subjected to intensive faulting and the faulted masses were tilted. In the part of the region north of the Apache Mountains the younger Precambrian formations were little deformed, but they were widely stripped of younger formations; additionally, in the southern part of the region great parts of the Apache strata were stripped away and lesser remnants were buried beneath continental deposits of the Cenozoic.

The Apache group, the Troy quartzite, and the extensive diabase intrusions are important hosts of metal deposits in east-central Arizona. It is therefore worthwhile that the effects of late Precambrian diastrophism versus those of Cenozoic time be distinguished.

Formations of the southern Arizona series cannot yet be correlated with those of the Grand Canyon series. But comparable lithologic features in the younger Precambrian rocks of both areas suggest that correlations of the Troy and Apache formations with the Unkar group of northern Arizona, and perhaps ultimately with the Pahrump series of the southern Death Valley region, may be feasible when more information is available for these sequences.

Introduction

Geographic settings of the outcrops of younger Precambrian rocks

Younger Precambrian formations crop out in two regions in Arizona. In northern Arizona exposures of the Grand Canyon series, totaling less than 100 square miles in area, are restricted to outcrops in the depths of the Grand Canyon. The Troy quartzite and the Apache group of stratified formations and large coextensive intrusions of diabase, the subjects of this report, crop out much more widely in the southeastern part of the state, principally in a belt 20 to 60 miles wide that extends from the foot of the Mogollon Rim, at the north, southward 170 miles about to the latitude of Tucson (see fig. 1). Additionally scattered outcrops of

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Figure 1.--Distribution of outcrops of younger Precambrian rocks in Arizona in relation to the three principal physiographic regions: (1) plateau region; (2) mountain region; and (3) desert region (Physiographic regions after Ransome, 1919, p. 27; 1923, fig. 2.).

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these formations exist in an area 50 to 65 miles south of Phoenix. Although the early descriptions of the Grand Canyon series form a part of the classical geologic literature of Arizona and this series is the more widely publicized, the geology of the Troy quartzite and the Apache group is much better understood, owing the economic interest in the area of their outcrop.

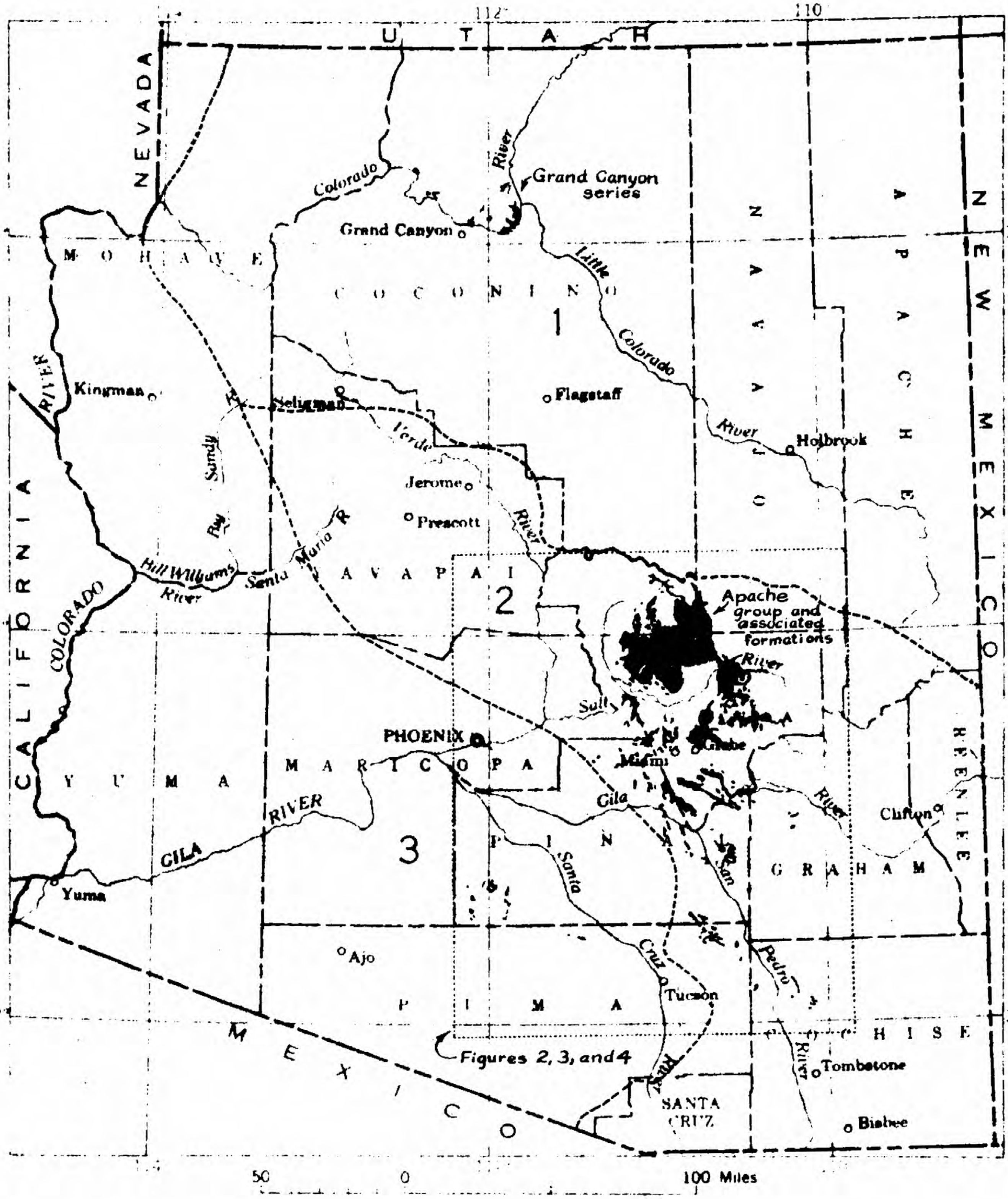

FIGURE 1. - Distribution of outcrops of younger Precambrian rocks in Arizona in relation to the three principal physiographic regions: (1) plateau region; (2) mountain region; and (3) desert region. (Physiographic regions after Ransome, 1919, p. 77; 1923, [illegible])

Outcrops of the younger Precambrian strata of southern Arizona are distributed as shown on figure 2; the coextensive intrusions of

Figure 2.--Map showing outcrops of younger Precambrian strata and coextensive diabase intrusions in southeastern Arizona

Precambrian diabase are also shown in as much detail as the scale of the map will permit. Many intrusions of diabase that outcrop as narrow bands are not shown separately from the stratified formations. The quadrangles in which studies of the younger Precambrian rocks have been made or are in progress are also indexed on this map. Figure 3, which encompasses

Figure 3.--Index map showing positions of some paleogeographic features of the younger Precambrian formations.

the identical area shown on figure 2, is presented as an index of the geographic positions referred to in the text and the positions of some of the principal paleogeographic features of the younger Precambrian and Cambrian terranes are indicated in as much detail as present knowledge will allow. The scale of figure 3 is inadequate to show a few minor geographic features referred to, but a general location of these features is stated in the text, and all omitted features may be found on the topographic quadrangle maps published by the Geological Survey.

To provide a terrane setting for his original descriptions of the Apache formations, Ransome (1903, p. 14-16; 1916, p. 133-135) divided the state of Arizona into three principal physiographic regions; the plateau region to the northeast, the desert region to the southwest, and the mountain region between (see fig. 1). Physiographers ordinarily divide the state into two provinces: through the central part of the state the boundary between the Basin and Range province on the south and the Colorado Plateau province on the north by general usage coincides with the Mogollon Rim (Fenneman, 1931, p. 381-383). In later description Ransome (1923, p. 1) redefined the boundaries of the mountain belt, and since that time it has been included with the desert region or considered a part of the Mexican Highland section of the Basin and Range province (Butler and Wilson, 1938, p. 9 and pl. 1). For the area of interest here this relegation of the mountain region entirely to the Basin and Range province is unfortunate, because the structural aspects of the northeastern half of Gila County are characteristic of the Plateau province and quite different from those of the Basin and Range province.

Principal distinguishing features of the Plateau province are the approximate horizontal disposition of the layered rocks and their lateral continuity, with only relatively minor structural disturbance. But in the adjacent Arizona portion of the Basin and Range province faulting and the resulting lack of continuity have been important factors in the development of topographic forms. The ranges and intermontane basins owe their relative difference in relief primarily to faulting. During Cenozoic time younger Precambrian, Paleozoic, and Mesozoic rocks were eroded from large parts of the basin areas and the basins were filled with gravel, other fluviatile sediments and lacustrine deposits. Volcanic materials were interbedded with the basin fills and extensively covered any of the older formations exposed during Cenozoic time. Faulting may be of considerable complexity. In some places, as in the Dripping Spring Mountains and in the Globe-Miami mineral district, the rocks underlying considerable areas are broken into small blocks by faults of several trends, so that individual outcrop belts can be visualized as a huge mosaic of tilted fault blocks (Ransome, 1919, p. 76-79), which are covered in part by erosional debris that accumulated during and since the elevation of the ranges.

Using these structural characteristics and related features as criteria for distinguishing between Plateau structure and Basin and Range structure, a fairly definite line can be drawn through Gila County separating the two structural provinces. Beginning north of Roosevelt Lake in the basin occupied by Tonto Creek, this boundary can be traced along the southern foot of the Sierra Ancha, thence east-southeast along the north flank of the Apache Mountains through a point about 15 miles north of Globe, and then southeast along the foot of the southern escarpment of the Natanes Plateau (see fig. 3).

From this boundary between structural provinces north to the Mogollon Rim the younger Precambrian formations are widely exposed (see fig. 2). In this Plateau portion of the Mountain region differential erosion of resistant sedimentary rock units that alternate with less resistant horizontally disposed layered formations has resulted in mesa-canyon topography -- similar to topographic forms of the Grand Canyon, but on a smaller scale. Owing largely to the relief of this area -- commonly 2,500 to 4,500 feet locally -- and to the continuity of exposures, observations of the stratigraphic details of the Apache group, the petrologic details of the diabase and the relations of the Precambrian formations to Paleozoic formations are enhanced. Furthermore, structural features that predate or are a consequence of diabase intrusion can be distinguished readily from features resulting from Laramide or later orogenies. Conclusions drawn from this area can be more fully substantiated than those drawn from observations elsewhere in the region of younger Precambrian outcrops. Within the Plateau province outcrops of the younger Precambrian rocks are surmounted to the north and east (northeast of the Sierra Ancha and east of Canyon Creek, and east of U.S. Highway 60 south of the Salt River) by a thick cover of Paleozoic and younger formations. Northwest and west of the Sierra Ancha the Apache formations are entirely missing owing to pre-Paleozoic and Cenozoic erosion.

The Basin and Range portion of Arizona that includes outcrops of younger Precambrian formations (figs. 2 and 3) can be visualized as mainly of small mountain ranges, ridges, and hills isolated from one another by narrow to broad, generally linear basins that are mainly floored on older Precambrian rocks. These basins, for large parts of the region, are filled with continental deposits of Cenozoic age, to depths of as much as a few thousands of feet. Thus outcrops of younger Precambrian rocks, mainly exposed in the mountain areas, may be widely separated and their structural relations to younger and older formations are ordinarily obscured by Laramide and later deformation. North of the Gila River the intermontane basins are narrower and the younger Precambrian rocks crop out more abundantly and continuously than elsewhere within the Basin and Range province. Owing to exposure in and near the great copper-mining districts of Ray, Globe-Miami and Superior, the formations of the Apache group were defined in this part of the region and to the present time the most complete descriptions of the rocks are from this area. Southeast of Winkelman as far as Arizona Highway 86 younger Precambrian rocks undoubtedly exist in somewhat greater amount than suggested on figure 2, but most of the occurrences postulated in the Galiuro Mountains and contiguous ranges east of the San Pedro River are beneath a thick cover, chiefly of Cenozoic volcanic formations. If complexly deformed erosional remnants of the Apache group on the north slopes of the Santa Catalina Mountains are excepted, west of the San Pedro River only sparse outcrops of the younger Precambrian rocks, exposed in small mountain masses widely separated by broad alluvial plains, exist. Most stratigraphic sections in these exposures are incomplete. Except in the Dragoon quadrangle, and in the Mammoth and Holy Joe Peak quadrangles where geologic studies are currently in progress, little attention has been afforded to details of younger Precambrian geology in areas south of the latitude of Winkelman.

Scope of the report

The writer was first introduced to geologic problems of the Apache group while investigating an iron deposit near the head of Canyon Creek during a period of 4 1/2 months in the winter of 1941-42. From mid-October 1942 through May 1944, and again in July and August 1946, my principal assignment was the examination of asbestos deposits that occur in the Mescal limestone of the Apache group and adjacent to diabase intrusions. This work was concentrated mainly along the canyon of the Salt River (within the Blue House quadrangle) and in and northeast of the Sierra Ancha, but included investigations of small areas as far south as the Mescal Mountains. During these studies observations were restricted largely to the upper formations of the Apache group. Incidental to other assignments in the period 1950-1954, I had the opportunity of making observations throughout the north-south range of outcrops of the Apache formations. In 1953 the McFadden Peak and Blue House Mountain quadrangle were assigned for detailed geologic mapping, but little work was done until July 1954. Between July 1954 and June 1956 parts of the McFadden Peak quadrangle were mapped. In March, April and May 1957, the writer and C. T. Wrucke mapped by reconnaissance methods a strip along U. S. Highway 60 as a contribution to the new geologic map of Arizona, currently (1961) being compiled. This mapping included a belt 10 to 25 miles wide, extending from north of the Salt River south almost to Globe, and added considerably to our understanding of features seen in detailed mapping of other areas. During 10 days in June 1960, in preparation for this report, I took the opportunity of seeing some of the Apache rocks that outlie the areas of most abundant outcrop. In this reconnaissance outcrops as far west as the Vekol Mountains in southwestern Pinal County and as far east as Mt. Turnbull in Graham County were examined. At least passing observation has been made of some exposures of the younger Precambrian rocks in most areas of known outcrop. The one large area of outcrop to which, unfortunately, almost no field study has been given lies west of 111th meridian and north of Roosevelt Lake.

Early in my experience it became quite apparent that details of stratigraphy and lithology of the Troy quartzite and Apache formations were little known: generalizations, derived from observations in small areas, were not wholly applicable throughout the region; and pre-Paleozoic structural events as they affect the present-day disposition and distribution of the younger Precambrian formations have been almost totally unrecognized. These lacks in our knowledge have affected, at least initially, the interpretation of features associated with mineral deposits in or near the younger Precambrian rocks. Examples of misinterpretation persist to the present.

Some instances from my own experiences -- examples for which the reinterpretations are yet not generally appreciated among those interested in the mineral deposits -- are illustrative. Hematitic iron ores occur locally in the Mescal limestone of the Apache group. The gross form and distribution of the first such ore body studied -- the largest body discovered to date -- suggested that particular concentrations of the ores were localized along fault zones, and that the ores might have had a source in emanations from a diabase intrusion (Burchard, 1931, p. 59-60). Under such seemingly reasonable premises my colleague, A. P. Butler, Jr., and I first labored in devising exploration programs and attempting to estimate ore reserves. Only later did we appreciate that the diabase intrusions post-date these ores, and that many of the faults, which might be visualized as localizing channels, post-date even the diabases. Only from observations made after studies of the iron ores terminated was it fully appreciated, as will be detailed later, that the ores are not in anyway hydrothermal or metamorphic phenomena, but are genetically associated with the development of the surface of unconformity that separates the Mescal limestone and the overlying Troy quartzite in some areas. As another example, the stratigraphic and lithologic controls for emplacement of the asbestos deposits were partly misinterpreted and the regional distribution of the deposits not understood, until it was appreciated that the chert content of the Mescal limestone was variable from place to place, that the formation has been pervasively dedolomitized on a regional scale as a consequence of diabase intrusion, and that many faults of the region are genetically associated with the diabase intrusions and, with the diabase, predate the Paleozoic formations. Other, and perhaps more significant examples could be described. During the various observations of the younger Precambrian rocks, the need for better understanding of such problems has been a prime consideration.

This report might include -- if the data were available -- numerous detailed descriptions of the Apache formations in the many areas of outcrop, and the fairly complete paleogeographic synthesis of younger Precambrian geology that such descriptive data would allow. For the lack of adequate information, however, the report must be considerably less ambitious in scope, and is written to present: (1) such general interpretations of the regional geology of the younger Precambrian rocks as are now possible, and (2) to provide sufficient background data that these interpretations can be applied in deciphering the geologic settings of several mineral-bearing districts in southeastern Arizona.

In order to provide uniformity and precision, not previously available in most descriptions of the younger Precambrian formations, in this report descriptive terms are standarized by use of the "Rock-Color Chart"(Goddard and others, 1948) and the Wentworth grain-size scale. In describing stratified rocks the terms proposed by McKee and Weir (1953) are used. For the sake of generalization, deviations from these standards of terminglogy are occasionally made; these deviations will be apparent from the content of the individual statements.

Acknowledgements

During many discussions since 1941, both in the field and in the office, E. D. Wilson, geologist of the Arizona Bureau of Mines, has provided information from his exceptional knowledge of the regional geology of Arizona. As a consequence of his close association with N. H. Darton in the early 1920's, Dr. Wilson was able to direct me to several of the seemingly critical localities noted by Darton; in addition he guided me to areas in northern Gila County where relations at the base of the Apache section are particularly well exposed, and to three localities in the Mescal Mountains where Cambrian fossils had been collected. During the 1940s, B. S. Butler of the University of Arizona offered much helpful comment. N. P. Peterson, during the course of his studies for the Geological Survey since 1943 of the Globe-Miami mineral belt, has advised me of many aspects of the geology of that area. During 1954-56 H. C. Granger and R. B. Raup studied uranium deposits in the Dripping Spring quartzite and from this work, which is being prepared for publication by the Survey, furnished much detail on the stratigraphy of the formation. To them goes the credit for collecting uraninite and galena specimens from which age determinations were first derived. C. T. Wrucke, who aided in detailed mapping of the McFadden Peak quadrangle in 1955-56 and in the reconnaissance mapping along U. S. Highway 60 also did considerable petrographic examination of diabases collected during these studies and that of Grangerand Raup. For his careful observations and stimulating discussions I am very grateful.

From attempts to understand the geologic events recorded in the younger Precambrian rocks of northern Gila County, the principal conclusions of this report concerning the relations of the Troy quartzite and the great intrusions of Precambrian diabase to the overlying Paleozoic formations were evolved as working hypotheses (Shride, 1958). But my postulations concerning relations to the Cambrian formations would not have been substantiated by tangible evidence without the generous cooperation of Geologic Survey colleagues, who currently are or recently have mapped areas in southern Gila County and farther south. In and west of the Little Dragoon mountains, as early as 1947, J. R. Cooper observed Bolsa quartzite resting unconformably on diabase sills that inflate the Dripping Spring formation. In 1956 S. C. Creasey orally provided information on Cambrian sandstones and quartzites and on the lithology and thicknesses of the lower formations of the Apache group for several localities from the Mescal Mountains south to Mammoth. These descriptions suggested that the relations noted by Cooper might be found in the areas north of Mammoth. In May 1958 M. H. Krieger guided me to two localities east of Winkelman, where fossiliferous sections of Cambrian quartzites and sandstones rest unconformably on diabase sills, which inflate the Dripping Spring formation. Later Mrs. Krieger, in the course of mapping the Holy Joe Peak quadrangle southeast of Winkelman, was the first to recognize critical single outcrops in which the Bolsa quartzite can be demonstrated to unconformably overlie both Troy quartzite and large diabase intrusions into the Troy quartzite. In May 1960 she directed me to these outcrops. In November 1960 a field conference with these and other interested geologists in attendance was held to review some of the problems of Cambrian-younger Precambrian stratigraphy. In addition to observing areas in northern Gila County and south of Gila

County, interformational relations were viewed in the Christmas quadrangle, southeast of Globe with the guidance of C. R. Willden, in the Superior quadrangle under the guidance of D. W. Peterson, and in the Inspiration quadrangle under the direction of N. P. Peterson. During this conference A. R. Palmer assisted greatly in the interpretation of Cambrian formations by providing counsel on paleontologic aspects.

Thus, for the interchange of factual information on the nature of the Cambrian-Precambrian boundary and for critical discussions of my interpretations of the features of this boundary, I am indebted to many individuals. Much remains to be learned about the geology of the younger Precambrian and the Cambrian formations; though others have contributed information, they should not be held responsible for the significance that I attribute to various features. From work now in progress, undoubtedly modifications of the generalizations herein presented will be forthcoming.

Previous nomenclature and correlations of the younger Precambrian rocks

The Apache group was originally defined by Ransome (1903, p. 28-39) as the result of field work done in the vicinity of Globe during parts of 1901 and 1902. The group was named from exposures in the Apache Mountains north of Globe (fig. 3), but was described as most completely and representatively exposed on Barnes Peak 7 miles northwest of Miami and there divisible, in ascending order, into the Scanlan conglomerate, Pioneer shale, Barnes conglomerate, and Dripping Spring quartzite. In the intricately faulted terrane of the Globe-Miami mineral belt the Mescal formation that intervenes between the Dripping Spring quartzite and the Troy quartzite is unusually thin, highly siliceous, and inconspicuous in outcrop, or where of more typical limestone lithology the Mescal is remnant as widely separated small outcrops in which the relations to both quartzites are not readily apparent. As a consequence the Troy and Dripping Spring quartzites were first grouped together under the name "Dripping Spring quartzite", and Ransome supposed the Mescal to be a part of a sequence of Devonian and Carboniferous age, which he termed the "Globe limestone." In 1910 and 1911 while mapping the Ray quadrangle, which encompasses a large part of the Dripping Spring Range and the northwest part of the Mescal Mountains, Ransome (1911, p. 747) recognized that a cherty dolomite formation capped by basalt separates the two thick quartzitic formations. He restricted, therefore, the name "Dripping Spring" to the lower arkosic quartzite and declared the name "Globe limestone" obsolete. Still later (Ransome, 1915, p. 380-385) the intervening carbonate formation was named the Mescal limestone, and the upper quartzite-sandstone formation was designated the Troy quartzite and the uppermost formation of the Apache group.

Recognizing the need for information about the Apache formations on a regional scale, in 1912 Ransome made reconnaissance traverses northwest of Globe to Roosevelt Dam and along the west side of the Sierra Ancha, and visited outcrops of the Apache group that had been recognized by C. F. Tolman in the Santa Catalina Mountains north of Tucson. While studying the mineral deposits of the Bisbee district in the southeastmost part of the state in 1902, Ransome (1904, p. 28-35) had defined the Bolsa, Abrigo, and Martin formations, some knowledge of which is germane to the present discussion. His data on these formations were supplemented by observations in the Tombstone district prior to 1912. From this broad background descriptive data were summarized and formations of the several localities provisionally correlated by Ransome (1916) in U. S. Geological Survey Professional Paper 98-K. In large part that report deals with the stratigraphy of the Apache group, and the Troy quartzite, and T remains to the present time -- 45 years later -- the most comprehensive summary of the geology of the younger Precambrian rocks in southern Arizona.

Between 1919 and 1922, as a part of the excellent reconnaissance mapping that culminated in the first geologic map of Arizona, N. H. Darton outlined the distribution of Apache outcrops virtually as known today. And scattered throughout his supplementary "Résumé of Arizona geology" (Darton, 1925) are bits of descriptive detail that aid in interpreting younger Precambrian geology and in selecting critical localities for more detailed observation. For areas north and east of the Sierra Ancha and for the Mescal and Hayes Mountains southeast of Globe, in particular, Darton made brief notes of gross lithologic features or of interformational relations that have not otherwise been described.

Because the Apache group appeared to be in conformable sequence with the overlying fossiliferous Paleozoic formations and seemingly exhibited no greater degree of metamorphism than these Paleozoic rocks in the relatively small areas that Ransome studied in detail, he (1919, p. 49-50, pl. 13) provisionally regarded the group as Cambrian and possibly Ordovician and Silurian in age. Darton as early as 1910, however, had regarded the Apache group as approximately equivalent to the Grand Canyon series of northern Arizona, and of Algonkian age -- a designation now replaced in provincial usage by "younger Precambrian." And later, Darton (1925, p. 32-37) was the first to describe the erosional unconformity that exists between the Troy quartzite and the underlying formations. From his many notes it is obvious that in northern Gila County, where exposures of the Apache rocks are most extensive, Darton observed the unconformity to be a rather subtle feature, ordinarily recognizable as representing an erosional hiatus only because the basal conglomerate of the Troy commonly includes pebbles of chert and basalt; the chert was obviously derived from the Mescal and the basalt from the flows that intervene between the Troy and Mescal formations in many places. But Darton (1925, p. 32-34, fig. 76) also observed in the eastern part of the Mescal Mountains that units presumed to be the Troy quartzite rest in angular unconformity on the Mescal and Dripping Spring formations, and stated that in the southeasternmost part of the range the Troy ultimately extends across a surface on pre-Apache granite. The unconformities of the different areas were supposed, with some misgivings (Darton, oral communication, 1948), to represent one hiatus; this supposition will be refuted in a later section of this report. Furthermore, Darton (1925, p. 32-36, 48-50) found primitive brachiopods, which were then regarded as Upper Cambrian forms, in sandstones overlying the quartzitic part of the Troy. He also recognized an unconformity, not observed by

Ransome, at the top of these sandstones and below the overlying Martin limestone of Devonian age. Darton tentatively -- and rightly, as is now known -- correlated the fossiliferous sandstones with the Upper and Middle Cambrian Abrigo limestone, which exists farther south in Arizona. But Stoyanow (1930; 1936, p. 474-478), from similar observations made slightly later, correlated this sandstone plus the underlying quartzites that comprise most of the Troy with the Bolsa quartzite, which underlies the Abrigo formation in the southeastern part of Arizona. Darton (1932) agreed that this correlation was plausible, and Ransome (1932, p. 6) conceded that the formations underlying the Troy should provisionally be correlated with the Grand Canyon series. During the past three decades, as a consequence, the Troy has been regarded as Middle Cambrian in age and not a part of the Apache group; the formations below the Troy are now considered younger Precambrian. Recently Lochman-Balk (1956, p. 539-544), after viewing the Troy-Martin relations in the Salt River Canyon with me and comparing the published descriptions of the Bolsa and Abrigo, has noted that the Troy should be considered Precambrian in age. To the present writing, in which concepts somewhat different than those of Lochman-Balk are presented, the correlation of the Troy quartzite of central and southeastern Arizona with the Bolsa quartzite of southeastern Arizona has been accepted generally (see Gilluly, 1956, p. 24-25).

In the Tortilla and Dripping Spring Mountains, south and southeast of Ray, respectively, Ransome (1919, p. 53, 56) found small bodies or dikes of diabase that intrude sedimentary rocks as young as Pennsylvanian in age. He therefore regarded the large diabase sills of these areas, the Globe-Miami district, Roosevelt Dam, and the Sierra Ancha to be post-Pennsylvanian, and as probably emplaced during the Mesozoic era. Darton (1925, p. 254-257) took exception to Ransome's correlation, and suggested that the small diabase dikes were feeders for some of the Tertiary or Quaternary basalt flows of the region. Darton noted extensive sills intrusive into the lower part of the Troy in some areas, and also that the Martin limestone of Devonian age is locally in sedimentary contact with such sills. But he also noted (Darton, 1925, p. 36) that "the Troy lies unconformably on the Dripping Spring quartzite and the Mescal limestone, as well as on the great sills of diabase which invade these two formations", and with some indecision designated the diabase Precambrian(?) on the geologic map of Arizona (Darton and others, 1924). This designation has been given little heed, though in the Little Dragoon Mountains, east of Tucson, Cooper (1950, p. 31 and fig. 13) has found the Bolsa quartzite of Middle Cambrian age is nonconformable on diabase sills. M. N. Short and his associates (1943, p. 38-39) after observing, in the vicinity of Superior, diabase intrusive into the Troy quartzite but not into the Martin limestone regarded the diabase as post-Middle Cambrian and pre-Upper Devonian. Still others, in conversation, have suggested that the extensive diabase intrusions may be of two or more ages. Most commonly the sills have been regarded as an early phase of the widespread igneous activity that accompanied the Laramide orogeny during Late Cretaceous or early Tertiary time (see Peterson, 1954; Peterson and others, 1951, p. 35-36).

in descriptions of small areas, published and unpublished, many other workers have contributed information that could be used in a regional understanding of the younger Precambrian geology, but few have attempted to so integrate their findings. Some of the more significant observations, which have proved helpful in the present study, can be enumerated here. Wilson (1922, p. 307; discussion amplified 1939, p. 1151-1153) was the first to note lapout relations of the Apache group in its northernmost area of outcrop. Much later Castil (1954)[1] indicate that the Pioneer shale

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[1] Castil, R. G., 1953, The geology of the eastern half of the Diamond Butte quadrangle, Gila County, Arizona: Univ. Calif. (Berkeley) doctoral thesis (unpublished), includes the fuller descriptions of the overlap relations.

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and part of the Dripping Spring quartzite lapout in the northwest part of the Sierra Ancha, and noted for the first time that much of the so-called shale of the Pioneer formation is actually a tuffaceous sedimentary rock. He correctly postulated that the ash-bearing beds would prove to be of considerable lateral extent. Somewhat earlier, Peterson (Peterson and others, 1951, p. 13-15) in local use had substituted the term "Pioneer formation" for "Pioneer shale" because much of the formation, and especially the lower part, is an arkosic sandstone. Wilson (1928, p. 30) was also the first to suggest that the stromatolitic limestones in the Mescal have value as a stratigraphic marker. Although Ransome and Darton had observed the white sandstone herein designated the "middle member" of the Troy quartzite, Burchard (1931, p. 54-57) was the first to indicate that it is a sandstone lithologically quite different from any other in the stratigraphic sequence. From exposures along the northern reaches of Canyon Creek, Burchard designated this unit the "Chediski white sandstone member" of the Troy quartzite, and regarded it as the basal member of the formation. As traced southward across the Salt River, the more conspicuous definitive features of the sandstone disappear, and farther south in the areas afforded geologic studies because of economic interest this part of the Troy has not been separately distinguished. Therefore the name "Chediski sandstone" has not been used except in the restricted area where originally defined. Although in this and most other localities the "Chediski sandstone" is the basal member of the Troy, it is not such everywhere, as will be shown.

Principal conclusions of this report

As a conclusion of this study, redefinition of the Apache group and the Troy quartzite is deemed worthwhile; furthermore, it should be more generally understood that the lithologic terms of the formation names, which are widely accepted as descriptive, do not accurately indicate the lithology of most of the units. The present lithologic terms do provide, however, some descriptive flavor, which might otherwise be lost, and therefore should be retained. The redefinitions presented in this paper are summarized on table 1.

Table 1.—Pre-Devonian formations of southeastern Arizona.

The principal modifications of Apache nomenclature are as follows. The Scanlan and Barnes conglomerates, because they are everywhere too thin to be separately mapped and because of their genetic relations to the overlying clastic sediments, are here considered merely as basal conglomeratic units (not members) of the Pioneer shale and the Dripping Spring quartzite, respectively. This is in contrast to the formational status previously accorded the conglomerates. Two lithologically distinctive members are characteristic of the Dripping Spring quartzite. The boundary between the Mescal and Dripping Spring formations is now recognized as an unconformity, and is placed a few feet stratigraphically below the horizon ordinarily chosen in the past as the contact. And the Mescal limestone is divided into three members; the uppermost member has not been recognized previously as a part of the formation. Although before this investigation an unnamed basalt unit was considered the uppermost formation of the Apache group, two basalt units are now known; one is within the Mescal formation in a few areas, and one overlies the Mescal in many areas. The problem of nomenclature imposed by a basalt unit and an unconformity within the Mescal formation is recognized. But practical considerations, discussed on pages 176-177, suggest the usage presented in table 1. The Troy quartzite is now known to be locally almost three times thicker than in sections previously accepted as typical. In areas of such development the formation is divisible into three distinctive members; the top two members have been noted previously, but no heed has been taken of them as definitive units. From consideration of regional stratigraphic relations to the Bolsa quartzite of Cambrian age, the Troy quartzite is herein designated as younger Precambrian in age.

Some quartzite-sandstone units that vary locally overlie the Troy quartzite and in past descriptions have been included with the Troy, are herein restricted from the formation and considered equivalents of the Bolsa and Abrigo formations. Correlations of the past, in which quartzites of the Mescal Mountains and areas farther south were designated "Troy", must now be reconsidered. Actually such redefinition will affect to only a slight degree, however, the designations on large scale maps that have been published. Possibly the Ray quadrangle is the only published map on which the geology depicted will be seriously modified as a result of this revision in nomenclature.

Where basal sandstones of the Martin limestone of Devonian age locally attain appreciable thickness, generally they have been erroneously included with the Troy quartzite. Recognition of the Devonian sandstones as separate entities is significant in reconstructing the evolution of geologic features of the region but, again, no great cartographic errors have been made in outlining geologic units on the geologic maps that have been published.

Better appreciation of the structural disposition of the widespread diabase intrusions and recognition of the relations of the basal Paleozoic formations to the Apache formations now allow reconciliation of the various views as to the age of the diabase. The evidence accumulated during this study indicates overwhelmingly that all the diabase intrusions of any significant bulk are younger Precambrian in age. Small basic dikes, which have been correlated erroneously with the larger intrusions, probably do post-date Pennsylvanian strata; such dikes are extremely rare.

Regional stratigraphic data now permit some reconstructions of the structural evolution of the younger Precambrian formations. It should be more generally recognized that: (1) deformation preceded diabase intrusions, but deformation which accompanied the intrusion of diabase sills was more widespread; and (2) the resulting structures were truncated and the younger Precambrian rocks were deeply eroded prior to the earliest Paleozoic sedimentation of record in any given area. Such recognition could aid greatly in unraveling the structural history of several base-metal districts of southeastern Arizona.

Rock types of the older Precambrian basement

Rock types of the older Precambrian terrane that underlie the Apache group are of special interest as sources for the clastic sediments of the Apache and Troy sections. Feldspar and quartz are principal constituents of some of the Apache and Troy units; others are largely quartzose. The pebbles of the younger Precambrian conglomerates generally are largely of metamorphosed quartzite derived from some older terrane; even the bulk of the granules, pebbles, and cobbles in the conglomerates of the Cambrian formations and the rare basal conglomerates of the Devonian sections may include much pre-Apache quartzite.

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 the Apache Mountains. Detailed studies (Ransome, 1919, p. 35-37; Gilluly, 1956, p. 11) indicate that the Pinal consists mainly of metamorphosed sedimentary rocks, but it does include minor intercalations of volcanic rocks. Basaltic units are now represented by amphibolites, and intrusive and extrusive rhyolitic units have been noted (see Anderson, 1951, p. 1334-1335).

Slightly schistose and nonfoliated sedimentary and volcanic rocks, at least in part probably equivalent to the Pinal schist, underlie the Apache group north and west of the McFadden Peak quadrangle (Wilson, 1939; Gastil, 1958) and crop out widely in the Mazatzal Mountains farther west. The sedimentary rocks are mainly shales (now shales, slates, and phyllites), thin- to thick-bedded quartzites, and fine-grained to conglomeratic sedimentary units comprised largely of volcanic debris, which ranges from basalt to rhyolite in composition. The volcanic rocks are dominated by rhyolite in the form of flows, tuffaceous deposits, and agglomeratic accumulations, but basalt flows and basaltic agglomerates are common. The southernmost remnant of such rocks, within the area of Apache outcrop, is continuously exposed in a belt, 1 to 4 miles wide and about 10 miles long, that extends along the south side of the Salt River northeast from the mouth of Cherry Creek (see fig. 4). Much farther northwest, in the vicinity of Jerome, Prescott and Bagdad the Yavapai series, lithologically similar except that quartzites and conglomerates are missing from the sections, crops out through considerable areas (see Anderson, 1951; Anderson, and others, 1955, p. 7-12, Anderson and Creasey, 1958, p. 8-45, for descriptions). Relatively small bodies of pyroxenite, gabbro, diorite, and rhyolite were intruded into all those rocks prior to widespread invasion by granitic magmas.

In the areas of Apache outcrop, all these rocks were strongly deformed, then invaded by granitic masses of batholithic dimensions and deeply eroded prior to deposition of the Apache group. The granitoid masses range from diorite to granite in composition, but the more recent studies suggest that they are dominantly of quartz monzonite or granodiorite. The principal examples that have been studied in some petrographic detail are the Oracle granite, which is widely exposed on the north flank of the Santa Catalina Mountains and northward to a latitude at least 10 miles beyond Mammoth, and the Ruin granite and Madera diorite of the Globe-Miami mineral belt and vicinity. The Oracle granite, at least in part (see Peterson, 1938, p. 8-9), is a quartz monzonite, as is the Ruin granite (Peterson, 1954). Peterson (1954) describes the Madera diorite of the Globe quadrangle and of the type area in the Pinal Mountains as a granodiorite; another example is the Johnny Lyon granodiorite of the Dragoon quadrangle (Cooper and Silver, report in preparation).

Much of the quartz monzonite is a rather distinctive coarse-grained rock. Where I have had reason to particularly note the rock in northern Gila County, an abundance of large grayish-orange pink to pinkish-gray microcline phenocrysts, which enclose numerous small grains of plagioclase and biotite, make the texture seem coarser on casual observation than is actually the case. These poikilitic subhedral phenocrysts, ordinarily 1/2 to 1 inch in length but commonly as much as 2 inches, are conspicuously scattered through a coarse (1-8 mm) hypidomorphic groundmass of pinkish potash feldspar, almost clear quartz, pale greenish-yellow or yellowish-gray chalky plagioclase, and black biotite. The interior two-thirds of the plagioclase grains generally are so altered to sericite as to make satisfactory identification impossible. The outer one-third of these grains is zoned; in the specimens that I have examined microscopically the zoned rims are within the compositional range of albite. The plagioclase of like "granite" from localities north of Miami has been described as oligoclase (Ransome, 1903, p. 74). Much of the biotite, which is the form of thick books, has been altered to chlorite. Sparse crystals of brownish-black sphene can be noted megascopically in many specimens. Apatite and minute grains of hematite(?) are accessories observed under the microscope. Variant facies of the quartz monzonite exist — apparently in relatively small volume — but have not been studied by me; an orangish-pink aplite dike facies seems to be the most common. Quartz monzonite of this general description is the dominant basement rock as far south as the Hayes Mountains; and from Ray southward the length of the Tortilla range a similar rock is said to prevail (Ransome; 1919, p. 37).

Outcrops of the older Precambrian formations, differentiated only as to rock types separately significant in furnishing detritus now recognized in the Troy and Apache formations, are shown on figure 4.

Figure 4.--Sketch map showing outcrops of the older Precambrian terrane, subdivided as to rock types that would contribute different detrital materials for late Precambrian sedimentation. Map of same area as figures 2 and 3.

The proportions of rock types seen in outcrop are very likely representative of the terrane exposed just before Apache sedimentation began. From the map the dominance of granitic rocks and the sparsity of quartzites in the pre-Apache terrane is readily appreciated. For adjacent regions, which must have furnished the bulk of materials incorporated in the Apache group and the Troy quartzite, the proportion of granitic rocks to other rock types must have been of like order.

Granitic detritus is abundantly represented in the Pioneer, Dripping Spring, and Troy formations. Some of the coarser microcline fragments seen in all these formations have poikilitic inclusions of chalky plagioclase and otherwise are texturally identical to the Ruin and Oracle granites. Without doubt the granitoid masses furnished arkosic detritus through a long period of time.

If the highly quartzose units are excepted, seemingly the Pinal schist and lithologically equivalent formations — depicted as nonresistant rocks on figure 4 — furnished little of the detritus that was incorporated in the younger Precambrian formations. Dubiously, some of the argillic materials of the siltstone units in the Apache group could have been derived from such sources. Some of the denser rhyolites and the rare thin units of jasper in the older Precambrian formations can be recognized in the conglomerates of the younger Precambrian. These materials, however, are scarce detritals. The resistant rhyolites that have been recognized as detritals, for the lack of separate depiction on the available source maps, have been included with the nonresistant rocks on figure 4. Such rhyolites exist only in the northwestern part (north of Salt River) of the area shown on figure 4.

Many of the quartzites that antedate the granites were thoroughly sheared or sheeted then firmly recemented, and are fine-grained, highly vitreous, and are of brown, red, or purple hues distinctive from those of the later quartzites. The bulk of the quartzite gravels in the conglomeratic units of the Apache group and the Troy formation are readily identified as detritals petrographically like the quartzites now exposed in the older terrane. Reddish-brown jasper pebbles found in the coarser fractions of the older quartzites are also conspicuous though minor constituents in the Apache group and Troy quartzite. For lack of detailed information only the larger known occurrences of the older Precambrian quartzites are shown on figure 4; additional thin units are fairly common in some of the pre-granite formations — especially in those that lie north of the Salt River. Outside of the region shown on figure 4 the outcrops of Pinal- or Yavapai-type rocks that have been studied are largely devoid of quartzite, and probably few that have not been studied include quartzite. As will be detailed later, during the accumulation of most of the younger Precambrian sediments the older Precambrian quartzites within the area of figure 4 were buried by the basal units of the Apache group, and could not have been sources of much of the quartzite debris found in the several formations. A lithologic terrane similar to that shown in figure 4 must have been exposed somewhere outside of the region of the present Apache outcrops.

Pre-Apache unconformity

As already implied, a profound unconformity separates the older Precambrian rocks from the overlying sedimentary formation of the Apache group. Intense deformation of the Pinal and related formations, followed by granitic intrusions of batholithic proportions signify a considerable orogeny that marked the end of older Precambrian rock accumulation in central and southeastern Arizona. The mountains formed during this orogeny were leveled to a plain during the long period of erosion that preceded Apache sedimentation.

The basal formations of the Apache group were deposited on a surface of remarkably little relief. In degree of smoothness and characteristics of the underlying regolith this surface is wholly comparable to the equivalent surface exposed in the Grand Canyon, which classically has been referred to as a peneplain. In a definitive study of the Grand Canyon examples, Sharp (1940) concluded that the similar surface is a product of subaerial erosion only slightly modified by marine erosion.

For most of the region the pre-Apache surface exhibits local relief of only a few feet at most; recognizable instances of even this slight relief are rare. From the Sierra Ancha southward variations in the thickness of the Pioneer shale — the basal formation of the Apache group — suggest that the monotony of the planar surface may have been relieved slightly by broad low hills rising above the general plain or by broad basins cut below it. If so, the lapping out of basal beds of the Pioneer is too subtle to be noticed in tracing continuous outcrops distances of one to five miles. Such lapout is obvious, however, north and northeast of the Sierra Ancha, where the Pioneer shale (as indicated on fig. 3) and at least the lower half of the Dripping Spring quartzite pinch out by lapping against a high in the basement surface. Where the Pioneer is very thin south of the Sierra Ancha (see fig. 3 for localities), a more subtle lapping relation might be noted if outcrops were continuous enough so that lateral thinning of the monotonous sequence of mudstones could be seen.

It seems more than coincident that the lapout north of the Sierra Ancha occurs where the basement formations of a large area are almost wholly of sedimentary and volcanic rocks (Gastil, 1958) rather than terrane prevailingly of less resistant granite. In the sequence of this area steep-dipping ribs of quartzite and other relatively resistant rocks, in bodies too small to be shown on figure 4, are common. The structural grain of these rocks is northeast. The line of lapout trends northeast (see fig. 3) and crudely parallels the southeastern boundary of a belt of older Precambrian rocks that was unusually resistant to erosion (see fig. 3). Present-day remnants of the Apache formations surmount some of the higher parts of this belt of metamorphic rocks, but no remnants exist on the northwest side of the belt, though undoubtedly Apache formations once existed much farther in that direction than the present outcrops. Whether the belt of lapout represents a large elongate monadnock rising above the general pre-Apache surface, or a broad high of regional extent is not known. This is the only area in which a prominent deviation from the general planar surface can now be demonstrated. In other areas the schistose rocks generally are not fortified by ribs of resistant rock.

The granitic rocks immediately beneath the pre-Apache unconformity were disintegrated prior to deposition of the Apache group. As now remnant, 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-oxide minerals, shows ill-developed bedding structures indicative of transportation, but in most places it represents residual material in virtually original position. The arkosic regolith merges gradationally downward with the underlying unweathered massive granite. In many areas — even where the basement rocks are not granitoid — this arkose and hematitic silt dominate the matrix of the overlying Scanlan conglomerate.

Where the unconformity truncates the Pinal schist or related metasedimentary or metavolcanic rocks the foliation or other layered structures may be somewhat obscured immediately adjacent to the erosion surface, but no other effects of pre-Apache disintegration are particularly notable in the field. In places a thin rubble zone, ordinarily only a few inches thick, intervenes between the basal conglomerate of the Apache group and solid rock of the older Precambrian. Where such rubble was derived from schist and exhibits crude bedding, it may include abundant angular to subrounded fragments of white vein quartz in a matrix of small schist fragments. The quartz undoubtedly was derived from veins that locally are abundant in the schist and in adjacent granitic rocks. Ordinarily such quartz and schist fragments are also abundantly incorporated in the conglomerate that overlies the regolith zone; indeed, more often than not, all such detritus is included in the conglomerate which then rests on a clean-swept surface of schistose rock.

Younger Precambrian rocks

Sections of individual formations of the Apache group and the Troy quartzite are most extensively remnant in the Sierra Ancha. Farther south, owing to unconformities within the sequence and unconformities that separate the younger Precambrian strata and included diabase intrusions from the overlying Paleozoic formations, sections generally are less complete. Within the part of the region that now lies between the Gila and Salt Rivers, the Troy quartzite was greatly thinned and locally the Mescal limestone and even the upper part of the Dripping Spring quartzite removed by pre-Paleozoic erosion, but generally though this area the Apache section remained virtually complete at the time the Paleozoic formations were deposited. South of the Gila River the upper part of the Dripping Spring quartzite and all of the overlying formations and included diabase intrusions were removed from large areas prior to Cambrian sedimentation; even in this southern area, however, most of the Mescal formation and moderately thick sections of the Troy quartzite were preserved locally in blocks down-faulted prior to post-Troy erosion. Still other aspects of distribution that contribute to the incompleteness of stratigraphic sections are described in the discussions of the individual formations.

The thicknesses of the formations of the Apache group shown on Table 1, and as discussed throughout the report unless otherwise specifically stated, are the thicknesses typically remnant below the Troy cover, and not those thinned post-Troy erosion. The thickness of the Troy formation in any given area is in large part dependent on the degree of pre-Paleozoic erosion in that area. Therefore the cited thicknesses of Troy are those remnant beneath a cover of Paleozoic rocks.

Thin to thick diabase sills are coextensive with the Apache formations and invaded the group in almost every locality where an appreciable part of the section can be viewed; many Apache sections are displaced by sills at several horizons. Consequently, in many areas complete sections of even one formation must be pieced together from two or more partial plates of the formation encompassed between sills. Actually, I am aware of only a few places where the entire Apache section can be viewed from bottom to top, and none of these sections are uninterrupted by diabase sills. Generally, to arrive at an overall thickness for the Apache group, one must piece together partial sections from several localities within a restricted area.

South of the Plateau portion of the mountain belt, as described earlier, the Troy quartzite and the Apache formations were much faulted during Tertiary time and commonly are partly obscured by a cover of Cenozoic rocks. In what is now the Basin and Range portion of the area of outcrop pre-Paleozoic erosion also caused greater and more varied thinning of the section than to the north. Within this part of the region, therefore, estimation of the thickness of the Apache group is even more difficult.

In the vicinity of the highest parts of the Sierra Ancha, where the Troy quartzite is at least 1,200 feet thick, the total thickness of the younger Precambrian sequence can be extrapolated as between 2,650 and 2,800 feet. Elsewhere remnants of the Troy are generally less than 600 feet thick, so other statements of total thicknesses that include the Troy have little meaning. In the Plateau portion of the region, where entire sections have been pieced together with considerable confidence at several localities, the Apache group ranges from 1,250 to 1,600 feet in thickness. North of the Sierra Ancha, of course, these thicknesses must be too great by roughly a factor of two, because locally the Pioneer shale and at least the lower member of the Dripping Spring quartzite are lapped out. In the Basin and Range area the 1,600-foot maximum may be applicable to some sections as far south as the Apache group crops out, but other sections not affected by pre-Paleozoic erosion appear to be as little as 1,100 feet in aggregate thickness of the pre-Troy formations. Some sections, because of pre-Troy erosion, are even thinner. The thinner sections are those in which the overall thickness was reduced largely by (1) thinning of the basal Pioneer formation, (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 many areas, if the intrusive diabase sills are included, the aggregate thicknesses of younger Precambrian layered rocks are roughly twice those noted above (i.e., 2,000-5,000 feet).

Stratigraphic units, whether members of formations or even lesser units, of the younger Precambrian section are remarkably consistent in lithology and other features throughout the broad region of outcrop. Some features are unusual in their uniformity of distribution. Furthermore some major variations, such as variations in thicknesses of the formations, that do exist are largely the effects of superimposed geologic processes rather than original. Because of the common aspect of uniformity and because of the interrelations of features, interpretations of the origins of the formations are deferred until all are described. Only some of the minor features are interpreted as to origin in the following descriptive sections. Similarly, because the age and correlation of the younger Precambrian formations can be considered in better perspective after their geometric relations to the intrusive diabases and the Paleozoic formations are described, such discussion is also held for a later chapter.

Apache group

Pioneer shale

The Pioneer shale was a favored formation for inflation by diabase, and the Pioneer with included sills is exposed in steep to gentle slopes. Commonly these slopes are heavily mantled by the resistant rock debris from overlying formations, so the boundaries between the two rock types are difficult to distinguish. Thus complete sections of the Pioneer are rarely exposed. For the following descriptions I have relied much on observations of excellent exposures at the south end of the Sierra Ancha and in the canyon of Cherry Creek, where the formation ranges from about 250 feet to slightly more than 500 feet in thickness.

Such thicknesses are greater than those that have generally been considered typical of the formation. The thicknesses widely accepted as typical are those stated by Ransome (1916, p. 136 and pl. 25). He estimated the average thickness of the Pioneer in the Ray quadrangle as about 150 feet, but also noted a regional range of 100 to 250 feet. The places where the Pioneer shale is known to lap out or is consistently thin are shown on figure 3. As previously noted, north of the Sierra Ancha the Pioneer laps out completely. In the vicinity of Coolidge Dam (the dam that ponds San Carlos Lake—see fig. 3) the formation is thin or missing, and where observed at several places along the San Pedro Valley downstream from Mammoth is generally less than 30 feet thick. As far as is known the Pioneer shale has not been observed anywhere east of the line that depicts this belt of thinning on figure 3. Although as little as 150 feet of the Pioneer formation exists in many areas south of the Salt River—as along the south escarpment of the Natanes Plateau and for several miles farther south, and at places (according to Ransome, 1916, p. 136) in the Dripping Spring and Tortilla Mountains — other than as noted above, no particular pattern of thinning has yet been discerned. Indeed, as far south as the Apache group crops out the Pioneer exhibits maximum thicknesses of 300 to 500 feet. As examples, Cooper and Silver (report in preparation) measured a 306-foot section in the northwestern part of the Dragoon quadrangle, S. C. Creasey (written communication, April 1961) has measured thicknesses of 450 to 500 feet in the southwest part of the Mammoth quadrangle, and Carpenter_/ estimated a thickness of 400 feet in the Vekol Mountains/.

_/Carpenter, R. H. 1947, The geology and ore deposits of the Vekol Mountains, Pinal County, Arizona: Stanford Univ. doctoral thesis (unpublished).

The Scanlan conglomerate, at the base of the Pioneer shale, typically is represented by 1 to 8 feet of sub-angular to well-rounded pebbles or cobbles in a matrix that reflects 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 fine- to coarse-grained arkosic debris derived largely from the older Precambrian granite. White quartz pebbles are common, and in places compose the bulk of the granules. Sparse granules and small pebbles of reddish-brown jasper can be seen in many outcrops. Pebbles of volcanic rocks or of schistose rocks, like those of the older Precambrian terrane of northwestern Gila County, are few but can be noted on careful inspection of many exposures. Where the conglomerate overlies the Pinal schist it commonly includes fissile fragments of the schist in abundance, and the matrix may in large part be comprised of dark-colored shaly debris from the schist; even in this setting, however, abundant angular grains of feldspar and quartz from the granite ordinarily are incorporated in the matrix. In nearly all exposures of the conglomerate the fines fraction of the matrix is highly hematitic material similar to the mudstones or sandstones of the overlying section. In places pebble-free lenses of grayish-red mudstone or sandstone are included in the conglomerate. Ordinarily the contact between the conglomerate and overlying beds is sharp, but there are many examples of transitional conglomeratic intervals of a few inches or, less commonly, of a few feet at the top of the conglomerate. And rarely thin lenses of conglomerate or of sandstone that includes scattered pebbles exist as much as 40 feet above the basal conglomerate.

In thickness and in pebble characteristics, the basal conglomerate varies locally. In many places it is represented by only a few inches of granitic debris in which granules or small pebbles of quartzite or white quartz are sparsely distributed. In other places, as at Roosevelt Dam, the conglomerate is as much as 30 feet thick and is mainly of close-packed cobbles. In a given exposure the conglomerate may include boulders as much as 1 foot in diameter and be largely comprised of cobbles 4 to 6 inches in maximum diameter; within one-half mile along the outcrop cobbles may be few and the conglomerate is largely of pebbles 1 to 2 inches in diameter. Similarly along such a length of outcrop the thickness of the conglomerate commonly ranges from a few inches to 8 feet--or, in extreme examples, to 20 feet; and in one locality the conglomerate is comprised mostly of well-rounded pebbles, but in an adjacent area includes an appreciable content of subangular gravels. In many localities the pebbles and cobbles are disc-shaped, and in some are disposed in imbricate relations. No consistent direction of imbrication, which would indicate a general direction of currents during deposition, has been noted. Thus regional variations in pebble size, composition, and disposition or in conglomerate thicknesses, which might suggest distance or direction from sources, seem to be no greater than the variations noted in a relatively small area. A prominent exception to this generalization is described at the end of this section.

The conglomerate ordinarily crops out as narrow ledges or, where thick, as small cliffs. Only in very rare circumstances are outcrops broad enough to be mapped without excessive cartographic exaggeration, and rarely has the conglomerate been shown on maps, except by a diagrammatic symbol. As the Scanlan genetically is merely the coarse-textured basal unit of the Pioneer shale, it is here proposed that the conglomerate be so regarded, and that it not be afforded formation or even member designation. Although thin, the unit is widely distributed. Furthermore the term "Scanlan conglomerate," because it brings to mind a well known, naturally distinctive unit, will continue in popular use by those concerned with the Apache rocks. Therefore, the name "Scanlan conglomerate" should be retained to informally designate the bed or beds that mark the base of the Pioneer shale.

The Pioneer lithology that catches the casual observer's eye, because it is strikingly different from that of other formations of the Apache group, is a grayish-red siltstone or a silty mudstone that commonly includes abundant grains of fine sand size. These rocks are minutely laminated or cross-laminated (see fig. 5) in beds 6 inches to 3 feet thick.

Figure 5.--Delicate cross-lamination in tuffaceous siltstone of the Pioneer shale. Black specks in center of reduction spots are limonite pseudomorphs after pyrite. Note enlargement.

Small assymetrical ripplemarks are sparsely seen in some sections. The rock is firmly indurated but generally is closely jointed, both normal to and parallel to bedding, so that it erodes to receding slopes. Characteristically small disk-shaped flakes, 1/4 to 1 1/2 inches in maximum dimension and with one or more rounded surfaces and angular edges, spall loose from the outcrops and abundantly litter the slopes. Typical of these siltstones and mudstones are abundant, yellowish-gray to light brown reduction spots, as much as 1 inch in diameter but generally 0.1 to 0.2 inch across (fig. 5). Minute cubic grains of limonite, obviously derived from pyrite, can be seen in the centers of some of the smaller bleached spots. In some localities, bluntly terminated lenticular zones of similar bleaching, which may be as much as 1/2 inch thick and several inches in diameter, parallel the bedding, or less commonly, the high-angle jointing. These larger bleached spots may characterize certain beds of a section, and seemingly are most prevalent in the more quartzose beds. The mudstones and siltstones and like-colored sandstone beds, only slightly coarser in grain size, comprise the most of the Pioneer in many areas. These rocks, because of their strong tendency toward a fissile habit of splitting, are the so-called "shales" of the Pioneer.

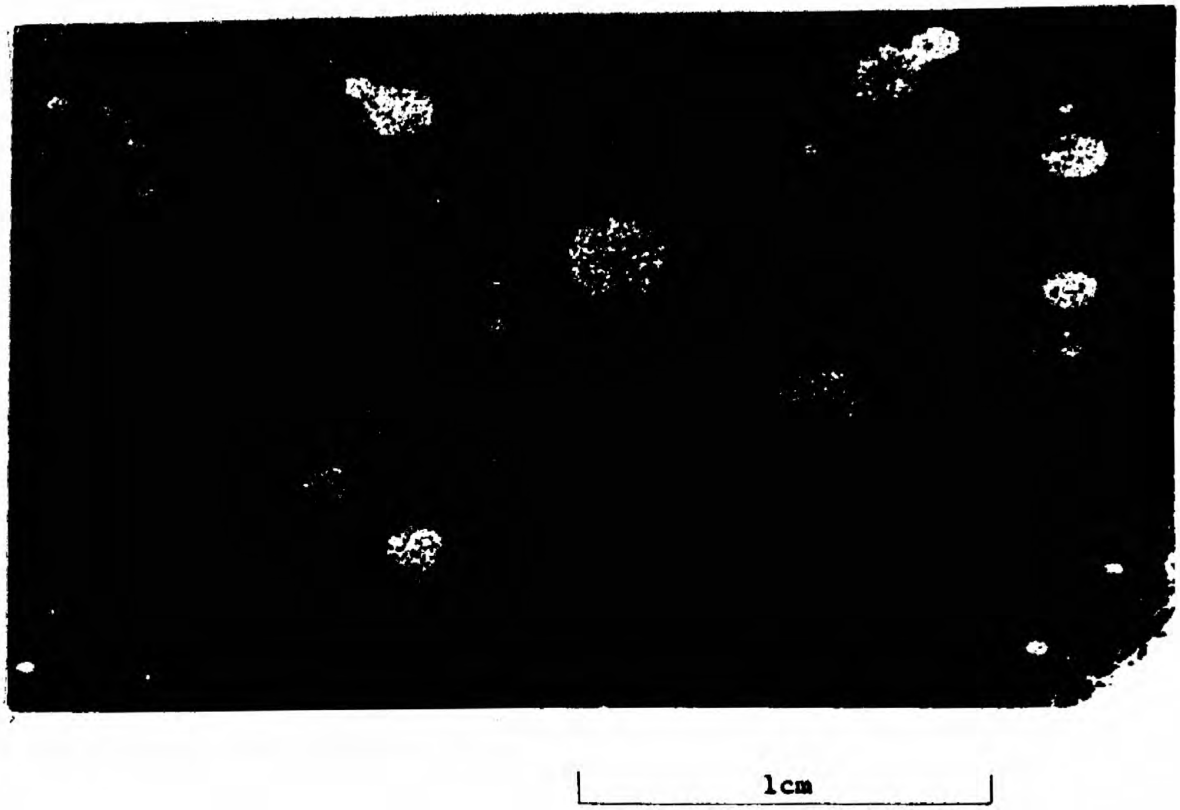

FIGURE 5. DELICATE CROSS-LAMINATION IN TUFFACEOUS SILTSTONE OF THE PIONEER SHALE.

Black specks in centers of reduction spots are limonite psuedomorphs after pyrite. Note enlargement.

Under the microscope these "shales" are noted to be comprised of angular grains of feldspar and quartz of silt and fine sand size, set in a highly hematitic matrix of small, irregularly lenticular devitrified glass shards and clay-size material (fig. 6). From a study of X-ray

Figure 6.--Photomicrograph of tuffaceous siltstone of the Pioneer shale. Wispy grains are relicts after glass shards; more or less equidimensional clear grains are quartz; like clouded grains are feldspar; dark areas are heavily dusted with hematite. Plain transmitted light. From specimen of figure 5.

powder patterns, Gastil (1954), who was the first to recognize the tuffaceous nature of the shaly-splitting units, suggested that the devitrified shards are now largely of finely divided muscovite and very fine-grained chalcedony. The fine-grained matrix materials exhibit some optical similarities to the individually distinct shards, and probably much of the matrix that is not microscopically resolvable is of like mineralogic composition. Robert L. Smith, who has studied tuffaceous rocks extensively in the course of his duties with the Geological Survey, has stated (oral communication, August 1956) that the rod-like and bifurcated cross sections of the shard pseudomorphs and their other features suggest ash derived from acid eruptions—or from eruptions that were at most no more mafic than andesites. He thus upholds Gastil's designation of the pumiceous materials as rhyolitic in type. In the typical mudstone the silt-size mineral grains, other than pseudomorphs of glass fragments, are of feldspar and quartz; which comprise roughly one-quarter of the rock. In some of the thin sections

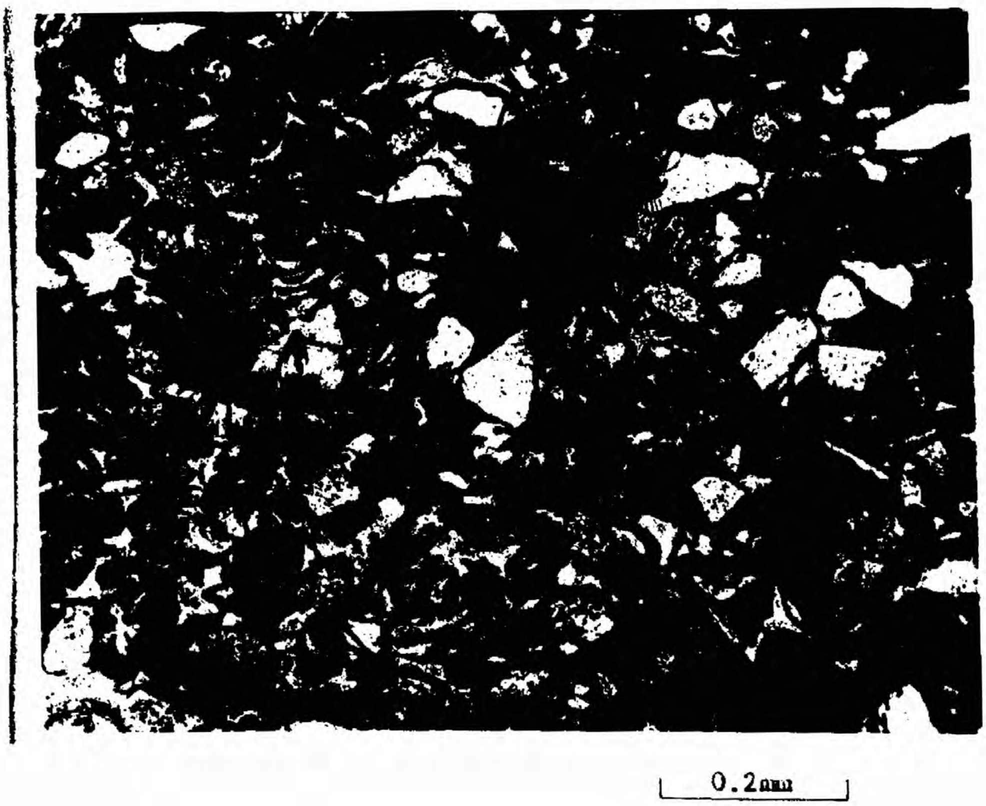

FIGURE 6. PHOTOMICROGRAPH OF TUFFACEOUS SILTSTONE OF THE PIONEER SHALE.

Wispy grains are relicts after glass shards; more or less equidimensional grains are quartz; like clouded grains are feldspar; dark areas are heavily dusted with hematite. From specimen of figure 5.

viewed this coarser fraction is largely of feldspar, and a large part of the feldspar is plagioclase. No particular attempt has been made by me to determine the proportions of potash and soda feldspars. Minute flakes of muscovite abundantly spangle freshly fractured hand specimens of some units. Thin sections from specimens in which mica is not particularly noticeable to the unaided eye show sparse to fairly abundant shreds of muscovite. All platy grains—muscovite plates, the shard relicts, and tabular cleavage fragments of feldspar—tend to be alined parallel to the delicate bedding. But the bedding is more obviously marked by the hematite, which is thickly dusted throughout the rock but particularly opaques the finest grained parts of the matrix.

In some localities hard beds, which appear in outcrop to be very fine-grained impure quartzite, are interlayered with the usual tuffaceous mudstones. These beds generally are dusky red purple or blackish red--or, in other words, somewhat darker and more purplish than the ordinary mudstones. In these beds and the adjacent mudstone beds, minute flakes of mica are particularly apparent. In thin sections the detrital grains are noted to be slightly larger than those of the mudstones--but still of very fine sand size. Quartz makes up a large portion of such grains, the matrix is generally less abundant, and devitrified glass shards are fewer, but otherwise the rock is not greatly different from the mudstones described above. In all such specimens viewed under the microscope, planar lamination but no cross-lamination has been observed. In some specimens shards are sparse, but without exception they have been found when searched for. This quartzitic variety of the mudstone has been noted throughout the region, but is seemingly much more prevalent south of the latitude of the Apache Mountains than north of that latitude. It seems particularly abundant in most southerly occurrences of the Pioneer.

Although the above-described rock types are commonly considered representative of the formation, where the Pioneer is more than 150 feet thick ordinarily they comprise, by themselves, only the upper one-half to two-thirds of the formation. The lower part of the Pioneer is generally of fine- to very coarse-grained units of arkose or feldspathic sandstone intercalated with mudstones like those higher in the formation. These coarser-grained units, which are thin- to thick-bedded and are cross-bedded, may constitute most or only a small part of the lower half of the Pioneer. In a few places in the McFadden Peak quadrangle fine- to medium-grained arkose composes as much as 50 feet of the uppermost part of the formation. Except for hematitic fines that impart a prevalent reddish cast, some of these sandstones are not readily distinguished from similar beds in the lower member of the Dripping Spring formation. J. R. Cooper (oral communication, 1958) has noted similar sandstones immediately below the Barnes conglomerate in the Dragoon quadrangle. Because this sandstone unit, generally 25 feet or less in thickness but locally as much as 50 feet thick, so strikingly resembled the arkoses above the Barnes conglomerate Cooper preferred to include it with the Dripping Spring quartzite on his map.

Coarse- and very coarse-grained arkoses are usually confined to the basal 25 to 40 feet of sections 250 feet or more in thickness. Rarely thin beds of quartz sandstone or even of conglomerate of quartz granules that are light gray in color may be included in such units; otherwise these coarser clastic beds are generally the grayish-red of the tuffaceous sediments in the upper part of the section.

Exceptionally the obvious interbeds of fine-grained mudstones are lacking and the basal arkoses are thick-bedded and massive-cropping. These constitute 40 to 90 feet of the basal part of the formation, and even in thin section are not notably different from the flesh-colored medium-grained arkoses of the lower member of the Dripping Spring. Individual beds may be well-sorted, but as a unit such basal arkoses of the Pioneer are probably not sorted as well as a similar thickness of the lower member of the Dripping Spring. Lacking other distinguishing characteristics, the basal arkoses of the Pioneer can be differentiated from those of the Dripping Spring by thin seams of brownish-black or blackish-red siltstone that separate a few beds. These seams of dense exceedingly tough siltstone, are 1/10 inch to 8 inches thick in their thickest parts and are discontinuous. Without exception known to me they separate a few of the Arkose beds. In the few thin sections that have been cut from these siltstones pumice fragments are conspicuous. Such hard dark-colored layers of siltstone have been seen even in Scanlan conglomerate that is overlain by 60 feet of massive arkoses, indicating that ash falls were occurring during the earliest stages of Pioneer deposition. In a few localities very coarse-grained and partially pebbly arkoses, obviously derived from the underlying granitoid rock of the immediate vicinity constitute the basal few tens of feet of the Pioneer. A notable section is exposed on the north flank of Pioneer Mountain, in the Ray quadrangle near the type locality. There the basal 45 to 50 feet of the Pioneer is constituted of very coarse, ill-sorted pinkish-gray arkosic debris from the underlying Madera diorite. Such arkoses have no counterpart in the Dripping Spring quartzite. Some of the lightest colored basal sandstones of the Pioneer are highly quartzose, and are more like the sandstones of the middle member of the Troy than of any unit in the Apache group. Such sandstones are probably rare; I have seen only three small outcrops of such lithology.

Lateral variations have been noted only in the northwestern part of the region. As the lapout area of northwestern Gila County is approached closely the Pioneer formation coarsens, and particularly in its basal part varies greatly in lateral distances of a mile or two. In the northwest corner of the McFadden Peak quadrangle and the southeast corner of the Diamond Butte quadrangle, for example, the basal 60 to 90 feet of the Pioneer is a very coarse-grained, cross-bedded feldspathic sandstone. This unit includes beds of granule conglomerate and many lenses of pebble conglomerate. The remainder of the section is medium- to coarse-grained gray arkose, and except for dusky red weathering surfaces and a tendency to be more friable in parting is very like overlying units of the Dripping Spring. The Scanlan conglomerate in some places is represented by a few inches to 3 feet of granitic debris with generally only scattered small pebbles of quartzite; occasionally this debris includes a quartzite boulder more than 1 foot in diameter. In nearby areas 3 to 8 feet of typical cobble conglomerate may mark the base. Short distances to the east and south normal sections of the Pioneer, as described in preceding paragraphs, exist.

These aberrant coarse facies of the Pioneer that lack the siltstone beds apparently mark areas in which the pre-Pioneer surface was high, and in nearby areas the Pioneer may be thin or missing. Where the formation is thinned or the Dripping Spring is actually the basal formation of the Apache group, the basal conglomerate may be unusually thick and comprised of closely packed, well-rounded cobbles or boulders. Such a variation is seen in tracing the basal units 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. From about the place where the Pioneer thins completely out the basal conglomerate begins to thicken appreciably and become coarser, so that at Potato Butte—where possibly as much as 100 feet of the Dripping Spring quartzite is missing by lapout—the basal conglomerate is 85 to 110 feet thick. At Potato Butte the conglomerate is comprised largely of well-rounded cobbles 3 to 6 inches in diameter, but the basal 50 feet includes abundant cobbles 6 to 8 inches in diameter and occasional boulders up to 1 1/2 feet in diameter. This unusually thick conglomerate, which here is the lateral equivalent of the Scanlan conglomerate plus the Barnes conglomerate, mantles an irregular surface cut on granite; local relief of as much as 35 feet can be observed along an outcrop length of 200 feet. In this particular area the granite has been swept clean of the arkosic debris usually seen below the pre-Apache unconformity. Apparently such thick accumulations of basal conglomerate are confined to localities where notable local relief is characteristic of the pre-Apache surface. Further west, where hills of the pre-Apache surface project up through thick sections of Pioneer and into the Dripping Spring, the basal two-thirds of the Pioneer is unusually coarse, includes many conglomeratic lenses, bedding is discontinuous, and medium-scale cross-stratification is conspicuous. The siltstones are seen only in the upper part of the Pioneer. North and east of Young, where the Pioneer is thin or missing and the pre-Apache surface is virtually planar the conglomerate is commonly only 6 to 8 feet thick, and in places is almost nonexistent.

Still another aspect of the Scanlan-Barnes conglomerate is worth note because the coarse constituents are not well-rounded or sorted. The most accessible exposure is 9 1/2 miles due north of Young along the north wall of the canyon of Haigler Creek.[1] At this locality the older

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[1] Specifically, this example is at and immediately west of the hairpin turn of the Chamberlain Trail road, where that road turns northward to follow the crest of the ridge that lies east of Gordon Canyon. The outcrop can be found by traversing the Chamberlain Trail 3.2 miles northward from its crossing of Haigler Creek. According to the U. S. Forest Service map of the Tonto Forest (1950 edition), the outcrops are in the W1/2 sec. 6, T. 10 N., R. 14 E.

Precambrian quartzites dip 55°-60° ESE, and the upper member of the Dripping Spring quartzite that here overlaps these quartzites dips about 15° easterly. Where the pre-Apache unconformity is exposed in the roadcut that traverses the canyon wall it is an irregular surface with relief of several feet. Above the unconformity is a breccia, 20 feet thick, of angular blocks of the older Precambrian quartzite in a matrix of medium- to coarse-grained sandstone. Some of these blocks, which are as much as 2 feet in diameter, have traveled only a few feet from their source, and can be matched back into the irregularities on the surface of the older quartzite from which they were plucked. A few hundred feet to the west, along the outcrop, the breccia gives way to a conglomerate of poorly rounded boulders, some of which are 2 1/2 to 3 feet in diameter. West of the breccia exposure about 600 feet the conglomerate is 40 feet thick. The matrix of this conglomerate is an arkose much like that which comprises most of the lower member of the Dripping Spring; this arkose, which is quite unlike the upper member in texture and bedding, also comprises several feet of the section above the conglomerate. Such facies of the conglomerate are rare, and to the best of my information all are confined to the area of lapout north of the Sierra Ancha. Wilson (1939, p. 1151-1152) has described other examples in the vicinity of that noted above.

If other lateral variations than those attributable to metamorphism exist in the Pioneer elsewhere such variations are very difficult of recognition. With the exception of sections in the area of lapout in northwest Gila County, medium to coarse sandstones of the Pioneer formation appear largely at the base of locally thick sections. These basal sections can probably be interpreted as local accumulations on the topographically low portions of the pre-Apache surface. And the differences in lithology noted in earlier paragraphs are probably local variations in sedimentation.

Where the Pioneer formation is thin in the several exposures along the San Pedro valley, the lithology is quite in contrast with that of the northwestern area of thinning, in that no coarsening of the basal part of the section is seen. Commonly the Pioneer of these localities is mostly the hard dark-colored particularly quartzose siltstone. The Scanlan conglomerate bed is generally very thin or represented only by scattered pebbles in the basal few inches of the formation. Possibly such sections represent deposits on a surface that was relatively high but of very slight local relief. Alternatively, these could be ordinary sections in which the upper part was eroded away prior to deposition of the Dripping Spring quartzite.

The contact, described in the next section, between the Pioneer and the overlying Dripping Spring quartzite is everywhere sharp and readily distinguished.

Dripping Spring quartzite

The Dripping Spring ~~formation~~ quartzite, except where thinned along the lapout area north of the Sierra Ancha or where truncated by later unconformities, is 550 to 700 feet thick and everywhere is characterized by two distinctive members of roughly equal thickness. The lower member, which includes at its base the Barnes conglomerate unit, is largely of thin- to thick-bedded, massive-cropping flesh-colored arkose with subordinate units of light-colored feldspathic quartzite. In contrast, the upper

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As used 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.

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member is dominated by thin-bedded and thin-parting dark-colored, dense arkosic siltstones, intercalated with minor units of thin-bedded fine-grained feldspathic quartzite and arkose. The formation is so firmly cemented throughout that the rock fractures across the component grains. Thus, in the past the composition and textures of the various units of the Dripping Spring have been given little heed, and the entire formation has been characterized inexactly as quartzite.

From the latitude of Young south as far as the Gila River, the Dripping Spring is generally completely represented. The section described in table 2, measured 2½ miles west of Cherry

Table 2.--Representative section of Dripping Spring quartzite.

Creek and near the south boundary of the McFadden Peak quadrangle, (see fig. 2) is probably representative of the Dripping Spring throughout this part of the region. Only in a few scattered localities between the Gila and Salt Rivers was the upper part of the formation stripped away during the intervals of erosion that preceded deposition of the Troy quartzite, the Bolsa quartzite or the Martin limestone. But south of the Gila River, in the area presently encompassed by the San Pedro River drainage, small to large parts of the Dripping Spring section were eroded prior to deposition of the Bolsa quartzite. Consequently complete sections of the formation are rarely remnant south of latitude 33° N. Fairly thick and presumably complete sections do exist, however, in the Vekol Mountains and adjacent ranges 50 to 60 miles south of Phoenix (L. A. Heindl, oral communication, 1959). As already noted, the Dripping Spring thins by lapout north of the Sierra Ancha so that in at least a few localities the upper member rests directly on the older Precambrian terrane. In texture and composition the several stratigraphic units of the formation seem not to be grossly variant except in the immediate vicinity of this northern area of lapout.

Lower member

The basal conglomerate of the formation, the Barnes bed, consists of well-rounded granules, pebbles, and cobbles moderately well to very well cemented in a matrix generally of arkose, but in places of feldspathic sandstone. Pebbles and cobbles of light to dark gray, and grayish-orange to grayish-red vitreous quartzite are dominant; pebbles of quartz and reddish-brown jasper are common, and pebbles of volcanic rock (largely rhyolite) can be noted in some outcrops. The gravel constituents are largely spheroidal or ellipsoidal in shape, but in some localities a large portion are discoidal. The discoid gravels ordinarily are not imbricately alined, and where imbrication does exist no particular direction of imbrication has yet been noted. The matrix ranges from very fine-grained to very coarse-grained, and is ill-sorted and generally much coarser-grained but otherwise is quite like the arkose of the overlying basal three-quarters of the lower member. Commonly lenses of pebble-free arkose exist within the conglomerate.

The Barnes conglomerate varies considerably in character in short distances along the outcrop. In places the conglomerate is represented only by scattered granules and small pebbles in the basal few inches of the Dripping Spring arkose; at the other extreme the conglomerate may consist of closely packed cobbles and be as much as 40 feet thick. Along one-half mile of outcrop a range of thicknesses from 2 to 15 feet is not uncommon. Generally, the Barnes conglomerate is 5 to 30 feet thick, and seemingly is a more persistent unit than the somewhat similar Scanlan conglomerate. In a given vertical exposure the gravels of the Barnes ordinarily are dominated by a particular size range, but this size does not necessarily persist far laterally; one outcrop may be mostly of large cobbles and an outcrop one-half mile away mostly of small pebbles. In some localities where the conglomerate is thin (1 to 5 feet) white quartz pebbles comprise most of the gravels; such occurrences have no great lateral continuity and do not appear to characterize any particular area. In short, the variations in composition, size, shape, rounding, sorting, and thickness of the gravels appear to be no greater regionally than the variations that can be observed in a small area.

In the area of lapout north of the Sierra Ancha, as previously noted, the basal conglomerate of the Apache group where it directly underlies the Dripping Spring may locally be more than 100 feet thick and largely of cobble and boulder gravels. Close to the area of lapout, where the Barnes is still underlain by at least thin sections of Pioneer, however, the gravels are not notably coarser nor the conglomerate thicker than elsewhere. This lack of a notable variation close to the area of lapout is well illustrated by an exposure along the canyon of Bryant Creek, 7 miles southwest of Young (see Diamond Butte quadrangle). There the Pioneer and most of the lower member of the Dripping Spring locally lap against a small hill of older Precambrian quartzite and volcanic rock. A zone of cemented rubble, 4 to 12 feet thick, of angular quartzite blocks mantles this particular fossil hill; thin layers of angular blocks extend along a few bedding planes in the Dripping Spring for distances of a few tens to a few hundreds of feet away from the hill, and occasional lone erratics are seen even farther away. In proximity to the hill, however, the Barnes conglomerate consists of one and in places two 6- to 10-inch layers of pebbles; where there are two layers they are separated by about a foot of arkose. In a few places no conglomerate was seen along the contact between the Pioneer and Dripping Spring formations. In these exposures the gravels are locally dominated by white quartz pebbles, which are unusual in that they are largely angular to subrounded; however, quartzite and rhyolite pebbles are well rounded and overall the conglomerate is not greatly different from many examples seen in areas farther south.

The transition from the Barnes conglomerate to the overlying non-pebbly arkose is generally only a few inches thick, but transition zones of pebbly arkose several feet thick have been seen in several localities. A few feet or even a few tens of feet of the arkose that immediately overlies the conglomerate is ordinarily, however, slightly coarser-grained than the rest of the lower member.

The contact between the Barnes bed and the Pioneer is everywhere sharp, but in many places is slightly undulatory. A few scattered instances of slight angular discordance between the Barnes and the Pioneer, and of channels 1 to 2 feet deep in the top of the Pioneer have been seen. Unfortunately, the upper part of the Pioneer is almost everywhere a monotonous sequence of siltstones, commonly poorly exposed, and lacking in distinctive marker units; if a regionally angular unconformity did mark the top of the Pioneer, it would not be readily recognized. In a few places the upper 6 inches to 3 feet of the Pioneer, normally grayish to blackish red, is bleached greenish gray or almost white. No fragments of the Pioneer have been observed in the Barnes or in the basal arkoses. Only in the northernmost areas of outcrop does the arkosic lower member of the Dripping Spring have a purplish cast that might be attributed to fines derived from the Pioneer. In general the two formations seem conformable.

In view of its obvious lithologic and genetic affinity to the Dripping Spring quartzite, and because it is generally thin and locally even absent, the Barnes conglomerate should be regarded simply as the basal bed of the Dripping Spring quartzite. As a stratigraphic unit the Barnes should be considered a convenient marker bed where present, but it should not be afforded formational or member status.

The lower member of the Dripping Spring quartzite is of resistant, massive-cropping, hard arkose and feldspathic quartzite. The lower two-thirds to three-fourths of the member--excluding the Barnes--is of pale brown to pale reddish-brown or reddish-orange, generally fine- to medium-grained, thinly-laminated to thick-bedded, firmly cemented arkose, which weathers grayish-orange pink to moderate brown. Roughly the upper one-quarter to one-third of the member is of light gray to pale red feldspathic quartzite. In many places small pebbles of light gray chert and white quartz are sparsely scattered through this quartzite. Although the pebbles are not seen in every outcrop, and where seen are not in every bedding unit, scattered pebbles can be considered characteristic of the quartzite. Outcrops of the quartzite tend to be "pock-marked", owing to the weathering free of less firmly cemented aggregates of sand. In texture and bedding the quartzite is quite like the underlying arkose.

North of the Sierra Ancha, in the area where the formation thins by lapout, at least some sections of the lower member are largely of medium- to coarse-grained arkose, rather than the fine to medium textured arkose seen in most sections. Elsewhere no notable lateral differences in texture have been seen. Throughout the area of distribution, an individual unit of arkose is fairly well sorted. That is, at a given locality a particular unit of several beds may be mostly of medium-grained sandstone; that above or below may be uniformly fine-grained, and little vertical gradation of sand sizes can be seen within a unit.

The entire lower member is comprised of tabular beds which individually exhibit cross-stratification. Many of the beds are only 4 to 6 inches thick, but those that comprise the bulk of the member exceed 4 feet, and in some areas several beds are 12 to 20 feet in thickness. The cross-bedding is generally of the straight or concave (tangential) type (see McKee and Weir, 1953), and largely of low to moderate angle and small to moderate scale. The bedding is only poorly etched into relief by weathering, and for most outcrops the internal cross-stratification is so inconspicuous that the type and scale are difficult to determine. Nor is the thin- to thick-bedding conspicuous. In most areas 3 to 5 partings of consistent stratigraphic position etch out strongly in every exposure of the member; the rest of the bedding partings are seen only on close examination. Only rarely are the thinnest bedding units defined by partings. The casual impression of the usual exposure of the lower member is of very thick-bedded tabular units with little or no internal structure.

In the rare areas that the lower member has been deeply decomposed bedding partings between the tabular units and the cross-bedding within these units etches out in striking relief. In several such exposures in the northern part of the region and in rare examples elsewhere cross-bedding of medium to large scales (5 to 40 feet between planes of truncation) is seen in wedge-shaped sets of strata that comprise the thicker bedding units. The dip of these crossbeds in some areas exceeds 20 degrees, but in other places it is 10 degrees or less. Perhaps such wedge-shaped sets of crossbeds, particularly those of low dip and large scale, are more prevalent regionally than is now appreciated, but the bedding features of the member are generally so obscure that such a characterization cannot at present be made with confidence.

Rare channels, a few inches to a foot or two in depth and a few feet wide, exist between bedding units. And if the most prominent bedding planes are exposed in plan view, well-preserved ripple-marks may be abundant in the thin siltstone or silty sandstone seams that occasionally mark such inter-bed surfaces. Exposures are rarely adequate, however, for observation of such features.

In general the upper one-half of the lower member crops as a cliff, and the lower half as a steep slope of ledges. In areas of extreme relief the entire member, including the Barnes conglomerate, may form one cliff. The boundary between the massive lower member and the thin-parting upper member is everywhere sharp; commonly the thin-bedded basal siltstones and arkoses of the upper member erode differentially leaving a bench, a few feet to a few tens of feet wide, that prominently marks that contact between the members.

Upper member

The upper member of the Dripping Spring ~~formation~~ quartzite is strikingly different from the lower member in that it is of thinly stratified units, which form thinly and conspicuously parted outcrops. The lower two-thirds of the member is dominantly of dark-colored siltstones, and the upper one-third dominantly of highly feldspathic fine-grained beds that superficially resemble quartzites. Quartzite-like beds are singly or in sets interbedded with the portions of the member that are mostly siltstone; similarly thin beds and seams of the dark-colored siltstone are numerous in the portions mostly of quartzite-like rock. Thus, the siltstones cause the entire member to be relatively dark in color and to contrast strongly with the flesh-colored strata of the lower member. The upper member crops out as a slope broken intermittently along the contour by ragged ledges or small cliffs. Thus in topographic expression, also, it contrasts greatly with the massive-cropping lower member.

In many areas, as along the canyon of the Salt River, most outcrops of the upper member are lightly coated or even crusted with reddish-orange to dark reddish-brown limonite, and from a distance the rusty-colored and thinly parted outcrops are readily distinguished from those of any other member or formation in the younger Precambrian sequence. As a consequence of slight differences in porosity, texture, composition, and exposure these coatings impart a surficial color banding. Such banding, which parallels the gross bedding features but not necessarily the minor features and may be pronounced, has been described (Ransome, 1916, p. 138) as typical of the lower member and is generally accepted as characteristic of the entire formation. The banding is only typical of the upper member and is not seen in every unit of this member in all areas. Only occasionally is a somewhat comparable thin color striping seen on outcrops of the lower member.

The strata that outcrop and fracture like quartzites are largely very fine-grained and subordinately fine-grained, though rare beds of medium or even coarse grain do exist. On fresh fracture these sandstones are grayish-orange pink to dark yellowish brown. Many specimens that appear megascopically to be largely very fine- or fine-grained prove under the microscope to include silt and clay size material that comprises 30 percent or more of the rock. Only for the coarser-grained rocks, with minor amounts of fine matrix, can the feldspar content of these sandstones be estimated. Most such specimens are arkoses with a minimum of 30 to 50 percent of clastic feldspar grains; relatively few beds are feldspathic quartzites, with feldspar content approaching that (25 percent) of an arkose. Therefore these strata will be referred to collectively as arkoses.

Much of the siltstone, whether fresh or weathered, is very dark in color. In many areas, however, the siltstone weathers deeply to light colors--yellowish gray to moderate brown--and with the interbedded fine-grained arkoses assumes a coarsely porcelaneous texture on freshly broken and natural surfaces, so that the two rock types cannot be readily distinguished. Some modification, especially of color, by weathering persists to depths ranging from a few feet to a few tens of feet below the outcrop. Adjacent to fractures, joints or other partings the leached siltstones commonly are strongly impregnated with a grayish-red limonite stain. Excavations or mine workings into several of the bleached and stained outcrops indicate that most of the leached strata would prove to be medium to dark gray if they could be observed below the zone of weathering.

The siltstones are highly feldspathic and slightly to moderately micaceous. As noted by Granger and Raup (1959, p. 425, 439), the potassium content of the siltstones is abnormally high for a clastic sedimentary rock. The K_2O analyses of specimens from six widely separated outcrops, of the principal siltstone unit in the lower one-third of the member, ranged from 10.7 to 14.6 percent K_2O, and the average of the six specimens was 12.3 percent. Fine-grained pyrite is abundantly disseminated throughout the siltstones, is the main cause of their dark color, and is the mineral that alters abundantly to limonite. Small amounts of carbonaceous material also contribute to the color, and in some places thin seams of fine-grained graphite have been noted (see Granger and Raup, 1959, p. 442-443). Pyrite also occurs in much lesser amounts in the arkoses, principally along joints, fractures, and stylolites.

The siltstones are uraniferous, and locally include sufficient uraninite and secondary uranium minerals so that during 1950-56 the upper member was intensively prospected for uranium ores, especially in northern Gila County. Although the entire member is abnormally radioactive, an appreciable part of the radioactivity is attributable to the high potassium content of the strata, according to Granger and Raup (1959, p. 438-439).

Because of the variations in surficial aspects depending on exposure--for instance, the difficulty in distinguishing porcelain-like siltstone from similar appearing fine-grained arkoses--the tops and bottoms of stratigraphic subunits of the upper member are difficult to consistently define and precisely correlate from one locality to another. Furthermore, the upper member of the Dripping Spring is one of the units of the Apache group pervasively displaced by diabase intrusions, so subunits cannot be traced laterally and their original lithology may be obscured by metamorphic effects. Subunits of the member seemingly vary in thickness and are somewhat differently positioned in the vertical sequence as viewed in different localities. Obvious gross lithologic features and bedding structures, however, do characterize certain groups of beds. Perhaps future detailed stratigraphic studies will indicate a widespread consistency in the subunits of the member. At present only generalizations can be made. The section described in table 2 is probably representative of at least those sections north of the Gila River.

Five gross units are readily recognized in most sections of the upper member. The three most conspicuous are an uppermost unit dominated by arkose, which comprises the upper one-quarter to one-third of the member, and two relatively thick units of siltstone, which comprise much of the lower two-thirds of the member. The two siltstone units are separated by, and the lower siltstone underlain by units dominated by beds of arkose. Both principal siltstone units also include thin tabular or sub-tabular beds of arkose or thin sets of such beds.

The lowest arkose unit comprises the basal 20 to 50 feet of most sections. These arkoses are generally very fine-grained and are notably micaceous in handspecimen, which is not necessarily true of the higher arkoses, and they have a tendency to weather more to reddish hues than do the higher arkoses. Individual strata are thin to very thin, commonly are separated by thin seams of slightly finer grain, and are thinly parted on the outcrop. The basal several feet of the unit commonly includes as much siltstone as arkose. Consequently the upper part, with fewer siltstone seams, crops as a ragged or hackly fractured ledge or small cliff that overhangs or tops the lower part, which forms a steep slope that recedes back from the bench that marks the contact between the upper and lower members.

Somewhere in the interval 70 to 140 feet above the base of the upper member, or about one-third of the way up in the member, the two prominent siltstone units are separated by a unit, 10 to 35 feet thick, that is mostly of very fine-grained to medium-grained arkose and feldspathic quartzite. The basal 20 to 40 feet of the upper siltstone unit commonly includes many thin interbeds of similar arkose and the top of this arkose unit can be particularly difficult to define. These interbedded siltstones and arkoses and the arkose unit commonly form a cliff exposure in the canyons. Away from the canyons the arkose unit may form the only ledge exposed in the middle part of the member.

Commonly aggregates of sand selectively weather free of the outcrops of the coarser-grained beds of this unit, leaving abundant pits ordinarily less than one inch in diameter but in some examples as much as four inches across. Such "pock-marks" are also found in other sandstones of the upper member, but seem to typify this arkose more regularly than any other.

In many areas the two prominent siltstone units are thickly littered with flaggy and slabby rubble, but in much of northern Gila County both siltstone units crop out as steep smooth dark-colored slopes, slightly convex upward, that are virtually free of rock debris and vegetation. Apparently, where considerable sulfate is being released by weathering of the highly pyritic siltstones, vegetation cannot flourish. And much of the siltstone exfoliates in small chips that, with the coarse debris from the uppermost quartzitic arkose unit, readily travel downslope and off the steep siltstone outcrops. Such barren slopes seem to be particularly prevalent where a thick diabase sill is in proximity to the upper member of the Dripping Spring. In many places in northern Gila County, the observer can readily visualize the relative positions of the various units of the upper member from distances of 1 to 3 miles, owing to the conspicuous parallel rock scars that mark the two principal siltstone units. In areas farther south, particularly at lower elevations where the vegetation seems to be more tolerant and all outcrops tend to support less vegetation, the siltstone exposures are not so conspicuous. But they commonly are still crudely marked by a relative lack of cover.

Thin-bedded fine- to medium-grained arkose ordinarily dominates the upper 100 to 130 feet, or roughly the upper one-third of the member. In some places, if the strata have been accurately differentiated in the field, as little as 60 feet of the upper part of the member is arkose. Freshly broken surfaces of this rock range from grayish-orange pink to light brownish gray or pale yellowish brown, and the rock weathers to rusty yellowish brown colors rather than the dusky red hues of the lowest arkose unit. Thus in hand specimen or from a distance this upper unit is lighter in color than the other arkose units of the member. Generally the lower 50 feet of the unit includes beds of slightly coarser grain, contains fewer siltstone layers, and is thicker bedded and less feldspathic than the remainder of the unit. In many localities the upper 5 to 30 feet of the unit includes numerous siltstone layers or is mostly of siltstone. The lower one-half of the unit commonly is a cliff-former or crops as a very steep slope of ledges; the upper half crops as thin ledges on a moderate to very steep slope. Where this arkose is exposed in the bottom of a canyon, ordinarily the entire unit is a cliff-former.

The siltstones and arkoses are stratified and cross-stratified in thin tabular beds and irregularly pinching and swelling beds, which range from a fraction of an inch to 3 feet in thickness. Thin laminae are characteristic (fig. 7).

Figure 7.--Typical lamination and irregular bedding in upper member of Dripping Spring quartzite. Siltstone of unit 8, table 2. Photograph by C. T. Wrucke.

The arkoses are somewhat thicker bedded than the siltstones, and those beds that comprise the lower 50 feet of the upper arkose unit are generally the thickest-bedded of all. Mudcracks and ripple marks may be observed sparsely in one locality but abundantly in the same stratigraphic section nearby. These features and stylolites are particularly abundant in but are not restricted to the siltstones. The stylolites commonly have an amplitude of less than 1/4 inch, but amplitudes exceeding 1 inch have been noted. Many seen in unleached siltstone are filled with pyrite and some include carbonaceous material. Those in the leached rock invariably are loci for abundant limonite. Large channels, which cut through the siltstones and any included arkoses and are filled with similarly interbedded siltstones and quartzites, have been noted in a few localities north of the Salt River, but are rare. Those seen by the writer, in the lower siltstone unit and in the arkose unit that separates the two siltstone units, range from 20 to 30 feet in depth and 200 to 250 feet in width. Granger and Raup (1959, p. 424) have noted a channel "about 50 feet deep and 700 feet wide at the outcrop", which apparently cuts into the lower member.

FIGURE 7. TYPICAL LAMINATION AND IRREGULAR BEDDING IN UPPER MEMBER OF DRIPPING SPRING QUARTZITE.

Siltstone of unit 8, table 2. Photograph by C. T. Wrucke.

Additionally the lower siltstone unit exhibits small scoured channels, which are filled with silty to fine-grained arkose. These scour-and-fill structures, described briefly by Granger and Raup (1959, p. 437-438 and fig. 55), are particularly striking where exposed in cross sections (fig. 8). Such filled

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Figure 8.--Scour-and-fill structures and compaction features in siltstones of the upper member of the Dripping Spring quartzite. Canyon of Cherry Creek in NE¼ sec. 10, T. 7N., R. 14E. (unsurveyed).

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channels seemingly are rare in one section but very abundant in a stratigraphically equivalent section not far distant. Because the scour-and-fill phenomena are unusual in their trends and distribution, and therefore their possible genetic implications, they warrant particular note.

Cross sections of the channel fills range from about 1 inch to as much as 8 feet in width; most range from 6 inches to 2½ feet in width. The depth of many of the smaller channels is roughly the same as the width, but the depths of the larger channels are commonly one-third to three-quarters of the width, and some of the wider fills are even lenticular or plate-like in cross-section rather than crudely elliptical. Figure 8 illustrates a variety of the cross-sectional forms displayed by the fills, and a typical range of sizes in a given outcrop.

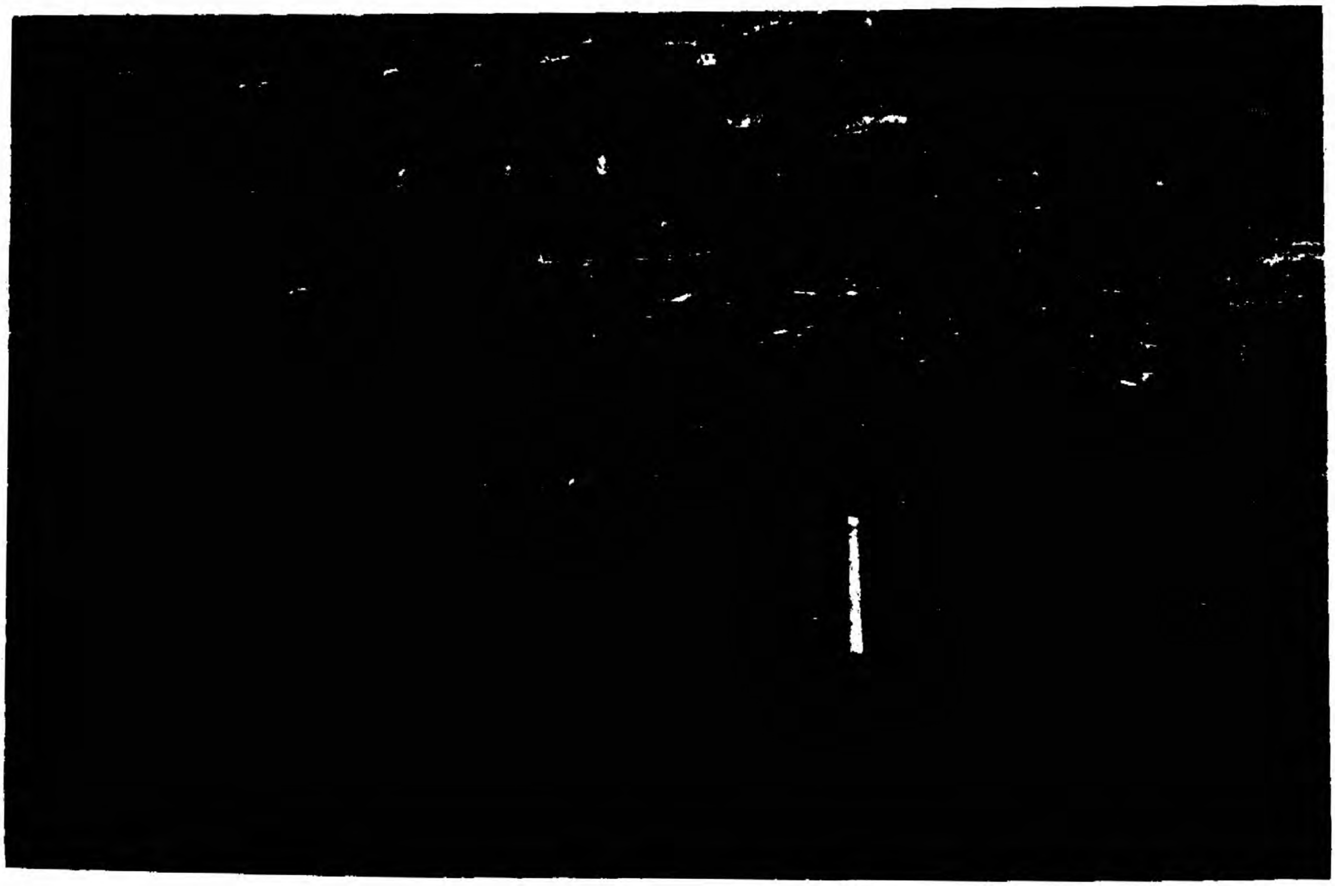

FIGURE 8. SCOUR-AND-FILL STRUCTURES AND COMPACTION FEATURES IN SILTSTONES OF THE UPPER MEMBER OF THE DRIPPING SPRING QUARTZITE.

Canyon of Cherry Creek in NE¼ sec. 10, T. 7N., R. 14E. (unsurveyed).

The scour-and-fill features ordinarily are not well exposed in longitudinal section in the same outcrop that exposes cross sections. Therefore the ratio of lengths to diameters of the fills, which are crudely cigar-shaped, is not well documented. From two different localities, for several "cigars" ranging from 6 inches to 2½ feet in diameter the lengths of the fills were determined as 20 to 30 times their widths.

A unique feature of the cigar-like fills is consistent near-parallel alignment, areally and stratigraphically. The fills of a given outcrop commonly show a variation in axial trends of only 10° to 15°, and throughout a large area they exhibit a range of trends only slightly greater. For example, in the McFadden Peak quadrangle, an area of about 250 square miles, the axial trends fall within the range N. 5°E. - N. 25° E. At the latitude of Globe the axes measured were of N. 15°W. to N. 10° E. trend; wherever seen throughout the region all were of northerly direction.

The lower surfaces of the fills are mostly rounded and trough-like in cross section; the upper surfaces arch gently upward or are almost flat. The fillings are invariably lighter colored and slightly coarser-grained than the siltstones in which the scours were eroded. The dense arkose that fills the scours may not show perceptible bedding, but if seen the stratification of these cores is undisturbed, horizontal and finely laminated. The siltstones surrounding the cores show marked compaction phenomena. The host strata along the lower sides of a fill are truncated; additionally they commonly warp abruptly downward where they terminate against the core material (fig. 8). In many examples the thinly laminated host siltstones are so intricately folded where they impinge against the cores that details of the minute folds are difficult to trace. The siltstone strata arch gently over the cores and are not truncated against them. The effects of compaction commonly persist, through the laminated or thinly bedded siltstones, for distances of several inches or even a few feet above and below the fills. Secondary shaly partings, developed abundantly in weathered exposures of the siltstones, emphasize the draping effects of compaction.

In the fraction of an inch of siltstone that is arched immediately over many of the sandstone cores shrinkage cracks commonly exist in abundance. These mudcrack-like features are somewhat irregular in outline, but grossly all are crudely alined normal to the axis of the cores. Possibly these shrinkage cracks also are partially or wholly effects of compaction.

Where the scour-and-fill features are particularly abundant the "cigars" commonly impinge one on another. An earlier-deposited fill was partly scoured away before the deposition of a parallel second core, and both of these may have been modified in the formation of a third core. Invariably, such multiple cores are separated by thin seams of siltstone. Several examples of multiple cores exist in the view shown in figure 8, but the intervening siltstone seams are not sufficiently weathered out so that the individual cores can be readily seen in the photograph. In some beds, also, separate fills may be joined along the bedding by thin tabular units of sandstone. The bottom of such connecting sandstone units is commonly undulant in cross section (see fill at bottom of pick handle, fig. 8).

In longitudinal section the scour-and-fill structures are inconspicuous, as is shown in fig. 9. In fact, unless outcrops

Figure 9.--Scour-and-fill structures of Dripping Spring quartzite viewed in longitudinal section. Both ends of pick are against scour fills. Same locality as fig. 8.

are examined closely and the cores identified and traced laterally, these features commonly will be overlooked. As a prominent joint set in the upper member ordinarily parallels the "cigars", longitudinal sections are more commonly exposed by erosion than cross sections. Therefore, the scour-and-fill features ordinarily are not noted in the abundance that they exist. Many examples of apparent undulant bedding, such as that depicted in figure 7, are actually exposures of outcrops along a vertical face that is neither normal nor parallel to the trend of the scour-and-fill structures. All irregular bedding cannot be ascribed, however, to such features.

The scour-and-fill structures commonly exist through stratigraphic intervals of 20 to 100 feet, and may characterize the entire lower siltstone unit. They have not been noted in the upper siltstone unit. They have been seen as far south as the latitude of Superior. I have not given particular heed to this part of the section farther south, and other geologists have not reported them. Very likely the scour-and-fill structures do exist much farther south, but whether they exist throughout the region of outcrop of the Dripping Spring remains to be determined.

FIGURE 9. SCOUR-AND-FILL STRUCTURES OF DRIPPING SPRING QUARTZITE VIEWED IN LONGITUDINAL SECTION.

Both ends of pick are against scour fills. Same locality as figure 8.

The thin-bedded strata of the upper part of the Dripping Spring quartzite have usually been regarded as gradational into the basal limestones of the overlying Mescal formation. As will be shown in the next section, however, the contact between the Dripping Spring and the Mescal is everywhere sharp and represents an erosional unconformity.

Mescal limestone

The Mescal limestone, as herein redefined and described, includes several features and variations that have not previously been considered integral aspects of the formation. The Mescal was described by Ransome (1919, p. 42-43), who defined the formation, 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 these layers 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, the formation has been typified as a hard, thin-bedded, cherty, dolomitic limestone, paraphrasing Ransome's early descriptions.

The Mescal limestone is divisible into three readily distinguished members: a thin- to thick-bedded lower carbonate member (180-270 feet thick); a stromatolite-bearing, in part massive-cropping middle member (40-130 feet thick); and a siliceous argillite upper member as much as 100 feet thick. In many areas south of the Gila River, the Mescal ~~formation~~ limestone is entirely missing owing to destruction during one or more of the erosional episodes that resulted in the unconformities that bound overlying formations of Precambrian or Paleozoic age. The middle and lower members persist wherever the Mescal is remnant, though a part of the middle member is commonly missing owing to pre-Troy erosion. The upper member is apparently absent everywhere south of the latitude of Roosevelt Dam.

The carbonate members of the Mescal are comprised, in different areas, of three distinctly different lithologies, two of which reflect modifying geologic processes after the carbonate rocks were lithified. The original carbonate units were cherty dolomites, which are now very subordinate in quantity in most areas. Next, as a consequence of deep leaching before and during the time of deposition of the Troy sandstones, the dolomitic members were completely silicified in some areas and at least in slight degree rendered more siliceous throughout the region of present outcrops. The larger known areas in which the cherty dolomites were converted completely or almost completely to ferruginous chert are outlined on figure 9. The last--excepting surficial weathering phenomena of late Cenozoic time--and regionally most pervasive modification is the metamorphic effect caused by emanations from diabase sills, which inflated the Mescal almost everywhere. This metamorphism, the last Precambrian event recorded in the rocks, caused the sections of cherty dolomite and dolomitic sections only partially silicified to be converted to sections wholly or almost wholly of silicate-bearing calcite limestone. Bedding characteristics were in part obscured by the metamorphism. Therefore the metamorphosed carbonate sections are of different aspect than the unmetamorphosed sections, and are described separately in the following pages.

Few extensive outcrops of cherty dolomites are known. Unmetamorphosed Mescal occurs through an area of about 30 square miles that extends north and west from Sombrero Butte along the east side of Cherry Creek (see McFadden Peak quadrangle for location). Much smaller areas of cherty dolomite exist in the vicinity of Bull Canyon at the south end of the Sierra Ancha (McFadden quadrangle), immediately north of McFadden Peak in the central part of the Sierra Ancha, along the upper reaches of Haigler Creek 10 miles north of Young, in the vicinity of 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. The following descriptions are drawn largely from these examples. Farther south only a few very small areas of unmetamorphosed cherty dolomite have been recognized. Those known to the writer 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 that a few scattered outcrops may exist in that range. Most other exposures of the Mescal are partly or entirely of metamorphic limestones--ordinarily the latter.

Stratigraphy of unmetamorphosed sections

The sequence of units shown in table 3 is quite

Table 3. Section of unmetamorphosed Mescal limestone.

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representative of the sequence as seen regionally. The lower member of this particular section is possibly the thickest, however, of any section seen to date. The middle member of table 3 is somewhat thicker than in the average section, but thicker sections exist. The section is further atypical, as will be described later, in that it was measured in one of the three small areas where a basalt flow separates the middle and upper members of the Mescal.

Lower member

The lower member of the Mescal, where not metamorphosed, is composed of dolomite; most, but not all, beds of this member include sparse to abundant chert as thin, irregular, discontinuous bands that parallel bedding. The dolomite is dense to fine-grained and ranges from yellowish brown through pale red to grayish red; brownish hues dominate most sections, but some are mostly of reddish dolomite. The chert ranges from white to black in color; that most abundant is light to medium gray. In general the dark hues of chert are in the lower half of the member; upward in the lower and middle members the chert generally is light in color and more massive, and may occur in relatively thick, irregular bands, nodules or lenticular aggregates of nodules that only grossly mark bedding. The dark-colored chert is that formed contemporaneously or nearly contemporaneously with the deposition of dolomite; much of the light-colored chert is that formed contemporaneously with the development of karst topography and other features of leaching. The chert etches out of the dolomite on weathering resulting, as others have noted previously, in a very rough irregular banding of outcrops. Throughout the region individual units of the lower member are distinctive in distribution and habit of included chert and in bedding features; so that the stratigraphic position of a given outcrop can be determined within narrow limits.

According to Ransome (1916, p. 128) in the upper part of the Dripping Spring quartzite "the beds become thin, flaggy, and rusty, with a tendency to grade into the Mescal limestone." Generally, the boundary between the formations has been accepted as gradational, and some indecision has existed as to the stratigraphic horizon that should be used as a cartographic boundary between the formations. Actually the two formations are separated, for all practical purposes everywhere, by a distinctive sandstone unit. Furthermore, the base of this sandstone unit marks an unconformity, and the sandstone is genetically a part of the Mescal limestone ~~formation~~.

This basal sandstone probably averages 5 to 6 feet in thickness; in a few places it is only a few inches thick or does not exist, in others it is as much as 15 feet thick. The sandstone is ill-sorted: it is generally medium-grained, but coarse to very coarse, well-rounded grains of clear vitreous quartz and smoky opalescent chert are characteristic as abundant individual grains and irregular aggregates. Such grains are also sparse to abundant in the basal few feet to few tens of feet of the overlying carbonate section. In places the sandstone is thinly cross-bedded, in other places slump structures are common, and in still other areas the unit is massive and essentially structureless. The composition varies. Ordinarily the unit is a moderately well cemented arkose, but in places it is a hard quartzite virtually free of feldspar. The sandstone is slightly dolomitic--or calcitic, if metamorphosed adjacent to diabase. Reddish-brown limonite abundantly mottles and veins the outcrops of arkose.

The contacts between this sandstone and the overlying and underlying units are both sharp. In many places, but only along short lateral intervals, the sandstone overlies a breccia, 6 to 20 inches thick, of angular chips of siltstone or quartzitic arkose, obviously derived from the uppermost beds of the Dripping Spring ~~formation~~ quartzite after it was lithified. This breccia has a matrix of dolomitic sandstone. Rarely angular fragments from the Dripping Spring are found in the sandstone overlying the erosional unconformity or even in the dolomite that overlies the sandstone.

The large distinctive grains of quartz and chert, the carbonate content and poor sorting cause the separating sandstone to be quite different from any unit in the Dripping Spring. These features, the breccia that locally marks its base, and the feldspathic material that represents--at least in part--eroded and redeposited material from the Dripping Spring indicate that the sandstone unit should be considered the genetic equivalent of a basal conglomerate of the Mescal limestone. Wherever, north of the Gila River, the contact between the formations has been critically examined the above features have been noted. In places the thin-bedded siltstone or silty arkose that ordinarily is found at the top of the uppermost arkose unit of the Dripping Spring appears to be thin or missing. Perhaps, therefore, significant erosion did occur after the Dripping Spring was lithified and before the Mescal was deposited. In some critical areas where the upper part of the Dripping Spring seems to be thinned, however, the outcrops are poor and subject to more than one identification. Furthermore, some apparent lateral variations observed in the upper member may be attributable to variations in sedimentation. Our understanding of such variations is wanting. Pending a better understanding of the stratigraphy of the upper member of the Dripping Spring, the unconformity between the formations probably should not be visualized as representing a profound hiatus--nor should the possibility be ignored.

Almost everywhere the poorly sorted sandstone unit crops out as a massive, rounded ledge, which conspicuously separates the Dripping Spring and the Mescal ~~formations~~. Where the sandstone is thin, as along much of the canyon of Salt River east of Canyon Creek, the unit may not be readily recognized due to a cover of debris shed from overlying units. Because the unit is everywhere thin and crops as a ledge, as a practical matter, no appreciable cartographic error is made if the base of the overlying carbonate unit is mapped as the contact between formations.

Immediately above the basal sandstone of the Mescal is a breccia of dolomite, which is generally 18 to 40 feet thick but may be thinner or much thicker. Typically this unit is comprised of cherty dolomite blocks, ranging in dimensions from small chips to slabs 8 by 10 by 30 feet, in a structureless matrix of pale red or greyish-red to grayish-orange silty dolomite. The blocks are grossly similar to the cherty dolomites of the next higher unit. And in areas where the breccia unit is thick, certain slabs can be correlated--by peculiarities of the individual chert bands, or the sequence of banding or bedding features--with flat-lying beds that overlie a thin unit of the breccia in adjacent areas. In fact, in places along the upper margins of the breccia some slabs, surrounded by typical matrix material, are separated only a few inches from their original bedding position. The blocks, in which the bedding may be approximately horizontal or in diverse orientations up to vertical, may be sparsely distributed in the matrix or closely spaced with comparatively little matrix. Small fragments are most abundant in the basal part of the breccia, and the matrix material may comprise a large part of this portion of the breccia. Very large blocks may be found anywhere in the breccia, but generally are most numerous near the top. No voids exist in the breccia. The matrix of the basal few feet of the unit commonly includes abundant coarse grains of quartz and chert like those that characterize the basal sandstone; and in many places such grains are sparsely distributed through the matrix of the greater part of the breccia.

The breccia ordinarily weathers to a smooth relatively gentle slope; the fragments and blocks of dolomite do not etch into relief and are distinguished from the matrix only on careful observation. Furthermore the slopes collect much debris. Because of poor exposures and because weathering tends to obscure rather than define the breccia, but does emphasize the silty and sandy nature of the matrix, in most exposures the breccia would be identified as a dolomitic mudstone. For this reason the fine-grained clastic rocks of the Dripping Spring have been visualized as gradational into the Mescal. On metamorphism the sedimentary structures in the breccia blocks were partially obliterated, and the quartz grains and chert were partly or completely converted to silicate minerals. Unless excellent exposures exist, this material also would be identified in the field as a metamorphosed mudstone. Everywhere, good exposures on cliffs or in fresh cuts through the unit, whether it is metamorphosed or not, belie this identification and reveal the breccia characteristics. The origin of this breccia will be considered after some features of the next two distinctive stratigraphic units are described.

In the canyon of the Salt River, between the mouth of Canyon Creek and U. S. Highway 60, the breccia is thin or missing, and in the one basal section seen by the writer in the Vekol Mountains there is no breccia. Everywhere else, if this part of the section has been afforded careful attention, the breccia has been recognized. In and near the Sierra Ancha it is about 100 feet thick in a few places. The overlying cherty dolomite unit is relatively thin where the breccia is thick, and vice versa. Data collected to the present, however, are inadequate to generalize that the thicknesses of the two units are strictly in inverse proportions. In the McFadden Peak quadrangle these two units comprise roughly the lower one-fourth of the lower member.

The breccia is transitional upward, by less brecciation and less diversity in orientation of the blocks, into flat-bedded, massive-cropping, thin- to thick-bedded (1-8 feet) pale brown, or pale red to grayish-red dolomite that includes abundant chert, which usually is dark-colored. Very irregular layers of brownish-gray to black chert are 1/20 inch to 4 inches thick and appear, in the prominently etched outcrops, to comprise 20 to 60 percent of each bed. In most places this very cherty dolomite crops out as a slope; in steep-walled canyons the unit commonly forms cliffs.

The next overlying unit is of thin to thick (1-6 feet) beds of dolomite that alternate with beds of cherty dolomite of comparable thicknesses. Some beds include appreciable chert only through the upper one-third to one-half of their thicknesses. The cherty beds are much like those lower in the section. In unmetamorphosed sections that have been studied this 20- to 30-foot thick group of beds is a striking marker due to the alternation of cherty and chert-free beds. The abundance and distribution of silicate minerals in some metamorphosed sections suggest that locally most of the beds included chert, and that such an alternation of beds may not be typical of all sections. This unit ordinarily occurs within the interval 50-85 feet above the base of the Mescal.

A few chert-free beds of this unit exhibit peculiar markings, not yet identified. As viewed on bedding planes these markings look like clusters of reeds or blades, which lie parallel to the bedding surface and radiate out from a center. Where an individual "cluster" is completely exposed the reed-like impressions do not radiate through 360° of arc but subtend two opposite 20° to 90° arcs of a circle, like a sheaf of straw gathered tightly at the middle. The long dimension of the sheaf-like markings ordinarily ranges from 2 to 8 inches, but forms as much as 2 feet across have been seen. The markings everywhere are so abundant that the "sheaves" lap one on top of another. In any given bed of a given area the individual sheaves are fairly uniform in size--for example, 2 to 3 inches. But in a different bed in the same area the markings may be uniformly of a different size--say, 6 to 7 inches.

The markings ordinarily are not conspicuous in the outcrops of unmetamorphosed dolomites, but where the enclosing beds have been converted to calcitic limestones the markings weather out very conspicuously. The metamorphic limestones tend to part readily, or even be shaly; and if such host beds are pulled apart on the splitting planes, every plane of easy separation is marked by these imprints.

In a cross-sectional view of the host beds the features ordinarily are defined only by discontinuous and slightly irregular films of silt arranged subparallel to the principal adjacent bedding planes. In some 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. In cross-sectional view one "sheaf-like cluster" is difficult to distinguish from another.

These markings exist throughout beds or sets of beds, ranging from 1 to 6 feet thick, that are separated by dolomite or cherty dolomite beds of comparable thicknesses but free of the features. Locally, asymmetric ripplemarks are abundantly preserved in the intervening dolomite beds; ripplemarks have been noted in other intervals of the lower member but are comparatively rare. The beds that include the sheaf-like markings ordinarily are in the lower 10 to 15 feet of the unit, but they have been seen throughout the unit. In unit 4 of the section described in table 3, as an example, the markings characterize 6 beds that are interspersed throughout the 29-foot unit. Wherever this stratigraphic interval has been carefully examined north of Globe the markings have been found. Farther south no search has been made for them.

Identification of these peculiar markings has been sought, without success. At first the features were noted only on weathered outcrops of silicated limestone beds, and it could be speculated that they were the remnant impressions of clusters of some silicate mineral now weathered away. Recognition of the impressions on freshly exposed parting planes, as well as in completely unaltered dolomite beds made this speculation untenable. Some consideration has been given to the possibilities that they represent imprints of ephemeral ice crystals, which might have formed on exposed mud-flats, or imprints of inorganic crystals such as evaporite salts, which are now leached away. No crystal forms quite like these markings have come to my attention. The "sheaf-bearing" beds apparently exist as stratigraphic entities throughout an area of at least 2,500 square miles. Perhaps the uniformity in size of the "sheaves" in a given bed of a particular area, and their different but uniform size in a different bed or different locality, might be explained as impressions of organic remains--probably plant remains. Such organisms might have been particularly sensitive to environmental factors, such as water salinity or temperature, so that at a given place and time they developed only in certain sizes. The stromatolites of the middle member, which undoubtedly are of organic origin, exhibit analogous characteristics that must reflect environment. Paleontologists have not advised me of organic imprints of these sheaf-like forms. The origin of the markings remains an enigma.

The unit last described and the cherty dolomite unit that intervenes between it and the dolomite breccia locally exhibit slump features or slight brecciation. Bedding laminae may be slightly distorted or actually disrupted, and bedding details are "blurred" or vague, as though slumping occurred before the carbonate mud was completely lithified. In other outcrops breccia fragments have sharp outlines but, unlike the basal breccia, are in a matrix of dolomite indistinguishable from that of the fragments. One of the commonest manifestations of such disruption is seen in the dark chert bands, which appear to be pulled apart as angular blocks and the narrow spaces between the blocks filled with dolomite. The resulting band of chert is actually a train of small angular blocks that parallels the bedding. In many instances the blocks are only slightly or not perceptibly reoriented.

Commonly the slump and minor breccia features are confined to one bed and are not necessarily found in the beds immediately above or below, though at a locality not far away the adjacent beds may include similar features. Furthermore, in a few areas where the unit of alternating cherty dolomites and chert-free dolomites exhibits numerous examples of slumping and brecciation, the top of each disrupted bed is marked locally by an intraformational conglomerate, a few inches thick, of subangular granules or small pebbles of chert similar to that in underlying beds. The sharp edges of these chert gravels generally are only slightly rounded by abrasion. Pebbles of dolomite, which are rarely seen, ordinarily are well-rounded but of poor sphericity. In some examples in which the chert fragments show no abrasion, it can be recognized that they were derived from the bed on which they lie, indicating that the chert bands existed in the underlying bed before the next bed of cherty carbonate was deposited. Furthermore, the excellent preservation of delicate bedding details in the beds and in the reworked dolomite sand matrix of the conglomerates suggests that the beds initially, or almost initially, were comprised of dolomite. If the beds had been laid down as limestones, which were dolomitized and recrystallized at a still later time, the details of original bedding probably would have been obscured.

The lenses of intraformational conglomerate have been noted only in a few localities in northern Gila County, but where they are conspicuous the beds of the interval 30-50 feet below exhibit numerous--though perhaps widely spaced--examples of local slumping or brecciation. Thus there is some suggestion that the cherty dolomite beds described to this point, immediately after lithification or perhaps in part during lithification, slumped in varying degrees. But, because the sea floor on which the carbonate muds were being deposited was above current or wave base, any irregularities produced by this settling were smoothed out by truncation of the last deposited bed or beds, and the eroded materials were redeposited as lenses of chert gravels in a clastic dolomite matrix.

Because certain peculiar features of some of the dark-colored cherts provide additional information on the environment of sedimentation, and ~~bacuase~~ because they suggest an interpretation of the above described slump features and the odd basal breccia of the Mescal, I digress here to consider the significance of these features.

Many of the layers of dark gray to black chert, in the stratigraphic units described to this point, include innumerable molds and casts that can be reasonably interpreted as pseudomorphs of halite (common salt) crystals. In most of the cherts the molds are equidimensional pits, 1 to 5 millimeters across, so rounded in outline that the cubic forms of halite crystals cannot readily be distinguished. Or the cubic outlines are distorted and could be visualized as almost rhombic. Such pits, ordinarily filled with brown dolomite, are sparse in some chert layers and constitute more than half the volume of other layers. On weathered outcrops, in which the fillings have been leached out, the highly pitted chert bands are suggestive of vesicular basalt. Such distinctive chert layers characterize the lower one-third of the lower member everywhere. Although most of the pits do not show outlines diagnostic of an original halite filling, on careful search cubic outlines can be found in most chert layers, and in some areas some of the cherts copiously exhibit cubic molds. More diagnostic, however, are larger sharply outlined molds and casts, which typify fewer chert layers but can be seen in almost every well-exposed section. These molds and casts, which range from 5 millimeters to 30 millimeters in diameter, have the hopper-shaped skeletal forms typical of halite crystals in many salt deposits (see, for descriptions, Shrock, 1948, p. 146-148; Dellwig, 1955, p. 89-95; Brooks, 1955).

Early-formed dolomite is commonly associated with evaporite deposits, therefore the recognition of salt crystals reinforces the premise that the dolomites are essentially primary.

In all features the basal dolomite breccia of the Mescal suggests the collapse of consolidated and almost consolidated cherty dolomite beds into a mud or mush-like material that once underlay them. Or visualized another way, owing to a lack of support from below a natural stoping action occurred; as this stoping progressed upward, slabs and blocks of relatively competent strata were dropped into a matrix so incompetent that it flowed readily to fill all voids. A soft mud-like base, mostly of dolomite, that would persist in unconsolidated form while higher dolomite strata were lithified, then loaded by still higher strata to cause collapse, is difficult to conceive. Moreover, the intraformational conglomerates not far above the breccia indicate that the brecciation took place while carbonate muds were still depositing and before a great thickness had accumulated. And the distribution of the lesser local slump and brecciation features in relation to the intraformational conglomerates of that part of the section indicate that like collapse occurred bed by bed as the strata accumulated. By like tokens, the breccia did not form after the entire dolomite section was deposited, lithified, uplifted, and then exposed to the leaching process that will be described on later pages. The variable thickness of the basal breccia and the lesser degrees of reorientation and movement of the slabs and blocks in the highest parts of the breccia negates the possibility that the breccia is of sedimentary origin. The breccia is a stratigraphic entity throughout an area of at least 4,000 square miles. As such it is not likely a tectonic breccia, because no traces of related tectonic structures are seen in beds higher and lower in the Apache sequence.

The occurrence of halite and its stratigraphic distribution suggests an explanation for the breccia and the slump features. The pseudomorphs of halite are particularly abundant in the cherts of the blocks and slabs in the breccia and in the cherty dolomite that overlies the breccia in most places. Fewer of the chert layers in the unit of alternating beds of dolomite and cherty dolomite include casts and molds, and only an occasional chert layer in still higher beds shows these pseudomorphs. The breccia and slump structures are distributed in a somewhat similar pattern. Slump features, like those noted higher in the section, are common in the larger blocks of the basal breccia, and where the breccia unit is thin are even more striking in contorted strata that are lateral equivalents of the breccia. Fifty to one hundred feet up in the section contorted strata are not as abundant, and healed fractures are the more common minor structural features. The basal breccia, in typical 15- to 40-foot thicknesses, is mainly confined to the interval in which halite was especially abundant. In its thickest expression of about 100 feet, the breccia includes blocks from the unit of alternating dolomite and cherty dolomite, which was probably deposited in a less saline environment than the underlying units. It is here postulated that the basal few tens of feet of the carbonate section once included a much greater amount--and perhaps variety--of evaporite salts than is now apparent.

With the freshening of the sea waters as higher strata were deposited, these salts were largely leached away, leaving a carbonate mush into which the higher strata collapsed. In most places only a few tens of feet of the overlying strata were broken, but in some places at least 100 feet of the section was brecciated. The minor slump features noted in the highest units described to this point were formed at different times, because slumping that affected one bed did not necessarily affect the next bed higher, which also may exhibit slump features. As these strata accumulated the sea waters were probably alternately highly saline and less saline. Whether the main breccia was formed as a consequence of recurrent leaching is not known. In any case, the unit to be described next was not affected by slumping, and it is likely that the basal breccia had formed by the time the deposition of the lower part of this next higher unit was complete.

The fourth distinctive stratigraphic unit crops as a cliff, topographically the most prominent outcrop of the lower member--where it is not metamorphosed--and comprises most of the middle one-third of the member. The basal 15- to 25-foot part of this cliff former is of dense, virtually structureless, clayey dolomite, which may crop as ledges but as a rule is covered by debris from the cliff. Except for detrital granules and small pebbles, this dolomite is free of chert. The cliff portion generally is comprised of thin beds, but includes beds as much as 5 feet thick. These beds are alternately of very cherty dolomite (chert as much as 80 percent) and of moderately cherty dolomite (chert 10 to 25 percent). On weathering the chert is particularly emphasized, and most of it etches out as wispy or coalescing, irregular and highly porous bands, many of which are less than 1/4 inch thick. A surface coating of black oxides is characteristic of these weathered cherts.

Although the early-formed chert, which is mostly dark-colored, does exist in higher carbonate units of the Mescal, this is generally the highest unit in which it is the prevalent chert. This also is the highest unit in which pseudomorphs of halite crystals have been seen. The chert of still higher units is generally light gray to brownish gray and exists in considerably different amounts in different sections, so that the dolomites of the upper one-third of the member are difficult to typify as to chert content. As will be shown, this lighter colored chert is largely of secondary origin. Except where the entire Mescal section has been subjected to later silification such chert is not common in the lower two-thirds of the member.

Next above the cliff-forming cherty dolomites is a unit composed of 1- to 8-foot beds of dolomite with occasional interbeds that include chert. This unit, 50 to 70 feet in thickness, crops as prominent ledges on a slope. It differs from the lower two-thirds of the member in that the dolomite tends to be in light hues of brown, rather than dark shades of brown or red, and the chert is mostly light gray and weathers whitish. Exceptions to these generalizations are known. Chert generally is in relatively small amounts in the lower one-third or one-half of the unit, but in some localities is very abundant in the upper part. Typically, chert makes up the tops of the beds. Laterally the chert content varies greatly. In places only the uppermost few inches of a few beds are cherty; in other places all beds of at least the upper half of the unit include chert, and two or three thin layers of chert may coalesce laterally to one zone of chert as much as 10 feet thick. Such lateral variation may, in extreme examples, occur in a distance of a few hundred feet. Chert nodules, ranging from a fraction of an inch to 2 feet in longest dimension, are abundant in some of the most cherty sections.

Lenses of sandstone, which parallel or are discordant to bedding, are rare in the Mescal section but exist in greater abundance in this unit than in any other. These lenses, their origin and spatial relations to concentrations of chert are considered further in a later section of this report.

The uppermost group of beds, a unit 10 to 30 feet thick, is very like the group just described except that it is thin-bedded and slabby partings are prominent in the slope-forming outcrops. The light-colored chert layers range from paper-thin laminae to layers 6 inches thick. Compared with other units of the lower member, chert ordinarily comprises little of the volume of this unit.

Although the beds of upper two units of the member are laterally variable in lithology, largely as a result of late silicification, the sequence of beds is remarkably uniform from the Vekol Mountains to the Mogollon Rim.

The contact between the middle and lower members, except where both members are remnant as Karst breccias, is everywhere sharp and easily determined. The dolomites below the contact are flat-bedded and are thin beds with a slabby parting; those immediately above include relicts of algae colonies and form massive outcrops.

Algal member

The middle or algal member, in most areas 80 to 100 feet thick, consists of two units: a lower, thick-bedded, cliff-forming unit characterized by fossil forms (stromatolites) relict of the gross outlines of colonies of algae, and an upper, slope-forming, thin-bedded unit largely devoid of algaloid structures. The algal structures grade out upward in the basal 10 feet of the upper group of beds. The algal colonies are not in mounded or true reeflike masses (bioherms), but are in a regularily-bedded blanket-like deposit known as a biostrome (Cumings, 1932; Link, 1950). North of the latitude of Globe the biostrome is fairly uniformly 50 to 60 feet thick, and generally comprises the lower one-half to two-thirds of the member. Farther south like thicknesses probably exist, but in many places the biostrome is only 30 to 35 feet thick. Throughout the region the upper unit is locally only a few feet thick or is missing, owing to one of the episodes of erosion that preceded the deposition of the Troy quartzite.

Where not metamorphosed the middle member is composed of grayish-red to yellowish-brown dolomites; the brown dolomites weather brownish to yellowish gray. Light gray to gray chert, in many places tinted by minute flecks of red hematite, occurs--generally sparsely--throughout the member. In each unit the chert content increases upward. In the massive biostrome small, irregular lenses of chert, a fraction of an inch to a few inches in thickness, are typical. Toward the top of the upper unit closely spaced chert nodules may locally comprise zones 1 to 4 feet in thickness, and in most localities the top few inches to few feet of the member is almost completely silicified. The contact with overlying rock units is everywhere sharp.

In much of the area between the canyon of the Salt River and the abandoned mining village of Chrysotile (fig. 3) the uppermost bed of the member is almost entirely of chert and displays conspicuous structures similar to those referred to as cone-in-cone structures (Pettijohn, 1949, p. 185-356). A like bed displaying cone-in-cone structures, ranging from 2 to 20 inches in thickness, has been seen elsewhere but not necessarily at the exact top of the member. Apparently during silicification a particular bed was especially susceptible to the formation of these structures. And in the Salt River-Chrysotile area this resistant bed was a base for the erosion that preceded later sedimentation. Where the upper unit is thin the bed with structures is missing; where the unit is thick this distinctive bed is as much as 15 feet below the top of the member.

From the Salt River south as far as Globe bedding planes within the massive lower unit are rarely less than 6 feet apart. In other areas, bedding planes are from 6 inches to 6 feet apart, and typically are at intervals of about 4 feet. The beds are thinnest toward the top of the algal unit.

Fossil forms such as those characteristic of the lower part of the member preferably should be termed "stromatolites," rather than "fossil algae." As well-stated by Rezak (1957, p. 129), "fossil algae preserve recognizable organic micro-structures that enable the examiner to determine their true biologic relationships"; although it is recognized that stromatolites "have come into existence through the work of certain lowly forms of algae," all that ordinarily remains of these algae are "large headlike masses" or the gross forms (stromatolites) of the original colonies. Johnson (1946, p. 1089, 1098) has suggested that a given form genera and species of stromatolite, which is typified by certain megascopic features of form, may have resulted from the life processes of several biologic species or even of several genera and species of algae that lived in constant association.

Stromatolites are present throughout the basal unit of the middle member of the Mescal, and exist laterally wherever this part of the section was not eroded prior to deposition of younger rock units. In plan the structures appear as flattened hemispheres or inverted saucer-like disks outlined by concentric laminae. The "saucers" are 2 to 20 inches in diameter. Figure 10A shows a typical

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Figure 10.--Structures typical of the stromatolite beds of the middle member of the Mescal limestone. Stromatolites identified as *Collenia frequens* Walcott. From cliff above Regal mine, Blue House Mountain quadrangle. A, View of upper surface; B, Cross-sectional view.

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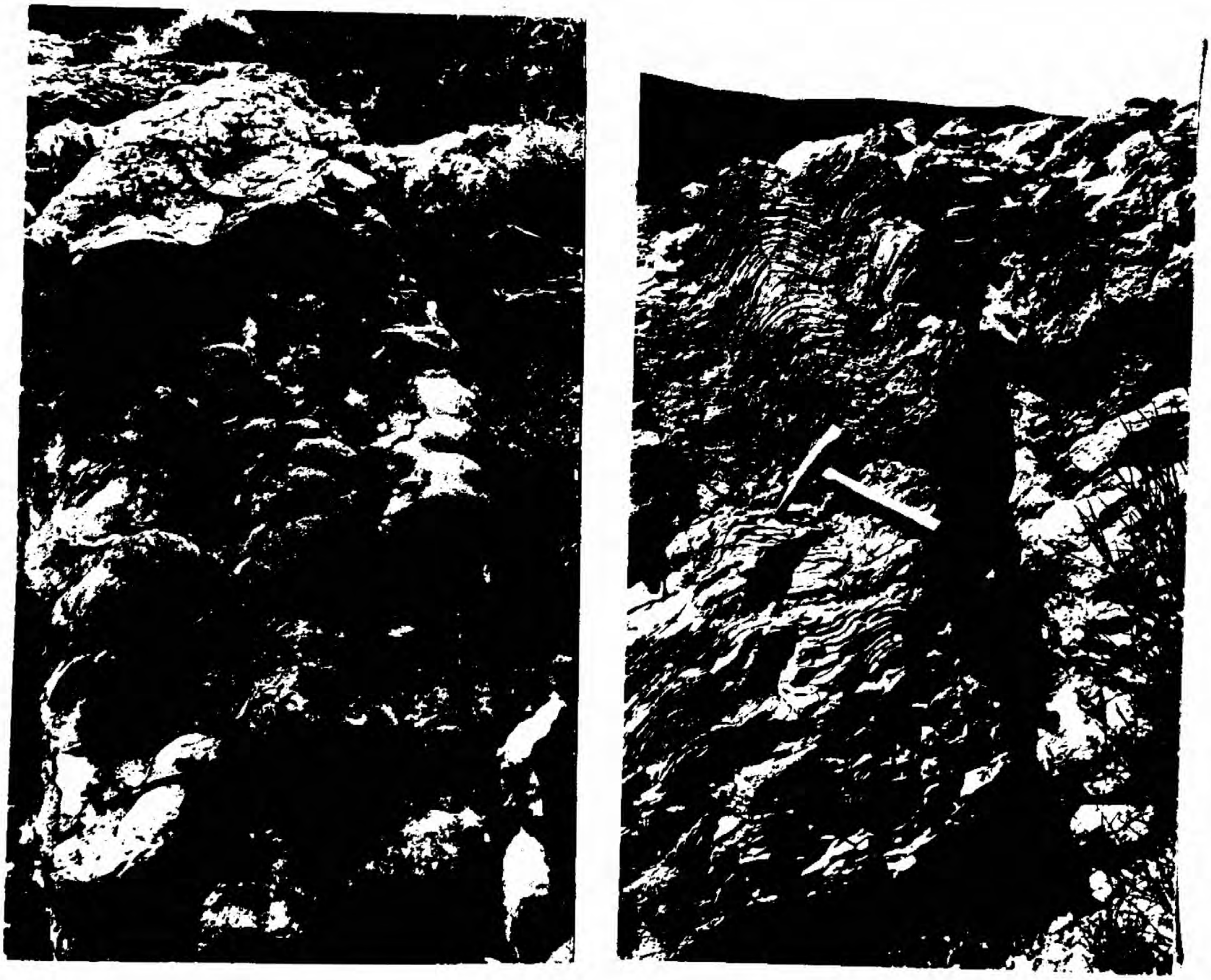
grouping, in plan, of the individual colonies. This picture represents a common expression of the larger colonies, for which exfoliated outcrops are common. Plan views in which the concentric rings are exposed are more typical of colonies of smaller diameter, especially those exposed in waterworn outcrops. In cross section the laminae, marked by chert or silt, etch out prominently on weathering as flattened "sine waves" (fig. 10B). The amplitude of the "waves" ranges from 1/4 inch to 5 inches. Ordinarily the structures are 5 to 10 inches in diameter and 1 to 1 1/2 inches in amplitude, and in any given outcrop are fairly uniform in dimensions. Richard Rezak (oral communication, April 19, 1956)



A. VIEW OF UPPER SURFACE

B. CROSS-SECTIONAL VIEW

FIGURE 10. STRUCTURES TYPICAL OF THE STROMATOLITE BEDS OF THE MIDDLE MEMBER OF THE MESCAL LIMESTONE.

Stromatolites identified as Collenia frequens Walcott.
From cliff above Regal Mine, Blue House Mountain quadrangle.

of the Geological Survey has identified these forms as _Collenia frequens_ Walcott, and has noted (Rezak, 1957, p. 133, 136-140) that they are common stromatolites in the younger Precambrian Belt series of northwestern Montana.

The basal bed, 4 to 6 feet thick, of the lower unit exhibits a different stromatolite form. In cross section the gross structures are crudely conical in form, with the apices of the cones downward. The axes of the cones ordinarily are inclined at angles of 40 degrees to 80 degrees to the bedding surfaces. The laminae within each cone-form are convex upward, and individual laminae increase in area upward. The widest part of the conical cross-sections is ordinarily 2 1/2 to 4 inches. No identification has been sought for this stromatolite form.

In recent years the stromatolite reef or biostrome of the middle member of the Mescal has been widely recognized as a distinctive unit by those who prospect for and mine asbestos, and it popularly is termed the "algal limestone." But the existence of the flat-bedded upper unit has not been appreciated generally because: its outcrops ordinarily are not conspicuous, or in places much of the upper unit was eroded away before deposition of higher strata, or the two units of the member are separated by a diabase sill. Regardless of preferred terminology and the lack of stromatolite forms in the upper unit, because the biostrome by being prominent seemingly dominates the middle member and is commonly called the "algal limestone," to more distinctively designate the whole of the middle member it will be referred to as the "algal member" in this report.

Upper member

Through much of northern Gila County an upper member, comprised mostly of argillite and subordinately of fine-grained mudstone, unconformably overlies the middle member of the Mescal. South of the latitude of Roosevelt Dam this member apparently is everywhere missing owing to erosion prior to deposition of the Troy quartzite (see fig. 3 for area of outcrop). Farther north to about the 34th Parallel, though locally missing for the same reason, the member ordinarily ranges from 25 to 100 feet in thickness; a thickness of 50 to 80 feet is typical for large areas. In three small areas remnants of basalt flows, like those that overlie the Mescal, intervene between the middle and upper members of the Mescal ~~formation.~~. Throughout most of the area north of the 34th Parallel the Troy quartzite rests directly on the middle or lower members of the Mescal.

In the Sierra Ancha and eastward to Canyon Creek this upper member is dominated by yellowish-brown to reddish-orange, minutely laminated, but flaggy- to massive-parting, siliceous argillites interbedded with subordinate units of medium gray to black argillite, which are somewhat less siliceous and commonly exhibit a shaly parting. The sediments originally deposited apparently were entirely of materials of less than silt size, and may have been of fine clay sizes. But now these argillites are very hard and dense rocks that everywhere exhibit incipient recrystallization, which obscures their original mineralogy and texture. The least recrystallized examples are abundantly spotted with minute aggregates, 0.2 to 3 millimeters in diameter, of mica or amphibole plus other minerals so fine-grained as to be difficultly resolvable under the microscope. The upper and lower boundaries of the member were particularly favored horizons for the intrusion of diabase sills. And adjacent to the intrusions many of the argillite beds may be coarsely mottled throughout with aggregates of similary fine-grained minerals.

The average radioactivity of the argillites, as determined by airborne scintillometer in and near the Sierra Ancha, is higher than that of all other units of the Apache group except the upper member of the Dripping Spring quartzite, which it roughly equals in radioactivity (Gingleby and Mead, 1955, p. 9). This characteristic led to considerable prospecting of the member during 1954-55, but the prospecting apparently resulted in the discovery of only a few insignificant showings of uranium-bearing material. One typical specimen of comparatively little-recrystallized argillite, submitted for analysis by semiquantitative spectrographic methods, contained potassium in excess of 10 percent (Neuerburg and Granger, 1960, p. 766). Judging only from the one sample, the argillites may have an abnormally high potassium content comparable to that of the upper member of the Dripping Spring. And this content, in large part, may account for the abnormal radioactivity.

North of the Salt River the basal unit of the upper member is composed of chert and typically is a few inches to 10 feet in thickness. This unit in some places is a laminated or thinly bedded chert; more commonly it is a breccia of small angular chert fragments in a matrix of chert, and in still other places it is a conglomerate of well-rounded chert pebbles in a matrix of silty chert. In a few small areas the conglomerates are as much as 40 feet thick. This chert unit overlies an erosion surface on the middle member. The chert breccias and conglomerates apparently were derived by the breaking up and reworking of the bedded chert, and are not mainly of detritus derived from the underlying member. Locally, however, small amounts of chert from the underlying algal member are incorporated in the basal few inches of these rocks. In many areas, the chert unit crops as a prominent ledge and is a striking marker separating the upper and middle members. The basal chert unit is absent in only a few places north of the Salt River, but elsewhere is commonly absent. Where it is missing the contrast between the argillic beds of the upper member and the much silicified carbonate beds at the top of the middle member is still striking.

In a few places in the Sierra Ancha one or two limestone units, 6 inches to 8 feet thick, are enclosed within the argillite. The best exposures known to the writer are in the vicinity of Asbestos Peak at the south end of the range (see McFadden Peak quadrangle for location); this limestone, 2 to 6 feet thick, is about 20 feet above the base of the member. At two other localities in the central part of the Sierra Ancha (near Workman Creek Falls and on the east cliff wall of McFadden Horse Mountain, McFadden Peak quadrangle), where the member is about 60 feet thick, thin beds of limestone exist in the top few feet of the member. These limestones apparently are lenticular, because they have been seen in only a few sections. Nowhere have these carbonate beds been seen in a section free of the metamorphism caused by the diabase intrusions. Judging from the forms and distribution of included aggregates of silicate minerals, the limestone beds in the unmetamorphosed state were silty and moderately cherty dolomites.

East of the longitude of Canyon Creek and south of the Salt River the upper member is not everywhere present, and almost everywhere it is inflated and displaced by diabase sills. Therefore details of the stratigraphic section in this area are not as well known as farther west. The lowest 20 to 30 feet of the member tends to be of shaly or crumbly, greenish-gray to dark gray mudstone, which crops poorly and in most places is concealed by debris from overlying siliceous argillites. These argillites, which are like those described above, comprise the rest of the member. Calcareous or siliceous concretions, as much as 3 inches in diameter, occur at least locally in the nonresistant friable mudstone. The basal chert bed of the member exists only locally and is thin. But the uppermost 1 to 4 feet of the underlying algal member are almost everywhere completely silicified and reddened by abundant hematite, and might be mistaken for this resistant basal bed.

Late-formed chert in the Mescal

In at least two extensive areas, outlined approximately on figure 3, and in several small areas the dolomite members of the Mescal were completely or almost completely converted to chert in Precambrian time. And in several areas of a square mile or less all the middle member and the upper part of the lower member were silicified. Also, some individual dolomite beds throughout the region were at least partially silicified. This chertification was part of a process of leaching and solution of the dolomites that resulted in the formation of karst topography in certain areas and in genetically related massive collapse breccias in other areas. The deposition and distribution of silica and associated concentrations of hematite can be best described and understood if the sections that include the lesser amounts of secondary silica are considered first.

Some features of the late chertification are best described in relation to the erosional unconformity between the upper and middle members of the Mescal. In most areas the upper member seemingly is concordant with the underlying carbonate beds, and the only indication of unconformity is the silicification of the uppermost strata of the algal member. Similar concentrations of chert along contacts between formations have long been recognized (Leith, 1925) as suggestive of an unconformity. In a few localities, pebbles and angular fragments, obviously derived from the topmost silicified beds of the algal member, but now incorporated in the basal strata of the argillite member, further substantiate an erosional hiatus. Truncation of the beds of the algal member by the unconformity can be recognized in a few areas, if the contact is observed along a considerable length. The thinning caused by this mechanical erosion and the silicification immediately below the contact are certainly subordinate phenomena, however, compared with the solution thinning and silicification that affected some areas of dolomitic rock.

In widely separate localities, massive bed-like bodies of sandstone or quartzite are sparsely scattered through the carbonate section; most are in the upper one-third of the lower member. For a considerable time these "beds" were seen only in limestone terranes, where relations are obscure owing to metamorphic effects. These sandstones were puzzling because they exist only in certain areas, do not occur everywhere at the same stratigraphic horizon, most are devoid entirely of features that can be construed as sedimentary structures, and all were apparently lenticular. Recently these "beds" and related sandstone "dikes" have been observed in well exposed outcrops of dolomite, in forms readily recognized as filled solution cavities. Especially in the central part of the McFadden Peak quadrangle, rubble-filled sinkholes and solution enlargements of joints have now been observed in the algal member, and in this area some solution openings and sandstone fillings have been traced downward almost to the bottom of the lowest member. Adjacent to the sinkholes, solution cavities along bedding planes are abundant phenomena.

The solution features pre-dated, at least in part, the deposition of the argillite member, as is indicated by the geometric relations of some cavities to the erosional unconformity. Some chert incorporated in the basal beds of the upper member is similar to chert spatially concentrated in the vicinity of the sinkholes. The collapse breccia of a few sinkholes is in an unbrecciated matrix of argillite like that of the upper member. Also, solution voids along bedding planes are commonly rendered prominent by a thin filling of reddish-orange argillite (see descriptions of units 7 and 11, table 3). Locally the bedding of the basal part of the argillite member is undulatory and crudely approximates in cross section the configuration of underlying dolomite beds that subsided locally as a result of partial leaching. Such undulations were apparently formed while the argillites were still plastic; they have not been found away from sinkhole areas.

Collapse and probably additional solution continued during the extrusion of the basalt flows that overlie the upper member; certainly such effects occurred during the deposition of the lower parts of the Troy formation. Angular blocks of argillite, broken after lithification, are noted in some sinkholes. Rare apophyses of basalt that extend downward from an overlying flow and partly fill tabular voids in the carbonate rocks have been seen. The prevalent filling, however, is of sandstone identical with that of the middle member of the Troy. In a few examples--those best exposed are 1 to 2 miles southwest of Gunsight Butte in the north-central part of the McFadden Peak quadrangle--the roofs of sinkholes collapsed, allowing plug-shaped masses of the lower and middle members of the Troy to drop and fill the sinkholes. Within these masses most vestiges of bedding were destroyed, but brecciation is not apparent in the sandstones of the middle member. Therefore the collapse must have occurred while these sandstones were unconsolidated. Comparatively few bedding-plane cavities are filled with arkose like that of the lower member of the Troy. The quartzitic fillings are sandstones that were indurated adjacent to diabase intrusions.

The chert content is much higher in beds that exhibit abundantly the above described effects of solution than in stratigraphically equivalent sections that show little leaching.

Cherts of two generations occur in the dolomites of the Mescal. As already described, the older cherts--best represented by the dark-colored, thin layers or the irregularly, wispy layered dense cherts particularly characteristic of the lower two-thirds of the lower member--were formed penecontemporaneously with the dolomites. Chert of the later generation is mostly light in color, much of it is flecked with hematite, and it ordinarily--but not necessarily--occurs in thicker and more irregular masses than the older and denser chert. Because the later chert occurs in greatest quantities in dolomites that exhibit solution phenomena, and because in detail particular concentrations border and extend outward from solution cavities, this chert formed after the lower and middle members were completely lithified. In sections mostly of dolomite the younger chert is largely confined to the middle member and the upper one-third of the lower member.

Where the later chert is particularly concentrated in such sections, lens-shaped aggregates of ellipsoidal nodules of chert in a matrix of chert and carbonate locally take the place of one or more carbonate beds. On outcrops the carbonate of the matrix and the relict unreplaced carbonate within the nodules weathers away, so the nodules appear to be "punky" and in a vuggy matrix. Some beds were more susceptible to leaching and silicification than others. Some include little secondary chert; other beds exhibit numerous discrete nodules of such chert. The chert nodules or aggregates of nodules are ordinarily concentrated in the upper portions of individual beds. These nodular chert zones are those that vary considerably in thickness, and a sequence of beds that includes such chert may also show striking lateral variations in thickness. Typical variations might be as follows. In an ordinary section a 20-foot sequence of sparsely silicified beds may be capped and underlain by 1-foot zones of nodular chert. If traced a mile one direction the nodular zones of secondary chert virtually pinch out and the equivalent sequence may be 22-24 feet thick. Thus overall there is only a modest variation in thickness and in chert content. But in the opposite direction the chert and solution thinning of the intervening dolomites may increase to a considerable degree, so that in a distance of a mile the two chert zones may coalesce and the equivalent stratigraphic interval is only 6 to 10 feet thick. In exceptional instances, as noted earlier, such a change of chert content and thickness can be seen in a lateral distance of a few hundred feet. In such places some brecciation, caused by slump on solution of the dolomite, can usually be observed in individual beds. Voids in the secondary chert zones may or may not be filled with Troy-type sandstone. Owing to solution and settling the carbonate members, and especially the upper one-third of the lower member, thin considerably in some areas.

Fragments of early chert incorporated in the secondary chert can be recognized in many places. Such recognition is made difficult because on leaching much of the older chert was bleached to a light color and rendered somewhat porous. Thin bands of the early chert were completely bleached, and the borders of thick bands or thick fragments were bleached. Without doubt a considerable part of the volume of the zones of secondary chert represents relicts of the early-formed cherts mechanically concentrated by the removal of the dolomite. Indeed, in some examples of completely silicified sections individual beds, on close inspection, are seen to be comprised of angular fragments of the early chert in a matrix of the late chert. In a notable occurrence along the south 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 the basal 30 feet of the algal member is so perfectly preserved that detailed stratigraphic correlations are readily made. This sequence and in general the individual beds, however, are only about one-third the thicknesses of the equivalent intervals in nearby little silicified sections.

Where the carbonate section is metamorphosed, ordinarily no distinctions can be made between the early and the late chert. Silicate aggregates that pseudomorph the zones of closely-packed nodular chert, characteristic of the late chert, are readily recognized by their gross outlines.

All gradations from dolomite sections that include little secondary chert to those that are almost entirely of chert exist. In a few places the algal member was completely silicified during the development of solution cavities. The resulting unit is a massive-cropping silica-cemented chert rubble in which angular blocks of silicified stromatolites can be recognized. Occasional blocks of unaltered dolomite can sometimes be found in this breccia. The best examples in which relations to the surrounding carbonate terrane can be observed have been seen in the central and northwest parts of the McFadden Peak quadrangle. There a zone of rubble, containing recognizable relicts of the entire middle member, may be as little as 20 feet thick; but one-half mile away a little leached equivalent dolomite section may be as much as 100 feet thick. In such areas of silicification commonly the uppermost 20 to 30 feet of the lower member is similarly silicified, and the siliceous breccia of the middle member merges with that of the lower member, obscuring the usual sharp contact between members. Cavity fillings of sandstone or quartzite occupy the place of a few of the uppermost beds of the relatively little leached dolomite section immediately below such rubble zones. Sandstone fillings are probably more abundant in such settings than in any other.

Also, completely silicified sections of both dolomite members in which the entity of individual beds is preserved grade laterally into sections comprised entirely of a heterogeneously agglomerated collapse breccia. The latter commonly is a coarse breccia very like that seen in individual sinkholes, and it was probably formed--with long continued leaching--by the coalescence of sinks developed in the dolomite terrane. In areas adjacent to those in which karst topography was being formed beds were leached less actively, but were gradually thinned and silicified to such a degree that large-scale collapse, typical of the areas in which sinkholes abound, was no longer possible. Thus, locally within karst areas that generally now display only rubbly remnants of the Mescal, the stratigraphic sequence of whole units is still preserved as at Shell Mountain.

The most extensive areas of chertification, and those in which silicification of the dolomite members is virtually complete, are those in which the upper member of the Mescal and the basalt that overlies it were removed by pre-Troy erosion. Additionally, they are areas where the middle member of the Troy, not the lower member, rests directly on the middle or lower members of the Mescal. The converse, that the carbonate rocks displaying such relations to the Troy are everywhere thoroughly silicified, however, is not true.

The largest known area in which the dolomite members of the Mescal were pervasively silicified is encompassed by that portion of the drainage basin of Canyon Creek north of the 34th Parallel (see fig. 3). This belt of silicification extends northwest as far as the headwaters of Haigler Creek, and in places westward along Haigler Creek considerable thicknesses of the Mescal are silicified. In this belt the Mescal was thinned by erosion prior to deposition of the Troy, so in places the middle member of the Troy rests on the lower member of the Mescal. Dolomite comprises a large part of some exposures in this belt, but almost invariably such sections exhibit much collapse breccia like that formed in the sinkholes. In many places the only part of the section that was not subjected to brecciation as a consequence of late solution phenomena, was the older massive breccia that is the basal dolomite unit of the Mescal formation. This breccia lacked the bedding planes that were favored channelways for solution, and apparently was therefore somewhat less affected.

Another large belt of silicified Mescal, exposed for a width of 1 to 3 miles, extends from U. S. Highway 60 southeast about 13 miles along the southern margin of the Natanes Plateau (fig. 3). In this area the Mescal was considerably thinned by pre-Troy erosion, and locally the middle member of the Troy even rests on the Dripping Spring quartzite.

Immediately east of and just within the northeast corner of the Globe quadrangle, through an area possibly not exceeding 6 square miles, only 60 to 120 feet of thoroughly silicified Mescal intervenes between the Dripping Spring and Troy quartzites. Although there are other areas of like extent, this one exhibits particularly well some notable features. The thinning of the carbonate members in this area appears to be largely a solution phenomena. Little large-scale jumbling of breccia blocks, as occurred in the karst areas, is seen and stratigraphic units representative of all of the lower member and part of the middle member can be recognized. In contrast, exposures of the dolomite members 1 to 2 miles to the east are little silicified and the sections there exceed 300 feet and possibly approach 400 feet in thickness. This silicified belt includes comparatively little hematite, whereas abundant hematite is characteristic of large parts of the two cherty belts noted above. Nor is the basal part of the overlying Troy formation abundantly reddened by detritial hematite as is common in the more northern areas. In several other areas all or considerable parts of the Mescal section were silicified; these silicified sections exhibit features like or intermediate to those exhibited in the three areas described above.

Where the carbonate members are thoroughly brecciated and silicified the breccia, particularly in its upper part, includes much sand in its matrix. Also the matrix, remnant blocks of dolomite and the beds of the basal 10 to 30 feet of the Troy commonly are highly hematitic. The basal one-half to two-thirds of the silicified Mescal formation may or may not be darkened by ferruginous material.

Because sand in the breccia matrix increases upward, because the basal conglomerate of the Troy may be largely of locally derived angular breccia fragments, and the abundant hematite tends to obscure internal structures of both formations, in a few places the contact between the Troy and the silicified Mescal is somewhat difficult to determine. The angular pebble conglomerate at the base of the Troy, however, generally includes some well-rounded pebbles of foreign derivation and shows stratification indicative of lateral transport. Thus in most areas it is readily distinguished from the angular collapse breccia of the underlying Mescal.

Where the original bedding of the dolomites was pseudomorphed in the silicified Mescal, the bedding of the uppermost 10 to 70 feet of the formation may be relict only in trains--as seen in cross section--of little reoriented angular blocks of chert, separated by an abundant matrix of hematitic siltstone or sandstone but lying in broken layers that reflect the original bedding. Delicate details of original bedding are commonly preserved in the individual blocks by thin films of silt or by porous laminae in which minute aggregates of specular hematite are abundant. If the blocks are relicts of the stromatolite-bearing unit of the middle member, the original curvature of the stromatolites as seen in cross section commonly was accentuated on silicification. In some relict beds the blocks are separated horizontally by little matrix; in other beds individual blocks are separated by 2 to 4 inches of matrix material. Commonly, alternate beds are characterized by blocks of different aspects of size, separation, or reorientation. The individual chert fragments low in the brecciated section ordinarily have rims of reddish-brown color. Many blocks high in such breccias are stained reddish brown to grayish red throughout, and could be termed jasper rather than chert. Near the bottom of the 10- to 70-foot intervals the blocks are rudely domino-shaped and may be more than a foot in maximum dimensions. Upward in the sections their dimensions decrease, the angular edges may be rounded, and generally the blocks are more widely separated in the matrix. Right up to the contact with the overlying Troy, however, the relict bedding may be well preserved. Thus the rounding of the blocks and their wider separation must be a solution phenomenon and not a consequence of abrasion and lateral transport by current or wave action.

Hematite concentrations in the Mescal

As a geologic phenomena related in space and time to the concentration of secondary chert, hematite was concentrated at the top of the Mescal carbonate section. Some of these concentrations are large enough and of high enough iron content to have been prospected to some degree in the past decades (Burchard, 1931; Stewart, 1947). And since 1958 the deposits have been very actively prospected.

The iron deposits are local concentrations of hematite, which occur along the upper boundary of and in most instances entirely within the silicified Mescal formation, in the settings described in the preceding section. Earthy but dense hematite and unctuous specularite in massive layers, which parallel the relict bedding of the enclosing formation, constitutes most of the ore-grade material. The earthy hematite is dominant in the lower grade ores. Small botryoidal masses of hematite are sparse but widely distributed. Micaceous, splendent specularite commonly lines small vugs and comprises veinlets in the hematite layers and in layers of hematitic rock that are interlensed with the ore. The ore material is everywhere siliceous. Hematite and silicia (SiO_2), the latter largely in the form of chert and detrital quartz grains, are roughly in inverse proportions and together generally comprise 90 to 98 percent of material that assays more than 25 percent iron.

The best known and to the present (1961) the largest deposit known is exposed on the east and west walls of Canyon Creek near its junction with Swamp Creek, a minor tributary 6 miles south of the Mogollon Rim.[1]

[1] I am greatly indebted to A. F. Butler, Jr. (written communication) for the dimensional data summarized in these notes on this deposit.

In this deposit, the ore-grade material is in a zone 20 to 70 feet thick, in which the layers of hematite are intercalated with and gradational into layers of highly hematitic argillite(?), siltstone, silicified sandstone, and light-colored to dusky red chert. This zone crops out almost continuously for a north-south length 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. The middle part of zone, generally 5 to 25 feet thick, contains the highest content of iron and roughly two-thirds of this zone, wherever exposed, has an iron content of more than 30 percent. The higher-grade portions of the zone are discontinuous lenticular bodies; individual lenses with an iron content of more than 50 percent range from 2 to almost 40 feet in thickness in the exposures of the Swamp Creek deposit. The high-iron portion of the zone grades downward into layered chert breccia identical with that described in the previous section. Tabular chert blocks or layers of little reoriented chert fragments similar to those described, except that the layers or fragments are in a matrix of silty or sandy hematite, are sparsely to abundantly interlensed throughout the "iron-formation". These layers are fewest in the high-grade middle part of the zone. Thin layers of highly hematitic sandstone, thoroughly impregnated with silica, commonly constitute a considerable portion of the upper part of the zone.

Lenses of hematite-bearing rock, too small and thin to constitute potential ore, occur locally in silicified Mescal strata below the main 20- to 70-foot ore zone. The basal few feet of the overlying middle (Chediski sandstone) member of the Troy and thin lenses up in the member locally include concentrations of hematite, which approach ore grade in the vicinity of the Swamp Creek deposit.

Several other iron deposits, petrologically similar to that at Swamp Creek, exist in the large area of silicified Mescal that outcrops along the upper reaches of Canyon Creek. Outcrops of some of these are as widely distributed, or even more widely distributed. But none of the outcrops--as far as is now known--exhibit lenses with an iron content of more than 40 percent that are as continuous or thick as those at Swamp Creek.

Lesser concentrations of hematite-bearing rock occur at the top of the Mescal south of the 34th Parallel, but rarely are lenticular units of almost pure hematite seen. Instead, thin chert layers or thin layers of little reoriented chert fragments lie in a matrix that is largely of hematite. Commonly this matrix is of coarsely micaceous specularite, rather than earthy hematite. The belt of siliceous Mescal that is exposed along the south rim of the Natanes Plateau includes local concentrations of such material, some of which are several feet thick. This is the southernmost area known to the writer in which such concentrations have outcrop areas of several acres. But thin layers and lenses of hematite a fraction of an inch to a few inches thick, can be found at least sparsely distributed through the upper part of the Mescal wherever an appreciable thickness of the section is silicified. Rare lenses of hematite, a few inches thick and as much as 2 feet across, are even found in dolomite remnants within partially silicified sections or in dolomite beds immediately below the zone of silicification. And hematite abundantly flecks the chert that immediately underlies, everywhere, the unconformity that marks the top of the middle member of the Mescal--whether the remainder of the carbonate section includes such secondary chert or not.

North of the 34th Parallel the basal several feet of Troy sandstone immediately overlying the iron deposits commonly includes so much hematite that the relations of the Troy to the underlying hematitic Mescal are obscure. South of the 34th Parallel, and even in many places north of it, water-worn rounded pebbles of hematite are locally abundant in the basal beds of the Troy, and occasionally are in conglomerate lenses tens of feet above the base of the formation. From their distinctive lithology, there is no doubt that these siliceous hematite pebbles were derived from the underlying Mescal formation. Tentatively, in past discussions of the iron deposits, the diabase intrusions that everywhere post-date the Troy quartzite, have been cited as possible sources of hydrothermal solutions that deposited iron along the unconformity between the Mescal and Troy formations. Even without the evidence afforded by the hematite pebbles in the Troy, in many places geometric relations between diabase bodies and the iron-bearing zones clearly indicate that the intrusions post-date the hematite deposits.

Mutual spatial relations indicate that the secondary chert and the hematite in the Mescal had genetic aspects in common. The hematite deposits are here regarded as fossil lateritic accumulations, which in part may have been mechanically reworked and transported short distances along the pre-Troy erosion surface. It can be speculated that the dolomites of the Mescal or the basalt flows, which once everywhere overlay the Mescal, were the source rocks from which residual concentrations of iron were derived.

In many areas the dolomites of the Mescal, particularly those of the upper part of the section, are grayish red and obviously include an appreciable content of iron. The red dolomites ordinarily contain considerable volumes of secondary chert. A few conspicuous examples that do not include such chert are vuggy or even of punk-like texture, and obviously have been sites of much leaching. Spatially, thick sections of reddish dolomites are in the vicinity of prominent solution features, such as sinkholes. The brownish dolomites, which include relatively little iron, are much more abundant than those that are red. In view of this and the geometric relations of the two types of dolomite to solution features, probably the red dolomites are hematitic as a consequence of the process that resulted in iron concentration rather than being a source of iron that was concentrated.

The basalt flows are the more likely source of iron. Even the least altered basalts are abundantly dusted through with hematite. In most of the basalt the texture is largely obscured, even as viewed in thin sections, owing to the abundance of hematite. Specularite is one of the most common fillings for vesicules and other voids in the basalts. Where remnants of the flows exhibit pre-Troy soils, these decomposition products are almost indistinguishable from the reddish clay- or silt-size material that makes up the matrix of the cherty collapse breccias or the basal parts of iron-ore zones in many areas.

By analogy with iron-sparse counterparts, the iron-rich zones of chert breccia, in which the fragmented layers of chert reflect bedding of the original dolomite sequence, are solution breccias in a matrix probably composed of lateritic residues of the overlying basalts. These residues filtered downward to fill the spaces left by the solution of dolomite and the partial collapse of the remnant chert layers. Some sinkhole breccias have a similar matrix. The thicker layers of hematite and the interlensed siltstone, sandstone, and layers with fragmental chert in some of the larger iron deposits are rather discrete units, and may in part be bedded materials. Therefore in places the upper parts of the residual accumulations may have been winnowed and even transported short distances by wave or current action to form some of the layered deposits of hematite. More probably, however, additional studies will show that the layering is more apparent than real over sizable areas, and such a mechanism need not be called on to explain the layering in local areas. In any event, silica was abundantly available as a cementing medium during accumulation of the iron, and the deposits were lithified before the encroachment of the sea in which the conglomeratic sandstones of the basal Troy were laid down.

Lithology of the metamorphosed Mescal

Diabase sills ordinarily inflated the lower member of the Mescal along one to five horizons, and diabase also commonly was intruded along the contact between the algal and argillite members; as a consequence the areas in which the Mescal can be found in a totally unmetamorphosed state are few and small. Areas in which the entire carbonate section or a large part of it were completely silicified are also relatively few and restricted. It seems well in this discussion of stratigraphy of the Mescal, therefore, to describe -- at least in general terms -- the metamorphic limestones that comprise the Mescal of most areas. Brief reference has already been made to some metamorphic aspects of the upper (argillite) member, and in this section no additional note of that member is made. The highly siliceous and hematitic modifications of the lower two members were rarely intruded by diabase, and metamorphic changes were minor. So only the metamorphic modifications of the cherty dolomites are of concern here.

The lower and middle members of metamorphosed sections are mostly of calcitic limestone; most of the chert was altered to silicate minerals. In this lithology and in details of color, splitting, and topographic expression the silicated limestone strata of the metamorphosed sections are in striking contrast to the steep slope-forming or cliff-forming, yellowish-brown to dusky red, cherty dolomite strata of the unaltered sections. As viewed from a distance the metamorphosed lower unit of the algal member is gray and remains a cliff-former; but strata of the upper unit of the middle member and most of the lower member characteristically present a chalky white appearance, seemingly are mostly thin-bedded or even shaly, and are mostly slope-formers.

Mineralogic changes in the carbonate sections were brought about largely through recombination, induced by heat from the diabase intrusions, of the constituents in the cherty dolomites. The silica and small amounts of alumina in the strata combined with the magnesia of the dolomite to form very fine-grained mixtures of silicates, in aggregates that pseudomorph the original concentrations of chert and other impurities. The carbonate relict after this process of dedolomitization is the calcite that comprises the bulk of the reconstituted strata. With few exceptions these calcitic limestones are very fine-grained and of sugary texture. The principal silicates through most of the section are diopside, tremolite, talc in small amounts, and serpentine. The first three silicates, with calcite, occur in mixed aggregates that pseudomorph the original concentrations of chert. Dense serpentine, which replaces these minerals and chert and is by far the most abundant of the silicates, generally also is pseudomorphic; but in part this mineral also occurs along fractures and faults. Thus, the silicates have the same stratigraphic distribution and are roughly of the same volume as chert in the pre-metamorphic dolomites. These silicates are largely light-colored minerals, which are whitish-weathering. They do not etch out prominently on the outcrops, as does chert; therefore, their abundance in the metamorphic limestones is generally underestimated. Where secondary chert existed in zones 1 1/2 feet or more in thickness, the zones of silicate minerals may include relicts of unreplaced chert. Otherwise chert is largely missing in the metamorphic limestones. Where the dolomites included an abundance of argillic or silty material, the strata were metamorphosed to a dark greenish-gray soft rock, which apparently is a very fine-grained aggregation of serpentine, chlorite(?) and calcite. Other metamorphic minerals occur in the metamorphic limestones, but in amounts too small to modify the overall lithology of interest here. Further description of the metamorphic minerals is deferred to the section entitled "Metamorphism associated with diabase."

The metamorphic limestones of the lower two-thirds of the lower member--those equivalent to the basal breccia and the overlying cherty dolomite beds up to and including the strata that form the major cliff in unmetamorphosed sections--are nonresistant rocks that erode to moderate or gentle slopes. The basal sandstone of the [illegible] appears little different in outcrop, except that calcite rather than dolomite occurs in the matrix and limonite stains on the weathered outcrops are generally more abundant. The argillic, sandy matrix of the overlying dolomite breccia ordinarily is a greenish-gray rock that is massive and the included dolomite slabs may exhibit prominent shaly partings. The dolomite of the blocks in the breccia were metamorphosed to calcite and the borders of the blocks obscured by gradation into a metamorphic product like the matrix. The dark chert bands that occur in the interior of blocks more than a foot thick commonly exhibit surprisingly little reconstitution to silicates. The strata that overlie this breccia and constitute the rest of the lower two-thirds of the member were converted to limestones that are parted at intervals ranging from paper thin to 30 inches; most partings are at intervals of 2 to 12 inches. Chert was ordinarily completely reconstituted, but the silicate minerals in this part of the section will not be noticed except on close inspection. The dolomites that included the least chert or included chert only in thin wispy bands are those that metamorphosed to shaly limestones. The massive-bedded silty dolomite unit (unit 6 of table 3) that underlies the main cliff-former midway in the member, for example, is everywhere a shaly limestone in metamorphosed sections. Those beds that included the most chert in the form of dense bands are the limestones with the fewest partings. The thin partings strongly suggest bedding, but most are bedding cleavages imposed during metamorphism. The thicker "pseudo-beds" are commonly very friable where moderately weathered, and are separated from similar "beds" by a fraction of an inch to several inches of shaly limestone. The layered rocks of this part of the section are subject to slope creep, and only locally are outcrops of a section adequately in place so that all details can be appreciated. A few weak ledges of the more resistant limestones usually crop out on the lower portions of slopes. Otherwise this part of the member is generally covered by debris from higher in the formation.

The upper 50 to 70 feet of the lower member 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 and of the stratigraphic sequence of beds are mimicked more faithfully in these limestones than in strata lower in the member. Some concentrations of secondary chert between some of the thicker dolomite beds were metamorphosed to dense aggregates of silicates, which literally "weld" the separate beds together. As a consequence some silicated limestone "beds" of this part of the section appear more massive than their dolomite counterparts. Ordinarily, however, some silty material was concentrated along the solution planes between beds, and on metamorphism the dolomite adjacent to such material was rendered shaly. The shaly-parting limestone that separates the thicker crystalline limestones becomes chalky and etches out on weathering. The upper part of the lower member therefore tends to outcrop as rounded ledges on a steep slope. If the basal breccia of the Mescal was particularly thin, as in exposures along the Salt River Canyon east of Canyon Creek, the lower part of the section exhibits similar units of thicker-parted limestone. Otherwise relatively thick-parting limestones are not seen in metamorphosed sections of the lower member.

The metamorphic limestones of the lower unit of the middle member reflect faithfully original details of bedding and stromatolite structures, and the unit remains a massive cliff-former. The upper unit metamorphosed to limestones much like those in the upper 50 feet of the lower member. Chert is rarely observed except in the top few inches to few feet of the metamorphosed member, where partially serpentinized relicts are abundant.

For the carbonate members of the Mescal not silicified to a great degree two extremes of lithology have been described: a cherty dolomite type and a low chert calcite limestone type. Gradations exist, of course, between the two types. On the fringe of a metamorphic belt pinkish-gray or brownish dolomitic limestones can be observed grading into whitish calcitic limestones, and cherty limestones are locally intercalated with silicate-bearing limestones. Such transition zones are narrow and rarely seen. Most descriptions of the Mescal that typified the formation as of dolomitic limestones and limestones possibly were based on observations of partially metamorphosed sections. More likely they were descriptions of silicated limestones in which the whitish-weathering silicates were not recognized. Observations of metamorphosed limestones also influenced descriptions that characterize the Mescal as thin-bedded. Most sections are entirely of one lithologic type or the other: that is, entirely of thin- to thick-bedded, brownish to reddish cherty dolomite, or entirely of silicate-bearing, whitish-weathering limestone that is largely thin-parted or even shaly.

Where metamorphosed and unmetamorphosed examples can be compared in adjacent areas, the thicknesses of the carbonate members seem to decrease toward the area of metamorphism. All stratigraphic features--such as degree of solution and chertification or of erosion at the top of the sections--being comparable, no metamorphosed section has been seen that is as thick as the equivalent unmetamorphosed sections. The lower member of cherty dolomite shown on table 3, 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. This and other comparable examples suggest that metamorphosed sections are 15 to 20 percent thinner than equivalent cherty dolomite sections. Other processes also contributed to lateral thinning, and present information is too incomplete to prove that dedolomitization during metamorphism caused such shrinkages in stratigraphic thicknesses. Evidence of loss of volume on dedolomitization has recently been reaffirmed (Cooper, 1957, p. 582-588) for a comparable instance of metamorphism, and the possibility of such shrinkage certainly should be entertained for the Mascal occurrences.

Thickness

Owing to erosion after lithification and thinning of the dolomite members by solution and perhaps by metamorphism, the Mescal exhibits considerable variations in thickness. Furthermore, because parts of the Mescal were abundantly displaced by intrusions of diabase, estimates of the aggregate thickness of the three members are difficult to make in many areas. If only those sections affected least by erosion and solution are considered, apparently the carbonate members were once fairly uniform in thickness throughout the region.

Limestone or dolomite sections of the lower member in the Sierra Ancha and east to Canyon Creek range from 200 to 270 feet in thickness. Southeast, in the Salt River Canyon and south as far as Chrysotile the lower member ranges from 130 to slightly more than 200 feet in thickness. Interestingly, in this area the carbonate strata are thoroughly reconstituted everywhere. Further south little silicified or eroded sections exceed 100 feet and typically are 120 to 230 feet thick. Everywhere the thickest sections are those that exhibit little or no metamorphic effects.

Generally the thickness of the middle member is dependent on the extent to which it was eroded, either prior to deposition of the upper member or later, before deposition of still younger formations. Along the north edge of the McFadden Peak quadrangle the upper flat-bedded unit of the member is mostly missing, so that the thickness at that latitude is 30 to 60 feet; farther south in the quadrangle thicknesses of as much as 105 feet have been measured. Along the canyon of Salt River east of Canyon Creek, most sections probably exceed 60 feet; thicknesses of 85 to 100 feet are common; and one section of 130 feet was measured near the Regal mine, which is on the south rim of the canyon and 5 1/2 miles west of U. S. Highway 60. Farther south the upper unit is thin or missing in many areas, and the member is ordinarily 50 to 70 feet thick. The stromatolite unit is everywhere persistent, and it is remarkably consistent in thickness through broad areas. Near Superior it is 30 feet thick, and in its southernmost observed exposures 75 miles to the southwest in the Vekol Mountains it is 35 feet thick. Farther north the biostrome unit is generally 50 to 60 feet thick, and in places is as much as 80 feet.

Thicknesses of the upper argillite member depend on two variables: the amount of slumping, due to solution of the underlying dolomite strata, and the amount of erosion prior to deposition of the Troy quartzite. In some areas—ordinarily of one square mile or less—slumping of the dolomite members caused basins in which basal beds of the argillite member accumulated in greater thicknesses than in adjacent areas. Local increases in thickness of not less than 30 feet are attributable to this cause. Throughout large areas in which a basalt capping is remnant on the upper member, thicknesses are fairly uniform except for those places in which underlying strata subsided. A basalt cover persists through most of the McFadden Peak quadrangle and the upper member is typically 45 to 60 feet thick in the quadrangle; locally, however, it is at least 100 feet thick. As already noted the upper member is missing outside the area outlined on figure 3, and locally missing within that area, owing to pre-Troy erosion. Whether the member ever extended far beyond the outlined outcrop limits is not known.

In a few areas, as along the south rim of the Natanes Plateau south of Sawmill, in a small area 2 miles north of Globe, and also in an area possibly of considerable extent at the southeast end of the Mescal Mountains, the Mescal formation was eroded entirely and the Troy quartzite rests directly on the Dripping Spring quartzite. In some areas it was also eroded prior to deposition of either the Cambrian or the Devonian formations. In areas of abundant silicification the formation is commonly less than 100 feet thick. In the McFadden Peak quadrangle, where the lower members are largely of carbonate and the upper member generally comprises a part of the total thickness, the formation ranges from 350 to 420 feet in thickness. Along the canyon of the Salt River in the Blue House Mountain quadrangle, where the upper member is thinly remnant, from correlations of partial sections the formation is estimated to range from 250 to 350 feet in thickness. From the Natanes Plateau south at least to the Mescal Mountains the combined two lower members are generally 250 to 350 feet thick. Farther south little eroded occurrences are probably of the same order of thicknesses. Along the east face of the Vekol Mountains, for example, the formation is about 300 feet thick. In several southern areas, however, the Mescal is missing by pre-Troy or pre-Bolsa erosion.

Stratigraphic nomenclature

Obviously the lower and middle members of the Mescal limestone are distinctive units that can be grouped together as a formation. But in view of the erosional unconformity between the upper and middle members and because of the distinct lithologic differences between these two members, many geologists would be disposed toward designation of the upper member as a separate formation. Such formational designation would be well within the principles of stratigraphic nomenclature accepted by American geologists (Ashley, and others, 1933, p. 427-439). Indeed, in the only previous published note on the argillite unit, formational status has been proposed by N.E.A. Hinds (1935, p. 32), who says "At and near Roosevelt Dam, there are present locally above the vesicular basalt a minimum 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" 1/

1/ Hinds here used the term "member" in a formational sense. The section described by him is incomplete, and due to deep weathering modifications it is not outwardly typical of sections seen elsewhere. Nor was it possible at $\boxed{\text{the}}$ Roosevelt Dam to appreciate that basalt flows overlie the argillite in typical complete exposures.

Genetically there is some slight, but not compelling, reason for including the argillite unit in the Mescal formation. The few lenticular limestones that occur within the unit are very like some of the limestone beds at the tops of the lower and middle members. This suggests recurrences of the environment in which the carbonate sediments were deposited.

Although there are adequate bases for defining the argillite unit as a formation, it seems preferable to designate the unit as the uppermost member of the Mescal limestone--as a practical expediency. The primary considerations in this reasoning are: the unit is characteristically thin; it crops mainly in cliffs, as a cartographic unit too narrow to depict except on large-scale maps; and almost certainly it will not be recognized and separately defined except within an area of about 800 square miles in northern Gila County. East of the area of outcrop outlined on figure 3 the Apache group is deeply buried by younger formations. West of the longitude of Roosevelt Dam, where additional outcrops of the argillite member may yet be recognized, few remnants of the Apache group exist. North and south of the area of outcrop of the upper member shown on figure 3, observations have already indicated that additional outcrops are unlikely. Thus the argillite unit is best considered a member--only locally remnant--of the Mescal limestone.

Then, because the Mescal includes an argillite unit, in places is a chert formation, and originally was mainly of dolomite strata, some might prefer to apply a different lithologic term in redefining the formation. Limestone though of metamorphic rather than sedimentary origin as previously conceived, makes up the bulk of all exposures. Other lithologic facies are of little significance volume-wise. Therefore the designation "Mescal limestone" may well be retained.

Basalt flows

Throughout much of the region of Apache outcrops the Troy quartzite is separated from the upper or middle members of the Mescal by one or more flows of basalt. Generally this basalt formation ranges from 40 to 90 feet in thickness. Within areas in which the flows are ordinarily a part of the section basalt may locally be missing; and in a few extensive areas, as north of the 34th Parallel and along the Salt River Canyon east of Canyon Creek, the basalt is largely absent owing to pre-Troy erosion. But in a few areas, as along the mountain front 1 to 3 miles south of Superior, the basalt exceeds 200 feet in thickness; in one place 10 miles east of Globe a basalt sequence 375 feet thick was measured. Boundaries between individual flows are not everywhere readily distinguished, but where they have been positively recognized individual flows are 50 to 125 feet thick. In the usual section only one flow is seen, but where the basalt formation is thick four flows have been distinguished; with less certainty a fifth flow may be present in a few sections.

In the Sierra Ancha and east to Canyon Creek the basalt rests on the upper member of the Mescal limestone, is present in thicknesses ranging from 20 to 100 feet, and generally is not less than 30 feet thick. As already noted the basalt is generally missing in the Salt River Canyon, but 5 miles south in the vicinity of Chrysotile outcrops are again prominent. About one-half mile south of Chrysotile is a section, slightly more than 100 feet thick, that is unusual in that two flows are separated by 10 to 12 feet of silty limestone. Only in one other area, between Cherry and Canyon Creeks, has a sedimentary unit been noted between flows; there a few inches of laminated argillite locally separates two flows. Southeast from Chrysotile along the Natanes Plateau the basalt is generally missing, but in an area 16 to 20 miles to the southeast one flow is locally remnant in thicknesses of at least 110 feet. Southward in the Apache Mountains the basalt is missing, but along the low chain of hills that extends 12 miles southeast from the Apache Mountains the basalt is commonly more than 250 feet thick. Farther south, where complete sections of the middle member of the Mescal exist, the basalt ordinarily is also remnant and exceeds 50 feet in thickness in most sections that have come to my attention.

The basalts are generally grayish-red to blackish-red or brownish-black on both fresh and weathered surfaces. If the color is modified by weathering, pale brown to dark yellowish-brown hues prevail. Hematite is abundantly disseminated through the rock in every locality, so that a blackish-red color dominates most exposures. An intersertal or intergranular texture, defined by laths of plagioclase 0.1 to 0.3 millimeters in length, is obvious in most handspecimens. The rock has been altered to such a degree that microscopic identification of the original plagioclase has not been possible. Under the microscope the original minerals prove to be altered aggregates of albite, calcite, serpentine, quartz, and chlorite; in some specimens magnetite, epidote, and an amphibole (hornblende?) are common. Needle-like crystals of apatite are abundant in some of the basalt.

Particularly striking in many outcrops are sparse to abundant tabular phenocrysts of plagioclase. These crystals, in a groundmass of typical texture and grain size, are plates that range mostly from 0.5 to 2 millimeters in thickness and from 5 to 20 millimeters in diameter. Crystals as much as 40 millimeters (about 1 1/2 inches) across have been observed, and rarely the phenocrysts comprise as much as 40 percent of the rock. The basalt of some outcrops is not porphyritic, that of others is only partly porphyritic. In particularly hematitic examples the phenocrysts are commonly moderate to dusky red and not conspicuous. The vesicular portions of some flows include the greatest abundance of phenocrysts.

The basalt is characteristically amygdaloidal. Amygdules are especially abundant in the tops and bottoms of flows. Partly filled elongate vertical vesicules (pipe vesicules) are observed locally in the bottoms of flows. Flow breccia may be conspicuous at the tops and bottoms of flows, and is observed at a few places within flows. Rarely the basal few feet of a flow is a breccia that includes fragments derived from the upper (argillite) member of the Mescal. Variolitic basalt, a few inches to a few feet thick, rarely comprises the basal part of the lowest flow. Vesicules and voids in the breccias are filled by calcite, quartz, specular hematite, and an unidentified grayish-green to grayish-olive, soft mineral. In many outcrops the quartz amygdules have a thin rind of an unidentified pale green to grayish-green material. In a few localities blue-green oxide copper minerals coat weathered joint and fracture faces.

All structural features of the basalts indicate that they formed as subaerial flows; no features suggest submarine flows. In some areas the flow breccias that ordinarily top a flow were eroded away, and a planar surface of erosion was formed before extrusion of the next flow. Such planation and the two isolated examples of sedimentary strata between flows suggest inundation of the region by waters between outpourings of lava. But these waters need not have been marine waters.

Basalt that separates the upper and middle members of the Mescal has been observed only in three localities: (1) along the southern front of the Sierra Ancha, through an area of not less than nine square miles; (2) south and east of the junction of Ash Creek and Cherry Creek (north-central McFadden Peak quadrangle), over an area of about one square mile; and (3) in the vicinity of Roosevelt Dam. This basalt differs in no aspect from that described above, except that the porphyritic variety has not been observed in these occurrences. The lower basalt of the southern Sierra Ancha and of the Roosevelt Dam localities is as much as 50 feet thick, and that near Ash Creek is as much as 110 feet thick.

The Apache basalt described in the geologic literature is said to separate the limestone portion of the Mescal from the Troy; south of the latitude of Chrysotile most of the basalt is in this relation to the two formations. Thus the basalt at this position in the Apache group in the vicinity of Globe, Superior, and Ray, in the Mescal and Dripping Spring Mountains, and even as far southwest as the Vekol Mountains, could correlate with the lower basalt of the Sierra Ancha. At present, however, this correlation cannot be made with any assurance. Obviously the upper basalt overlies the argillite member of the Mescal along its southern and thinner extremities, and could also be the basalt unit that overlies the middle member farther south.

Typically the upper basalt crops out as prominent blackish ledges or cliffs. For this reason, and owing to their position just below the Troy quartzite, these outcrops are readily recognized as the basalt formation. In some areas, and especially where diabase was intruded into the basalt or along a stratigraphic horizon near the flows, the basalt is readily disintegrated on weathering. In these instances, or if talus cover is almost complete, basalt may be difficult to distinguish from diabase.

The following field criteria for differentiating the Apache basalt from the diabase have proved useful. (1) The tabular plagioclase phenocrysts of the basalt have no counterparts in the diabase. Some coarse facies of the diabase include plagioclase crystals even larger than the phenocrysts of the basalt, but the plagioclase of the diabase is lath-shaped and intergrown with large grains of other minerals; in contrast, the phenocrysts in the basalt are invariably tabular crystals in a fine-grained groundmass. (2) The amygdules of greenish minerals in the basalt might be confused with certain aggregates of greenish minerals in some diabases, but the quartz, quartz-hematite, or hematite amygdules are characteristic of the basalt and very resistant to weathering. (3) In many places the above criteria are not of use because the basalt is not appreciably porphyritic or amygdaloidal; in these instances weathering products can be useful. The basalt decomposes to a fine, powdery, dark reddish-brown soil; locally the diabase does form a similar soil, but the typical soil derived from diabase is granular and fragments of plagioclase laths are abundant in a yellowish-brown clayey matrix. The typical diabase soil can usually be found in a brief search. Because certain of the amygdules of the basalt are resistant, they tend to concentrate in basaltic soils, even where not obvious in adjacent outcrops, and these relicts can similarly typify the basalt soils. (4) The finer-grained diabase--that most likely to be confused with basalt--commonly breaks down into a pebbly rubble (fig. 13). The abundant ellipsoidal pebbles of this soil represent the host crystals of pyroxene in the ophitic-textured diabase. The small plagioclase laths enclosed in these pebbles tend to weather chalky white. In contrast, a soil derived from basalt includes only scattered pebbles and fragments of irregular form rather than rounded outline, and the feldspar laths in these pebbles, if distinguishable, are almost invariably weathered reddish-orange to yellowish-brown. The basalt pebbles when broken do not, moreover, show the flashing cleavage faces of pyroxene characteristic of the diabase pebbles. The writer is aware of a number of misidentifications of the basalt as diabase; some have been perpetuated in the geologic literature. Distinction between the rock types will seldom be in doubt if the above field tests are applied.

Troy quartzite

For reasons that will become obvious in the discussion of stratigraphic relations of basal Paleozoic formations to the Troy quartzite, the Troy is herein redefined as Precambrian in age, rather than Cambrian. Only in a few small areas have Cambrian strata, herein restricted from the formation, been mapped with the Troy quartzite as defined and mapped originally by Ransome (1915, p. 384-385; 1919, p. 44-45; 1923), or as described in greater detail from his later observations (Ransome, 1916, p. 139-141,154). Therefore, this redefinition does not modify or only to a slight degree modifies the cartographic depiction of the Troy on most published geologic maps of areas that include the formation. The few areas that should be modified are mostly south of the Pinal Mountains, within the belt where the Troy is overlain by Cambrian strata (see fig. 3). In many references or brief stratigraphic descriptions, applicable to the part of the regions south of the latitude of Globe, the Bolsa quartzite of Cambrian age has been correlated with the Troy or--sometimes along with still younger Cambrian strata--has been described as Troy quartzite. These descriptions, in future reference should be reconsidered in light of the redefinition. In many details, the Troy formation is quite different from grossly similar Paleozoic strata with which it has been correlated. So that the differences can be enumerated at a later place (table 5) the Troy is described in some detail in the following pages.

The Apache group was differentially warped on a broad scale and eroded to different depths in various parts of the region before deposition of the Troy quartzite. In general the pre-Troy unconformity was planar, but in at least a few places local erosional relief on this surface exceeded 100 feet. Additionally in the northwest part of Gila County the eastern part of a structural basin a few hundred feet lower than the general surface can be reasonably interpreted as existent at the start of Troy deposition. Although variations in elevation of the surface of deposition affected the thicknesses of Troy accumulation as shown on figure 11, post-depositional

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Figure 11. Reconstruction of north-south geologic section after the Troy quartzite was deposited. Note particularly pre-Apache surface, and pre-Troy surface. Section extends from present Mogollon Rim south to Dragoon Mountains.

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events were much more effective in determining the thicknesses of the formation now available for view (compare figs. 11 and 16). The Troy quartzite was differentially uplifted, largely owing to inflation by diabase sills, and variously eroded prior to deposition of Cambrian strata; and in the northern part of the region the Troy was again exposed to erosion before Devonian strata were deposited. As a consequence the formation is missing below the basal Paleozoic strata in some areas and is at least 1,200 feet thick in others. Thicknesses are of a considerable range in some relatively small areas; those of specific parts of the region are considered in the course of description.

Details of stratigraphy and lithology of the Troy quartzite are somewhat different in different parts of the region of outcrop. Lateral variations are well enough understood to be certain that all are variant aspects of one formation. South of Globe less information on the lithologic changes has been accumulated than in the area to the north. In the Sierra Ancha the Troy can be readily divided into three members: the lowest member is a reddish arkose, the middle member a white sandstone, and the uppermost a grayish quartzite.

The section described on table 4 is one of the most complete yet

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Table 4.--Section of Troy quartzite typical of the Sierra Ancha.

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found in which the usual features of the three members are readily seen.[1]

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[1] One of two measured sections of the Troy described by Ransome (1916, p. 154) was measured on Baker Mountain, 2½ miles southeast of that described here. The Center Mountain section of table 4 and that of Baker Mountain are very similar, but the Baker Mountain section is not as well exposed, and the upper 200 feet or more of the section exposed on Center Mountain is missing from Ransome's locality owing to pre-Devonian erosion. Actually, with some misgivings apparently, Ransome (1916, p. 151) recorded a third measured section at Roosevelt Dam. That section is only 160 feet thick and is probably partly or wholly of Devonian sandstone.

As indicated in the table, each member is of distinctive lithology and each includes sedimentary structures that are characteristic. The differences between members are most conspicuous in the part of the region north of the Salt River; therefore the stratigraphy of that area is considered first in describing each of the three members.

Arkose member

The lowest member of the Troy is a fine- to medium-grained, firmly cemented arkose. Most bedding units are thick or very thick, fine-grained and of remarkably uniform sorting. Most of the arkose is pale red to grayish-red on both fresh and weathered surfaces, but light to pale brown colors, as reported in table 4, are common also. In handspecimen this rock is somewhat like the arkoses of the lower member of the Dripping Spring formation. But the latter arkose is ordinarily of an orange-pink or reddish-orange hue, is somewhat less uniform in texture, and is so firmly cemented that it breaks with a quartzitic fracture--across rather than around individual grains. In outcrop the two arkoses are strikingly different. The lower member of the Troy in areas of low relief erodes to gentle slopes, and in areas of high relief to steep slopes studded with rounded to angular ledges or even to vertical cliffs. Regardless of the form of outcrop, large- or very large-scale, low- to high-angle cross-stratification is very conspicuous through much of the member, owing to erosional etching or prominent parting along the inclined planes. Sets of cross-strata range from a few tens to several hundreds of feet in length between horizons of truncation. In contrast, the cross-stratification of the lower member of the Dripping Spring is mostly of small to medium scale; individual sets are generally less than 15 feet in length; and owing to firm cementation, weathering rarely etches out the cross-stratification for easy observation.

Outcrops of the arkose member of the Troy now exist through an area of only a few hundred square miles at most, and abundant outcrops--mostly within the McFadden Peak quadrangle--exist through an area of less than 150 square miles. In the highest parts of the Sierra Ancha, within 2 to 4 miles of the section shown on table 4, the member is 400 to 480 feet thick. To the north, east, and south it thins rapidly by lapping against successively higher parts of the pre-Troy surface. Within a radius of 12 to 15 miles in these directions from Aztec Peak the middle member becomes the basal member of the Troy. The arkose member can be traced into areas in which it is less than 100 feet thick. But, owing to Tertiary or Recent erosion, the Troy is largely missing along the line of complete lapout. Therefore the limit of the member as shown on figure 3 is an extrapolated line: to the east of the line isolated remnants of the middle member rest directly on Apache strata, to the west outcrops of the lower member are known. North of the Salt River and west of the 111° Meridian erosional remnants of the Troy ~~formation~~ are also few, and are widely scattered. Unfortunately none of these exposures has been examined. Judging from the pattern of thickening, the lower member might be as thick or even thicker than the thickest sections in the high parts of the Sierra Ancha. West of Tonto Creek, of course, all younger Precambrian formations are missing. So generalizations concerning the paleogeographic significance of the arkose member must be derived largely from outcrops within the McFadden Peak quadrangle.

Where thickest, the arkose member generally rests on the upper basalt of the Apache group, and the basal part of the member is not conglomeratic or is only sparsely pebbly. Below this contact, in most places, the upper part of the basalt is firm and no fossil soil remains between the two formations. In some places, however, a residual fossil soil, a few inches to a few feet thick, mantles the basalt. Some of this basaltic soil with partially decomposed fragments of the basalt was incorporated in the basal few feet of the arkose, causing the beds of this interval to be dark in color. Along the canyon of Cherry Creek in the northern part of the McFadden Peak quadrangle, in an area probably exceeding 20 square miles, the basalt is thin or missing and the Troy rests mainly on the upper member of the Mescal. Along much of this and like contacts and wherever the arkose member is considerably thinned a conglomerate or conglomeratic sandstone, 10 to 20 feet thick, exists immediately above the unconformity or is within the arkose 10 to 20 feet above the contact. Included pebbles of the underlying formations of the Apache group--especially fragments derived from the basalt, the argillite member of the Mescal, and the ferruginuous silicified parts of the carbonate members of the Mescal--distinguish this conglomerate from the Barnes and Scanlan conglomerates, which may exhibit a similar arkosic matrix.

Most sections of the lower member exhibit very large-scale cross-stratification practically throughout. So that the description cited as typical in table 4 will not seem at variance with this statement, it requires explanation. Where the arkose member is thickest, its lower parts are comprised of tabular beds that generally range from 3 to 10 feet in thickness (see units 1 and 2, table 4). These beds do not display the extremely large-scale crossbedding. But comparable beds do not exist, except in the basal few feet, where the member thinned appreciably by lapout. So an arkose bedded like that of units 3 and 4 of the Center Mountain section comprises very large parts of most sections. The very large-scale cross-stratification, in all outcrops examined, dips 7° to 22° eastward. The prevalent angles of these cross-strata are 18° to 20°.

In a few places exposures along the canyon of Cherry Creek, particularly in the north half of the quadrangle, the lower part of the arkose member is notably different. There channels, generally broad and shallow but rarely narrow and steep-walled, were eroded into and in places several tens of feet below the basalt prior to deposition of the Troy. Bedding structures are not well developed in the arkose that fills these channels, and the arkose of most channels is poorly sorted and sparsely conglomeratic. The granules, pebbles, and cobbles of the conglomeratic sandstone are mostly of older Precambrian quartzite, like those in the Scanlan and Barnes beds. But pebbles of volcanic material--mostly rhyolite--and sparse pebbles derived from the underlying formations of the Apache group are also conspicuous. The channel-filling conglomeratic arkose is a few feet to at least 100 feet thick.

Chediski sandstone member

North of the Salt River the middle member of the Troy is mostly a pinkish-gray, loosely to moderately well cemented, very poorly sorted, porous sandstone, which erodes to rounded slopes surmounted by rounded knobs and hoodoos bizarrely sculptured by erosion. Outcrops viewed from a distance are mostly whitish, in striking contrast to the reddish or brownish hues of the lower member. This coloration and the massive outcrops studded with distinctive erosional forms, also contrast with the somewhat darker colored cliff-forming quartzites of the upper member.

In 1929 E. F. Burchard (1931, p. 54, 56-57) recognized the white sandstone as a very distinctive stratigraphic unit, and termed it the "Chediski white sandstone member" of the Troy quartzite. The type locality was designated as Chediski Mountain, a prominent mesa on the west side of Canyon Creek 23 miles north of the Salt River. The member forms an almost horizontal, conspicuous and typical outcrop along the east face of this mesa.[1]

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[1] The term "chediski" as used locally by the Apaches means "long white rock." Therefore this name is particularly appropriate for the member.

Burchard described only the outcrop characteristics of the member, and south of the Salt River the member is not as distinctively whitish nor as singular in its erosional forms. Therefore this part of the Troy has been distinguished as a separate member only in the vicinity of the iron deposits near the headwaters of Canyon Creek. Less obvious features of the member set it apart everywhere, however, and it could well be designated the <u>Chediski member</u> or <u>Chediski sandstone member</u> throughout the region of Troy outcrops.

Between the Salt River and the latitude of Young the middle member is generally about 250 feet thick. Farther north the member probably thins. Along the northern reaches of Canyon Creek thicknesses variously reported range from 20 to 270 feet; apparently most sections are less than 150 feet. Whether the thinner intervals everywhere represent sections overlain by the upper member of the Troy or some sections are those truncated prior to deposition of Devonian strata is not known to the writer.. Along the Mogollon Rim sections that are only 75 to 100 feet thick are capped by the upper member of the Troy.

North of the Salt River the sandstone is mostly of subrounded to well-rounded, but pitted and frosted clear quartz grains, which range from fine to very coarse in size. Thin lenses of white quartz granules are common. Sand grains of medium- to coarse-grain sizes dominate most bedding units. The matrix of the sandstone is of sericite and a whitish clay; in much of the member sericite is very abundant. Clay aggregates that have the outlines of cleavage fragments of feldspar are occasionally noted. And rarely cleavage fragments of orange-pink feldspar are seen. Throughout the region feldspar can generally be found if the coarser-grained beds are searched carefully. In such beds grains of granule size are sometimes seen, but generally feldspar grains and angular clay aggregates apparently relict after feldspar are only a fraction of the size of the quartz grains. Under the microscope potassium feldspar can be seen in some sandstones that megascopically appear devoid of the mineral. Chemically the sericitic sandstone is probably equivalent to a feldspathic sandstone.

The contact between the middle member and the lower member is sharp, and where seen in cliff exposures obviously truncates the cross-strata of the lower member. This surface commonly shows local irregularities ranging from a few inches to as much as two feet. In many places outcrops of the middle and upper members are so subdued in profile that these relations are not manifest.

Where it overlies the lower member, the basal beds of the middle member may be pebbly, as illustrated by unit 5 of table 4, or may not be pebbly. Fragments from the lower member have not been identified in the middle member. Where the Chediski sandstone is the basal member of the formation the basal 2 to 40 feet of the member is a pebble to cobble conglomerate or a very conglomeratic sandstone. Ordinarily the conglomerate is less than 20 feet thick. Throughout most of the region the gravels of this conglomerate are very like those of the Barnes conglomerate, except that sparse pebbles of basalt and silicified dolomite from the Mescal can ordinarily be found. Northward the rounded pebbles and angular fragments from these Apache units and from the argillite member of the Mescal exist more abundantly in the conglomerate. And north of the Salt River the gravels locally are almost monolithic, being of one or the other of these units. Such almost monolithic gravels are especially conspicuous where the conglomerate overlies a terrane of thoroughly silicified, highly hematitic dolomite rubble, and the gravels are largely of such rubble in a matrix of hematitic sandstone.

In the Sierra Ancha and north and east of that range the sandstone of the 25- to 60-foot interval immediately overlying the basal conglomerate is sparsely pebbly and includes lenses of granule, pebble, or occasionally even cobble conglomerate. Unit 5 of table 4 is fairly typical of many such intervals. This unit is massive and generally crops as a gentle slope only partly exposed. Bedding features are vague and defined mostly by discontinuous thin lenses of very coarse sandstone and conglomerate, which suggest irregular but grossly horizontal stratification. Discontinuous lenticular beds of cross-stratified sandstone are seen rarely. Pebbles in this unit are in great part of fine-grained vitreous quartzite like the quartzites in the older Precambrian terrane. Quartz pebbles are also abundant and in some areas predominant; a part of these differ from quartz pebbles higher in the section in being reddish or brownish rather than white. Pebbles of other lithologies, noted in the basal conglomerate, are generally rare in this interval. Some of the quartzose gravels, especially those of the larger sizes, exhibit polished facets characteristic of wind-worn pebbles or ventifacts (Bryan, 1931, p. 29-36). In some sections ventifacts are fairly common; in others they are rare.

In sections exceeding 200 feet, roughly the upper one-third of the Chediski member is cross-stratified on medium to large scales. The lower two-thirds of the member, excepting the basal portion just described, locally exhibits like cross-bedding but mainly is a very massive unit in which bedding features are obscure.

Pre-consolidation slump features characterize the massive middle unit of the Chediski member. Textural variations in the sandstone cause the slump structures to be conspicuous on close view. The sandstone is largely of medium-grain, but it is very poorly sorted. Before slumping it apparently included some thin lenses and thin more or less tabular beds mostly of coarse sand. The most abundant manifestations of slumping are contorted or swirl-form mixtures of variously sized sands, as though the sands of the different layers had been gently stirred together. More prominent in topographic expression are parts of the differentially sorted sand layers that survived slumping as crudely lenticular masses. Commonly such lenses are roughly saucer- or bowl-shaped in cross-section; others are intricately contorted and thickened and thinned by folding. Where the sands slumped no more than a foot or two and abundant vestiges of original bedding persist, several saucer-shaped lenses may be alined laterally so that a crude scalloped effect is seen in cross section. The coarser-grained lenses of slumped sandstone are generally more firmly cemented than the surrounding sandstone. These lenses etch out on weathered surfaces, to accentuate the slump features.

Seemingly the slump structures in this unit formed while the sandstone was being deposited. Tabular subunits, ordinarily 10 to 50 feet thick, are defined by inconspicuous horizontal (or originally horizontal) planes. In a given outcrop these planes may show little modification by slumping. A few of the massive slump-marked subunits are separated by prominently cross-stratified beds, 1 to 2 feet thick, uniformly tabular and sharply bounded top and bottom. These planar features suggest that a subunit of sand was deposited, slumped, then was eroded at the top before the next like subunit was laid down. Furthermore, particularly where a bed of cross-stratified sandstone intervenes between massive slump sandstones, the relations suggest that once a set of beds slumped no additional soft-rock deformation occurred. Though grossly applicable, the latter supposition is partly refuted by some details. If the thin cross-stratified beds and the planes that separate tabular masses of slumped sandstone are traced laterally a few hundred feet, almost invariably both features are seen to be eradicated locally by slumping. Thus in one outcrop the slumped beds appear as fairly well defined units 10 to 20 feet thick, but in a nearby outcrop equivalent sandstones are separated by planes as much as 50 feet apart.

North of the Salt River the lower one-quarter to one-half of the slumped unit commonly includes pebbles or cobbles very sparsely distributed (see unit 6 table 4). Outcrops have been seen in which single cobbles, 6 to 8 inches in diameter, are isolated at intervals of 15 to 50 feet along the exposure, in sandstone that is otherwise not pebbly. The upper part of the unit is generally pebble-free.

Probably more characteristic of the lower part of the contorted sandstone unit is a dark color. In many areas the basal 10 to 80 feet of the Chédiski member is grayish-red to grayish-red purple. A part of the sandstone of these colors is mottled or streaked by reduction spots generally of the whitish hues of the greater part of the member. In some examples the mottled part of the sandstone is light brownish gray to light greenish grey. Commonly the coarser-grained streaks and layers, or the boundaries between sandstones of different grain size are mottled in preference to the remainder of the unit. Thus the swirl-form slump structures, even if not etched out by erosion, may be strikingly outlined by differences in color. In some areas the basal pebbly sandstone unit is largely devoid of red-purple coloration, owing to reduction of iron oxides. In many such examples the lower part of the overlying slumped unit retains the color, so that in cliff exposures a band of reddish sandstone a few tens of feet above the base of the member is conspicuous.

In areas of high relief the contorted sandstone unit is the part of the member that erodes to fluted cliffs or hoodoos. But in areas of low relief such erosional features are wide-spaced and rounded massive outcrops are typical. These rounded outcrops are smooth except for V-shaped furrows or etched-out joints, which penetrate the outcrop to a depth of only a few inches at most. The furrows generally occur in two sets of about equal spacing, and together the sets form a reticulate pattern. One outcrop may be marked by furrows 6 inches to 1 foot apart; another by furrows 2 to 3 feet apart. Such outcrops have some similarity in appearance to rock walls built of crudely squared but tightly matching blocks. In relatively few outcrops the "block" outlines are of 5 or more sides, rather than 4-sided. The reticulate patterns of furrows particularly characterize subunits in which the contorted layers are not too different from the enclosing sandstone in grain size.

Approximately the upper one-third of the Chediski member is D' sericitic sandstone, which is coarse- to very coarse-grained but otherwise very like the sandstone of the lower units. This sandstone is in thick beds, which are grossly tabular and internally are tangentially cross-stratified on medium to large scales. The dips of the cross-strata generally exceed 10° and commonly approach or slightly exceed 20°. In any one outcrop the directions of dip are various; no dominant direction of dip can be ascertained from information collected to date. In some areas the beds are mostly 3 to 10 feet thick; in others most beds range between 10 and 20 feet in thickness. Although in a given outcrop individual beds appear to be tabular, if traced laterally many prove to be subtly wedge-shaped or lenticular. The bedding structures of some outcrops are not seen except on close inspection (unit 8 of table 4 is an example), but in most outcrops the bedding features etch out in bold relief. In many areas the contact between this strongly cross-bedded unit and the underlying contorted sandstone unit is so poorly exposed that the nature of this boundary is indeterminate. In at least a few areas the boundary between the two units is planar and sharp, and probably represents an intraformational surface of erosion.

In typical sections of this upper cross-bedded unit perhaps half of the beds are sparsely pebbly throughout. Most of the pebbles are of white quartz; pinkish quartz is common, and reddish-brown jasper pebbles are rare. Pebbles of other compositions are practically nonexistent. The pebbles are well rounded and as much as 2 inches in diameter; most are of diameters less than 3/4 inch. Pebbles gradually become more abundant upward, and in the upper half of the unit are particularly concentrated in the upper few inches of individual beds. Only rarely do concentrations of pebbles exist at the bottoms of beds. In the upper 25 to 35 feet of this unit characteristically the uppermost 2 to 6 inches of each bed is very conglomeratic. And the uppermost 1 to 6 feet of the member is everywhere a conglomerate or conglomeratic sandstone. The contact between this conglomeratic bed and the upper member is sharp and planar. The positions of the pebble layers suggest that the finer constituents of the top parts of each sparsely pebbly bed were winnowed away leaving behind only the coarser fractions. The relative increase upward in numbers of pebble layers and in concentrations of pebbles in the layers indicates that such winnowing became increasingly more effective and prevalent as successive beds were deposited.

North of the Salt River the upper crossbedded unit is the only part of the member, for practical purposes, that is appreciably quartzitic. The lower part of the unit is commonly very friable, but this is not everywhere true. Upward the clay or sericitic matrix decreases in amount, in general the grain size of individual beds increases, and the sandstones are more firmly cemented by quartz. In some sections a bed or two of quartzite can be found in the upper part of the unit. Occasionally all voids of a coarse-grained lens in the contorted sandstone unit are completely filled by quartz. Where the matrix is somewhat less abundant than usual, anywhere in the member, clusters of minute needle-like quartz crystals--seen only under a microscope--may penetrate the clay-sericite matrix; these needles are not attached to the clastic quartz grains. Indeed, excepting the minor quartzitic examples noted, a singular characteristic of these sandstones is that the individual sand grains do not exhibit--in the slightest--overgrowths of quartz.

East of Canyon Creek in the vicinity of the canyon of Salt River some of the features of the middle member so conspicuous in the Sierra Ancha gradationally become less obvious, and the member becomes dominantly quartzitic. The easternmost exposures, known to the writer, of the middle member in its typical light-colored friable form are along the lower reaches of Cibecue Creek. Farther east and south onto the Natanes Plateau quartzitic sections range from pinkish-grey or light brownish-grey to greyish-red. The darker colors, which reflect an abundance of hematite in the matrix, are dominant. The fine- to coarse-grained quartzite and quartzitic sandstone beds are crudely tabular and range from 1 to 20 feet in thickness. Medium -scale tangential cross-stratification is conspicuous in some beds. Small pebbles and granules are notable through much of the member. Pebbles do exist sparsely scattered through the quartzites, but they are most abundant as lenticular concentrations, generally 3 to 6 inches thick, bordering or separating beds. Some conglomeratic lenses as much as 3 feet thick have been noted. The conglomerate lenses commonly fill channels in the underlying beds. The middle part of the member is the least pebbly. The massive-cropping quartzitic strata of this area are quite like those that Ransome (1916, p. 139-141 and pl. 27A) described and illustrated as typical of the lower and middle parts of the Troy in the Ray quadrangle.

Farther south but within the structural bounds of the Colorado Plateau, principally in exposures in the vicinity of the mining camp of Chrysotile and southeast along the southern margin of the Natanes Plateau, the member is again dominantly of sandstone. These sandstones are in a sequence very like that of the Sierra Ancha, and lithologically are grossly comparable except that most are highly hematitic and greyish-red to dusky red in color. Only west of U. S. Highway 60 do light-colored sandstones form an appreciable part of the member.

These sections of reddish sandstone and those of reddish quartzite, in the adjacent Salt River Canyon area, possibly are thinner and were deposited on a surface topographically somewhat higher than those in most areas to the northwest and south. Because parts of the sequence were displaced by diabase intrusions and therefore details of stratigraphic sequence are obscure, and because the upper part of the member was eroded away in some parts of this northeastern area prior to deposition of Devonian strata, such generalizations concerning lateral variations in thickness cannot be made with great assurance. In some isolated outcrops the Chediski member is at least 200 feet thick, but in some other areas where complete it is probably considerably thinner than 200 feet. Where seemingly thinnest the member laps across different parts of the Apache formations; in places the Troy rests on thin remnants of the algal member of the Mescal; in other places it rests on the lower member of the Mescal or even on the Dripping Spring quartzite. In this part of the region the lower part of the Chediski member is distinctly more tabular bedded than the lower part of the member in the Sierra Ancha, and cross-stratification is common. Also the interval that displays contorted bedding is thinner and less conspicuous in outcrop. The upper prominently cross-stratified unit of the member, however, is comparable--except for coloration--to that described for sections to the northwest.

Farther south, throughout the portion of the region structurally a part of the Basin and Range province, the Chediski member is largely quartzitic in certain areas of a few square miles, and is partially quartzitic through still broader areas. Quartzite sections particularly notable are in the vicinity of the abandoned mining camp of Troy--near the type locality of the formation--and in and near the northeastern part of the Globe quadrangle, where Ransome probably first studied extensive outcrops of the Troy. In both of these areas considerable parts of the quartzite section are light colored and except for feldspar and pebble contents are grossly very like the upper quartzite member. Through most of the region south of the Natanes Plateau, however, the member is mostly of sandstone.

Thicknesses of the Chediski member have not been determined in many places in the southern part of the region. Wherever, between the Apache Mountains and the Gila River, the member is overlain by the upper member and therefore remnant in full, thicknesses of 250 to 300 feet appear to be common. In a few places, thicknesses more than 300 feet may exist, suggesting that the member thickens slightly southward. However, in its southernmost exposures along the San Pedro River, in the vicinity of Holy Joe Peak and Aravaipa Creek, thicknesses of 180 and 210 feet have been measured (M. H. Krieger, written communication). Perhaps variations noted in the southern part of the region reflect relief on the pre-Troy erosion surface, rather than a significant regional variation in thicknesses of accumulation.

South of the Natanes Plateau the Chediski member is best described by comparisons with the conspicuous sandstones north of the Salt River. The sandstones of the southern facies are compositionally grossly similar to those of the northern area. Well-rounded quartz grains, ranging from fine sand size to granule size comprise the bulk of the clastic materials. The matrix is similarly dominated by sericite and clay. Orange-pink feldspar is much more abundant in the southern facies, and in the coarser-grained beds occasional granule-size grains of potassium feldspar are conspicuous. These commonly are compound grains that include quartz, and are very similar to the coarse detritus derived from the granitoid rocks that underlie the Apache group. In some localities coarse muscovite, which fills interstices between quartz grains, is very abundant in a few beds. Such muscovite is so rare in the northern facies as to be atypical. For example, in the section shown on table 4 muscovite was seen very sparsely in only a few of the coarser-grained lenses near the top of unit 6.

The southern sandstones are mostly yellowish-grey or light-brownish to medium dark grey or pale red, and although not particularly dark-colored overall they do not contrast with the adjacent stratigraphic units as do the whitish outcrops of northern Gila County. Some beds of medium light grey to medium dark grey color contrast with adjacent beds of brownish or reddish hues. The dark-grey sandstones owe their color to abundant minute aggregates of black hematite, which partly fill the voids between quartz grains. Similar aggregations of blackish hematite and the consequent grayish hues of sandstone beds have not been noted in the northern area, although the reddish sandstones of that area--especially in the basal part of the member--do owe their color to abundant hematite. In some places in the southern part of the region the basal 25 to 120 feet of the member also ranges from pale red to grayish-red or grayish-red purple. For instance, in the extensive exposures of the member in the range of hills that extends southeast from the Apache Mountains the basal 50 to 75 feet of the sandstone is commonly of reddish hues. In most areas, however, sandstones of dark reddish color are subordinate.

poorly sorted. As any given section is traversed stratigraphically individual beds are noted to be quite variable in the degree of sorting exhibited. Although poor sorting is also characteristic of sections farther north, in contrast, about the same degree of sorting characterizes stratigraphic units 30 to 100 feet thick. South of the Natanes Plateau this contrast in sorting is emphasized because granules, pebbles and even cobbles, compositionally in variety like those restricted largely to the basal conglomerate north of the Salt River, are seen throughout the member. In the northern area the upper half of the contorted sandstone unit is virtually pebble-free; in the southern area pebbles are fewest within the comparable stratigraphic interval, whether that interval is comprised largely or only in small part of slumped sandstones. Angular to rounded fragments derived from the resistant formations of the Apache group are not confined to the basal conglomerate, as in the northern area, but can be found sparsely throughout the lower half of the member. Very rarely, in lenses of conglomerate, fragments of schist and pebbles of granitoid rocks like those that underlie the Apache group are seen. Also, within the same part of the member, some quartzose pebbles derived from pre-Apache formations show polished facets and shapes like brazil nuts--features that characterize ventifacts. Such pebbles are rare, and have not been noted in many places. In most sections, however, some pebbles show flat faces and rounded edges separating the faces, as though wind-worn pebbles of typical "brazil-nut" forms had been subjected to subaqueous abrasion that rounded the sharp ridges between adjacent facets. The coarser constituents of the upper part of the member show regional similarities. In the northern part of the region virtually all pebbles of the cross-stratified upper unit are of white quartz; in the southern part of the region like quartz pebbles are predominant in the gravels of the upper part of the section.

South of the Nutanes Plateau the stratigraphic interval 100 to 150 feet above the basal conglomerate of the Chediski member is of crudely tabular beds, which mostly range between 6 inches and 4 feet in thickness. Farther north roughly the upper half of the equivalent interval exhibits massive units of contorted sandstones, and in the lower half individual beds are not as distinctly separable. Some of the tabular beds are tangentially cross-stratified on small to medium scales; others are horizontally stratified. The upper parts of many beds were channeled prior to deposition of the next higher bed. Lenses of conglomerate or conglomeratic sandstone, 6 inches to 3½ feet thick, commonly fill the channels. As seen in cross section, many slightly undulant partings between beds are marked by trains of granule-size gravels. In other examples somewhat coarser gravels sparsely litter such planes, and in extreme examples, individual cobbles, 3 to 8 inches in diameter are noted at intervals ranging from a few inches to several feet along the bedding planes. Owing to vagaries of intraformational erosion, some beds wedge out as traced laterally. And gravel lenses or layers, separated by one or more beds of sandstone at a particular locality, may coalesce as traced laterally along the outcrop.

In the southern part of the region the sandstone unit that exhibits slump structures generally is probably not as thick as the contorted sandstone unit in the northern areas. In some southern areas such units are not more than 30 feet thick, though in others slumped beds aggregating at least 125 feet have been seen. At least in some areas the tabular beds that exhibit slump structures are somewhat thinner (8 to 15 feet) than in the northern part of the region; slumping may only slightly modify the original sedimentary structures and individual beds of contorted sandstone may be conspicuously defined, top and bottom, by layers of pebbles. Such generalizations cannot be made for all southern examples, however, because in some localities slumping occurred to such a degree that individual beds within a massive outcrop are difficult to distinguish. The reticulate weathering pattern noted for the northern outcrops is also a common feature of the southern outcrops of contorted sandstone.

Tabular beds of cross-stratified medium- to very coarse-grained sandstone typify the upper part of the member in the southern part of the region, as in the northern part. Unfortunately, I have not taken specific heed of this part of the section in very many areas. And if the sandstones of the northern and southern parts of the region are especially different in comparison, as are lower units, I am not aware of the differences. Southward from the latitude of Globe coarse grains of detrital feldspar are obvious in all of the outcrops of these beds that I have seen. Everywhere in the southern part of the region this uppermost unit tends to be quartzitic. If overgrowths of quartz completely fill voids between the detrital quartz grains, this part of the section forms massive light grey outcrops in which bedding features are obscure. Grossly these outcrops of quartzite are not very different from the quartzite of the upper member, and locally the contact between members may be difficult to define.

Quartzite member

The upper member of the Troy is everywhere of quartzite beds. The slightly different units in the vertical sequence are probably rather uniform in characteristics through the region. These quartzites lack unusual features that would particularly distinguish them from any other sequences of fairly clean quartzites. From observations over the region no great amount of detail can be added to the description of table 4, which cites one of the thickest sections--if not the thickest--in northern Gila County.

Although light grey or yellowish gray hues are dominant in some localities, and in some sections beds of dark values of greyish-red or greyish red-purple are conspicuous, generally the quartzites are greyish pink to pale red. Weathering accentuates the reddish colors. And where the formation is weathered deeply, detached slabs of thinner parting portions of the member may be rendered somewhat porous and be stained throughout in various shades of rusty brown. Such extremes of weathering discoloration are seen mostly where the slabby parting upper part of the member is exposed on mesa tops. In other exposures or on mesas where lower parts of the member are exposed such decomposition is not ordinarily seen. Indeed, the colors of the greyish quartzites, which include very little iron, ordinarily are essentially unmodified on weathering.

The quartzites are mostly medium- to coarse-grained and remarkably free of detrital grains other than clear quartz. Compared with any part of the middle member, individual beds and commonly sequences of beds several tens of feet thick are uniformly of a very narrow range of grain sizes. Rare beds include scattered grains of orange-pink feldspar. Granules and small pebbles, or concentrations of such gravels are extremely rare, and for practical description can be considered non-existent. Beds that include feldspar or pebbles are somewhat more common in the latitudes south of the Globe. In hand specimens the quartz grains appear to be angular and through the bulk of the member the grains fit tightly against one another. Thin sections show that these are well-rounded detrital grains locked together by overgrowths of quartz that generally fill completely the voids between original grains. A thin dust of hematite(?), which marks the rounded outlines of the original detrital grains, and minute aggregates of hematite, sparsely distributed and tightly held along the boundaries between overgrowths, cause the reddish colors in the quartzites.

Outcrops of some localities are abundantly pitted, owing to the weathering free of aggregates of sand that are not as firmly cemented as most of the rock. These "pock-marks," ordinarily not more than 1 inch in diameter, may be sites of thin limonite stains, coatings of black oxides or preferentially favor lichen growths, and be very conspicuous against a background of light-colored quartzite. They seem to be prevalent in certain beds. The "pock-marked beds" appear to have no use as stratigraphic marker units, except perhaps locally. In some localities they are found high in the member, in others low in the member; the bulk of the outcrops are devoid of these features. Only a lithologic correlation can be suggested. The "pock-marked beds" are ordinarily within a sequence of light-colored vitreous quartzite beds, most of which are so firmly cemented that the internal bedding structures and even some of the planes between beds are difficult to distinguish.

The quartzite of the basal several tens of feet of the upper member commonly is coarse-grained; some sections include beds of very coarse grain. Individual grains appear quite angular in hand specimen, and generally the interstitial voids are only partly filled by overgrowths. The porosity and coarseness of grain distinguish this unit from any other of comparable thickness in the member. Hematite is somewhat more abundant than in the rest of the member, and commonly this unit is greyish-red or greyish red-purple, but in places this unit is very light colored. Light grey or yellowish-grey quartzites seem to be common in the southern part of the region.

The overlying unit is 150 to 200 feet thick, medium-grained, tightly cemented, and ordinarily the most vitreous quartzite in the member. It is comprised of tabular beds, which range from a few inches to 6 feet in thickness but are mostly 1 to 2 feet thick. These beds are cross-stratified on small to medium scales, but in a great many outcrops the internal bedding details are obscure. The underlying coarse-grained unit exhibits very similar details of bedding. With the medium-grained unit it commonly crops as a steep ledge-studded slope, or in areas of much relief the two units crop as a simple cliff.

The contact between the medium-grained unit and the slabby parting unit next above is commonly marked by a topographic bench at the top of this cliff. As the upper part of medium-grained unit tends to part more readily and is somewhat less resistant to erosion than the lower quartzites, in many areas a step-like transition zone of ledges exists between the bench and the top of the cliff.

The highest unit of the upper member is also mostly medium-grained, but these quartzites are not so well sorted as those of the two lower units. Some beds are coarse-grained, and a scattering of coarse grains is common in the medium-grained beds. This unit differs from those lower in the section mostly in that it is thinly bedded, bedding structures etch out distinctly, and flaggy to slabby parting of the outcrops is everywhere characteristic. The beds are tabular and range from 2 to 30 inches in thickness; most are 3 to 12 inches thick. Some beds are horizontally stratified; most are tangentially cross-stratified on small to medium scales. Locally some of the more prominent partings between beds are undulant. Where this unit is exposed in thicknesses of more than 100 feet, it generally crops on a steep slope that merges downward with the bench previously noted and merges upward into ragged cliffs. The outcrops of the lower part of the unit are commonly covered by flaggy and slabby talus from higher in the section.

In the Sierra Ancha the upper member is readily recognized from a distance by its distinctive topographic profile of two prominent cliffs separated by a steep slope and a bench. In the higher parts of the range (within the west half of the McFadden Peak quadrangle) the slabby-parting upper unit of the member is remnant in thicknesses of 150 to 200 feet. Elsewhere north of the latitude of Globe the slabby unit is thin or missing, even where Paleozoic formations overlie the Troy, and the distinctive topographic profile is not characteristic. Farther south the slabby-parted unit or equivalent quartzites are generally missing owing to pre-Paleozoic erosion. Only in the Holy Joe Peak quadrangle, to the best of my knowledge, can definite exception be made to this generalization. Most exposures of younger Precambrian rocks in this quadrangle are of Apache strata or included diabase sills, both of which are directly overlain by Cambrian formations. But in certain small areas, between Aravaipa Creek and Holy Joe Peak, as much as 700 feet of Troy quartzite intervenes between the Apache strata and the Bolsa quartzite of Cambrian age (M. H. Krieger, written communication). Eleven miles to the west, near the mouth of Camp Grant Wash, comparable thicknesses of Troy probably exist but owing to Cenozoic faulting and poor exposures the relations between formations are not as readily seen. These various Troy sections are preserved in structural blocks that were depressed relative to the Troy of surrounding areas, and therefore survived the erosional planation that preceded deposition of the Bolsa quartzite. Near the head of Skeleton Canyon, about a mile west and northwest of Holy Joe Peak, the upper member of the Troy is slightly more than 800 feet thick, or about the same as the thickest sections in the Sierra Ancha. Overall these sections of the upper member are lithologically quite like those in the Sierra Ancha except that the upper 200 feet, corresponding roughly to the slabby parting thin-bedded uppermost unit, is much thicker bedded and forms massive outcrops. This part of the member is of light grey or yellowish grey quartzite in tabular beds, 1 to 8 feet thick, in which internal bedding structures are obscure.

Ransome (1916, p. 140) suggested that the Troy quartzite in the vicinity of Scott Mountain, 2½ miles northeast of Ray, could be as much as 1,000 feet thick. But, not realizing that local structural remnants such as those of Holy Joe Peak area could be much thicker than the average sections of the region, he rejected this thickness as unreal. Assuming that quartzites of Cambrian age do not comprise an unduly large part of the so-called "Troy" section in the vicinity of Scott Mountain, as is entirely possible, this might be another locality in the southern part of the region for examination of sections of the member in near maximum thicknesses.

Features that distinguish the Chediski and quartzite members

In typifying and illustrating the lower and middle parts of the Troy as massive-cropping conspicuously crossbedded and pebbly quartzite, Ransome (1916, p. 139-141 and pl. 27A; also 1919, p. 44) drew on observations, particularly from the vicinity of Globe and Ray, of the Chediski member where sections are especially quartzitic. The upper member does not include similarly conspicuous features. Furthermore, it is thin or missing in most of the areas mapped by Ransome. And sandstone sections of the middle member do not crop out as conspicuously as the quartzitic sections. Therefore neither the sandstones nor the pebble-free quartzites were accorded much prominence by Ransome.

It is quite understandable that Ransome did not distinguish the two members in the areas of his principal observations. South of the Salt River the Chediski member is not as obviously different from the upper quartzite member as in areas north of the river. Southern outcrops of the Chediski member are not as whitish in color; the sandstones are not as friable and in some areas are quite quartzitic; bedding features are more distinct, and outcrops are less massive; and pebble concentrations are less restricted in their stratigraphic distribution. Although not as prominently displayed in the southern part of the region, gross aspects of the Chediski member--such as slump structures and very irregular bedding--serve everywhere to differentiate the bulk of the member from the overlying quartzite member. The lower part of the upper member is coarse-grained, as is the upper part of the Chediski member. Where the upper part of the latter is quartzitic, and distinctive bedding characteristics are therefore largely obliterated, locally the boundary between the members cannot be distinguished in passing casual observation.

In most areas north of Globe the boundary between members is sharply and simply defined as a contact between sandstone and quartzite units. In addition the units above and below this contact have definitive characteristics. For consistency, if there is cartographic need to define the contact between members in those southern areas where a part of the Chediski member is quartzite, these definitive features are the criteria that should be used. The upper part of the Chediski member, even where a vitreous quartzite, includes some clay and sericite between quartz grains or is feldspathic; the coarse-grained basal unit of the upper member is virtually free of such constituents. Although the irregular beds or the internal large-scale cross-stratification of these beds may amply set them apart from the more uniformly bedded overlying quartzites, in examples vitreously quartzitic or in those much shattered by faulting these differences are not easily discerned. The quartzitic facies as well as the sandstone facies of the Chediski member are characteristically poorly sorted: some beds are not pebbly, but many are, and lenses or thin layers of pebbles separate beds or mark the tops of individual beds in the upper part of the member. Comparatively, the coarse-grained lower unit of the upper member exhibits much better sorting, and sets of beds encompassing a minimum stratigraphic interval of 20 to 30 feet typically show the same degree of sorting. Certainly a few of the coarser-grained beds locally do include grains of granule size or even thin lenses of granules. Such examples are so rare as to be of little practical concern. Fortunately, almost everywhere a conglomeratic quartzite or sandstone bed or a conglomerate bed, especially conspicuous because it is thicker and more persistent than any pebbly layer stratigraphically as much as 100 feet lower in the section, is the uppermost bed of the Chediski member. No comparable bed has been seen in the upper member. Rarely this conglomeratic zone is locally missing, but it will generally be found in searching the vicinity. As a practical expediency, the highest persistent conglomeratic layer in a sequence of coarse-grained quartzites probably can be taken as the contact between members, without being in error more than 20 to 30 feet stratigraphically from one locality to another in the southern part of the region.

Diabase

Form and distribution of intrusions

Diabase, in the form of dikes and sills, mainly inflated the sedimentary formations of the younger Precambrian, and in many areas the intrusions are restricted largely to the Apache group. But extensive diabase bodies were also emplaced in the older Precambrian formations. Paleozoic formations rest unconformably on the diabase. Only the broader outcrops of diabase are outlined on figure 2; all small intrusions and even some sills hundreds of feet thick exposed in cliffs or otherwise narrow in outcrop are omitted. The relatively few diabase bodies shown as outlying from Apache and Troy exposures are mainly intruded into older Precambrian formations; some are those exposed in erosional windows cut through Paleozoic or younger formations. The large diabase intrusions of southern Arizona are coextensive with the younger Precambrian sedimentary rocks. Cartographic limitations notwithstanding, figure 2 serves well to illustrate this.

The Salt River Canyon east of Canyon Creek and the Sierra Ancha are particularly good areas to observe the forms of diabase intrusions. Exposures are excellent and continuous for large areas, and the local relief is 2,000 to 5,000 feet so the habit of intrusions in different parts of the stratigraphic sequence are readily compared. Furthermore, details of sill habit differ somewhat from place to place, and these two areas include exposures of all variations that are recognized regionally. Although much the following description is drawn from observations in these two areas, the generalizations are applicable region-wide.

Sills greatly predominate in volume; dikes are relatively few and insignificant. Sills range from a few inches to 1,200 feet in thickness, and individual sills spread laterally for distances of a few feet to at least 20 miles. Thicker sills have been reported, but these are very likely composite sheets comprised of two or more separately injected sills. The sills are remarkably persistent; examples only one foot thick have been observed to crop out laterally for more than a mile. Dikes are relatively narrow; a range in width of a few inches to a few tens of feet would include most dikes. Some tabular bodies that connect a concordant sill at one horizon with a like sheet at another horizon are 50 to 800 feet in width; these masses can be considered discordant portions of the sills rather than dikes. Some dikes, 10 to 30 feet wide, can be traced for a few miles; these are rare. The sedimentary host rocks, which are virtually horizontal in the Plateau portion of the region and ordinarily dip less than 25° in the Basin and Range portion, tend to erode as cliffs and the less resistant diabase erodes as slopes. Therefore, the sills, almost ubiquitous with the Apache group, have extensive outcrops of great prominence in most landscapes that expose younger Precambrian formations.

The sills are remarkably concordant, in general, but each sill if traced carefully along its outcrop proves to be in part discordant. Many discordancies are in the form of high-angle, abrupt steps, up or down through the stratigraphic section. The height of these steps ranges from a few inches to several hundred feet and commonly varies along the strike of a given discordancy. A second common, but less obvious, type of discordance is low-angle; the contact between diabase and sedimentary rocks is a gently undulant plane that transects the bedding planes of the host rocks at a low angle. Most sills transgress from one stratigraphic horizon to another along a devious route, rather than by simple tabular dikelike connections. Discordant portions of sills that can be traced in cross section or along the strike for some distance--a few hundred yards to several miles--ordinarily are not simple high-angle or low-angle discordancies but a combination of the two types. That is, as a contact conformable with bedding is traced in section, it may cut abruptly up across bedding at a high angle, then flatten to conform locally with bedding or transect bedding at a low angle, steepen abruptly again, then flatten to make the connection between concordant portions of sills. Along the strike of a discordancy, also, the contact between diabase and sedimentary rocks may change from high to low angle. In this report, if the contacts of a tabular body are in general conformable on a regional scale with the enclosing sedimentary rocks, the term _concordant sill_ will be used. Where a tabular sheet is locally concordant but through large areas has contacts that transect the bedding of the host formations it will be referred to as a _discordant sill_, or where the relations to a concordant sill can be demonstrated as the _discordant part of a sill_.

Emphasis, here and elsewhere in the report, on the discordant parts of sills should not be construed to suggest that such features are everywhere characteristic. Though discordant contacts are numerous, they encompass only a small part of the total volume of diabase in sills.

The shape of the sills in plan cannot be characterized from present knowledge, because only rarely is the termination of a sheet observed, and the edges that have been seen obviously represent only a very small part of the perimeter of an intrusion. Some thin sills, clearly apophyses from thicker sills, wedge out gradually. Other thin sills and some thick sills terminate abruptly against high-angle faults. Also, sills cannot be readily described in plan because a given sill is not everywhere at one stratigraphic horizon and therefore is not exposed uniformly in relation to topography. A given sill may occur at one stratigraphic horizon through an area of a few tens of square miles, and via discordant connections exist in like volumes at horizons several hundreds of feet higher or lower in adjacent areas. Such extremes of stratigraphic position for different parts of one sill seem to be more characteristic of one area than another. The principal sill in the west half of the McFadden Peak quadrangle was emplaced throughout large areas along a horizon near the middle of the Dripping Spring quartzite, in other extensive areas it was intruded along different horizons in the Mescal limestone, and near Aztec Peak it was intruded along a horizon high in the upper member of the Troy quartzite. The largest bodies of diabase connecting the concordant portions of the sill are thick tabular parts of the intrusion that transgress the host rock mostly at high angles. Small step-like discordancies are abundant in this southeastern part of the Sierra Ancha. In contrast, east of

Canyon Creek along the Salt River Canyon, though a few "steps" of several hundred feet have been observed, most sills do not appear to deviate more than a few tens of feet from a given stratigraphic horizon in outcrop lengths of 5 to 10 miles; the many discordant bodies that connect concordant parts of the sills are mostly thin and of no great vertical dimension.

In the Salt River Canyon diabase inflated the section as multiple intrusions, but in the Sierra Ancha multiple injections apparently were rare. Where U. S. Highway 60 crosses the canyon three separate injections of diabase can be observed readily, and from widely separated observations the writer speculates that not less than five separate injections might be demonstrated. Throughout most of the McFadden Peak quadrangle only one stage of intrusion has been recognized; in a very few localities two injections have been seen. Probably multiple intrusions are not typical of this area, but are characteristic of the Salt River.Canyon and the area 5 to 10 miles south of the canyon.

The cumulative volume of diabase sills varies considerably in the region. In the Salt River Canyon the ratio of diabase to exposed pre-Paleozoic strata is about 1:1, or slightly greater. Because the sedimentary rocks tend to occur as narrow outcrops in plan and the diabase as broad outcrops, on a detailed map of the south half of the Blue House Mountain quadrangle the sedimentary rocks would appear as blocks and slivers engulfed in a sea of diabase. In the Sierra Ancha, in contrast, sedimentary formations greatly predominate over diabase in volume. The southeastern one-third of the McFadden Peak quadrangle between Cherry Creek and Canyon Creek, includes relatively few sills, most of which are thin and occur in the lower formations of the Apache group and in the underlying granitoid rocks.

Certain stratigraphic units were more susceptible to inflation by diabase sills than others. In most areas the Mescal formation was inflated at more horizons and now encompasses a greater volume of diabase sills than the other Precambrian formations. The Pioneer formation commonly was intruded by two or more sills, but these are usually thin. Sills are locally prominent in the Dripping Spring and Troy quartzites, and are few in the Dripping Spring and are even fewer in the Troy. Through much of the Sierra Ancha southeast of McFadden Peak one sill, ranging between 300 and 1,000 feet in thickness, splits the Dripping Spring. In many other areas in northern Gila County sills in the Dripping Spring are less than 200 feet thick, but south of the Apache Mountains sills 200 to 500 feet thick exist in this part of the section. Most sills in the Dripping Spring are in the upper member and, to the best of my knowledge, no sills of appreciable thickness or lateral extent occur in the lower member. In a few localities sills in the Troy formation are thick--as in the vicinity of Aztec Peak, where a sill 800 to 1,200 feet thick splits the formation; in several areas sills in this formation are between 30 and 200 feet in thickness, and in many places diabase intrusions are scanty or missing in the Troy. Diabase intrusions in the Troy have been observed particularly where that formation is now remnant in considerable thickness; if the Troy, and any included diabase, had not been removed over large areas by pre-Paleozoic erosion perhaps diabase would be seen more abundantly in the higher parts of the formation. Thin sheets of diabase are common, and in some areas widespread, in the upper 300 feet of the older Precambrian granite. These sheets approximately parallel the bedding of the overlying Apache group, and apparently inflate joints developed parallel to the pre-Apache surface. At depths more than 500 feet below the base of the Apache group diabase intrusions of appreciable extent are practically nonexistent. Thus where granitoid or schistose formations without included diabase are widely exposed, it seems reasonable to postulate that a considerable thickness of these formations was eroded and that remnants of the Apache group are not likely to be found in the vicinity. In summary, thin sills can be considered characteristic of the upper part of the pre-Apache formations, the Pioneer formation, and generally--but with prominent exceptions--of the Dripping Spring quartzite and the Troy quartzites ~~formations~~; the Mescal is characteristically inflated by several sills, which may be thick or thin.

Most sills were intruded along horizons defined by striking differences in the competency of the rocks above and below the horizons of intrusion. Good examples are the sills commonly in the Mescal limestone at or near the contact between the middle member--a thick-bedded, massive unit--and the lower member--a thinner-bedded, less competent unit. In a comparable example, sills ordinarily split the siltstone units of the upper member of the Dripping Spring adjacent to the subordinate, but more competent, feldspathic quartzite units within the member, or at a horizon not far above the contact between the siltstone member and the massive, quartzitic lower member. Diabase commonly was intruded along all unconformities in the Apache section except that between the Mescal and Dripping Spring formations. These occurrences are really special instances of intrusion along a horizon separating strata of contrasting competency. Where sills do not occur exactly at the contact between competent and relatively incompetent stratigraphic units, they are largely along horizons in the incompetent unit but within a few feet of the competent beds. Seldom is a sill observed in the middle part of an incompetent unit. Nor do many sills that occur along a horizon within competent units occupy these horizons for considerable lateral distances. Massive stratigraphic elements that do not part readily rarely are hosts for diabase. The basal breccia of the Mescal limestone, the later-formed massive solution breccias that comprises the Mescal in some areas, the massive middle member of the Troy, and parts of the lower member of the Troy that are cross-bedded on a very large scale are virtually free of diabase sills. Dikes or discordant dike-like connections between sills do penetrate these units.

With appreciation of these tendencies for localization of the sills and some knowledge of the horizons inflated in immediately adjacent parts of the region, if a discordant step in a sill is seen the various horizons at which the intrusion might again become concordant can be assessed, with some expectancy of accurate prediction. This approach has practical application in predicting sites of unexposed asbestos deposits, which are contact metamorphic phenomena in the Mescal limestone adjacent to diabase.

In the following enumeration of the stratigraphic positions of sills some typical examples of the regional consistencies and inconsistencies of stratigraphic position of the sills are noted. The northern part of the McFadden Peak quadrangle and a belt 2 to 4 miles to the north is an area of many discordant intrusions. In this area the principal sill, as much as 600 feet in thickness, was intruded mainly along the unconformity between the upper and algal members of the Mescal ~~formation~~; locally this sill steps up and follows the unconformity at the base of the Troy or splits and follows both horizons. A second sill, generally less than 200 feet thick, was intruded mainly at a position 40 to 50 feet below the top of the lower member of the Mescal. This and the higher sill are connected in a number of places by high-angle discordant portions of the intrusions. Diabase intrusions in the Dripping Spring and Troy quartzite ~~formations~~ of this area are relatively small and generally discordant at low to high angles.

In the southwestern part of the McFadden Peak quadrangle--south of McFadden Peak and west of the crest of the Sierra Ancha--the principal sill was intruded, in general, at a stratigraphic horizon low in the upper member of the Dripping Spring quartzite. Near Asbestos Peak, 4½ miles southwest of Aztec Peak (see McFadden Peak quadrangle), a second sill, 500 to 600 feet thick, separates the massive lower unit of the algal member of the Mescal from the thinner-bedded upper unit. These two sills merge as one sill along a belt of northwest-trending high-angle discordancies that crosses this southern fringe of the Sierra Ancha about one-half mile east of Asbestos Peak. From this line of discordancies east 2½ miles to the canyon of Coon Creek, which was eroded along another northwest-trending belt of discordancies, the horizon of intrusion is for the most part the unconformity between the upper and algal members of the Mescal. Along Coon Creek this major diabase sill abruptly thins by discordant steps and exists farther east only as minor apophyses high in the Mescal or along the unconformity at the base of the Troy quartzite. In the vicinity of Aztec Peak the same sill, here 1,000 to 1,200 feet thick, inflated a horizon high in the Troy. Along the west wall of Cherry Creek Canyon, between the latitudes of McFadden Peak and Aztec Peak, this sill is about 800 feet thick and inflated the upper part of the lower member of the Mescal. Most of the diabase outcrops seen in this part of the Sierra Ancha are of this major sill or of minor apophyses from it. Other sills exist in the parts of the McFadden Peak quadrangle above described but most are relatively thin and lower in the section.

Along the Salt River Canyon, and as far south as the village of Chrysotile, the relations of sills to enclosing sedimentary formations is somewhat different from those just described for parts of the McFadden Peak quadrangle in that: (1) much more diabase inflates the Apache formations of the Salt River-Chrysotile area; (2) multiple injections of diabase tended to penetrate along, or almost along, a common horizon; (3) diabase inflated the Mescal limestone at more horizons and more intricately than in the Sierra Ancha. As in the northern part of the McFadden Peak quadrangle, a sill commonly splits the Mescal formation 45 to 50 feet below the top of the lower member. In many places an apophysis from this sill or a different sill separates the algal member from the lower member or splits apart limestone beds of the uppermost ten feet of the lower member. In a few places this higher intrusion followed the first prominent parting that is a few feet above the base of the algal member. A third major sill occurs in the lowest 50-foot interval of the Mescal along a somewhat undulating plane. In places this sill was intruded just above the Mescal-Dripping Spring contact; in other places it is concordant for considerable distances at horizons immediately above or below a stratigraphic unit of relatively massive limestone beds that is 30 to 50 feet above the contact between formations. The two, and in many places three, major sills that split the lower member of the Mescal bifurcate in cross section, and are interconnected by discordant parts of the sills that transgress the bedding from one main horizon to another. Thus, for a large part of this area, the lower member of the Mescal is represented by thin plates or wedges of limestone, a few feet to more than 100 feet thick, interleaved with diabase sheets a few tens of feet to about 800 feet thick. The limestone plates are interrupted in many places by the discordant connections between sills. This geometric pattern can be contrasted with the single, geometrically relatively simple sill that inflates the lower member in the northern part of the McFadden Peak quadrangle.

The next horizon of extensive sill intrusion above the Mescal in the Salt River-Chrysotile area is the unconformity at the base of the Troy quartzite. The form of the intrusion at this horizon depends somewhat on the relation of the unconformity to the underlying rocks. The basalt that ordinarily overlies the Mescal is largely missing, and for part of the area the upper member and part of the algal member is missing owing to pre-Troy erosion. In much of this area, a simple tabular sill separates the Troy from underlying strata. Where remnants of the upper member of the Mescal exist apophyses from this sill may separate the upper and algal members or split the argillites of the upper member at one or several horizons.

Diabase intrusions occur within the Troy only in widely separated localities in the Salt River-Chrysotile area. In these occurrences the contacts between the sills and the quartzite tend to be undulatory. More high-angle discordant bodies than sills are seen.

Tabular plates of Mescal or Dripping Spring strata, seemingly completely isolated from one another and from thicker masses of strata by diabase, provide particular information on the mode of diabase intrusion. Examples of such plates 5 to 300 feet thick, at least 2 miles long and not less than 1/2 mile wide are exposed in the Salt River Canyon. Plates of such dimensions ordinarily are only reasonably inferred to be everywhere surrounded by diabase. Owing to the three-dimensional exposures provided by some close-spaced tributary canyons, however, like plates of lateral dimensions of only a few hundred yards are definitely known to be so isolated. Some plates wedge out, others end abruptly against a steep discordant contact. The bedding within most of the slabs virtually parallels that of strata above and below the enclosing diabase bodies. Where a discordant part of a sill terminates a plate the beds of the plate are noted to match perfectly in thickness and sequence the beds against the other wall of the discordant mass. As this implies, and as can be confirmed on close inspection, none of the sedimentary rock has been assimilated in the diabase magma. If viewed from a distance so that a considerable length of a given plate can be seen, no discernible warping may be apparent, even though the plate is only 10 feet thick and of the least competent strata. In some canyon walls only one such plate is seen, in other localities several plates are separated by tabular layers of diabase.

On casual observation these plates might seem to be huge xenoliths dropped into a sill before it solidified. Closer inspection, however, proves that most such plates intervene between sills injected at somewhat different times. As shown diagrammatically on figure 12, a simple sill--perhaps largely

Figure 12.--Diagrammatic sections showing relations of multiple injections of diabase to plates of host rock isolated between intrusions.

concordant--first inflated the section. A second intrusion, following the same general path as the first, locally deviated to pry from the wall of first sill a plate or plates of sedimentary rock like those described above. Invariably, if outcrops are of adequate geometry, large plates completely surrounded by diabase are seen to be of this origin. The second sill, as shown by its chilled margins and in some examples by displacement of the first sill, was intruded after the first sill had crystallized. Excellent exposures of one sill, with chilled margins, inflating another can be viewed along the part of U. S. Highway 60 that traverses the north wall of the Salt River Canyon. The contact between the sills of this particular exposure is generally planar, but locally undulant; everywhere it is practically concordant. And locally lenses only inches thick of limestone, the host rock for the two sills, indicate that the magma of the second sill was insinuated essentially along the contact between the first sill and its sedimentary host. In some areas there were at least three separate injections of diabase; the complex geometric pattern of displaced plates in some areas, which have not been studied, suggests that additional intrusions might be proved for these localities. No features have been seen that suggest injection of a later sill before the earlier intrusion had chilled throughout.

Some plates of sedimentary rock not widely separated by diabase from the main mass of sedimentary rock are shown, by lateral tracing, to be plates that connect with the main mass. The diabase separating the two plates of strata is a minor wedge-shaped apophysis from a larger diabase sill. This apophysis commonly represents a later diabase sill that wedges out, but this is not everywhere true.

The features described above have been seen on all scales, from examples in which the intervening sills are hundreds of feet thick to examples in which they are only inches thick. Everywhere the diabase was insinuated into the host rock without causing appreciable folding of the strata. Actually some very small-scale deformation, described on a later page, did occur.

Very rarely fragments of host rock that did drop into molten magma have been found. With very few exceptions these have been seen in varieties of intrusive rock other than the ordinary diabase that encloses the plates described above. The largest of these xenoliths that has come to my attention is 4 feet thick and about 15 feet in longest dimension. Such sizes are extremely rare; most xenoliths have dimensions in inches. The setting of these xenoliths is described in the next section.

Detailed mapping has shown that multiple intrusions of diabase are uncommon in parts of the Sierra Ancha, and reconnaissance in more southerly parts of the region suggests that such may be characteristic of other areas. From more casual observations it is certain, however, that multiple intrusions are typical of other large areas than that exposed in the Salt River Canyon. In much of the Basin and Range portion of the region, Cenozoic faulting obscures the geometric relations between diabase and the rocks that it intruded. Owing to pervasive alteration in Tertiary time, and consequent poorer outcrops, relations are particularly obscure in the vicinity of the base-metal districts of the region. Nevertheless, applying knowledge of the characteristic habits of the diabase obvious in the northern part of the region, reasonable interpretations of some puzzling structural relations involving diabase can be made.

A good example might be the relations of the diabase sill, commonly cited as the thickest known, in the Magma mine at Superior. Exposed by the mine workings are two sills reported to have a combined thickness of more than 3,000 feet; the upper sill is said by Short to be 2,000 feet thick (Short and others, 1943, p. 37). Described as xenolithic within the 2,000-feet 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." Aside from other considerations, "this attitude is difficult to understand if the blocks sank into a molten magma." In view of the demonstrable relations of such sedimentary blocks in many other places in the region and the complete absence of sizeable xenoliths within any known single diabase intrusion, Short's descriptions of the extent and attitude of the "xenoliths" provide the basis for an explanation of this example. Almost certainly the larger "isolated bodies of quartzite" exposed in the thick "sill" of the Magma mine are blocks or plates separating 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 in relatively restricted underground exposures, tends to support this interpretation. Some of the smaller "xenoliths" could be locally isolated by apophyses from the main intrusions, or more likely--because formations of the area are abundantly broken by Tertiary faults--many are probably offset fault segments. Of broader regional application, corollary to this interpretation, no sills thicker than the 1,200-foot examples in the Sierra Ancha have yet been confirmed.

Petrology

The diabase bodies are mostly of dark-colored, holocrystalline rock that is fine- to coarse-grained, but generally medium-grained, and is mainly of ophitic or subophitic texture. This rock will be designated the "typical" or "normal" diabase to distinguish it from the more feldspathic or "red rock" types: diabase pegmatite, aplite, and granophyre. These red rock facies are conspicuous because of lighter colors and differences in texture, but comprise only a minute part of the intrusive material.

Varietal types within the classification outlined above exist. Many thin sections of the normal diabase have been examined, but from field observations it seems likely that the following descriptions still may not adequately suggest all of the varietal types. Few thin sections of the red rock types have been studied. Furthermore, the spatial relations between some petrographic types have been seen in only a few places. Further study undoubtedly would result in better definition of the characteristics applicable to each of the petrographic types.

The normal diabase is a tough, medium gray to dark gray rock. The minerals readily visible to the unaided eye are plagioclase, black pyroxene, and in the coarser phases black ilmenite-magnetite and clear needles of apatite. In much but not all of the rock biotite is obvious, and rusty-weathering, yellowish-brown olivine can be seen in some diabase with the aid of a hand lens. In deuterically altered diabases a dark, greenish hornblende is readily seen. Large grains of pyroxene are particularly noticeable on firm outcrops, because cleavage faces flash in the light and mottle the surface of the outcrop. These pyroxene grains are crowded with laths of plagioclase; and on weathered surfaces these ophitic aggregates characteristically protrude, resulting in the knobby or warty surfaces typical of many diabases. The form of the plagioclase grains is much more apparent on weathered than fresh surfaces.

Normal diabase weathers light olive gray to moderate yellowish-brown as observed at close range, and can ordinarily be distinguished at a distance by the characteristic greenish or olive gray hue of the parts of slopes barren of vegetation. Spheroidal weathering to rounded, lumpy- or warty-surfaced boulders is common. The rock disintegrates to a light brown or yellowish brown granular soil that commonly, but not invariably includes abundant kernels (fig. 13) that represent the relatively resistant oikocrysts of pyroxene.

Much of the typical diabase disintegrates rapidly. Some outcrops, first observed as newly blasted, tough, hard rock faces in various aspects of exposure, and revisited at intervals, crumbled almost entirely to soil during periods ranging from three to ten years. Figure 13 shows

Figure 13.--Outcrop of diabase showing characteristic effects of disintegration caused primarily by weathering. Small residual pebbles are ophitic aggregates, little decomposed and of typical sizes.

a typical outcrop of such diabase that had been exposed in a roadcut about four years. Other diabase forms hard, little leached outcrops that disintegrate very slowly. These diabases ordinarily exhibit comparatively little deuteric alteration, and commonly olivine is readily recognized in such outcrops.

In many localities conspicuous light- or dark-colored ribs, ordinarily less than 1/4 inch in width project a small fraction of an inch to two inches above the weathered surface of the diabase. The light-colored, usually pinkish ribs represent concentrations of albite along joints or minute fractures in the diabase; the dark ribs are like concentrations of hornblende, biotite, and chlorite. These ribs ordinarily mark diabase that was deuterically altered to a considerable degree.

FIGURE 13. OUTCROP OF DIABASE SHOWING CHARACTERISTIC EFFECTS OF DISINTEGRATION CAUSED PRIMARILY BY WEATHERING.

Small residual pebbles are ophitic aggregates, little decomposed and of typical sizes.

Correlation of thin section studies with field observations indicate that the bulk of the normal diabase is composed largely of labradorite and clinopyroxene. Olivine may be absent, sparse, or abundant but exists in appreciable amount in so much of the rock that this facies of the normal diabase can be characterized, and has been (Ransome, 1919, p. 54), as olivine diabase. Albite, rather than labradorite, is the plagioclase of a texturally similar facies of the normal diabase that occurs in much lesser volume. Either pyroxene or amphibole may dominate the mafic mineral content of the albite diabase, which does not contain olivine. This diabase or diabase with the calcic plagioclase but with the pyroxenes partly altered to amphibole seem to have been described as hornblende or augite-hornblende diabase (Peterson and others, 1951, p. 34; Short and others, 1943, p. 36). All variations between olivine diabase and albite diabase have been seen in one sill.

Thin sections of little altered olivine diabase show that labradorite (An_{55}-An_{70}) comprises 45 to 65 percent of the rock, clinopyroxene 10 to 25 percent, olivine 5 to 20 percent, biotite and hornblende each comprise 1 to 5 percent, and ilmenite-magnetite, apatite and sparse grains of sphene aggregate 2 to 8 percent. Most of the plagioclase is subhedral; the laths in the usual medium-grained olivine diabase range from 1 to 4 millimeters in length, and commonly the outer one-third of the crystals is normally zoned. A small part of the clinopyroxene in some thin sections is clear pigeonite ($2V < 30°$), but clear or slightly brownish or pinkish augite is the more prevalent and in many specimens the only pyroxene identified. Grains of augite range from a fraction of a millimeter to 40 millimeters in diameter and are commonly 5 to 20 millimeters. In many thin sections these grains include plagioclase grains subophitically rather than ophitically, as would be judged from the hand specimen. An orthopyroxene, which is hypersthene in many instances and probably in most, has been observed in small amount in some specimens. The olivine, indicated to be about $Fo_{60}Fa_{40}$ in composition from refractive indices of a very few examples (see method of Deer and Wager, 1939, p. 21), occurs as small rounded grains or clusters of rounded grains. Biotite, most of which is strongly pleochroic from yellow to reddish brown, is widely distributed, but is an abundant constituent in comparatively little of the slightly altered olivine diabase. Pyrite is sparsely distributed through most of the normal diabase, and in some specimens chalcopyrite can be identified.

Much of the olivine diabase shows considerable effects of late magmatic or deuteric alteration. Olivine, the most unstable mineral in most specimens, is replaced by chlorite or by serpentine; the latter includes the trains of fine granules of magnetite commonly found in serpentinized olivine. In some specimens the olivine was also converted to talc, iddingsite, or bowlingite. In the diabase of many localities the outlines of the mafic minerals have been completely obliterated; olivine is absent, and from the outlines of alteration minerals it cannot be said unequivically that any are pseudomorphic after olivine. But from studies of suites of specimens of typical diabase, olivine is suspected to have existed in many localities in which it or pseudomorphic alteration products are not now apparent. Augite was ordinarily replaced by green hornblende and chloritic products; in some specimens stilpnomelane is abundant, and tremolite has been observed in some. The reddish-brown biotite, at least in part, occupies the earlier position of both olivine and pyroxene. Greenish biotite occurs in diabase so thoroughly altered that the original mafic minerals are difficult to identify. Epidote, in abundance, takes the spatial positions of both mafic minerals and plagioclases in the diabase of a few areas, but in most localities epidote is absent. The cores of the labradorite laths were commonly converted to fine-grained aggregates of sericite, clay minerals, and chlorite, and the rims were rendered turbid by alteration products generally too fine grained to be resolved under the microscope. In large parts of some specimens the turbid rims are albite. Prehnite replaced the plagioclase, especially the cores of the laths, of some of the diabase. Leucoxene is associated with skeletal aggregates of ilmenite-magnetite and is seen particularly in the coarser variants of the diabase.

Some parts of a given intrusive body exhibit little alteration and the primary mafic minerals are almost intact; in other parts the plagioclase and the olivine are much altered and the pyroxenes little altered. In still other parts feldspars and all primary mafic minerals are much altered. In the diabase of a given outcrop area of a few acres the degree and type of alteration may be relatively uniform or may vary considerably.

Discrimination between some albite diabase and diabase with labradorite as the plagioclase can be difficult in the field; under the microscope discrimination between turbid albite or microperthite and saussuritized labradorite is not readily made for some specimens. In clear-cut examples of albite diabase the plagioclase is entirely albite; the other principal minerals are augite and hornblende. Augite may be the dominant mafic mineral and be only moderately altered to hornblende, or the essential mafic may be a fibrous hornblende that is gray green in handspecimen but under the microscope is strongly pleochroic from yellowish green to blue green and is plumose in habit. Olivine or pseudomorphs after olivine are not seen in the albite diabases. Most of the biotite is reddish-brown in thin section; some is green. Accessory minerals are those observed in the olivine diabase. In most of the albite diabase the albite is grayish pink to orange pink in handspecimens, and the turbid laths or turbid rims of this pinkish mineral are in relatively few examples, microperthite. Alteration products partly or completely obscure twinning in the al ite. Prehnite is a more abundant alteration product of the plagioclas in the albite diabase than in the olivine diabase. The pinkish col ation of the plagioclase is not characteristic of all albite diabase; a ppreciable part of the albite as seen in handspecimen is white to ay. Albite is often suggested in the field, however, by clouded rims the grains and by vague, serrate boundaries characteristic of the al e laths.

In relatively uncommon examples of albite diabase difficultly identifiable relicts of labradorite suggest that the albite is the result of sodic metasomatism. Many thin sections, however, lack any indication of an earlier plagioclase.

Much of the rock that might be termed a hornblende diabase in the field and some of the albite diabase is coarser than the average normal diabase. Some is very coarse-grained and includes labradorite or albite laths 5 to 25 millimeters in length and augite crystals, which may be slightly to largely altered to grey-green hornblende, as much as 50 millimeters in length. In some of this coarser-textured rock the augite tends to fracture as curved blades; many of these grains are twinned. Where the hornblende is obvious in handspecimen it is fibrous and under the microscope much appears plumose. Biotite tends to be more abundant in such rocks than in the average olivine diabase. Apatite needles are larger and seemingly more abundant, and skeletal crystals of ilmenite-magnetite are readily seen in handspecimens.

As the diabase coarsens and some of the plagioclase is readily identifiable as albite, ophitic texture may not be readily observed in outcrops. But under the microscope the ophitic texture of such rocks may be obvious. The generalization may be valid, however, that much of the coarser albitic variety of the normal diabase tends to lack ophitic texture. This is true of an albitic variety that has a greater plagioclase content than is usual.

Albite and quartz, as graphic intergrowths on a microscopic scale, occupy interstices between plagioclase laths in some of the diabase that is seemingly of typical texture in the handspecimen. This micropegmatite does occur sparsely in otherwise non-albitic diabase and has been observed in little altered olivine diabase, but occurs most commonly and abundantly in albite diabases. Not all albite diabases, however, include micropegmatite. Intrusions that contain micropegmatite generally are lighter in color than most of the diabase; some specimens in which micropegmatite has been observed, however, are very dark in color because ilmenite-magnetite is extraordinarily abundant.

All variations of mineral content between that of olivine diabase and that of an albite diabase appear to exist in some sills. Thus diabase that lacks olivine, but has labradorite as the plagioclase may include abundant blue-green hornblende as does much of the albite diabase. Or albite diabase, without vestiges of labradorite, may be texturally similar to the ophitic olivine diabase; in such examples olivine is absent but the augite may be largely unaltered and hornblende sparse. The rims of the labradorite grains of some much altered olivine-type diabase have been replaced by albite or microperthite(?). Olivine diabase grades into diabase without olivine and into diabase with micropegmatite. All of these transitions are subtle and not easily determined in the field. In contrast, though these varieties do grade into the red rock types, the transition is relatively abrupt and ordinarily takes place through a zone a few inches to a few feet in width. Exceptionally the zone between normal diabase and large masses of the coarse pegmatite or granophyre is only a few inches in width, and some of the highly feldspathic dike-like masses--for instance, some aplites--have sharp boundaries with the normal diabase.

Gravity differentiation certainly occurred in some of the thicker sills of the Sierra Ancha, but the effects of gravity differentiation have not been observed in all diabases. In some thick sills certain layers are more feldspathic, others more olivine-rich than the average normal diabase. Olivine, however, is not largely restricted to a zone low in a sill, as for the classic well-studied Palisade diabase sill of New Jersey (Walker, 1940, p. 1067), but may be an varietal mineral through much of a particular thick sill. Where olivine is particularily abundant, commonly in the middle one-third of a thick sill or through a like interval slightly lower, certain fairly conspicuous layers may include coarser grains of olivine than the remainder of the sill. These layers are not sharply defined but are gradational into the rock above and below. Pigeonite has been identified only in specimens from the upper part of some sills, but because sampling for petrographic studies has not been adequate, pigeonite cannot be said to be restricted only to the upper parts of sills, as is true for sills in other regions. Enough hints of layering phenomena are known to indicate gravity differentiation has occurred in some sills and to suggest that future careful study of the mineral species and their relative concentration could result in recognition of subtle mineralogic differences in "stratigraphic" zones in the sills and a systematic sequence of layers. Suggestions of gravity differentiation have not been noted in sills less than 500 feet thick. Where suggestions of layering have not been observed, gravity differentiation may not have been effective or the effects of this process may be too subtle for field recognition.

Of the red rock facies the diabase pegmatite is the most abundant. Diabase pegmatite occurs most widely as irregular masses a few inches to a few tens of feet in diameter, and as lenses or crudely lenticular dikes a few inches to a few feet in width. Such masses may be found in any part of a sill, and occur in many areas throughout the region. In some sills, however, they are sparse or seemingly absent. Some roughly tabular masses, a few feet to a few tens of feet thick and a few tens to several hundreds of feet--and exceptionally a few thousands of feet--in outcrop length, roughly parallel and are proximate to sill boundaries, but are within the sills. Volume-wise such masses locally comprise the bulk of the rock of this type, but such masses are relatively few and have been noted only in certain parts of the region.

Characteristic of the pegmatites is an extreme coarseness of grain and an irregular or blotchy aggregation of light and dark minerals. That is, a cluster of mafic minerals may be surrounded by clusters of plagioclase that are separated by but include little mafic material, and vice versa. In a like manner clusters of coarse but stubby crystals of plagioclase may exist adjacent to aggregates of very coarse elongate laths of plagioclase. Plagioclase laths 1/8 to 1/4 inch across and 1/2 to 1 1/2 inches in length are common, and laths of like breadth and 3 1/2 inches in length have been seen. Augite grains 1/2 to 2 inches across are common, and crystals as much as 4 inches in maximum dimension occur. These large augite crystals, which surround plagioclase in subophitic fashion, commonly also include abundant skeletal grains of ilmenite-magnetite.

The diabase pegmatites varies from a dark-colored rock with abundant augite and ilmenite-magnetite through a pinkish-gray rock with abundant dark greenish-gray fibrous hornblende, but little other essential mafic mineral, to a coarse- to very coarse-grained yellowish- or pinkish-gray rock that is 95 percent or more of feldspar. A particular mass may include this entire range or be relatively uniform in mineralogy. The darkest pegmatitic type apparently approximates the mineral composition of the normal diabase that includes no olivine and no hornblende except as a deuteric alteration of the pyroxene. Probably the most abundant variety of diabase pegmatite is composed almost entirely of albite and hornblende. Most of the plagioclase of the pegmatitic facies is turbid, light gray to grayish-orange pink albite. Sphene is a conspicuous accessory in some pegmatites.

Another coarse-grained facies is granophyre. This facies has received little attention because it is rare, and because on weathering it decomposes so readily that representative specimens suitable for microscopic study are difficult to find. Owing to pervasive decomposition, seldom are outcrops adequate to determine the form of the granophyre bodies, but this rock type probably occurs mostly as irregularily tabular bodies a few feet, to a few tens of feet thick and a few tens to a few hundreds of feet in horizontal dimensions. Some pegmatite bodies include small masses of granophyre. More or less equidimensional grains of orange-pink feldspar, intergrown graphically with quartz, give the granophyre a predominantly orangish hue. Euhedral or subhedral quartz grains can ordinarily be observed with the aid of a handlens, and in some occurrences the graphic intergrowths of feldspar and quartz is readily seen in handspecimen. Under the microscope the weathered feldspar available for study is abundantly charged with hematite and limonite and so clouded as to make mineral identification difficult. A large part of the feldspar is albite, but potash feldspar may comprise a part of the rock. The feldspar grains ordinarily range from 1 to 10 millimeters in diameter. Some of the granophyre with a high content of quartz is almost lacking in mafic minerals; other occurrences include 10 to 25 percent of a rusty-weathering, fibrous gray-green amphibole and abundant thin plates of ilmenite-magnetite. Transitional facies between granophyre and normal diabase include augite.

The diabase aplite occurs principally as narrow veins, a fraction of an inch to 6 inches in width, in fine- to medium-grained diabase. These veins are characterized by vague, gradational borders. The transitional zones from aplite to normal diabase range from one-tenth of an inch to a few inches in width. Quite in contrast are rarer dikes of identical composition that are as much as two feet in width and have sharp boundaries with normal diabase. The aplite is insignificant in volume but is a conspicuous facies throughout the region because of the equigranular texture and pink color which contrast with the normal diabase. Aplite is not observed in every outcrop but does occur somewhere in most sills. Where innumerable veins of aplite occur the host diabase commonly is albitic.

The aplites are composed mainly of pink feldspar, which constitutes 80 to 95 percent of the rock. The anhedral equant feldspar grains typically range from 0.05 millimeter to 2 millimeters in diameter; in any given vein or dike the grain size is fairly uniform. Clay-like alteration products cloud and make determination of the feldspar difficult. Nor is it certain that the albite that has been identified in a few thin sections is representative of all occurrences, potash feldspar may exist. Most of the albite is not twinned. Typically the aplite is mottled with concentrations of gray-green hornblende and commonly with yellowish-green epidote; both of these minerals are interstitial to grains of albite. Epidote has been observed with a handlens, but generally is not readily distinguished from the amphibole except under the microscope. Epidote does not occur in all specimens, and probably is not typical of every aplite body. Small miarolitic cavities, lined with hairlike needles of hornblende, occupy rarely the position of the larger hornblende patches. Sphene is sparse but ubiquitous in the few specimens of the aplite veins that have been viewed microscopically, but has not been seen in the sharp-walled dikes. In the narrow transitional borders between aplite veins and typical diabase the equigranular texture of the former appears to be superimposed on the intergranular texture of the latter, resulting in a salt-and-pepper textural appearance. This transitional border looks much like the usual aplite except that the mafic mineral content is higher. In thin section these borders are seen to include abundant chlorite and brown biotite.

A quartzose aplite, which has been seen only in and close to the Sierra Ancha, occurs at or near the margins of diabase sills. Quartz that is micrographically intergrown with feldspar comprises a few percent to more than 50 percent of this aplite. In many though not all specimens minute rounded grains of diopside, seen only under the microscope, are scattered through the rock. Biotite and chlorite have been seen in some thin sections but, as a generality, quartz as a principal constituent and diopside as an accessory are the minerals that set this aplite apart from the ordinary non-quartzose aplite.

The quartzose aplites occur mostly as irregular bodies, crudely lenticular in form and a few inches to several feet thick. These bodies mainly parallel contacts between diabase and the upper member of the Dripping Spring, and in this association may have outcrop lengths of several hundred feet, but ordinarily are much less extensive. Like lenses of lesser extent border contacts between the argillite member of the Mescal and diabase, and a few very small bodies have like relations to the basalt. The quartzose aplites have not been seen in association with host rocks of other lithologies. Some quartz aplite layers form a transitional zone between hornfelsed host rock and other varieties of diabase. Secondarily these aplites occur as dikes, a few inches to 5 feet wide, that are in diabase sills but are not far from the sill border or are in the host formation.

Quartz-free aplite in similar spatial relations to the limestones of the Mescal formation may be predominantly pink but usually is pinkish-gray to greenish-gray. This aplite has a greater content of mafic minerals than the ordinary quartz-free aplite. Only a very few outcrops are known, and the writer has not examined thin sections of the gray aplites; other differences may exist.

With the exception of the sharp boundaried quartz-free aplite dikes, pyrite has been observed in all of the red rock facies. It is sparse to abundant and sporadically distributed in the pegmatites and in most of the aplites. Most of the quartzose aplite includes much pyrite; thus outcrops of this aplite and of some masses of pegmatite are rendered conspicuous by abundant limonite stain. Most of the granophyres also include abundant pyrite, the leaching of which is a factor in the pervasive decomposition of the outcrops.

Fine-grained to aphanitic borders, caused by rapid chilling of the magma against the host rock, are characteristic of the normal diabase. Most of these selvages are fine-grained, rather than aphanitic, and are thin. They are rarely more than 3 feet thick and most are 6 to 18 inches. The effects of chilling can commonly be observed for 15 to 30 feet away from the contact, however, because a slightly finer than average grain size prevails through this interval. Aphanitic borders are mostly confined to dikes and sills less than two feet thick. Commonly only scattered microlites of plagioclase and magnetite can be identified in such borders. Where multiple intrusions exist, the chilled contacts of a later intrusion against an early intrusion are similar in texture and thickness to the selvedges bordering a sedimentary host. This similarity and cross-cutting and inflationary relations of the later diabases indicate that the earlier intrusions had cooled and completely crystallized before the intrusion of the later diabase.

Rarely a part of a sill border does not exhibit chilling against sedimentary rocks. In these occurrences the grain size of the igneous rock--which is quartzose aplite, granophyre, diabase pegmatite, or rarely normal diabase that includes micropegmatite--decreases little or not at all as the contact with the host rock is approached. Nevertheless, except for the gradational boundaries of many quartzose aplites, the contacts between igneous and sedimentary rocks are sharp.

The red rock facies, without doubt, are various aspects of residual differentiates segregated from the diabase magma. Some sharp-boundaried dikes of aplite and relatively sharp-walled veins and lenticular masses of aplite and pegmatite were definitely intrusive and emplaced after consolidation of most of the diabase. Other segregations, entirely surrounded by and gradational into diabase of normal texture, do not suggest by their forms any migration from the sites of residuum accumulation. The spatial relations of the red rock facies indicate that none represent later and independent acidic intrusions from a different source than the diabase magma. Instead, these facies can be visualized as crystallized from accumulations of the volatile-rich residuum derived from the particular intrusion of which they are now a part. In some places only a meshwork of crystals had formed before the feldspathizing residuum accumulated to fill inter-crystal interstices. But for the bulk of each diabase intrusion consolidation was so complete that normal diabase was competent to sustain fractures along which some of the residuum was injected as discrete dikes.

Certain tabular masses of the quartzose varieties of the differentiates are only at the tops of sills, are particularly thick and extensive, and include abundant feldspathic xenoliths. The association seems particularly significant.

Xenoliths have been found only at a few sills, and are numerous in still fewer places. Only a small proportion of the xenoliths exceed one foot in maximum dimension; in most areas where they have been found, none exceed this dimension. Most xenoliths are of host rocks that also are characteristically shattered much adjacent to faults. The xenoliths most numerous are from the potassium-rich upper member of the Dripping Spring quartzite, and unless otherwise stated the following discussion assumes xenoliths of this source. Most xenoliths are found in the border portion of the discordant part of a sill. Where not obviously adjacent to a discordant contact, the xenoliths are at sites not far removed from a major discordancy, and there they are less numerous. Some discordant diabase bodies obviously were intruded along pre-existing faults or sheared zones, and it is not unlikely that most of the larger-scale discordant bodies were emplaced along such lines of weakness. As a corollary, along some faults the host rock had been shattered and along discordant, as contrasted with concordant contacts, was more readily plucked loose and dropped into the diabase magma as xenoliths.

Without significant exception the rock surrounding feldspathic xenoliths is one of the quartz-bearing red rock differentiates. Quartzose aplite is the most prevalent host of xenoliths, and in this association is in tabular masses that border a sill. Apophyses from such masses commonly extend as narrow dikes into the upper member of the Dripping Spring formation. From many such dikes thin sill-like tongues of similar aplite extend along bedding in the enclosing siltstones or arkoses, which in this setting close to diabase ordinarily are recrystallized to fine-grained hornfels. These tongues, as much as 6 inches thick but mostly less than 1 inch thick at their junction with a dike, extend a few inches to several feet away from the dike before wedging out. In some places it is difficult to demonstrate conclusively that these tongues did not inflate the sedimentary rocks. But many aplite tongues merge gradationally into the hornfelsed sediments, and within the aplite minute bedding details of the delicately laminated strata are pseudomorphed, indicating that dilation did not occur. Furthermore, where the dike crosses a bedding unit that includes several aplite tongues, the dike may be somewhat wider than elsewhere, suggesting that it in part was also of metasomatic origin. Within some tabular masses of aplite the relict bedding of the included xenoliths, all of which may have vague boundaries, seems without exception to parallel the bedding of the bordering sedimentary rocks. Such features suggest that the quartzose aplites all could be metasomatic in origin, and also raise some question as to the origin of certain granophyres that similarly enclose relicts of sedimentary rock.

Other quartz aplites tend to partially invalidate the above suggestion. Although inclusions of siltstone or arkose commonly have vague boundaries and most are recrystallized, somewhere in most tabular masses of aplite some xenoliths have sharp boundaries and in some places all are randomly oriented. Furthermore some dikes of aplite do dilate bodies of the coarser differentiates, typical diabase and the host sedimentary rocks. A few examples of xenoliths caught in granophyre are cut cleanly and offset by aplite dikes. Where some inflationary dikes do vary slightly in width as viewed along their length the variations do not seem to reflect the rock type traversed, as is common in dike-like bodies that are replacement phenomena. The quartzose aplite in some part is material formed in place by soaking feldspathic host rocks with magmatic fluids. But in considerable part it must represent a melt formed by reaction between xenoliths and some fraction of the diabase magma, and injected as discrete dikes into a variety of host rocks.

A crudely tabular mass of differentiates exposed on the north wall of the canyon of Reynolds Creek, 2 1/2 miles south-southeast of McFadden Peak (in the SE¼ sec. 7, T. 6 N., R. 13 E., McFadden Peak quadrangle), is particularly instructive in exhibiting relations of menoliths to certain facies of the diabase. Here an extraordinarily thick mass of differentiates, exposed continuously along the strike of a discordant contact for at least 1/2 mile, separates normal diabase from overlying sedimentary formations. The upper part of the differentiate body is comprised of quartzose aplite, ordinarily 30 to 40 feet thick but locally as much as 60 feet thick. In places the top of the aplite layer is against limestones a few feet above the Mescal-Dripping Spring contact; in other places it is against siltstones or arkoses 35 to 40 feet below this contact. Dikes of the aplite occur in the overlying sedimentary rocks and in the underlying differentiate. Downward the aplite layer merges, through a transition zone a few inches to a few feet thick, with a light gray, coarse-grained rock considerably more uniform in texture than the usual coarser-grained differentiates. This rock, not yet examined microscopically by me, might be termed hornblende granite in the field. This granitoid facies, generally 30 to 50 feet thick but locally about 80 feet thick, merges abruptly downward with a coarse-grained gabbroic rock, which texturally is more similar to some of the coarse-grained diabase sills than to any of the red rock facies. This gabbroic rock, through an interval of 15 to 20 feet below the contact with the granite, grades downward into ordinary medium-grained diabase. The contour of the canyon wall about parallels the strike of the discordant contact, which dips away from the canyon. Topographic reentrants into the mountain face expose down-dip parts of the intrusive contact where it flattens and becomes virtually concordant with the overlying sedimentary rocks. Where the contact cuts across the bedding of the overlying host at the greatest angle the zone of layered differentiates is thickest; as the angle of discordance decreases down-dip the zone thins; and where the contact is concordant in a few small exposures, normal diabase is in contact with the sedimentary formations.

The xenoliths, which are siltstone and arkose fragments from the Dripping Spring formation, have different aspects in different parts of the layered differentiate. Small angular xenoliths, heterogeneously oriented, are abundant generally through the aplite and the transition zone to the granite. Many are so thoroughly reconstituted that they are difficult to distinguish from their aplitic matrix. Some xenoliths as much as 1 foot long are recognized only by faint relicts of bedding structures. Two small blocks of fine-grained diabase, which suggest that a chilled selvedge of normal diabase once bordered the discordant contact, were seen at one place a few feet below the top of the aplite. Xenoliths are sparse through most of the granite layer; those that exist above the basal fringe of the granite ordinarily range from 6 inches to 2 feet in largest dimensions and are randomly oriented. Downward, in the basal 15 to 20 feet of the granitic layer, however, inclusions increase in numbers and many are large--up to 15 feet in length and 1 to 4 feet thick. The borders of some of these large xenoliths are vague and a narrow reaction zone, 1 to 6 inches thick, that borders some of them is texturally like some of the highly feldspathic diabase pegmatites. Many of the inclusions, though much recrystallized, exhibit fairly sharp borders against the granitoid host. The xenoliths are largest and greatest numbers in the basal 4 to 8 feet of the granite. Striking--owing to the random attitudes of the higher xenoliths--is the strong tendency of these plate-like inclusions to be oriented parallel to the bottom of the granite layer, as though they had settled to a floor. Xenoliths are rare in the underlying gabbroic facies. The few seen were small and exhibited narrow rims of pegmatitic texture.

In viewing this occurrence, the impression cannot be avoided that feldspathic fragments settled differentially, perhaps partly as a function of size (Lovering, 1958) and partly as a function of impedance offered by different strata of the crystallizing magma through which they settled. And in contaminating the magma, which at this site was probably somewhat enriched by late-stage residual material, the existence of xenoliths in different sizes, numbers, and surface areas in different parts of the crystallizing magma must have had a considerable effect on the characteristics of the rock that finally crystallized.

Here and elsewhere, particularly thick and extensive masses of the red rock facies are spatially associated with xenoliths. This particular mass, in which xenoliths are more abundant than in any other example seen, is the thickest single unit yet noted. Also, the granitic facies of this body is unusual, suggesting that a different degree of modification, as well as greater volume of change, was affected here. An outcrop about a mile to the southeast, in a like setting and probably along the same discordant contact, is more representative of the differentiates that visibly include xenoliths. There a layer of granophyre, not generally as thick as the granite above described, is separated from the host rock by only a few feet of quartzose aplite. The aplite includes appreciable numbers of xenoliths, but they are sparse if not missing in the granophyre.

Xenoliths seem to characterize the quartzose aplites, though they have not been noted in all bodies of the coarser-grained quartz-bearing differentiates. Conversely xenoliths that exhibit indications of assimilation have been seen only in the quartz-bearing varieties of the red rock facies. The only exceptions, which apparently are insignificant, are examples such as the small inclusions seen in the gabbro at the base of the differentiate section along Reynolds Creek; even these are proximate to quartzose facies. Additionally suggestive of localization owing to contamination, quartz-bearing facies have not been found except in proximity to highly feldspathic host rocks. Rare xenoliths of limestone or quartzite are simply included in the typical diabase; such inclusions are metamorphosed; but are sharp-walled and show no sign of assimilation. Large bodies of the quartzose aplites and the granophyres are adjacent only to feldspathic strata of the upper member of the Dripping Spring. Empirically, they might be expected also adjacent to the arkoses of the lower member or of the Pioneer formation, but these units are not hosts to the thick sills seemingly requisite for the formation of the red rock facies. Lesser bodies of the quartzose facies do exist along the walls of sills intruded into the potassium-rich argillite member of the Mescal and into the basalt flows.

The quartz-bearing facies, through conspicuous, comprise only a small part of the red-rock facies. Indeed, examples are comparatively so scarce that the quartzose varieties should be considered only minor varieties and atypical. And possibly a mode of origin applicable to these cannot be ascribed to the varieties that exist in greater amounts.

Common in many sills are the small bodies of pegmatite and the narrow dikes and gradationally bordered veins of non-quartzose aplite, all of which may be found in any part of a sill. Most abundant, and in aggregate possibly the most voluminous of all, are the narrow albite veinlets that rib the weathered surfaces of many diabase outcrops. These albite ribs are merely small-scale manifestations of the thicker veins and dikes of quartz-free aplite. The dark-colored hornblende veinlets of similar habit must be related to the albite veinlets in origin.

Many of the small randomly distributed pegmatite bodies are narrow and elongate. With the dikes, veins and veinlets of aplitic material, these bodies obviously were emplaced along joints, fractures, and faults opened after the diabase had crystallized enough to sustain fractures. Apparently volatile-rich material, residual when consolidation was almost complete, migrated to fill and crystallize in and near the fracture openings. The form and distribution of the small bodies of pegmatite and aplite indicates that they could not have been products of magma contamination, as speculated for the quartzose differentiates.

The thick extensive bodies of diabase pegmatite are found only in the upper parts of thick sills. These bodies also are not numerous, though they are much more common than the similarily disposed masses of the quartzose facies. Some tabular bodies as much as 20 feet thick border the upper contact of a sill, and normal diabase does not intervene between the pegmatite and the host rock. In other localities, a few feet to as much as 80 feet of ordinary diabase intervenes between the top of the sill and a thick body of pegmatite. In a few places two thick layers enclosed in normal diabase have been seen in the upper 60 to 90 feet of a sill. Where a zone or zones of pegmatite lie some distance below the top of a sill small irregular lenses of pegmatite may be abundant in the diabase that separates layers of pegmatite or separates a layer of pegmatite and the top of the sill. These lenses lie roughly parallel to the layers. The diabase that includes these lenses and that of the interval several tens of feet below the lowest extensive pegmatite mass ordinarily exhibits much deuteric alteration. Some pegmatite layers are fairly uniform in thickness along several hundred feet of outcrop; others pinch and swell or lense out locally. The borders of some layers are interrupted by subparallel apophyses, which are separated from the main mass by thin wedges of normal diabase.

Xenoliths have not been noted in the pegmatite facies. In a few places quartzose aplite layers intervene between a pegmatite layer and the top of a sill, and some pegmatite layers merge vertically or laterally with bodies of granophyre. Even in these occurrences, xenoliths have been noted only in the quartzose facies. In view of this lack, and in view of the forms and distribution of the smaller bodies and the association with deuterically altered diabase, the larger masses of diabase pegmatite are probably best visualized as segregations of residual melt concentrated in the upper parts of sills during the late stages of magma crystallization. Further, it might be suggested that the quartzose facies where formed only where partially crystallized diabase magma, in which the residuum was accumulating, was seeded by feldspathic xenoliths.

The quartzose aplite bodies invariably border the tops of sills; granophyre and related coarse-grained bodies, and the thickest and only widely persistent bodies of diabase pegmatite occur at or near the tops of sills. Not all sills include these sizeable masses, however, nor are they prevalent everywhere at the tops of sills in which they do occur. Where such masses exist the tops of sills tend to be of albitic diabase; elsewhere the upper parts of sills may or may not be albitic. The accumulation of residual fluids at certain sites seemingly was accompanied by the albitization of the normal-textured diabase in adjacent areas. Most of the larger masses of the differentiates are along or not far removed from a major discordant contact. In many places thin discontinuous dikes of pegmatite and dikes and veins of quartz-free aplite are numerous along the border of a steep discordancy. The mere generalization that particular concentrations of the red rock facies occur at the tops of sills is only partially definitive for the diabase province of southern Arizona. The writer would suggest that these concentrations are most prevalent near large feeder dikes to the thicker sills and along the tabular discordant connections between the concordant portions of thick sills which were intruded at different stratigraphic levels; the larger concentrations are near the tops of sills, but in addition are largely restricted to the vicinity of discordancies. Even in such structural associations, the red rock facies may be missing.

Although albite diabase is widespread at the tops of some sills, and is common in association with the red rock facies near discordancies, it also seems to occur outside of these geometric associations. Some thin sills, 250 feet or less in thickness, which geometrically are apophyses from the olivine-bearing diabase portions of thicker sills, appear to be entirely of albite diabase through areas of at least a few square miles.

There is a regional variation in the distribution of the red rock differentiates and coarse-grained normal diabase. The only known extensive bodies of the coarse differentiates and of the quartzose aplite are in the thickest sills of the part of the Sierra Ancha that lies within the McFadden Peak quadrangle and in those sills to the northeast in the northern one-third of the quadrangle. None have been seen or reported in the Sierra Ancha west of the 111° meridian, but they can be anticipated in the thick sills of that part of the range. In many sills throughout the region small masses of pegmatite and dikes of non-quartzose aplite are abundant, but in some sills--even some particularly thick examples--these masses are singularly sparse or absent. Although veins and dikes of aplite are probably as abundant as in any locality elsewhere, in the Salt River Canyon-Chrysotile area very few examples of the coarse differentiates have been seen. The diabase of many of the sills that are virtually devoid of the differentiates is apparently somewhat coarser than that of the average sill. In the Sierra Ancha, in contrast, sills that are coarse-grained throughout are atypical. Perhaps a lack of differentiates in the coarser-grained sills has some significance yet to be determined. Walker (1953, p. 54) has noted that the distribution of diabase pegmatites seems to bear a similar relationship to the grain size of the typical diabases in some sills of Virginia and England, and has suggested "that failure to segregate the volatile constituents and late minerals may cause a general instead of a local increase in grain-size."

The differentiates of the southeastern Arizona diabases are those said to be characteristic worldwide of the tholeiitic magma type (Kennedy, 1933, p. 242-247). But the low silica contents (43-49 percent) of the normal diabases that have been analyzed suggest that they should be considered as the olivine-diabase type. Until more analyses are available for comparison, however, probably no attempt should be made to classify the diabases of the Arizona province by magma type.

Metamorphism associated with diabase

The Apache and Troy strata, except for thermal metamorphic effects induced at the time of intrusion of the diabases, are metamorphosed no more than the Paleozoic and other younger formations that overlie them. Around centers of late Mesozoic or Tertiary plutonic activity, and especially in the large copper-producing districts, the younger Precambrian and superimposed formations are locally metamorphosed. Considering only the younger Precambrian rocks, most notably some of the finer-grained formations of the Apache group, such as the Pioneer shale, were rendered schistose adjacent to some granitoid stocks, and in some areas considerable volumes of diabase were hydrothermally altered. The younger Precambrian rocks very locally were modified so they do not display exactly the petrographic characteristics described in this report. Such modifications are very restricted in terms of outcrop area, however, and will not be considered further. On the other hand, because diabase intrusions are virtually ubiquitous in Apache strata, the effects of thermal metamorphism adjacent to the diabase intrusions are of regional scale. These effects are not discussed in detail here. Rather, only enough description is given so that certain metamorphic features, commonly noted in the outcrops, can be appreciated as effects associated with the diabase intrusions and recognized as not of other associations.

The lithologically diverse formations intruded by diabase are metamorphosed for different distances from the intrusions. The carbonate rocks were most susceptible, were modified to the greatest degree and for the greatest distance from the diabase; the argillaceous and highly feldspathic rocks were modified to a much lesser degree and extent; and the highly quartzose rocks show little reconstitution. The mineralogic changes are largely those considered typical of the process termed "contact metamorphism." The contact-metamorphic minerals do not exist in the massive concentrations that occur near igneous intrusions in many examples of contact metamorphism. The effects are much more widespread, however, than those of thermal metamorphism ordinarily attributed to diabase intrusion.

The changes were brought about largely through reconstitution, induced by heat from the diabase, of the original rock-forming minerals. The heat was probably carried largely by solutions emanating from the diabase; but in the carbonate rocks carbon dioxide, released on the dissociation of dolomite, was probably an important transporting agent. On the intrusion of diabase the dolomites were fractured extensively but on a very small scale. The formation of carbon dioxide and this fracturing, which is more pervasive in the carbonate rocks than in the other formations, probably accounts for the wider distribution of metamorphic products in the carbonate rocks than in other rock types. Some addition of constituents from the diabase certainly occurred, but whether or not the additive process had a large role in the metamorphism is, at present, uncertain. The metamorphic minerals are largely those that would result from recombination of the original constituents of the rocks.

The internal alteration, technically termed endomorphism, of the diabase intrusions was an essential part of the process that resulted in contact metamorphism (exomorphism) of the host rocks. Stated another way, late magmatic or deuteric residual solutions on passage through the diabase caused uralitization, saussuritization, and albitization of earlier formed minerals in the diabase, and on exit from the diabase these or related emanations had a part in the reconstitution of the host rocks. The effects of endomorphism of the diabase have been described under petrography of the diabase and will not be considered further.

The quartsose rocks that included little argillic or feldspathic material were the rocks least susceptible to reconstitution by emanations from the diabase. The quartzitic portions of the Troy rarely exhibit effects of metamorphism; though very locally and immediately adjacent to a diabase intrusion the quartzite member was rendered more dense and vitreous. Parts of the arkose member of the Troy that border diabase intrusions were in some areas indurated to a quartzitic rock. Rare wedges of this sandstone more or less engulfed in diabase were recrystallized through thicknesses of as much as 100 feet, the original bedding structures are largely obscured, and in isolated outcrops the quartzitic product can be distinguished from the lower (quartzitic arkose) member of the Dripping Spring only with difficulty. The lower member of the Dripping Spring was a well indurated massive unit prior to diabase intrusion, and was rarely intruded by diabase. Therefore there are few opportunities to examine metamorphic products in this unit. In a few places thoroughly sheared zones of this arkose were invaded by diabase. In such places the feldspars were thoroughly recrystallized, resulting in a hornfels in which discrete grains of quartz appear to be embedded in a dense megascopically textureless matrix of potash feldspar. Parts of such hornfels do include concentrations of pyrite.

The upper member of the Dripping Spring was more widely inflated by sills, and therefore shows more widespread effects of emanations from the diabase than the coarser clastic rocks. The siltstones and fine-grained arkoses of this member locally were recrystallized to hornfels. The rock that is coarsened in texture to a degree obvious in hand specimens is generally restricted to zones a few inches to a few feet wide that border the sills. In a few places, where the upper member of the Dripping Spring was intricately inflated by sills or split by a thick sill, sections of the siltstone as much as 100 feet thick are obviously coarsened in texture and include abundant light-colored aplite-like segregations, which are merely another manifestation of the metasomatic effects considered in the previous section, in the discussion of the quartzose aplites. These coarser segregations are thin layers or lenses that crudely parallel bedding or are narrow vein-like zones or irregular masses that transect bedding. All are gradational into a very fine-grained hornfels, megascopically little different in appearance from the original siltstone or arkose. The coarser hornfels is not an abundant metamorphic product, and is found only in relatively few places adjacent to some of the thicker sills, particularly along or near discordant portions of these intrusions. In comparison with the unaltered rock, most of the strata that are proximate to diabase intrusions are better indurated, and on weathering are more abundantly stained by limonite. Otherwise it is generally not readily discernible, except by microscope study, that this rock has been recrystallized. In general, the potash feldspar is coarsened in texture, and in places is replaced by albite. Commonly, in the coarser aplitic-appearing rocks quartz and feldspar exist as micropegmatitic intergrowths; the mafics of such rocks include minute anhedral, clear pyroxene grains and an amphibole. Biotite may be common. In some examples sphene is an abundant accessory. In freshly broken rock sulfides may be conspicuous.

Generally the pre-Apache rocks exhibit little modification adjacent to diabase intrusions. Only in the quartz monzonites presumed to be equivalent to the Ruin granite of the Globe-Miami area have these effects been observed. Weathered outcrops of most of the "granite" that is adjacent to diabase seem to be more abundantly stained by limonite than outcrops distant from diabase, but textural modifications are generally lacking. But where diabase was intruded into units of the reworked granitic debris that underlies the Scanlan conglomerate or inflated zones of thoroughly sheared "granite" a variety of striking metamorphic effects may exist. None of these have been studied in thin section. In one manifestation of the metamorphism relicts of the abundant large microcline phenocrysts of the coarse-grained "granite" are well rounded, and these egg-shaped nodules are in a fine-grained olive gray or dark greenish-gray matrix of feldspar, quartz, dark micas and other dark minerals not megascopically identifiable. In gross effect this rock suggests an "orbicular" diabase. In other occurrences relicts of the microcline phenocrysts are absent, but rounded blebs of bluish-gray opalescent quartz exist in a similar matrix. The quartz blebs, as much as 1 centimeter in diameter, are abundant and in some outcrops they coalesce in masses of very irregular outline. Varieties between these two metamorphic rock types exist. In some little reconstituted examples, which were derived from regolithic material, broken fragments of the feldspar phenocrysts remain angular, but the quartz grains are rounded. Other recrystallized regoliths, representing an arkose in which the constituents other than quartz and feldspar had been winnowed away, were reconstituted to an orange-pink rock in which abundant coarse grains of quartz lie in an almost textureless matrix of feldspar. The scarcity of all such metamorphic rocks should be emphasized; few examples are well exposed. Some that appear most like the diabase may actually be large xenoliths of sheared granite surrounded by diabase.

The tuffaceous mudstones and siltstones of the Pioneer formation generally show effects of metamorphism through intervals a few feet thick adjacent to thin sills, but where the Pioneer is intruded by thick sills it may be recrystallized to some degree almost throughout. The dusky red, finely laminated strata were metamorphosed to hard dense hornfels, in which the small-scale bedding structures are largely obliterated. Where best developed the hornfels is grayish orange to grayish green, and is mottled throughout with small gray spots, as large as 3 millimeters but typically less than 1 millimeter in diameter, which are segregations of finely divided dark minerals. The ferric iron of the original rock may be only partially reduced, therefore many outcrops of the hornfels retain a pinkish or reddish tinge. If the rock was bleached in the slightest as a conse uence of metamorphism the outlines of the shard ~~shared~~ relicts were obliterated.

With the exception of the carbonate members of the Mescal, the argillite of the upper member of the Mescal and the basalt were the lithologic types of the Apache group most susceptible to metamorphism. Adjacent to thick sills all of the argillite was indurated to a very dense, novaculitic rock, which ranges from reddish orange to almost black and includes abundant conspicuous metacrysts or knots representing aggregates of metamorphic minerals. In some beds the knots are elliptical, as much as 3 millimeters in diameter, and are largely aggregates of fine-grained dark-colored mica. Other beds are irregularly mottled with light brown to dark yellowish-brown aggregates of a very fine-grained mineral (or minerals), which do not exhibit cleavage. A third prominent modification, commonly reddish-orange, is thoroughly flecked with lath-like metacrysts, 1 to 10 millimeters in length, of greenish-gray amphibole. Parting planes of this rock resemble surfaces abundantly impressed with miniature bird tracks. Other, less common, variants of the hornfels exist. The Apache basalt was thoroughly albitized and locally veined by epidote adjacent to some sills. Where recrystallized the basalt texturally resembles the diabase aplite, but is mottled in color from greenish gray to reddish orange. The grayish mottles reflect concentrations of mafic minerals--mostly chlorite, biotite, and fibrous amphiboles(?); the orange mottles are concentrations of very clouded feldspar--probably mostly albite--abundantly dusted with hematite. In a few localities both the argillite and the basalt were converted to a rock which cannot be distinguished in hand specimen from the diabase aplites. This aplitic material is in the form of irregular masses, a few inches to 30 feet across, which gradationally interfinger with less altered rock along bedding in the argillite and planes of fracture in the basalt. This metamorphic product borders a diabase intrusion and the contact between the diabase and the host rock can be difficult to define.

The aureole of contact metamorphism is widest and mineral transformations most obvious in the carbonate members of the Mescal formation. The original dolomite and cherty dolomite beds were dedolomitized and converted to calcitic limestone that includes small masses of silicate minerals, as noted earlier. Where wide dikelike discordant portions of diabase sills cut across the carbonate strata but concordant sills do not inflate the section, zones as much as 2,000 feet in width, in the direction of bedding on both sides of the intrusion, were dedolomitized. Metamorphic halos a few tens of feet to a few hundreds of feet in width border dikes that are 25 feet or less in width and isolated from other diabase intrusions. In those localities in which a concordant diabase sill, 50 feet or more in thickness, inflates the carbonate section, the entire section is generally dedolomitized. The thinner-bedded lower member of the Mescal tends to be more thoroughly converted to limestone than the massive algal member. If the dolomites did not include extraordinary amounts of secondary chert, relicts of chert are not abundant, and in some localities are almost nonexistent except in the limestone beds most remote from the diabase intrusion. Where chert was particularly abundant, as in the thoroughly silicified beds common below the unconformity at the top of the algal member or in those localities where silicification of the lower member was widespread as an adjunct to leaching and the formation of sinkholes, the chert was only partially converted to silicate minerals.

The dolomitic rocks yielded a variety of products according to their original composition and the stage of metamorphism. The cherty dolomites were most commonly reconstituted to mixtures of tremolite, diopside, talc, serpentine, and calcite which occur in a host rock of fine-grained, sugary-textured calcite limestone. The dolomites essentially free of chert were altered to limestones that include little of the silicate minerals. Most of the metamorphic products are very fine-grained; and individual grains of the silicate minerals generally can be distinguished only under a microscope. Some mixtures of silicates are so fine-grained as to be difficultly resolved microscopically; most individual grains are so small that optical properties are difficult to determine. Diopside occurs as fine anhedral grains; these grains may exist as sparse individuals or closely aggregated grains in a matrix of calcite or other silicates. Tremolite ordinarily occurs as very fine needles or radiating aggregates of needles in similar disposition. In some specimens aggregates of fine-grained talc are associated with diopside and tremolite; so few examples of this mineral have been recognized that its paragenetic relations to the other silicates are poorly understood. Serpentine is far the most abundant silicate in the limestone and tends to obscure the interrelations of the earlier-formed minerals calcite, diopside, tremolite, and talc, which it replaced in some instances and veined and enveloped in other instances.

Diopside, tremolite and talc occur as aggregates in the form of irregular layers, lenses and nodules that reflect in distribution the original concentrations of chert. The form of the chert concentrations are rather faithfully reproduced in these aggregates, but there is some suggestion that the dimensions of the silicate masses are very slightly larger than those of the original chert masses. Where chert was originally very abundant, some aggregates of the early silicates include cores of recrystallized chert. In all thin sections that have been studied the details of relations between these silicates and the chert have been obscured, however, by the ubiquitous serpentine that envelopes these aggregates.

In hand specimens, masses dominantly of diopside are light gray to very pale orange; those dominantly of tremolite lack the orangish tint and range from light gray to greenish gray. Masses of both minerals are tough and difficult to break; those mainly of diopside are difficult to scratch, whereas those of tremolite are readily scratched or chipped by a steel blade. Weathering may bring out the fibrous or radiating structure of the tremolite aggregates; the diopside weathers to a smooth surface. The silicate minerals weather whitish, and are much less conspicuous than the chert, which etches out boldly in the dolomites. Commonly those unfamiliar with the Mescal formation fail to appreciate that the limestones everywhere include large content of silicate minerals.

The serpentine mineral most abundant in the Mescal limestones occurs in dense masses, is ordinarily soft, exhibits a waxy or greasy luster on freshly exposed surfaces, and occurs in a wide range of colors. The colors range from yellowish gray to brownish black in the brown hues, and grayish yellow green to dusky blue green in the green hues, to black. Pale green, grayish green, and grayish olive are common colors. Impurities of other silicates, calcite, and magnetite commonly cause variations in luster, color, and hardness. A late-formed serpentine, ordinarily grayish-orange pink but in rare exposures pale green, veins the earlier serpentine of some localities. Both of these serpentines are fibrous, but the fibers are too small to be resolved under the optical microscope. Selected specimens studied by optical, differential thermal analysis, electron microscope and x-ray diffraction methods have been identified as chrysotile (E. J. Young, written communication). Probably all of the dense serpentine is of this species. No antigorite has been identified in the limestones. Some serpentine layers include veins of megascopically fibrous chrysotile. In certain structurally favorable geologic settings (Shride, 1952) these veins are numerous, and comprise the asbestos deposits that have been prospected or mined at more than 150 sites in Gila County.

The ordinary serpentine occurs in zones, generally a fraction of an inch to 3 feet in thickness, that mostly parallel bedding in the limestone; it may occur in paper-thin wispy, discontinuous layers or in tabular zones locally as much as 8 feet thick, or concentrations may range from microscopic blebs to elliptical nodules 3 feet in maximum diameter. Most of the serpentine grossly pseudomorphs in form and distribution the original chert zones or the chert nodules. Appreciable amounts of serpentine, however, replaced the rock adjacent to joints, fractures, and faults without regard to the original concentrations of silica; and in some localities serpentine has replaced diabase adjacent to similar structural features. Serpentine is the most widely distributed of the silicate minerals and may be found in the parts of the dedolomitized carbonate formations most distant from diabase intrusions, whereas the early-formed silicates have not been recognized in such distant limestones.

The above described silicate minerals have been given particular attention because they occur in the limestones that include asbestos deposits. Other silicates, also mostly in very fine-grained aggregates, that exist in other limestones have been given little attention and most remain to be identified. The silty, and sandy dolomite matrix of the basal breccia of the Mescal is ordinarily converted to a dark greenish gray, soft rock, which apparently is a very fine-grained aggregate of serpentine, chlorite(?) and calcite. In some areas much argillic material was concentrated in the algal member during silicification; on metamorphism this clayey and siliceous dolomite yielded products not common in the rest of the formation. Gray green micaceous minerals, ordinarily in thick hexangle books as much as 5 millimeters in diameter, may abundantly spot this rock and places constitute as much as one-third of the rock. These metacrysts are in a matrix of calcite, which includes serpentine and other silicate minerals too fine-grained for microscopic identification. The coarser micaceous aggregates are of at least two mineral species: one apparently is a chlorite; others remain to be identified. Locally beds of apparently similar original lithology are now composed largely of grayish-yellow green to greenish-grey tremolite, which occurs as randomly oriented blades, as much as 1/4 inch wide and 3 inches long. Rarely, similar tremolite occurs as irregular veins of wood-like cross-fibers. Very small amounts of other silicates not general through the region occur in a few localities. An example is a dark yellowish-brown garnet, which has been noted in sparse amounts in only three localities.

Magnetite, although present in many localities, is not a prevalent contact-metamorphic mineral in the Mescal limestone, as it is in many areas where carbonate rocks are intruded by diabase. Locally magnetite was sparsely to abundantly disseminated through 20- to 100-foot thicknesses of limestone; a few such occurrences have an areal extent of several acres. The coarse tremolite or the micaceous metacrysts noted above may be abundant in these occurrences. Few magnetite bodies of this sort are known. Much more numerous are occurrences in which magnetite replaced fractured limestones within a few feet of a discordant diabase intrusion. Some of this magnetite is disseminated in the limestone but most of it exists as narrow veins or small pods. Rare veins, 6 inches to 3 feet thick, are several tens to a few hundreds of feet long. Generally magnetite is lacking, even in the limestones immediately adjacent to discordant diabase intrusions.

Thermal metamorphism associated with the diabase intrusions is general throughout the region. If this is not appreciated, some of the metamorphic effects may be attributed to local thermal metamorphism of post-Precambrian date. Where the effects are inconspicuous, especially in weathered outcrops, the original lithology may be misinterpreted or go unrecognized—as generally has been the case of the carbonate members of the Mescal. Except where rare small outcrops are isolated from contiguous less modified parts of a formation, the metamorphic effects are not such that a formational unit is likely to be misidentified.

Late Precambrian deformation

It has not been recognized previously that the Apache group and the Troy quartzite were folded and faulted prior to intrusion of the Precambrian diabases. The effects of this episode of deformation are overshadowed throughout the region by the dilationary effects that accompanied the intrusion of diabase, and in the southern part of the region are further obscured by deformation of post-Paleozoic age. Details of pre-diabase structures have been studied, to date, only in the portion of the region structurally a part of the Colorado Plateau. Ransome and many who succeeded him recognized the deformation, mainly expressed in high-angle faults and vertical displacement of blocks of strata, that occurred as a consequence of dilation by diabase sills. Generally, Paleozoic and younger formations were inferred, however, to have been similarly affected by this deformation. But if the intrusions are of Precambrian age, as accepted herein, such inferences are fallacious. This conclusion is amply demonstrated in many places by the geometric relations of Paleozoic formations to the diabase intrusions and to the structures associated with them. These relations are described in the next section of this report. The following descriptions of the structures of younger Precambrian age provides a background for that discussion.

Pre-diabase deformation

The first deformation, of several that affected the structural disposition of younger Precambrian rocks as seen today, was the subtle upwarping of the Mescal limestone and older formations before the Troy was deposited. To the best of present information, no faulting accompanied this folding and for practical purposes the pre-Troy formations remained virtually horizontal. Therefore this mild deformation, considered on earlier pages, will not be discussed further.

After the Troy was lithified, but before diabase intrusion, Troy and Apache strata were strongly deformed along widely spaced, narrow belts; between these belts the formations remained horizontal. The structural belts observed in northern Gila County are characterized by both folds and faults. None positively determined to pre-date the diabase are less than 5 miles in length, and some have been traced 20 to 40 miles before being hidden beneath a cover of rocks that postdate the structures. Some of the belts are sinuous in plan, but all are grossly of north or north-northwest trend. Certain of these were loci along which erosion has been particularly effective. Thus they ultimately defined the geographic position of some of the major elements of the drainage pattern tributary to the Salt River; the canyons of Cherry Creek and Canyon Creek are examples.

The somewhat sinuous structural belt followed by Cherry Creek along its course through the McFadden Peak quadrangle includes some features not seen in other belts, but also includes all features typically associated with these belts. A single fault or in places a zone, 1,000 to 5,000 feet in width, of subparallel faults has been traced along the west wall of the canyon of Cherry Creek, from a locality two miles south of the quadrangle to a locality two miles north of it. Probably the structural belt extends much farther to the north and south. South of the quadrangle and north within it, along a length of 10 miles, several faults, individuals of a faulted zone almost a mile in width, strike N. 5°-10°W. Through the central part of the quadrangle, for a distance about 8 miles, the zone of faults is very narrow or is represented by one fault that strikes N. 30° W. Along the northernmost 3 miles within the quadrangle and where observed north of the quadrangle a single fault, inflated by a wide and steeply discordant dikelike apophysis from a diabase sill, strikes N. 10° E. Thus, along a 23-mile length, major segments of this fault belt differ in strike by as much as 40 degrees.

Pre-diabase fault displacements within the Cherry Creek structural belt are noteworthy, but this faulting did not result in displacements noticeable on or either side of the belt. Stratigraphic horizons west of the canyon are, if one ignores local displacements caused by diabase inflation, at about the same altitude as equivalent horizons in the plateau east of the canyon. Away from the north-northwest trending belt on either side of the canyon the formations are virtually horizontal. At places along the length of and within the belt, however, stratigraphic displacement caused by a combination of faulting and folding locally exceeds 1,500 feet. The Troy quartzite in places, therefore, is in juxtaposition with the pre-Apache granite.

East of Cherry Creek, in a zone about a mile wide and persistent the observed length of the structural belt, the formations were folded into a monocline of westward dip. The transition from the horizontal attitude of the strata east of the monocline to the 3° W. to 10° W. dips characteristic of most of the monocline is gradual. As Cherry Creek is closely approached, however, the dip of the strata increases rapidly and where the monocline terminates against the fault or narrow zone of faults immediately west of Cherry Creek, dips of 20° to 30° are not uncommon; in a few places beds near the fault are almost vertical in attitude. The above description should be modified for the northernmost and southernmost parts of the McFadden Peak quadrangle, in that the monoclinal flexure is almost entirely on the west side of Cherry Creek in those areas. Along parts of the Canyon a part of the displacement downward to the west is caused by parallel or sub-parallel normal or reverse faults, of north or north-northwest strike, as well as by monoclinal folding. In other words, steep northerly-trending faults step narrow blocks of strata successively downward to the west in some localities. Thus in places along its length the eastern two-thirds of the Cherry Creek structural belt suggests one-half of a graben.

The west border of the monocline is approximately defined by the fault that extends the length of the Cherry Creek structural belt, or where the belt includes a zone of several faults the principal fault of the zone--ordinarily one near the west border of the zone--can be considered the defining fault. The principal fault and the subsidiary faults of the faulted zones are intricately intruded by dikes or dikelike discordant portions of sills along most of the length of the belt. Therefore the monocline characteristic of the eastern part of the structural belt generally terminates against a dikelike body of diabase rather than a fault.

Immediately west of the principal fault of the structural belt sedimentary formations in general dip steeply west along the entire length of the belt. Along considerable lengths of this fold the formations are vertical or nearly vertical in attitude; in many places the strata that are within a few hundred feet of the fault are even overturned steeply to the east, and in a few places the overturned formations exhibit an easterly dip of less than 50°. This intense folding affects only a narrow zone; 500 to 2,000 feet west of the principal fault (the east boundary of the fold) the formations are virtually horizontal and continue so westward into the Sierra Ancha. The Troy quartzite, the Apache group and the granite that underlies the Apache group are involved in this fold. As some of the formations are very competent and as the stratigraphic thickness (about 3,000 feet exposed) of this sequence was too great a thickness to be accommodated in such a tight fold, the overall thickness of the folded section is thinned by shearing in the narrow belt of intense folding. In places parts of formations are missing, and in a few places parts are duplicated along a shear zone that parallels in gross attitude the attitude of the steep fault that bounds the fold on the east. Only parts of this belt have been mapped; pending additional information, no interpretation of the origin of the overturned fold will be attempted.

Other north to north-northwest structural belts are similar, except in details, to that of Cherry Creek. Few are as wide or show as intense folding as the Cherry Creek belt. In general the belts are separated by intervals of several miles; within these intervals the formations, though locally displaced by high-angle faults, are practically horizontal. Reconnaissance along Canyon Creek indicates that this stream follows a north-northwest belt, similar to that in Cherry Creek, from the northern border of the Blue House Mountain quadrangle to the junction with Salt River, a distance of 13 miles; southward this belt can be traced an additional 19 miles before being covered by Cenozoic gravels that fill the trough occupied by Sevenmile Wash east of U. S. Highway 60. South of the Salt River, minor fault displacements apparently of Cenozoic age were noted locally, but there can be no doubt that the principal deformation along the 32-mile length noted was prediabase. A seemingly contiguous zone of faults extends along Canyon Creek at least 10 miles north of the Blue House Mountain quadrangle. Lacking knowledge of the structural geology of this area, the writer is not certain that the faults of this interval are mainly Precambrian in age. Another narrow belt, of north-northeast trend, has been traced from roughly the southwest corner of the McFadden Peak quadrangle 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(?), 300 to 1,000 feet in width. Narrow belts of sharply folded strata border the dike. South of McFadden Peak the position of the fault is marked by the discordant portion of a thick diabase sill. Casual knowledge, gained by crossing similar structural features at widely separated intervals, suggests that other like north- or north-northwest trending structural belts of considerable length probably exist east and west of the area encompassed by the McFadden Peak and Blue House Mountain quadrangles.

The narrow but persistent zones of intensely folded strata that border the described structural belts, and others of lesser length, suggest drag adjacent to major faults. But these zones are so intimately associated with dike-like bodies of diabase that some might visualize them as the result of forceful injection of diabase magma. Except for minor superimposed details, it is wholly unlikely that the folds can be explained in this way. Although open folds and associated bedding and thrust fractures are common adjacent to discordant diabase intrusions throughout the region, these features are of small scale and local distribution and in no way comparable to the folds and sheared zones observed on a regional scale. The usual tendency toward concordant intrusions, the great lateral extent of many thin sills, the general lack--except very locally--of distortion of even the thinnest layers of incompetent sedimentary rocks partly or completely enclosed in diabase, and the lack of brecciation of the host sediments along the borders of intrusions refute the possibility that the magma was a viscous mass, which punched its way along faults or other lines of weakness and aggressively folded the wall rocks away from the plane of intrusion. Rather the diabase magma must have been a very fluid material, which insinuated its way along splitting planes or fractures. This deduction, which is only approach possible for most occurrences, is supported by direct evidence. In a few localities within the structural belts diabase intrusions do not exist at the outcrop; and the complete lack of thermal metamorphism, especially in rock types particularly susceptible to reconstitution, indicates that subterranean bodies of diabase are spatially remote if present at all. Thus the major faults, the folds and their associated shear zones, which characterize the regional structural belts of northerly trends, existed prior to the intrusion of diabase.

In or adjacent to the belts other features, which do reflect inflation by diabase intrusions, are corollary to this conclusion. Along a part of the belt the sharp folds reflect the drag that would be expected from the vertical displacement along the principal fault of the belt. The direction noted along other lengths is opposite that expected. The latter elements of a belt, if adequately exposed in the vertical dimension, invariably are noted to be inflated by diabase sills, which terminate abruptly against the fault that bounds the fold or merge with the dike that followed the fault (fig. 14). These

Figure 14.--Diagrams showing effects of sill inflation on displacement along a pre-existing fault. Arrows indicate relative displacement.

sills caused displacement opposite to the direction of original throw (fig. 14B). Other sills caused additional movement in the same direction as the original throw (fig. 14C). Locally the fold adjacent to the fault may be discordantly split by the sill (see horizon y, fig. 14B), indicating that the fold was pre-existent. In some places, where sufficient vertical exposure is available so that offsets above and below the sill can be viewed, the folds below the sill dip in a sense compatible with the direction of displacement on the lower part of the fault; those above the sill dip in a sense reverse that of the stratigraphic displacement (fig. 14C). In such examples the stratigraphic displacement caused by diabase inflation was opposite and greater than that along the pre-diabase fault. Such features indicate, again, that diabase intrusion was not a factor in causing the folds.

It can be speculated, but is not yet conclusive, that the dikes so prevalent along the northerly trending structural belts represent the principal feeder conduits for the sills of the northern part of the region. These are only sites where thick dike-like bodies can be inferred to cross from well down in older Precambrian rocks up through entire sections of younger Precambrian strata. Elsewhere most thick tabular cross-cutting masses apparently only connect a concordant body at one horizon with another concordant part of the same sill at another horizon.

The areal distribution of sills that join with the possible feeder dikes is particularly noteworthy. Thick or extensive sills that exist on one side of a structural belt commonly terminate abruptly against the principal fault, or merge with but terminate against a wall of the dike that occupies the fault. On the opposite side of the belt--say the east side--sills may be insignificant or missing. Elsewhere along the strike of the belt the relations may be reversed: that is, thick sills may inflate strata on the east side of the structural belt but be absent on the west side. Furthermore, sills on one side of a structural belt commonly were intruded along different horizons than those on the opposite side. Perhaps such spatial relations of sills to the structural belts can be considered clues to the existence of such belts, especially details of structures largely obscured in areas farther south.

Whether the northerly trend of the pre-diabase structures, seemingly characteristic of the Plateau portion of the region, is typical of like structures farther south cannot yet be judged. It is of interest to note that the Cherry Creek belt, if projected south, could extend into the northeast corner of the Globe quadrangle; there the fault patterns mapped by Peterson (1954) are similar to patterns noted along the east wall of Cherry Creek Canyon. And Peterson has indicated that some of the northwest-trending faults do predate the diabase intrusions.

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. Although some faults antedate the diabase, and some of these owe a part of their displacement to diabase inflation, far more numerous are the faults that owe their dislocation entirely to diabase inflation. The latter are the Precambrian faults most commonly seen. Some interrelations between inflation faults and diabase sills are shown on figure 15, which is idealized from obvious examples in the

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Figure 15.--Idealized geologic section showing relations of faults to diabase intrusions and to pre-Paleozoic unconformities. Note vertical exaggeration. Diagram is greatly simplified; ordinarily two or more sills inflate Apache section and several minor discordant steps, too small to depict here, would exist along sill boundaries. See text for additional explanation.

Sierra Ancha and the Salt River Canyon.

For most of these faults the stratigraphic throw closely approximates or is equal to the thickness of the inflating sills. Sills commonly terminate against faults. Where a part of the stratigraphic section above and below the sills cannot be seen in an area of high relief or otherwise interpreted, the causative relation of faulting to diabase intrusion might not be fully appreciated. By wedging apart the strata a sill may have caused the fault against which it terminates. And the sill ordinarily exhibits a normal chilled border against the part of the fault that it contacts. In the vertical dimension some faults obviously terminate either at the top (see locality A, fig. 15) or bottom (locality D, fig. 15) of a sill. Because the throw of such faults equals the thickness of the sill, the entire displacement can be demonstrated to be the result of inflation (compare intervals x and x' fig. 15). The high-angle faults and related features observed on a regional scale occur in facsimile on small scales, as is abundantly seen in mine faces or cliff exposures. The throw along a fault may be only a few inches if the inflating sill is thin or on the order of a thousand feet if the sill is thick. Faults of like origin are particularly seen where sills step discordantly from one horizon to another (locality C, fig. 15). The footwall contact of a sill may be concordant but a discordant step and fault may offset concordant parts of the hangingwall (locality B, fig. 15). 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. 15). 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 locality A, fig. 15). Such reversals in throw can be observed in other circumstances, as those in which inflation acted in opposition to the direction of pre-diabase faulting (fig. 14), or as in the junction along strike of two inflationary faults of opposite throw. Where more than one sill inflated a sedimentary section or multiple sills were intruded along approximately the same horizons the relations between intrusions and faults can be more complex, in that a later sill offset the first sill and may have displaced structures formed adjacent to the first intrusion as well as caused additional faults.

In any given small area in the Colorado Plateau portion of the region a conjugate system of two sets of joints is dominant. If prominent inflationary faults, dikes or highly discordant parts of sills exist in the same area, ordinarily they occur in only one alinement, parallel to one of the joint sets. Subordinate faults and intrusions commonly parallel the complementary set of joints, but just as commonly follow the trend of some other set. Cursory analysis of many observations indicates that two distinctive systems of joints exist, and leaves the strong impression that two systems of the larger-scale structures also exist. Detailed mapping of the faults and other larger-scale features, however, suggests that the structures of the northerly trends may not be rigorously separable into two sets.

The conjugate sets of one system of joints strike N. 10° - 25° E. and N. 70° - 85° W.; the sets of the other system strike N. 5° - 30° W. and N. 45° - 60° E. These four sets characterize all strata of younger Precambrian age, and also exist in the diabase. The joints of an individual diabase body, though not as conspicuous, reflect the joint pattern of the adjacent host formation. In any particular area, one of these systems strongly dominates the joint pattern in the host rocks. As northerly trending sets of high-angle inflationary faults, and in some localities individual faults, are followed along strike they commonly deviate in trend from north-northwest to north-northeast. Structures of the northeast or west-northwest trends may or may not be coupled with the northerly trend that is complementary.

That is, in one locality a prominent northeast fault or discordant contact may exist in the vicinity of a north-northwest fault--its complement. In another area the trend of the northerly faults may be north-northeast, but instead of the expected complementary structures of west-northwest strike, again the northeast structures prevail. As noted earlier, the throughgoing pre-diabase structural belts may be sinuous in trend throughout the range of northerly strikes. Apparently these belts set the pattern for later northerly striking inflationary faults, which are haphazardly coupled distribution-wise with faults of either of the other prominent trends. Only the latter, of northeast and west-northwest direction, are individually definitive.

In the northern part of the region, individual inflationary faults range from a few tens of feet to more than 10 miles in length. Those striking other than northerly are particularly conspicuous because they cut across the prevaling structural and topographic grain of the area. An outstanding example of a fault of N. 70° W. strike has been traced westerly from the northwestern corner of the Blue House Mountain quadrangle to a point on Cherry Creek two miles north of the McFadden Peak quadrangle. The fault probably continues even farther, east and west. Throughout the known length of 11 miles the fault is occupied by the discordant part of a diabase sill. From a point about one-half mile south of McFadden Peak another fault extends 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 15. Eastward this fault ends abruptly against the principal northerly trending fault of the 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. Most faults of west-northwest to west strike are less conspicuous; these two examples indicate, however, that such faults are significant on a regional scale.

Inflation faults of N. 30° - 60° E. strike are numerous, but compared with faults of northerly strike and some of west-northwest strike they are apparently short and ordinarily of small throw. Many are conspicuous in the landscape because of discordant bodies of diabase that were intruded along them.

Small-scale monoclines, synclines, anticlines and domes have been given particular attention by me, because they were prime structural factors in the localization of the chrysotile-asbestos deposits. These folds, which represent mild drag in sedimentary rocks immediately adjacent to diabase intrusions, are numerous. Many are so subtle that they are recognized only by detailed mapping on mine scales, and they are in no way analogous to the folds of regional extent that pre-date the diabase. These small folds are seen particularly adjacent to discordant portions of intrusions. Though numerous, such folds do not occur everywhere along or near discordancies, and it cannot be assumed that they characterize discordant contacts.

In addition to the small-scale high-angle faults that are miniatures of those of regional scale, small bedding-plane faults, thrust faults and near vertical strike-slip faults are locally abundant in strata adjacent to discordant diabase bodies. They are particularly abundant where the folds, noted in the above paragraph, exist. The small-scale faults, generally of only a few inches to a few feet horizontal displacement, and the folds represent mild structural adjustments that occurred during and immediately following the emplacement of diabase, as is testified by their geometric relations to successive intrusions of diabase.

The faults and folds of small scale are most apparent in the least competent units of the stratigraphic sequence: the lower member of the Mescal limestone, the lower two-thirds of the upper member of the Dripping Spring quartzite, and the tuffaceous siltstones and mudstones of the Pioneer formation.

Presumably the Precambrian structural features, characterized above for the Colorado Plateau portion of the region, are similar and equally as abundant in the southern part of the region. Certainly, in widely separate parts of the latter area, several examples of large-scale younger Precambrian structures can be readily recognized. In the Basin and Range portion deformation that was superimposed in Cenozoic time, and possibly that of late Cretaceous time, particularly obscured the structures that antedated and were induced by diabase inflation.

Block faulting and tilting of the block faults were characteristic of this late deformation; folding was generally negligible. In applying geologic studies to the metal-mining districts of southern Gila County and surrounding areas, it would seem particularly worthwhile to distinguish the bulk of the faults of younger Precambrian age from those of more recent origin. An understanding of the habits of diabase sills will facilitate prediction of the disposition of the strata displaced by the intrusions. Of more significance, most faults of inflationary origin are literally without "roots". They are not faults, then, that are likely to be reopened; nor do they extend to such depths that they would likely tap sources of metal-bearing solutions. Thus most of these dilation faults were probably not significant in the localization of metal deposits, and they should be distinguished from later faults of seemingly similar geometry that were effective as channelways for ore-bearing solutions.

Relations of Paleozoic formations to Precambrian rocks

The unconformities that define the base of the Paleozoic sections of the region everywhere truncate the structures and variously displaced formations of younger Precambrian age. South of a line that extends from the south flank of the Pinal Mountains northwest toward but not to Roosevelt Dam and southeast to the vicinity of San Carlos Lake, as shown on figure 3, the younger Precambrian formations are generally overlain by the Bolsa and Abrigo formations of Cambrian age. As drawn, this line depicts the northern limits of Cambrian outcrops as now known. But in the vicinity of this line, also, the unconformity that separates Cambrian and Devonian formations farther south in turn truncates the unconformity that defines the base of the Cambrian system. In part this line does approximately define the intersection of the two unconformities; in part that intersection-- now destroyed, owing to Cenozoic erosion -- must be interpreted as being somewhere to the north. Certainly the intersection was nowhere more than a few miles to the north, however, because where Paleozoic formations are next seen the Martin limestone of Devonian age is the basal formation of the Paleozoic sequence. On north as far as the Mogollon Rim remnants of the Martin directly overlie the younger Precambrian formations.

These relations are shown on figure 16, which is an idealized

Figure 16.--Reconstruction of north-south geologic section at the end of Devonian time. Section extends from present Mogollon Rim south to Dragoon Mountains.

reconstruction of the north-south geologic section that must have existed at the end of Devonian time. This much simplified section can be visualized as a composite of the north-south sections, which could be reconstructed in a belt ten miles wide on either side of a line extending south from the Mogollon Rim through Young, then southeasterly along the crest of the Sierra Ancha to Globe and Winkelman, and along the San Pedro Valley to the Little Dragoon Mountains. Several sections would be required to show all the structural aspects of such a belt. By use of a composite illustration, features cited in various parts of this report are projected into and can be identified with the idealized section. For example, the Troy formation is unusually thick but of abruptly different thicknesses in the Sierra Ancha; it is thin or missing along the south rim of the Natanes Plateau yet is locally thick but not persistent along the San Pedro Valley. The reasons for such variations can be seen on the section.

To visualize the structural deformation that affected the younger Precambrian formations before Paleozoic time this illustration can be compared with figure 11. The latter is a similar reconstruction of the terrane after deposition of the Troy quartzite, but before it was deformed.

The significance of certain features depicted on figure 16 was not appreciated until recently. As a consequence the Bolsa quartzite has been considered correlative with the Troy quartzite. Actually, certain sandstone facies of the Abrigo have been identified with the Troy--though unknowingly, because they were presumed to be Bolsa strata. And in a few areas locally thick basal sandstones of the Martin formation have been termed Troy quartzite. What is presently known of north-south lateral variations of the Cambrian formations will be described in another report now in preparation. Therefore the stratigraphy of the Cambrian system is not detailed in the following. Rather, emphasis is placed on the structural relations of these strata to the Precambrian rocks, and on criteria for distinguishing between formations lithologically somewhat similar.

Relations of the Bolsa and Abrigo formations

As originally defined in the Bisbee area by Ransome (1904, p. 28-30), and as demonstrated by Gilluly (1956, p. 14-15, 24) and by Cooper and Silver (report in preparation) to persist northward into the northwest corner of Cochise County of interest here, the Bolsa quartzite is comprised of medium- to coarse-grained pebbly quartzite. This quartzite occurs in tabular beds that range from a few inches to 10 feet thick, but perhaps average 2 to 4 feet (Gilluly, 1956, p. 15); the beds ordinarily are cross-stratified. In the Dragoon quadrangle and south to Bisbee, as reported in the above references, the formation generally ranges between 390 and 480 feet in thickness. Locally, where deposited on a surface of some relief the Bolsa is thinner. A basal conglomerate is characteristic. In some localities the pebbles of this conglomerate compositionally reflect the lithology of the underlying Precambrian terrane. Upward, by an increase in sandy shale layers and in micaceous sandstones or quartzites, the massive-cropping Bolsa is transitional into the Abrigo limestone. Gilluly (1956, p. 15) describes this transition zone as generally about 50 feet thick, but notes that at least in one locality it is almost 100 feet.

Again for the belt of outcrops between Bisbee and the Little Dragoon Mountains, Ransome (1904, p. 30-33; 1916, p. 145-148) and Gilluly (1956, p. 16-25) have described the typical lithology of the Abrigo limestone, and have noted a range in thickness from 700 to 844 feet. Above the basal transition zone, limestone dominates the lithology. Dolomite becomes increasing abundant and sand increases toward the top of the Abrigo. Commonly a thin unit of quartzite or sandstone caps the formation. The Abrigo is distinguished by thin-bedding, beds marked by distinctive edgewise conglomerates, numerous units that exhibit irregular partings, and irregularly anastamosing layers of silty and sandy material that separate irregularly rounded lenses of sandy carbonate. Worm-trails and fucoidal markings are abundant. Beds of massive limestone form a few ledges, and occasional beds of quartzite, sandstone or sandy shale are variously distributed through the sequence. Northward from the type locality near Bisbee the amount of sand increases (A. R. Palmer in Gilluly, 1956, p. 20).

Ransome (1904, p. 30) originally considered the Bolsa quartzite as Middle Cambrian in age on account of its conformable relation with the overlying Abrigo, from which fossils then assigned to the Middle Cambrian 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 accepted generally. 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 concisely summarized by A. R. Palmer (in Gilluly, 1956, p. 20-24).

The main aspects of Bolsa quartzite and Abrigo limestone sections generally considered typical of southeastern Arizona have been considered above. Of more interest here, because of problems in recognition, are some atypical facies that exist farther north.

In the northern part of the Santa Catalina Mountains (near the head of Peppersauce Wash, in the southwestern corner of the Mammoth quadrangle) Stoyanow (1936, p. 476-477) recognized a lithologically different Abrigo section. This section, measured as 750 feet thick by Stoyanow, includes considerably more sandstone, shale, and quartzite than sections farther south. The lower 400 feet, separately designated the Santa Catalina formation by Stoyanow, is dominated by rusty-weathering, thin and irregularly bedded micaceous mudstones and quartzitic sandstones, which are intercalated sparsely with thin layers of dolomite. This unit, in the section measured by Stoyanow, is capped by a conspicuous ledge-forming unit of clean quartzite, which he termed the Southern Belle quartzite. The latter is 25 to 30 feet thick along the upper canyon of Peppersauce Wash, but is only locally existent elsewhere in Cambrian sections of the northern Santa Catalina Mountains (S. C. Creasey, oral communication, November 1960).

These units undoubtedly are gradational southward by loss of clastic constituents into sections dominantly of carbonate beds. As an example, the Santa Catalina formation of Stoyanow can be readily identified lithologically with the lowest of three members of the Abrigo recognized by Cooper and Silver (report in preparation) in the Dragoon quadrangle. From south to north in that quadrangle this lowest member becomes increasingly sandy. In the northwestern part of the Dragoon quadrangle, 35 to 40 miles southeast of the Peppersauce Wash locality, this lower member has similar bedding characteristics and includes mudstones and sandstones like those of Stoyanow's Santa Catalina formation, but includes considerably more brown-weathering dolomite. The lower member of the Abrigo in the northwestern part of the Dragoon quadrangle is capped by a unit of quartzite, which thins out southward and is everywhere inconspicuous compared with the Southern Belle quartzite of Stoyanow. On the basis of distinctive faunal zones, Stoyanow defined the Santa Catalina and Southern Belle units as Middle Cambrian and the overlying strata below the Martin limestone (Devonian) were defined as Late Cambrian. Similarly, fossils locally notable in the Dragoon quadrangle essentially define the boundary between the lower and middle members of that area as the boundary between strata of the Middle and Late Cambrian age (A. R. Palmer, oral communication, November 1960).

Stoyanow (1936, p. 465-481) recognized several faunal zones during his study of the Cambrian strata in the northern Santa Catalina Mountains and elsewhere in southeastern Arizona. Consequently he proposed that Ransome's terminology should everywhere be revised. Overlying the Southern Bell quartzite of Stoyanow in the Peppersauce Wash area is a unit, almost 300 feet thick, of thin-bedded sandy limestone with intercalated thin sandstone beds; intervening between this unit and the Martin limestone of Devonian age is a thin (20- to 25-foot) unit of sandstone and quartzite. Stoyanow restricted the designation "Abrigo formation" to the limestone unit and to paleontologically equivalent sections farther south. The capping sandstone unit he also designated as a separate formation, the Peppersauce Canyon sandstone. Other formational designations were also proposed for southerly limestone equivalents of the Peppersauce Canyon and Santa Catalina units. The proposed subdivisions, difficult of lithologic definition except locally, comprise an apparently unbroken sedimentary sequence. For reasons already concisely enumerated by Palmer (in Gilluly, 1956, p. 24), there is no objective way of distinguishing Stoyanow's formational units as fundamental map units 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, has generally been followed and is to be preferred.

Stoyanow's contribution in dating by faunal zones the clastic units in the Abrigo of the northern Santa Catalina Mountains is particularly significant, because it provides the background for recognizing more northerly Abrigo facies, which include an even greater content of the coarser clastics. Briefly stated, a quartzite unit similar to the Bolsa quartzite and in the relative position of the Southern Belle quartzite thickens northward, at the expense of thin-bedded fine-grained strata like those of Stoyanow's Santa Catalina unit. At Zapata Mountain, 1½ miles northwest of Holy Joe Peak, for example, rather typical Bolsa quartzite, 160 feet thick, is overlain by a 170-foot thin-bedded unit of Santa Catalina aspect, and this in turn is overlain by 110 feet of cliff-forming quartzite beds very like the Bolsa quartzite of the same area (M. H. Krieger, written communication). The upper 60 to 70 feet of the thin-bedded unit that underlies the Southern Belle lithologic equivalent, and at least 100 feet of the immediately overlying interval includes many ledge-forming quartzite beds like those in the principal quartzite unit. The uppermost 100 feet of the Abrigo at this locality includes carbonate beds, which are missing from lower parts of the Abrigo section. Where Abrigo sections have been observed north of the Holy Joe Peak quadrangle entire sections are comprised of Bolsa-like quartzite units and thin and irregularly bedded units of fine-grained micaceous sandstone and mudstone. Except for a dolomite cement sparsely noted in some of the thin-bedded units carbonate is missing from the most northerly sections.

The "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 surfaces are ridged and knotted with numerous worm casts," which Ransome (1919, p. 44) considered the "most characteristic material" of the thin-bedded upper quartzite of the Troy of the Ray quadrangle, undoubtedly have their equivalents in these northern facies of the Abrigo formation. The lithology that Ransome described is noted particularly in the thin-bedded sandstones just above the Bolsa quartzite.

Throughout the region of Cambrian outcrops north of the Santa Catalina Mountains primitive phosphatic brachiopods of Cambrian aspect, but not diagnostic of the Cambrian, are abundant locally. These brachiopods occur particularly in the uppermost beds of the Bolsa quartzite, in the friable sandstones immediately overlying the Bolsa and in like sandstones above the cliff-forming quartzite unit higher in the Abrigo formation. Fragments of trilobites occur rarely in the brown-weathering micaceous sandstones; those noted at scattered localities from the Santa Catalina Mountains north into the Mescal Range, during a 5-day field trip in November 1960, were definitely of Cambrian age (A. R. Palmer, oral communication). In collections of trilobites made earlier by C. R. Willden in the southeastern part of the Mescal Mountains (at Poverty Flat, 5 miles southwest of Coolidge Dam--see Christmas quadrangle) an association of bolaspidellid and marjumiid forms generally characteristic of the upper Middle Cambrian has been recognized (A. R. Palmer, written communication to C. R. Willden, August 5, 1960). This confirms the earlier suggestion by Darton (1925, p. 255) that "brown to greenish-gray shales and soft earthy sandstone" of this locality should be correlated with the Abrigo limestone. Specimens of <u>Billingsella</u>, a brachiopod that characterizes the uppermost parts of the Abrigo limestone in more southerly sections, were identified by A. R. Palmer in the Zapata Mountain section, indicating that at least that far north no great part of the Abrigo formation was eroded prior to deposition of Devonian strata.

The reasons for not delineating, in the past, the Bolsa from the Troy are now apparent. The increase in coarse clastics, particularly quartzite, and the decrease in carbonate content northward from the region of typical Abrigo outcrops cause the northermmost outcrops to be grossly similar to the Bolsa quartzite. Moreover, in many places the Bolsa quartzite rests on the Troy quartzite in seeming conformity and the contact between the two formations is not conspicuous. In other areas the Troy quartzite is entirely missing and the combined Bolsa-Abrigo sections, which occupy the same relative stratigraphic position, are approximately the thickness of the Troy noted elsewhere in the southern part of the region. For want of recognized diagnostic features that would distinguish Cambrian strata from those of the Precambrian, by default the two sequences have been considered equivalent.

Fortunately in places the Bolsa quartzite clearly rests in angular or erosional unconformity on the Troy quartzite, formations of the Apache group and the included diabase intrusions. In these examples, not subject to misinterpretation, criteria for distinguishing Precambrian sandstones and quartzites from those of Cambrian age can be recognized. These diagnostic features are summarized on table 5, which also

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Table 5.--Criteria for distinguishing quartzites and sandstones of the Bolsa and Abrigo formations from those of the Troy formation.

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highlights some features not mentioned in the preceding generalized descriptions. Details of relative positions in the sequences of quartzite and sandstone, differences in textures of the sandstones, the common occurrence of Scolithus in the Cambrian quartzites, and of worm casts in the fine-grained Cambrian sandstones are particularly noteworthy. By the use of such criteria, for instance, 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.

Northward from the Santa Catalina Mountains the Bolsa quartzite generally is thinner than in the areas farther south, but in other aspects it is not notably different. In places it is not as well cemented as in the usual southerly outcrops, and in these places it might be described as quartzitic sandstone or a sequence of alternating quartzite and sandstone beds. Northward, also, the Bolsa rests on a somewhat irregular erosion surface, which variously truncates the formations of younger Precambrian age. Such features are especially well illustrated in the mapping now (1961) being done by H. H. Krieger in the Holy Joe Peak quadrangle. Within a two-mile radius west and north of Holy Joe Peak the Bolsa rests on diabase, the upper and middle members of the Troy quartzite and the lower member of the Dripping Spring quartzite, which were variously displaced by diabase intrusions. The part of the pre-Bolsa erosion surface that is on diabase is somewhat lower than the parts that truncate the more resistant quartzites. Four miles to the northwest at Brandenburg Mountain the Bolsa laps out and the Abrigo is thinned considerably where the Cambrian strata lap from a surface on diabase to higher surfaces of monadnocks eroded on the Dripping Spring and Troy quartzites. Near the north edge of the quadrangle, at a locality about 8 miles east of Winkelman, the pre-Bolsa unconformity has a slight angular relation in truncating sills intruded into the upper member of the Dripping Spring quartzite. Where State Highway 77 crosses the crest of the Mescal Mountains in the northeastern corner of the Ray quadrangle the angular discordance is somewhat more subtle. Near the highway the Bolsa is seemingly conformable with the middle member of the Mescal; one-quarter mile to the west a few tens of feet of the Apache basalt intervenes between the two formations; within the next mile to the west the contact cuts across the lower member of the Mescal and well down into a diabase sill intruded into the lower member. Incidentally, the outcrop of quartzite above this unconformity is the one from which Stoyanow (1936, p. 475) collected the brachiopods that are in part the past basis for designating the Troy as Cambrian in age. Cooper and Silver (report in preparation) describe the extreme in relief noted to date for the pre-Bolsa surface. For the northeast part of the Little Dragoon Mountains they describe the surface as generally plateau-like, but surmounted by hills as much as 100 feet high and cut by valleys 150 to 200 feet deep. The Bolsa rests variously on the Dripping Spring quartzite, the Pioneer shale and on diabase sills intruded into these formations. Although in many places Cambrian strata seem conformable with underlying formations, these examples indicate that discordant relations can be found wherever, from north to south, Cambrian rocks overlie younger Precambrian formations.

South of the Dragoon quadrangle the Bolsa everywhere rests on older Precambrian formations. Diabase intrusions into the older rocks have not been reported from this area, a considerable part of which has been mapped in detail (Gilluly, 1956). From this, I would surmise that the interval of older Precambrian rocks that ordinarily is host for sheets of diabase-- and therefore a considerable thickness -- was stripped away before deposition of the Bolsa quartzite.

In some additional localities it may ultimately be shown that Abrigo strata rather than Bolsa quartzite directly overlie the Precambrian formations. Krieger has already clearly defined the Brandenburg Mountain examples by mapping. Cooper and Silver have noted that the Bolsa thins to as little as 14 feet on the surface of relief in the Little Dragoon Mountains. Along the west edge the Inspiration quadrangle and in places 1/2 to 1 1/2 miles north of Superior (localities seen under the guidance of N. P. Peterson and D. W. Peterson) thin-bedded, friable finely micaceous sandstones, more like parts of the Abrigo formation farther south than like the Bolsa quartzite, rest directly on formations of the Apache group.

Wherever Cambrian strata overlie diabase the latter is decomposed through thicknesses generally ranging from 2 to 25 feet. In this interval below the unconformity the diabase is platy or shaly parting and generally is grayish red. Ordinarily there is no indication that any of the regolith below the contact was reworked and transported. In a few places the decomposed diabase is greenish gray; in such outcrops, in particular, the original textures of the diabase are observed to be relict up to the contact with the Cambrian strata. Although diabase soil incorporated in the basal part of the Bolsa commonly causes several feet of the strata to be dusky red, fragments of diabase in the basal conglomerate of the Bolsa are rare. In a few localities the diabase debris below the contact is bedded. In an unusual example 2 miles southeast of Superior diabase debris, mixed with an appreciable amount of well-rounded coarse quartz sand, is conspicuously cross-stratified through a vertical interval of 25 to 30 feet. Here the usual platy-weathering decomposed diabase underlies this unit of reworked material.

In the northern Santa Catalina Mountains a locally thick unit of distinctive grayish-purple sandstone conformably underlies the quartzite formation now known to be Bolsa rather than Troy quartzite. Stoyanow (1936, p. 477) relegated this unnamed unit to the Apache group. S. C. Creasey has shown the writer a locality where the basal 2 feet of the purple sandstone includes angular fragments of diabase, which is here intruded into the Dripping Spring quartzite. Creasey (oral communication, November 1960) additionally noted that in adjacent areas this sandstone rests on an unconformity that truncates both Apache formations and the pre-Apache formations and the pre-Apache granite. Like sandstone has not been recognized elsewhere in the region. The unit in the Santa Catalina Mountains could represent a local basal "pocket" filled with clastic material reworked from the reddish diabase debris, or possibly it represents reddish or purplish detritus from the Pioneer shale or from highly ferruginous silicified remnants of the Mescal dolomites.

Relations of the Martin limestone

The Martin limestone generally overlies the Cambrian formations in the southern part of the region and, except as absent because of post-Paleozoic erosion, everywhere overlies the younger Precambrian formations in the northern part of the region. The Martin is mostly of limestone and dolomite but does include beds of shale and sandstone; regionally the formation is quite variable in lithology (see Huddle and Dobrolvolny, 1952, p. 73-76, 83-85; Gilluly, 1956, p. 26-29). The Martin has generally been considered wholly of Late Devonian age. But recently Teichert and Schopf (1958, p. 213-215) have suggested that plant remains, collected from beds a few feet above the base of the carbonate portion of the formation, indicate that 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."

The hiatus between the Cambrian and Devonian is represented, in southeasternmost Arizona, by a disconformity. Channeling is not apparent in the underlying Abrigo limestone (Gilluly, 1956, p. 25-26). As the unconformity is followed northward from Cochise County, however, it is evident that pre-Martin erosion was increasingly more effective in stripping the existing formations. And in the northern part of the region channels of considerable extent were cut below the general plane of the erosion surface. Between Canyon Creek and Payson in particular a surface of moderate relief exists, so that in different localities different strata of the Martin formation lap against the Precambrian formations. In most places south of the Pinal Mountains Cambrian strata probably intervene between the Martin limestone and the Precambrian formations. But at least locally, as in the vicinity of the abandoned settlement of Troy in the Dripping Spring Mountains (see Ray quadrangle), the Martin rests on the Troy quartzite or on formations of the Apache group.

As a generality, north of the line that defines the northern limit of Cambrian outcrops (see fig. 3), more of the younger Precambrian sequence is remnant than in areas south of that line, and the Martin limestone ordinarily overlies the Troy quartzite. In some areas of as much as a few square miles, however, the Martin limestone rests on the Mescal limestone or the Dripping Spring quartzite, or on diabase sills intruded into these formations. Along the foot of the Mogollon Rim, where the Apache group was already thin as a consequence of lapout on a pre-Apache high, and northwest from Haigler Creek the pre-Martin unconformity progressively truncates lower strata in the Apache group. Ultimately at Christopher Mountain, northwest of Young, the Martin laps against older Precambrian formations. Farther west at this latitude younger Precambrian strata are not known to intervene between the Martin and the older Precambrian formations.

The pre-Martin unconformity truncates the younger Precambrian formations and the Precambrian structures that affected these formations in the same manner as the pre-Bolsa unconformity. In the Colorado Plateau portion of the region the structural relations are more readily seen, because the pre-Devonian unconformity is virtually a horizontal plane that can be traced for miles without disruption caused by post-Paleozoic deformation. Where U. S. Highway 60 transverses the north wall of the Salt River Canyon, up and down canyon from this locality for several miles, and east of Highway 60 on the south side of the canyon the unconformity--without offset--truncates diabase sills and the Mescal and Troy ~~formations~~ at different stratigraphic positions, dependent on the degree of vertical displacement of these formations caused by sills lower in the section. Where slivers or plates of Apache strata are included between sheets of diabase and were tilted slightly, the unconformity commonly truncates the strata at angles of a few degrees. Angular discordances of as much as 30 degrees have been seen, but in most areas the Martin strata appear to be concordant with the Apache strata.

A regolith zone quite similar to that described below the pre-Bolsa unconformity everywhere characterizes the diabases that immediately underlie the pre-Martin unconformity. Again, due to thorough decomposition of the diabase in this regolith, fragments of diabase in the basal beds of the Martin are rare. The basal part of the Martin ordinarily does not include as much regolithic material as its Bolsa counterpart, but in many localities the upper foot or two of the regolith includes abundant quartz grains like those in the basal sandstone of the Martin.

In the northern part of the region a dolomitic sandstone unit, a few inches to 20 feet thick, is characteristic of the Martin. In a few widely separate localities, this sandstone merges downward with a much thicker sandstone, which crops out prominently as cliffs or as steep ledge-studded slopes. These thicker sections of the basal sandstone fill the extensive channels noted previously. The channels range from a few feet to at least 250 feet in depth and from a few hundred feet to as much as one-half mile in width. At least two of the sandstone channels known to me exceed 5 miles in length. This sandstone generally has been casually included with the Troy, and therefore its relations to the diabase and the pre-Troy formations have been incorrectly interpreted.

The geometric relations of the sandstone-filled channels to the underlying formations clearly indicate that these sandstones should be considered Paleozoic rather than Precambrian in age. The paleochannels are generally broader and least steep-sided where they were eroded in diabase or other non-resistant rock. Where eroded into the Troy the channels tend to be steep-walled, and subrounded to angular blocks of Troy may be incorporated in the basal beds of the fill. Only in a few localities, where channels are in the quartzite member of the Troy, has a conspicuous conglomerate composed of gravels of cobble or boulder size been noted at the base of these sandstones. Such a channel, not less than 230 feet deep, cuts through the plate of Troy quartzite that forms the hanging wall of the thick diabase sill which underlies Aztec Peak in the Sierra Ancha. In places along the lowest part of this channel there is a cobble conglomerate of Troy quartzite gravels as much as 40 feet thick. In many places lenses of granules or small pebbles exist but are not conspicuous in the basal sandstones of the channel fills. In some localities the walls of the channels truncate the virtually horizontal bedding of the Troy at angles of 20° to 30°. Where a channel is in the Chediski member of the Troy, and the filling is largely of sand easily derived from this poorly cemented member and thus compositionally very like it, these channel relations may be the most distinctive feature that can be used in distinguishing the two units. Most of the so-called "Troy quartzite" exposed near the south abutment of Roosevelt Dam is probably Devonian sandstone derived from the Chediski member.

In some localities medium-grained sandstones dominate the channel fills; in others the sandstones are mostly coarse-grained or very coarse-grained; in every occurrence many beds are of very poorly sorted sands and scattered small pebbles and lenses of granular conglomerate are typical. The sandstones are generally very friable. Most are in tabular beds, a few inches to 10 feet thick, in which small- to large-scale cross-stratification etchs out prominently. Shaly-parting thin layers of sandy siltstone, local in areal extent, may separate some sandstone beds. Commonly a few beds are highly feldspathic, and rare beds are cemented by dolomite or calcite. In some places the sandstones range in color from almost white to dusky red; but generally, particularly as viewed from a distance, the sandstones are yellowish brown to reddish brown. In perhaps 4 out of 5 occurrences certain layers are conspicuously marked by abundant hematite-cemented concretions; these concretions are ordinarily in the upper parts of the fills. Thus, even in isolated outcrops, the overall aspect of these channel sandstones is wholly different from any sandstones of the Troy, Bolsa or Abrigo ~~formations~~. The way in which these sandstones grade into the basal carbonate beds of the Martin clearly indicates that they are genetic parts of that formation.

Geologic history of the younger Precambrian formations

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, bedding features or other internal structures. Lateral differences in thicknesses are largely the effects of erosion before a succeeding unit was deposited. For certain members the variations in thickness do reflect, to a degree, variations in the configuration of the basin of accumulation; the Pioneer shale and the arkose member of the Troy are particular examples. Some lateral lithologic changes, such as the variation from cherty dolomite to almost solid ferruginous chert or from dolomite to silicated limestone observed in the Mescal, are effects superimposed after deposition. As an example of the recognized variations, the bedding structures of the Chediski member of the Troy quartzite change considerably from north to south, but no marked variation in the overall lithology is apparent in the same interval. Admittedly other variations exist, but those that have been recognized seem of little paleogeographic significance.

The following summary of younger Precambrian events is presented to highlight some of the laterally persistent features and the lateral variations. Where possible, some of the genetic implications of these features is also interpreted, in the hope that recognition of like or related geologic phenomena may one day aid in correlating the Apache and Troy strata with those of younger Precambrian age in surrounding regions. The reasons for the interpretations are discussed more fully in the preceding sections, where the necessarily rather dogmatic expressions of this summary are also qualified.

Prior to deposition of the Apache formations the older Precambrian metavolcanic and metasedimentary formations and the granitic masses, which had intruded them in batholithic proportions, were deeply eroded, leaving an exposed terrane largely of granitic rocks as source material for later incorporation into Apache sediments. In an area not now exposed considerable volumes of quartzites, like those that now very locally underliethe Apache group, must also have been available as source rocks. The pre-Apachd surface is generally of low relief. Only in the northwestern part of the region of present Apache outcrops did the surface stand above the rest of the region, and exhibit local topographic features of notable relief. The surface probably represents a peneplain, additionally smoothed and swept clean of most of its residual rock debris by marine abrasion.

As the first sea of Apache time encroached across the region wave and current action must have been vigorous. In places the Scanlan bed of the Pioneer is composed of subrounded or angular gravels, which might well be almost entirely of local derivation. But the bulk of the gravels of this first transgression are well rounded and of highly resistant quartzite; some of these are of cobble dimensions and were deposited a minimum of several tens of miles from any quartzite outcrops likely to have been sources. During this first marine transgression across a prevalently granitic terrane a moderately thick blanket deposit of arkose much like that comprising the lower member of the Dripping Spring, might be expected to accumulate. Such arkoses do make up the lower part of the Pioneer in many areas, but even the cleanest beds of arkose--those comprised essentially of quartz and feldspar--are commonly separated by thin layers of tuffaceous siltstone, and in many areas fine-grained tuffaceous sediments immediately overlie the basal conglomerate. Thus voluminous and extensive falls of ash, from a source as yet unrecognized, may have literally flooded out the arkosic strand deposits of the encroaching sea. Because the bulk of the formation is of siltstone and mudstone delicately laminated and cross-laminated, a more probable explanation is that wave and current action were not vigorous enough to transport arkosic debris during much of Pioneer time. The lack of abrasion of the delicate glass shards tends to confirm this interpretation.

An erosional episode intervening between the deposition of the Pioneer and the deposition of the Dripping Spring must be inferred. The contact between the Pioneer formation and the overlying Barnes conglomerate is sharp and seemingly concordant. Rare shallow channels along the contact are the only direct indication of pre-Barnes erosion. Nevertheless, the Barnes conglomerate and the overlying massive-bedded arkoses, entirely lacking in tuffaceous material, strongly mark a break in sedimentation and herald a new cycle of sedimentation in a different environment. The thin, reasonably uniform layer of gravels comprising the Barnes conglomerate is not readily explained except as deposits of a transgressive sea. At the time the Barnes was deposited, local sources of gravels were buried by the Pioneer. Thus the Barnes conglomerate, even more than the Scanlan, forces the conclusion that the well-rounded and in part very coarse gravels were products transported long distances, and implies vigorous marine erosion at the start of Dripping Spring time. In a few areas, arkoses are found at the top of the Pioneer section. Possibly these are remnants of coarser deposits laid down during regression of the first Apache sea.

The lithology and bedding features of the lower (arkose) member of the Dripping Spring quartzite suggest vigorous erosion of a granitic terrane to provide clastics rapidly accumulated in a shallow transgressing sea. Rather uniform grain size through thick units of cross-stratified tabular beds implies considerable winnowing by tidal or wave action competent to spread the sorted sand widely in thick sheets. Toward the top of the member feldspar decreases, the grain size increases slightly and scattered small pebbles are included in the beds. Perhaps these features are hints of the termination of this period of shallow marine sedimentation.

The thin and irregular bedding, abundant scour-and-fill structures, occasional large-scale channels, and abundant ripple-marks and mudcracks, which characterize the upper member of the Dripping Spring, collectively indicate deposition at sites subject to shallow subaqueous erosion and at times to subaerial exposure. A feasible explanation for the consistent northerly alinement of the scour-and-fill features cannot yet be suggested. At least part of the material was repeatedly exposed or reworked and therefore aerated. Deposition must therefore have been rather rapid; otherwise carbonaceous material and the sulfates now represented in pyrite would not have been preserved. The muds of the upper member probably accumulated on broad tidal flats of a shallow sea that sporadically advanced on a granitic terrane of low relief.

The significance of the rather abrupt change from fairly well-sorted arkose of the lower member to relatively ill-sorted but finer-grained muds of the upper member is presently a subject of speculation. Perhaps the environment of the source area at the beginning of upper Dripping Spring time was modified abruptly. Or maybe the abrupt change marks a hiatus in deposition, for which other indicators have not yet been recognized.

The Dripping Spring sediments were indurated, and thereafter eroded, before the Mescal was disconformably deposited. The thin basal sandstone of the Mescal limestone was derived in large part by reworking of the upper part of the Dripping Spring quartzite, but is in part of rounded coarse sand derived from another source. Like coarse sand was additionally supplied during early stages of carbonate sedimentation in the Mescal sea; otherwise little clastic debris was contributed.

During the early stages of carbonate deposition circulation in the Mescal sea must have been restricted, to cause high salinity favorable to the precipitation of halite and perhaps other evaporites not now recognized. Such an environment would also favor the deposition of dolomite or the penecontemporaneous conversion of calcitic limestone to dolomite. Features indicative of restrictive barriers have not been seen; therefore such barriers must have existed outside of the present outcrop area of the Mescal. The sea floor was so shallow that wave turbulence was effective in partially and repeatedly reworking carbonate strata previously laid down. After about one-third of the lower member had accumulated dolomite beds, fairly well lithified, collapsed forming the peculiar breccia characteristic of the lower part of the Mescal. Salt pseudomorphs decrease upward in the section and particularly are fewest in beds about at the top of the interval of collapse. On freshening of the sea waters perhaps some leaching of the evaporites in the lower parts of the section occurred, causing the odd collapse breccia, which is variable in thickness but so widespread as to be essentially a stratigraphic phenomenon. The Mescal sea remained sufficiently saline thereafter, however, to favor the continued formation of dolomite.

During the deposition of the algal member of the Mescal the sea floor must have been remarkably stable and of uniform depth throughout the 15,000 square mile area now included in the area of Apache outcrops. Otherwise the stromatolites would probably occur in mounded masses (bioherms) rather than as a biostrome (see Link, 1950). Although possibly of different depth, the seafloor must have been similarly stable during accumulation of the upper one-third of the lower member. The beds of at least the upper 50 feet of this member exist in an almost invariable sequence recognized throughout the region.

The carbonate members of the Mescal were elevated above sealevel, at least briefly, and eroded before additional formations were laid down. At this time solution cavities, enlarged later, began to form. Basalt flows, now possibly relict only in very small areas, flowed out on this erosional surface, and were in turn largely eroded away. At least the northern part of the region then subsided and very fine-grained siliceous muds and subordinate carbonate muds accumulated in a quiet body of water. Following another episode of erosion several thin basalt flows were extruded as a very fluid magma and crystallized with remarkably similar characteristics throughout the region. No dikes or volcanic necks, from which these flows might have issued, have yet been recognized.

During the period of minor instability that followed the deposition of the carbonate members of the Mescal the Apache strata were broadly warped, and during the erosional planation that preceded deposition of the Troy quartzite the several formations of the Apache group were variously exposed. The enlargement of solution cavities in the dolomites of the Mescal had continued with each preceding episode of erosion, and with the pre-Troy episode, solution effects were widespread. Few, if any, of the dolomite sections of the Mescal completely escaped the effects of solution: in many areas only a few joints and bedding partings were widened by solution; in addition in some areas a few large sinkholes developed; and in fewer areas part or all of the dolomite section was converted to a massive rubble. In part such rubble formed by the coalescence and collapse of solution caverns; in part it resulted from the gradual thinning of beds by solution along bedding partings, and the subsidence of individual beds to fill the voids thus formed. Wherever solution features were developed concentrations of chert, in part mechanically concentrated residues, were formed. Incidental to these solution processes hematite, probably derived by the laterization of the basalt flows, was concentrated--particularly in certain areas. Collapse of the dolomites in the sinkhole areas continued during deposition of the Troy.

The first sea of Troy time transgressed across a surface that was generally planar but that did locally have some relief. At least 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. Arkosic sands, which had been fairly well sorted during some earlier stage in their development, were literally "poured" into this basin. The prevalent easterly dips of the large-scale cross-stratification suggests transport from the west. The arkoses visible today, however, likely represent only a small part of one fringe of a large deltaic deposit. Therefore, probably no great significance should be attached to bedding attitudes that can be observed in the lower member of the Troy.

At present it is not clear whether the sharp contact between the arkose member and the Chediski member of the Troy reflects a significant erosional hiatus, or merely a minor break in sedimentation. Throughout most of the region the base of the Chediski sandstone is marked by a conglomerate unit, which is similar to the conglomerates that commonly comprise the basal sediments of quartzose sandstones deposited in a transgressive sea. Regardless of interpretations of environment, comparisons of lithologies and sedimentary structures of the two members indicate that the Chediski member represents an episode in sedimentation quite different and perhaps separate from that of the lower member.

The sands of the Chediski member of the Troy are quite different from any deposited previously in the younger Precambrian sequence and no part of the member, except certain included gravels of small volume, could have been derived from these older strata. If the sands were derived from a granitic terrane, as their composition suggests, much rounding of quartz grains and destruction and winnowing of feldspar grains must have occurred before the sands became available for incorporation into a marine(?) deposit.

The excellent rounding and characteristic pitting (frosting) of the quartz grains, and the moderate- to high-angle cross-stratification, noted especially in outcrops of the region north of the Salt River, might if only casually viewed be considered suggestive of eolian deposition. The massive contorted sandstone units, which comprise the bulk of the member in the northern part of the region, then might be regarded as dune deposits that slumped when saturated by waters of an encroaching sea. Pebbles of characteristic ventifact forms certainly confirm an eolian stage in the formation of the sands; it should be recognized that ventifacts are sparse, show signs of later abrasion, or are seemingly missing in many of the southern sections of the Chediski.

Other features negate an interpretation that the Chediski sandstone is an eolian deposit. According to Pettijohn (1949, p. 232-236) modern dune sands tend to be fine-grained and well-sorted; apparently no diagnostic differences in size, sorting or sphericity have been established in comparing modern beach and dune sands; and fluvial deposits--of the environments that may be applicable here--show the greatest ranges of grain sizes. The coarse fractions of the Chediski sandstone are definitely atypical of eolian deposits. The southern outcrops of the Chediski member, except for the degree of rounding and the abundant pitting of the grains, are virtually lacking in features suggestive of eolian origin. Irregular bedding, channeling, and lenses of conglomerate between beds could be indicative of fluviatile deposits. Many of the conglomerate lenses appear to be lag gravels concentrated by the winnowing of the sands from the tops of beds on which the gravels rest. These gravels tend to exist in sheets rather than as local small channel fills, suggesting concentration on a sea bottom. While these gravels were being deposited the upper formations of the Apache group were being vigorously eroded somewhere outside of the region of present outcrops, or coarse Apache debris previously formed was being transported into the area. Pebbly sandstone is common in the northern part of the region, but discrete layers or lenses of conglomerate are common only in the southern part of the region. The southern sandstone sections, taken as a whole, include less clay and sericite in the matrix. The variation in these aspects is gradational from north to south.

All features considered, the Chediski sandstone is possibly best visualized as a deposit laid down in a transgressing sea, which encroached from south to north and was so abundantly supplied with coarse quartz-feldspar sand that only crude partial sorting could be accomplished. Great volumes of sand and subordinate amounts of small gravels, rounded, mechanically etched and partially sorted as to composition by the action of wind, must have been available for deposition into this sea. And the sea bottom must have been a site of considerable agitation, to cause spreading of individual beds as extensive sheets and to form a blanket deposit of the extent of the Chediski member.

The quartzite member of the Troy offers still a different paleogeographic aspect. Throughout most of the region its contact with the Chediski member is sharp. The only seeming contradiction of this statement is seen in the southernmost outcrops of the Troy, where the feldspar and clay content of the upper part of the Chediski decreases and this part is also quartzitic. 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 mineral constituents.

There is no sound basis for judging the additional thicknesses of younger Precambrian strata that once overlay the Troy quartzite in southern Arizona. If the Apache group and Troy are correlative with the Unkar group of the Grand Canyon, as Darton (1925, p. 36) surmised, thicknesses on the order of 10,000 feet may have been removed. Certainly an additional cover of appreciable thickness once existed; otherwise thick sills of diabase now seen in the uppermost parts of the thickest sections of Troy quartzite would not have been emplaced in their usual forms.

After lithification of the Troy, the younger Precambrian strata were locally faulted and folded; thereafter extensively intruded and displaced by diabase intrusions, then uplifted regionally and deeply eroded in Precambrian time. Despite this long history of deformation the bulk of the younger Precambrian strata were essentially flat-lying when the first Paleozoic seas encroached across the region, and for most of the region that has been discussed in some detail they remained almost horizontal until the Laramide revolution. During Tertiary time the younger Precambrian formations, with their cover of Paleozoic and Mesozoic formations, were extensively faulted and tilted throughout the Basin and Range portion of the region. Farther north, in the Plateau portion, these strata remain almost horizontal to the present time. Only since the Laramide orogeny has erosion again exposed the Troy and Apache formations.

The bedding features and other internal structures and for some units the composition indicate that the Troy quartzite and the sedimentary formations of the Apache group were deposited on a shallow fairly stable continental shelf. Formations and even member units of the sequence are individually quite different one from another, suggesting abrupt changes in the environment of deposition, in the environment of the source of materials, or in the mode of transport. Some of the abrupt changes are marked by unconformities; in retrospect perhaps additional unconformities are yet to be recognized. In any case, the sequence suggests accumulation of geologic units over a very long period of time.

Age and correlation of the younger Precambrian formations

In the southern part of the region the Troy quartzite, the Apache group, and the included diabase intrusions are separated from Middle Cambrian strata by an unconformity so profound their assignment to the Precambrian is quite reasonable. In the northern part of the region, though by analogy similar age designations can be made (Shride, 1958), direct stratigraphic relations prove only that the diabase is pre-Devonian.

The absolute age that can now be estimated is determined in relation to the diabase. Some question has existed as to whether the diabase intrusions are all of the same age, therefore a brief review of the critical diabase examples is worthwhile. For reasons stated in the introduction to this report, the age of the extensive diabase intrusions has been variously recognized or inferred to be (1) Precambrian, (2) questionably Precambrian, (3) post-Cambrian but pre-Devonian, (4) post-Pennsylvanian and probably early Mesozoic or late Paleozoic (the original age designation of Ransome), or (5) Late Cretaceous or early Tertiary. The last age designation, based on inferred relations in the structurally complex mineral-bearing belts in and near southern Gila County, has been that most widely accepted. In these areas, where exposures of contacts are commonly not the best and faults and shear zones commonly obscure relations, large masses of Paleozoic strata are seemingly "foundered" in some of the larger sills. Also large diabase bodies that underlie Paleozoic strata locally show step-like discordant contacts with the Paleozoic formations, and faults associated with these steps suggest offsets attributable to inflation by diabase. Such features alone are not diagnostic, as has been presumed, of intrusive relations.

Where such features have been observed by me the discordant contacts between the Paleozoic formations and diabase are either obvious fault contacts or are not exposed adequately so that their nature could be directly determined. Nowhere have diabase sills both overlain and underlain by Paleozoic formations been recorded; this lack of a sill habit is atypical--wherever large masses of diabase adjoin or engulf Apache or Troy strata, sills in these formations are apparent. Where Paleozoic rocks are reportedly intruded by diabase, the Martin limestone is the formation most commonly in contact with the igneous rock. The Martin is partly of cherty dolomite beds, which are comparable to the cherty dolomites of the Mescal and should be converted pervasively to silicate-bearing limestones if intruded by diabase. Where intruded by quartz monzonite, siliceous dolomites of the Martin have proved quite amenable to such silication (Cooper, 1957, p. 582-537). But the siliceous dolomites of the Martin--even where exposed within a few inches of diabase--are completely lacking in contact-metamorphic minerals. For some such examples, careful search for fully exposed contacts in adjacent areas has resulted in proof that the Martin rests in sedimentary contact with the diabase body that supposedly intrudes it. I have not found a single example of a chilled selvedge where a diabase body abuts against Paleozoic rocks, yet fine-grained selvedges may be quite apparent where diabase of the same body is against Apache rocks in the same area. The engulfed blocks of Paleozoic strata are everywhere most plausibly interpreted as blocks faulted against diabase, rather than blocks torn loose from their original position on invasion by a diabase magma. No large masses of diabase have yet been conclusively demonstrated to post-date any of the Paleozoic rocks, and abundant evidence exists that they antedate the Paleozoic formations in all reaches of the region.

Ransome (1919, p. 53, 56) clearly stated that small dikes of diabase cut into Paleozoic formations in only 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) for at least one locality confirmed Ransome's observations of dikes intrusive into Paleozoic strata, but he took exception to Ransome's correlation of these dikes with the large diabase sills. Darton suggested that the small dikes were feeders for some of the Cenozoic basalt flows of the region. Such dikes must be extremely rare. I have made a particular effort to confirm reported examples, and have yet to find in a Paleozoic host rock a dike that compositionally or texturally resembles diabase.

The Precambrian age of the diabases in the northern part of the region has recently been confirmed by lead isotope determinations. In 1955 H. C. Granger and others of the 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 veinlets in the Dripping Spring quartzite, but some of the veinlets sampled transect quartzose aplite dikes associated with the principal diabase sill of the Sierra Ancha. Traced laterally, this sill is readily observed to transect the highest parts of the Troy quartzite, and to be overlain by Devonian strata. On the basis of Pb^{207}/Pb^{206} ratios, determined from both galena and uraninite, L. R. Stieff (written communication to H. C. Granger, January 26, 1956; Neuerburg and Granger, 1960, p. 775-776) estimated the age of the occurrences to be 1,100 million years. It is of interest regionally that this age coincides with those determined similarly from uraninite occurring in the younger Precambrian Belt series of Idaho (L. R. Stieff, oral communication, May 1958). By independent determinations of Pb^{207}/Pb^{206} ratios in uraninite from Workman Creek (locality 3½ miles northwest of Aztec Peak, McFadden Peak quadrangle) and like ratios in zircons collected from the granitic differentiate on Reynolds Creek, L. T. Silver (1960; and oral communication November 1960) confirmed Stieff's estimate as a minimum age. Silver postulates an age of 1,200 million years or more for the diabase. Thus, throughout southern Arizona the large sills of diabase and minor intrusions positively apophyses from these sills predate the Cambrian considerably, and the Troy and older formations may be considerably older than 1,200 million years.

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 west 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, is about 7,000 feet thick; the Chuar, or upper group, is 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 Stoyanow, 1936, p. 1994-2000) have taken exception to Darton's suggestion. Still, our knowledge of the Grand Canyon series--or of any other younger Precambrian sequence in the southwestern United States--is so scant that regional correlation of individual formations is impracticable. But some additional bits of information that can be scrutinized as potential bases for correlations can be enumerated.

Formations of the Apache group crop out through a north-south interval of 160 miles, and throughout this interval the formations are remarkably consistent in gross characteristics. Moreover, certain sequences of beds consistently are of diagnostic lithology or have distinguishing bedding characteristics. The southernmost exposure of the Grand Canyon series is only about 135 miles distant from the northernmost outcrop of the Apache group (see fig. 1), so some of the distinctive characteristics of the Apache group and Troy quartzite should 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), provide hints that comparable features may be found in the same order of sequence in the younger Precambrian sections of northern and southern Arizona. Near the base of the Bass limestone, which 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 are counterparts of conspicuous layers in the lower part of the Mescal that include abundant molds of halite crystals. Descriptions of massive irregular chert nodules, jasper, intercalated reddish or purplish shales and thin arkosic sandstones, and undulatory banded cherty limestones, and their interrelations and metamorphic equivalents in higher parts of the Bass limestone are certainly suggestive of the solution phenomena and related secondary features common in the upper parts of the Mescal dolomite and limestone sections. Limestones of the Chuar terrane, if--dubiously--the descriptions are adequate, illustrate few features in common with the Mescal limestone. Beds of stromatolitic limestones are noted and illustrated for the Bass limestone, but their relative stratigraphic position has not been stated. Is it significant that where sections of the Bass are thin, and particular note of solution(?) features suggests erosion of the top of the formation, no mention of algal structures has been made? E. D. McKee (oral communication, August 1959) has stated that the algal structures in the Bass limestone are of Collenia aspect. Moreover, to the best of his knowledge, the stromatolites that occur in the relatively thin limestone units of the Chuar are all of "large biscuit-forms", entirely different from those found in the Mescal.

The Hakatai shale, which overlies the Bass limestone in thicknesses variously stated in the range of 500 to 530 feet, is collectively characterized as comprised of argillaceous shales, jasper that is dense and hard and of various colors, blue slate, and fine-grained pink or red sandstone, quartzite or mudstone--descriptions that could fit the upper (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 upper 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 argillite of the Mescal (H. C. Granger, oral communication, October 1960). In content of several trace elements the two specimens show remarkable agreement. In no elements, 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 in features that can be observed megascopically. But they are quite different from siltstones of the upper member of the Dripping Spring, which also contain unusual amounts of potassium. If such data are representative, distinctive chemical characteristics may be an aid in correlating younger Precambrian formations of different parts of the Southwest.

Some aspects of the Shinumo quartzite, which overlies the Hakatai shale and is about 1,500 feet thick, are certainly reflected in the Troy quartzite. Although Noble (1914, p. 51-53) did note lenses of conglomerate, he characterized the sandstones and quartzites of the Shinumo as composed of well-rounded, well-sorted fine grains[/] of

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

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quartz. From one of Noble's statements it might be inferred, with some doubt, 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, purplish or reddish in color. Walcott (1894, p. 511) and Noble made special comment on "curiously twisted and gnarled" layers of white or purple sandstones. None of the descriptions suggest that quartzites like those of the upper member of the Troy comprise a distinct separate unit of the Unkar section. After viewing a representative suite of specimens from the Troy section described in table 4, E. D. McKee (oral communication, August 1959) advised that white, purple and mottled sandstones of the texture of the Chediski member are conspicuous in the Shinumo quartzite, and that arkoses and quartzites like those of the lower and upper members of the Troy are also represented in the Shinumo. The large-scale cross-stratification so prominent in the arkose member of the Troy had not been noted by Mr. McKee, but he had seen slump structures very like those in the Chediski sandstone.

The Dox sandstone, which overlies the Shinumo quartzite and is 2,300 to 3,000 feet thick, is apparently highly micaceous and commonly friable or shaly parting and, except for its basal beds, seemingly has no counterpart in the southern Arizona sections. Although the basalts of the Apache group are thin compared with the 800- to 1,000-foot sections of basalt noted near the top of the Unkar group, Apache basalts undoubtedly were much thinned by pre-Troy erosion. The sedimentary formations, especially those of the Chuar terrane, that overly the Unkar basalts are mostly thinbedded non-resistant strata unlike any of the Apache formations. Correlatives of the Dripping Spring and Pioneer formations, of course, have not been suggested in the above resume. Nor have past descriptions hinted of unconformities that should be recognized if the ideas presented have merit. It would seem, nevertheless, with the collection of a modest amount of additional information, that a good case for correlating the Apache and Troy formations with the Unkar group might be made.

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 (Hewitt, 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 to 250 miles west of the westernmost outcrop of the Grand Canyon series and about 350 northwesterly from the Apache outcrop area.

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

Noble (1934, p. 174), who has worked extensively in both the Grand Canyon and Death Valley regions, has commented that parts of the Pahrump series are strikingly similar to the Grand Canyon series. Apparently the Crystal Springs formation, in particular, includes units comparable to those in the Apache group. In May 1952 I was privileged to view 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. The Saratoga Hills area, 35 miles northwest of Baker, California, is the locality of the most detailed stratigraphic descriptions of the series yet published (Wright, 1952). A thick feldspathic quartzite unit, not seen by me, makes up the basal part of the Crystal Spring formation. Overlying this quartzite, and midway in the section, is a purplish shale or siltstone unit, which appears in outcrop remarkably like the siltstones of the Pioneer formation of southeastern Arizona. Overlying this shale, in the Saratoga Hills, is a dark-colored, brownish-weathering fine-grained quartzite unit with some similarities to the upper member of the Dripping Spring quartzite. And the dolomite and chert unit, that makes up much of the upper part of the Crystal Springs formation, has many features, including stromatolite beds, in common with the Mescal limestone. It would be presumptious, however, to suggest any correlation of the Apache group and Pahrump series at this time. As Hewitt (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, Texas, Richardson (1909) long ago recognized that the Bliss sandstone of Cambrian age is unconformably underlain by a structurally little disturbed Precambrian section. Extrusive rhyolite as much as 1,400 feet thick lies immediately below the Bliss sandstone. The rhyolite in turn unconformably overlies a sequence dominated by beds of quartzite and sandstone, 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) has added to our knowledge of the Lanoria quartzite and he has recognized that it 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, named the Castner limestone by Harbour. If the thickness of included diabase sills is subtracted from the type section measured by Harbour, the Castner limestone is exposed through a thickness of almost 800 feet. However, the base of the Castner 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, which is 2,600 feet thick, seemingly is not particularly like the Troy quartzite.

Strata of younger Precambrian aspect also are exposed near Van Horn, Texas, about 100 miles southeast of El Paso. Before arriving at a tentative correlation of these strata with those of the Franklin Mountains, King (in King and Flawn, 1953, p. 125-131) carefully reviewed the pitfalls in equating the formations, and considered the problems of correlating the Precambrian strata of west Texas with those of Arizona. Even so, in light of his later information Harbour felt justified in taking exception to King's correlations of formations between the Franklin Mountains and Van Horn areas.

The information that can presently be brought to bear on the problems of correlation is given in full in the cited publications. Without detailing it further, I would suggest that if limestone formations were not common to both the Texas and Arizona sections, we would not be so inclined to provisionally equate the sections. And if the structural and plutonic events were given more emphasis in comparisons, we might be additionally inclined to equate the west Texas strata with those that underlie the Apache group in Arizona. At least until additional bases for correlation are available, probably little weight should be afforded lithologic similarities of the Precambrian formations in the two regions.

Diabase intrusions, definitely of Precambrian age, are common in all the Precambrian terranes discussed above. However, if the relatively thin Apache-Troy terrane is excepted, the intrusions are restricted in stratigraphic distribution 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; 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. Does the inconspicuous unconformity that marks the boundary between the Chuar and Unkar groups (Walcott, 1895, p. 325), or one of the like unconformities stratigraphically not far distant from that contact (Van Gundy, 1951, p. 954-957), have more significance than is now appreciated? Somewhat similarly, in the Death Valley region diabase intrusions have been recognized only in the Crystal Spring formation; no hint of diabase has been found in the Beck Spring dolomite; and abundant diabase debris, possibly derived from the lower intrusions, occurs in the conglomeratic member of the Kingston Peak formation (Wright, 1952, p. 14-15). Greenstones that possibly represent diabase sills seemingly are restricted to the Allamoore formation of the Precambrian sequence in the Van Horn area of Texas (King and Flawn, 1953, p. 73-79). In the Franklin Mountains several thin diabase sills intruded the Castner limestone. These may predate the Lanoria quartzite and definitely are older than the Precambrian 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 rank with the oldest of the younger Precambrian formations in the Southwest.

Red rock differentiates, which might provide material suitable for isotope dating methods, have been noted in the thickest sill in the Unkar strata. Such differentiates, if they exist in the diabases of the Death Valley region, have not come to my attention. Descriptions indicate that some of the thicker diabase bodies of that region are comprised of multiple thin sills, in which granitoid differentiates would not be expected. Sills of the Franklin Mountains are also too thin. Greenstones of the Van Horn area are not likely sources of suitable material.

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