Geology of Northern Arizona
WITH NOTES ON ARCHAEOLOGY AND PALEOClimATE

PART I · REGIONAL STUDIES

For
GEOLOGICAL SOCIETY OF AMERICA
ROCKY MOUNTAIN SECTION MEETING, FLAGSTAFF, ARIZONA

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1974
FOREWARD

The time is ripe for a summary of what is known and unknown geologically about an intriguing region! The hundreds of pages of text and figures which follow testify not only to the enormity of the task of understanding an area as diverse as northern Arizona, but also to the range of talents and techniques which have been brought to bear.

Much that is here is in print for the first time, and reflects research carried out within the last five years or so. The meeting of the Rocky Mountain Section has served as a magnificent impetus to collect our ideas, and the endless devotion and patience of the editors and their colleagues have made this splendid volume possible. We are all in their debt.

Augustus S. Cotera
Chairman, Rocky Mountain Section
27th Annual Meeting
Northern Arizona has attracted pioneers in North American geology since the days of John Wesley Powell; when northern Arizona was considered to be a part of our "last frontier." Perhaps Powell, Darton, Dutton, Walcott, Noble, Gregory, Gilbert, Davis, or Robinson would not have been surprised to learn that the geology of northern Arizona would play an important part several generations later in the geologic exploration of a new frontier—that of space. Largely because of the variety of geologic terranes, good exposures, and the presence of geologic analogs of lunar features, Apollo astronauts received training in ten different field sites in northern Arizona, including operational rehearsals for lunar geologic traverses.

Because of the importance of the geology of northern Arizona to both the history of the science and to an understanding of the geology of North America, papers were solicited to provide an up-to-date synthesis of this geologically varied region. The contributors represent a variety of sponsoring institutions. Most are affiliated with the U.S. Geological Survey, Northern Arizona University, or the Museum of Northern Arizona; other universities and research agencies are represented, as are state agencies and mining companies.

This group of papers provides a new overview of the stratigraphic, structural, volcanic, fluvial, erosional, and ecologic histories contained in the geologic record of northern Arizona. New geophysical and isotopic information has sharpened analysis of the regional geology. Paleomagnetic, potassium-argon, and carbon-14 data are applied to new time-stratigraphic reconstructions and correlations from the Precambrian to the present. Contributions involving the convergence of geologic, archeologic, and biologic evidence provide new insights into the past climates of northern Arizona and into the effects of the changing physical environments on the prehistoric cultural history of the region.

Some of the new data and reconstructions go far toward solving long standing problems. Just as important, these articles indicate areas where additional work is needed for a better understanding of the regional geology. This understanding is essential to the effective development of the resources of northern Arizona, and for sound decisions concerning the interaction between man and his environment.

The economic importance of the geology of Arizona is well known, and northern Arizona continues to produce significantly from its mineral and energy resources. Gold, silver, copper, uranium, asbestos, and building materials are, or have been, important products. Coal is plentiful in the northeastern part of the state, and several nearby oil and gas fields are active producers.

The geology of northern Arizona is as interesting and complex as it is varied. A major structural transition occurs in the vicinity of the Mogollon Rim in central Arizona, where the structural style changes from that of the Basin and Range to that of the Colorado Plateau. In the Grand Canyon is a classic exposure of the stratigraphic sequence from the Precambrian crystalline basement through the Permian Kaibab Limestone, and further to the east in Black Mesa the section continues...
upward through much of the Cretaceous Mesa Verde Group. For those interested in structural geology, there are great fault systems and many folds, including classic monoclinal flexures. Some of the fault systems are still active, as shown by recent seismic events.

Volcanic provinces are extensive in northern Arizona. An interesting variety of mafic volcanic rocks includes the alkalic basalts of the San Francisco volcanic field, the sodic monchiquites of the Hopi Buttes, and the potassic minettes of the Arizona-New Mexico-Utah border area. Andesitic, dacitic, and rhyolitic domes, flows, and tuffs are common in the San Francisco Mountain and in the Hackberry Mountain areas. The last volcanic eruption, which formed Sunset Crater and the Bonito flow, occurred less than 1,000 years ago.

Tertiary and Quaternary lake deposits are abundant and well exposed in the Verde Valley and in the Hopi Buttes area. Tectonic activity and volcanic outpourings have caused major changes in drainage patterns and in stream regimen. Complex sequences of surficial deposits and erosional surfaces occur throughout the entire region. Because of excellent exposures, surficial deposits can be subdivided and regionally correlated, and the ecology of the past thousands of years can be determined through the combination of archaeologic, biologic, and geologic studies. Although Arizona commonly is regarded as a desert state, northern Arizona is a high plateau with mountainous areas, and alpine glaciation occurred on San Francisco Mountain during the Pleistocene. The best documented meteorite impact crater is Meteor Crater near Winslow.

The variety of geologic features and extensive rock exposures make northern Arizona one of the most attractive areas in the world for geologic studies by both interested laymen and professional geologists. It is with this interest in mind that this volume has been compiled.

Thor N. V. Karlstrom
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Address all communications to the editors.

Additional copies of this volume may be ordered from the NAU Bookstore, Box 6044, Northern Arizona University Flagstaff, Arizona 86001
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A PRELIMINARY REPORT ON THE OLDER PRECAMBRIAN ROCKS
IN THE UPPER GRANITE GORGE OF THE GRAND CANYON

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ABSTRACT

Older Precambrian rocks of the Grand Canyon are divided into the Vishnu Group, the Zoroaster plutonic complex, and the Trinity and Elves Chasm gneisses. The Vishnu Group is composed of materials that originally were sediments and minor amounts of mafic igneous rocks, but are now schists and gneisses. These rocks were complexly deformed during metamorphism and were imprinted with a well-developed schistosity that is now almost vertical in most parts of the canyon. They were intruded during and after regional metamorphism by granitic plutons of the Zoroaster complex. Most of these intrusives are probably of magmatic origin, but the last episode of activity produced many dikes and sills of pegmatite and aplite and was likely a hydrothermal rather than a magmatic event. The Trinity and Elves Chasm gneisses contain sparse interlayers of rocks possessing the Vishnu lithology and in places are gradational at their boundaries into the Vishnu Group. Such relationships suggest that these rocks were originally sedimentary. However, over large areas, the gneisses are homogeneous and of quartzmonzonitic to quartz dioritic composition, and may therefore be orthogneisses. Their origin remains unknown.

INTRODUCTION

The exposures of schist, gneiss, and granite in the Upper Granite Gorge of the Grand Canyon provide one of the earth’s finest exhibits of Precambrian igneous and metamorphic geology. These outcrops, although limited in breadth, are virtually continuous for more than 40 miles along the canyon and are cut at nearly right angles to the major geologic structures. Until recent years, the inaccessibility of this terrane was a major deterrent to investigation. However, with the advent of inflatable raft travel on the Colorado River, access to the sheer walls and river-polished slabs on which the geology is so clearly displayed is no longer a serious problem.

The period of major orogeny is dated radiometrically at about 1700 m.y. Later, zones of intense shearing developed. Movement in these zones offset metamorphic grades as much as several kilometers vertically and also caused local retrogressive metamorphism.

The earliest substantial work in the older Precambrian terrane is that of Noble and Hunter (1916), whose field and petrographic data have provided a sound foundation for later work. The igneous and metamorphic geology of the Inner Gorge was the subject of a series of papers by Campbell and Maxson published during the 1930's (Campbell and Maxson, 1933, 1935, 1938; Campbell, 1936, 1937) and culminating with the geologic map of the Bright Angel quadrangle (Maxson, 1968). Radiometric dates for the area became available in the late 1950's and early 1960's (e.g., Aldrich and others, 1957; Damon and Giletti, 1961; Pasteels and Silver, 1965). We began a detailed study of the
petrology and structure of this terrane in 1971. Some of the results have been presented in a report (Babcock and others, 1974) and on the new geologic map of the Grand Canyon (Billingsley, 1974). The account given here is largely a preliminary report of this recent work.

The crystalline rocks of this region may be conveniently divided into three groups: 1) widespread schists and associated lithologies, named the Vishnu Group (Babcock and others, 1974); 2) a suite of granitic rocks, intrusive into the Vishnu Group, herein informally referred to as the Zoroaster plutonic complex; and 3) two similar units of gneiss, seemingly of different origin than either the Vishnu or Zoroaster groups, herein informally referred to as the Trinity gneiss and the Elves Chasm gneiss. The distribution of schists, gneisses, and granitic rocks is shown in figure 1.

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Many students from Western Washington State College have assisted on the project: Wm. Lingley and M. Walen as Master's degree students, and M. Bernardi, J. Bradshaw, B. Ellis, L. Lathrop, C. Lyon, and W. Walker as field and laboratory assistants. All have contributed substantially to the progress of this work. The manuscript has been reviewed by Prof. D. E. Livingston and D. Shakel, both of the University of Arizona, and their constructive comments are much appreciated. Financial support has been provided by the National Science Foundation (Grant GA-31232 to Babcock and Brown), the Museum of Northern Arizona, Western Washington State College, the University of Leicester, and the Natural Environmental Research Council of Britain (Grant to Clark). We are grateful to the National Park Service for their cooperation, advice, and invaluable logistical support. Finally, we acknowledge the commercial river touring companies, and their boatmen, who have shown remarkable interest in our project and given generously of their time and resources.
Figure 1.—Distribution of predominant rock types and degree of regional metamorphism in the Upper Granite Gorge of the Grand Canyon. The width of outcrop along the Colorado River is schematic and exaggerated. This map is preliminary and subject to revision with completion of work in progress.
The predominant lithology in the Vishnu Group is quartzo-feldspathic and micaceous schist. Relict sedimentary bedding is conspicuously displayed throughout the Canyon, except in areas of highest metamorphic grade. Crossbedding was reported by Campbell and Maxson (1933) in Lonetree Canyon, and we have found clear relict graded bedding in Grapevine Canyon (fig. 2). Such features, together with the relatively quartzose composition of many of the schists and the nature of their chemical composition (Boyce, 1972), indicate that these rocks were originally sedimentary. The bulk of the sediment was no doubt polymict sandstone, siltstone, and some shale.

Mafic schists and amphibolites were originally estimated by Maxson (1968) to be much more abundant than we have found them to be. Our mapping and petrographic study have revealed only two areas where amphibolite predominates over leucocratic schist: Clear Creek and Cremation Creek. The rocks there are thus now considered to be part of the Vishnu Group, rather than a separate formation—the Brahma schist proposed by Maxson (see also Ragan and Sheridan, 1970). Though only a subordinate lithology, mafic schists and amphibolites are widely scattered throughout the older Precambrian terrane. In many places they form thin (<1 m) interbeds within, and are gradational into, schists of obvious sedimentary origin. Thus, these mafic rocks themselves are best interpreted as metasedimentary also. Some bodies may represent lava flows intercalated with the original sediments, but this has not yet been proven. Other mafic units are clearly of igneous intrusive origin, cross-cutting the structure of the schist (e.g., at foot of Walthenburg Rapid).

Calc-silicate rock, presumably derived from an original sedimentary carbonate rock, is fairly common in the quartzo-feldspathic schist. It occurs as lesser and discontinuous beds, a few tens of centimeters thick. A unit of this rock several tens of meters thick is exposed in the vicinity of Horn Creek Rapid.

In areas of very high grade metamorphism, for example at Pipe and Bright Angel Creeks, the more leucocratic parts of the Vishnu consist largely of migmatitic gneiss (fig. 3). Mineral assemblages indicate that some of these rocks may have been hot enough to melt, although other possible origins for the migmatites cannot yet be ruled out.
Figure 2.—Graded bedding in Vishnu schist. The lighter part of each bed was originally sandstone, and is now quartzo-feldspathic schist. This grades upward into what was originally clay-rich sediment and is now porphyroblastic mica schist. In Grapevine Canyon.
Figure 3.—Migmatitic Vishnu gneiss, near Pipe Creek.
ZORASTER PLUTONIC COMPLEX

The Vishnu Group is interspersed with granitic bodies in many parts of the Upper Granite Gorge (fig. 1). These rock units, named Zoroaster Granite by Maxson (1968), are here designated the Zoroaster plutonic complex in recognition of their variable lithology and the fact that they all are intrusive igneous bodies. The rock bodies themselves vary widely in size, shape, composition, and degree of metamorphism. For this discussion, the rocks of this group are divided into large plutons, granitic dikes and sills, and pegmatite/aplite dikes and sills.

Large plutons

Ten large plutons have been mapped from canyon exposures. Exposure widths range from 1/2 to 8 km. The plutons are grouped into two broad compositional types: a) quartz-monzonite to granodiorite (rarely granite) relatively poor in mafic minerals; and b) granodiorite to quartz diorite (rarely diorite) relatively rich in mafic minerals. The plutons, their locations (along the Colorado River or in side canyons), and their compositional type (a or b) are as follows:

1. Zoroaster pluton; exposed from mile 84.6 to 85.7; type a.
2. Cremation pluton; exposed near mile 86.2; type a.
3. Phantom plutons; exposed in Phantom Canyon; type a.
4. Pipe Creek pluton; exposed in Pipe Creek and along the Colorado River downstream of Pipe Creek; type a.
5. Horn pluton; exposed below Horn Creek Rapid, near mile 90.6; type b.
6. Boucher pluton; exposed above Boucher Creek Rapid, near mile 96.0; type a.
7. Crystal pluton; exposed above Crystal Creek Rapid, near mile 97.6; type a.
8. 99 Mile pluton; exposed between Crystal and Tuna Creeks, at mile 99; type b.
9. Tuna pluton; exposed in Tuna Creek Canyon; type a.
10. Ruby pluton; exposed from mile 102.7 to 107.9; type b.

The Zoroaster pluton is described in some detail by Lingley (1973) and the Phantom and Crystal plutons by Walen (1973). Field, textural, and chemical analytical data all strongly suggest an intrusive igneous origin for these plutons. Xenoliths are locally conspicuous, and intrusive contact breccias occur along the margins of some of the plutons (fig. 4). Also, a contact metamorphic aureole is developed around some of the bodies. This is especially clear at the downriver contact of the Ruby pluton. Many of the granitic rocks show a metamorphic fabric. However, where this is absent, the plutons have igneous textures, such as normally zoned plagioclase crystals and hypidiomorphic texture. The chemical analytical data indicate that the plutons have typical igneous compositions that plot near the cotectic in the quartz-orthoclase-plagioclase-H₂O system.
Figure 4.—Intrusive breccia near margin of Tuna Creek pluton.
Some of the plutons have enormous elongate septa of Vishnu-type rocks extending through the middle of them. In the Zoroaster pluton, the septa are amphibolite, and one of these, some 5 m wide, extends more than 1,700 m through the pluton (Lingley, 1973). Foliation in both the Zoroaster pluton and amphibolite septa is parallel to the contact between the two rock types. Such septa are reasonably interpreted to be the result of syntectonic igneous intrusions. However, septa in the Phantom plutons and in the Crystal pluton must have orginated differently. These septa, also only a few meters wide and more than 100 meters long, are schists of metasedimentary origin as well as amphibolite. They could be huge xenoliths or roof pendants in the intrusive magma, but it is hard to imagine a mechanism of intrusion that would be passive enough to allow preservation of such bodies.

Judging from textures in the plutons, emplacement occurred during and possibly after regional metamorphism. The most strongly foliated of the plutons, and thus possibly the first to be emplaced, is the Zoroaster body. The pluton least affected by metamorphism is the Boucher pluton. This body is intruded into rocks of relatively low metamorphic grade (greenschist facies). It was emplaced either after metamorphism or low-grade metamorphism. Some parts of the plutons in Phantom Canyon are not foliated, even though the grade of regional metamorphism in this area is very high (upper amphibolite facies). Thus, intrusive activity here appears to have outlasted regional metamorphism. The Ruby pluton, exposed from some 8 km along the Colorado River, is emplaced into a high-grade metamorphic terrane at the upstream contact and is itself metamorphosed. However, at the downstream contact, the host schist is of low regional metamorphic grade (greenschist facies) and has been contact metamorphosed by the pluton. The pluton in this area is unmetamorphosed. Two possible interpretations of these relationships are that: 1) first, low-grade regional metamorphism affected the area; then, the pluton was intruded; finally, high-grade regional metamorphism affected the upstream part only; 2) the pluton was intruded during a single phase of regional metamorphism that affected the whole area. The intrusion occurred in a zone of metamorphic gradient. In the high-grade area the forces of regional metamorphism were strong enough to modify the pluton; at the low-grade contact, they were not, and the pluton metamorphosed the schist. There is no evidence elsewhere of two periods of regional metamorphism, and thus we favor the second hypothesis.

An age of 1,695 m.y. for the Phantom plutons and an age of 1,725 m.y. for the Zoroaster pluton have been determined by Pasteels and Silver (1965) using U and Pb isotopes.
Granitic dikes and sills

Dikes and sills of granitic material form a second type of rock body in the Zoroaster plutonic complex. Unlike the pegmatites and aplites (described below), these rocks are medium-grained, homogeneous bodies of quartz-monzonite to quartz diorite composition. They are particularly abundant in some areas of the highest grade of regional metamorphism (such as from about mile 77 to 79, and in Bright Angel Creek). The dikes may be genetically related to the larger intrusions, but there is no discernible spatial relationship. Some of the dikes are clearly intrusive, whereas others have diffuse contacts with the host schist and seemingly are of replacement origin (fig. 5).

Pegmatite/aplite dikes and sills

Dikes and sills of pegmatite and aplite (fig. 6) are found throughout the older Precambrian terrane and occur in all rock types. They are largely post-tectonic and later than the granitic plutons (fig. 7), although in some areas (e.g., Monument Creek) a metamorphic fabric extends from the host schist through the pegmatite. These layers range in size from a few centimeters to tens of meters in thickness and up to several hundred meters in length. In percent of the total rock of any given area (as large as a few hundred meters along the river) they range from 0 to 100. For many miles along the Colorado River 40 to 50 percent of the total rock is pegmatite (fig. 1). The greatest concentration of pegmatite is found in areas of high metamorphic grade, and virtually none is present in rocks of the greenschist facies.

Both the pegmatites and associated aplites are muscovite-bearing quartz-monzonites and granodiorites. Most of the pegmatites are strongly zoned, with concentrations of quartz and large K-feldspar crystals in the core (fig. 8), and smaller grains of plagioclase, quartz, K-feldspar and muscovite in the edge zones. Grains are up to 50 cm across and commonly elongate normal to the margin of the pegmatite, forming a comb structure. All rock types in contact with this granitic material show effects of K-metasomatism (see section on metamorphism), and a grade of contact alteration within the amphibolite facies, as indicated by the new growth of epidote and oligoclase in the contact zone. Commonly the host schist shows no evidence of forceful displacement by pegmatite intrusion. Where there are numerous scattered pieces of schist in an area of predominant pegmatite, all have the same orientation; only rarely has the pegmatite pried apart or rotated blocks of schist.

Clearly the pegmatitic material did not come from the host but was introduced from outside. This is indicated by the constancy of pegmatite composition regardless of host, and by the contact effects. The temperature during emplacement was appropriate to the amphibolite
Figure 5.—Diffuse contact of granite and schist, suggesting that the granite has replaced rather than intruded the schist. Hance Creek.
Figure 6.—View across Inner Gorge from Cremation Creek, showing pegmatite dikes crossing amphibolite.
Figure 7.—Typical cross-cutting relationships: granite cuts schist, and pegmatite cuts both schist and granite. Inner Gorge, near Cottonwood Creek.
Figure 8.—Pegmatite dike with quartz core and giant K-feldspar crystals. The crystals are approximately 50 cm long and extend from the edge of the dike into the core, where faces in contact with quartz are well developed. Near mile 78 on the north bank of the Colorado River.
facies. The emplacement medium allowed zonal and comb structures to form, and giant crystals to grow, and was thus probably predominantly an aqueous fluid rather than a silicate melt. The emplacement appears to have been a relatively passive process whereby large crystals grew freely into fissures filled with some type of alkali-silicate-bearing aqueous fluid.

THE TRINITY AND ELVES CHASM GNEISSES

The Trinity and Elves Chasm gneisses crop out from miles 91.5 to 92.9 and 112.4 to 118.7, respectively. These gneisses were first observed by Noble and Hunter (1916), who suggested that they might be older than the Vishnu.

The gneisses are mostly quartz monzonite to quartz diorite and are homogeneous in composition throughout large areas. However, a distinctive feature of both units is that they contain (well within the main body of gneiss) layers of Vishnu-type rocks. Quartz-rich schist and calc-silicate rock occur as discontinuous layers 5 to 50 cm thick (fig. 9). Amphibolite occurs both as concordant layers and in cross-cutting dikes. Of the two gneisses, the Trinity has a greater amount of the Vishnu-type metasedimentary rock, and the Elves Chasm unit has more amphibolite dikes. Both units are cut by numerous pegmatite/aplite dikes and sills.

The contact of the Trinity gneiss with the Vishnu Group is sharp and folded in places (well exposed on south bank of Colorado River at mile 92.3). However, in Monument Creek the contact is gradational in a distance of 50 meters from the gneiss through a zone of migmatites into Vishnu schist. At Walthenburg Rapids, the transition from schist into Elves Chasm gneiss is gradational through a distance of 100 meters. It is marked simply by a gradual transition from mica schist through coarse garnet-amphibole schist into garnet-bearing quartz-feldspathic gneiss, and finally into typical granodioritic Elves Chasm gneiss.

The age and origin of these rock units remain unknown, but work on these problems is underway. Two hypotheses reasonably consistent with the field and petrographic data are: 1) that the gneisses represent metamorphosed dacitic and andesitic extrusive deposits, in part intercalated with the Vishnu sediments; 2) that metasomatism on a regional scale produced the gneisses from schists of the Vishnu Group.
Figure 9.—Elves Chasm gneiss with interlayer of Vishnu-type calc-silicate rock. Near mile 114.
METAMORPHISM

The schists and gneisses of the Grand Canyon were formed primarily by regional metamorphism. Later, they were locally recrystallized by the intrusion of granitic plutons and local shearing. The metamorphic grade ranges from greenschist to upper amphibolite facies, but rocks of middle to upper amphibolite facies predominate. The mean grade of various rocks at different locations in the canyon is shown in figure 1. Metamorphic crystallization took place during and after regional penetrative deformation. Separate episodes of mineral growth and rock deformation can be deciphered for local areas, but as yet, no regional sequence of multiple metamorphic events can be recognized, and for this reason all the metamorphic activity is assigned to one broad period of orogenesis.

Mineral assemblages and metamorphic grade

Metamorphic mineral assemblages are summarized in table 1. More assemblages exist than are listed, but all those not shown are subassemblages of those listed. Based on these assemblages, the metamorphic grade is interpreted to range from the greenschist to upper amphibolite facies. The best indicator of grade in this terrane is the mineralogy of micaceous and quartzo-feldspathic rocks. Phase relations in such rocks are summarized in figure 10. The inferred sequence of reactions is much like that derived from other studies (cf. Albee, 1972). Some areas on the phase diagrams are left blank because the assemblages which would ordinarily plot there have not yet been found. These blank areas reflect a general absence of Al-rich rocks in this terrane. The reactions shown on figure 10 all involve muscovite, quartz and H\(_2\)O in addition to the minerals depicted on the AFM diagrams. Other prograde reactions have occurred, but these have not yet been documented in detail.

Some of the assemblages of table 1 contain one more phase than should theoretically be present for divariant equilibrium. Numbers 10 and 14 are two such assemblages. Ten is present at Clear Creek and has been described by Lingley (1973); 14 is common in Monument Creek and in 94 Mile Creek and has been reported by Walen (1973). Both may represent disequilibrium, or they may be regarded as equilibrium assemblages containing both the reactants and products of reactions that are not strictly discontinuous as written in figure 10 but proceed over a limited P-T range (Evans and Guidotti, 1966).

The maximum metamorphic grade in Grand Canyon is indicated by the breakdown of muscovite. No area has yet been found where the muscovite to K-feldspar-sillimanite transition has occurred. In Bright Angel Creek, muscovite is generally absent, but the K-feldspar-sillimanite assemblage is not found there either. In the vicinity of Monument Creek, all combinations of sillimanite, K-feldspar, and muscovite are found. Our tentative conclusion is that such areas attained a maximum
Table 1.—Mineral assemblages in metamorphic rocks of the Grand Canyon. All assemblages contain quartz and plagioclase. Common accessory minerals are: tourmaline, rutile, sphene, zircon, apatite, iron oxides, and sulfides. gfs = quartzo-feldspathic schist, am = amphibolite, cs = calc-silicate rock, mig = migmatitic gneiss, Zp = Zoroaster pluton.

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Figure 10.—Summary of assemblages and inferred reactions which are indicative of metamorphic grade in the Grand Canyon. G = garnet, B = biotite, C = chlorite, St = staurolite, S = sillimanite, Cd = cordierite, M = muscovite, Q = quartz, K-feld = K-feldspar. All assemblages on AFM diagrams contain quartz, muscovite, and $H_2O$ besides the minerals shown. Numbers in the AFM diagrams correspond to assemblages in table 1. Blank portions of the diagrams are incomplete because the appropriate assemblages have not yet been found.
degree of metamorphism at pressures and temperatures close to the experimentally determined reaction line for muscovite + quartz = sillimanite + K-feldspar (Evans, 1965). Another constraint on the P-T conditions is the occurrence of muscovite in granitic bodies that were penecontemporaneously intruded during this high-grade metamorphism. These bodies must have been emplaced at pressures above that of the muscovite + quartz breakdown curve. Further limitations on the possible P-T conditions of metamorphism are provided by the absence of kyanite and by the presence of sillimanite and cordierite in this terrane. Comparison of these relationships with experimental data (Evans, 1965; Richardson and others, 1969; Holdaway, 1971; Seifert and Schreyer, 1970; Luth and others, 1964) suggests a pressure of 5 ± 1 kb and a temperature of 675 ± 25°C for the highest grade of metamorphism. Such conditions would correspond to a maximum depth of burial of about 17 km and an average geothermal gradient of about 40°/km during this metamorphism.

**Metamorphic structures**

The predominant structural expression of metamorphism is a subvertical S₁ schistosity which in places is axial-planar to F₁ folds (fig. 11) of variable plunge and orientation. A lineation (L₁) formed by the alignment of elongate minerals is parallel to F₁ fold axes. The F₁ folds recognized are as much as 20 meters in amplitude. Larger folds probably exist but have not yet been observed. Excellent examples of F₁ folding can be found at the mouths of Clear Creek and Travertine Canyons, as well as in Boucher, F. J. (mile 102.7), Copper, and Walthenburg Canyons (fig. 12). F₁ folds are especially well developed in Walthenburg Canyon, where they occur on all scales up to about 15 meters in amplitude. The largest can be seen on both sides of the creek bed below a 15-meter waterfall and are fairly simple upright isoclines of "similar" style whose axes plunge 40° towards the east. The folding is defined by quartz-rich bands in a quartz-biotite schist. Both the schistosity, S₁, and the folds have been weakly refolded by a second phase of deformation, F₂.

Post-F₁ structures are common and are easily recognizable because they deform the predominant schistosity (S₁). In places a new strain-slip cleavage, S₂, is developed parallel to the axial planes of the second generation folds, F₂ (fig. 13). S₂ is marked by both a realignment of existing minerals and the growth of new platy minerals. Crenulation lineations, L₂, are parallel to the axes of F₂ folds. At some localities (e.g., Clear Creek, Boucher Creek, Walthenburg Canyon) more than one phase of post-F₁ folding can be recognized on the basis of overprinting relations, and kink bands commonly are the final expression of deformation. Large-scale F₂ folds exist and are relatively easy to map on the basis of attitudes of refolded S₁. A structure that deforms S₁ into a dome-like pattern in the area from Zoroaster Creek to Cremation Canyon has been recognized by Linglory (1973).
Figure 11.—$F_1$ fold hinge showing strong development of schistosity ($S_1$) through the hinge zone, parallel to the fold axial plane. Walthenburg Canyon.
Figure 12.—Large recumbent $F_1$ fold in side canyon on the south bank of the Colorado River near mile 102.7. The horizontal schistosity here is not typical for the Grand Canyon as a whole. Note hammer for scale.
Figure 13.—$F_1$ folds refolded by $F_2$. $S_2$ is weakly developed parallel to the axial planes of the $F_2$ folds. Walthenburg Canyon.
The latest episode of deformation before deposition of the younger Precambrian rocks is marked by zones of intensive shearing and local recrystallization of igneous and metamorphic rocks. Many of these shear zones remained active during later Precambrian and Paleozoic times.

Across two of these shear zones, at Vishnu Creek and at mile 95.7 (between Hermit and Boucher Creeks), the grade of regional metamorphism changes abruptly. In both localities the shear zones have brought high-grade rocks on the east against lower grade rocks on the west. Assuming that the degree of metamorphism reflects formation of the rocks at different crustal levels, a vertical movement of at least several kilometers can be inferred for these faults, with the east side up relative to the west side.

Metamorphic history

In many areas of the Canyon mineral growth occurred in multiple pulses, as shown by metamorphic textures. For example, in Clear Creek two generations of biotite occur, each defining a separate foliation (Lingley, 1973). At Hermit Creek, muscovite is developed both along the $S_1$ foliation, and as giant porphyroblasts (5 cm across) cross-cutting the foliation (fig. 14). Near Boucher Creek, low-grade schists contain rotated garnets that formed during development of $S_1$, and late chlorite porphyroblasts that overprint $S_1$. In general, mineral crystallization occurred during and outlasted penetrative deformation. The main schistosity throughout the area, $S_1$, appears to be the product of a single episode of deformation and mineral growth. Whether this was everywhere simultaneous is unknown. Separate later pulses of mineral growth and rock deformation, recognizable for small areas, have not yet been correlated on a regional basis. Pending further work, our conclusion is that only one major episode of regional metamorphism, with minor local fluctuations, has affected the area.

Contact metamorphism around some of the granitic rocks has caused growth of micas on earlier regional metamorphic assemblages. The amphibolites and schists in contact with pegmatite have become greatly enriched in mica; biotite has replaced hornblende in amphibolites, whereas muscovite has grown in place of plagioclase in the schists. This alteration is obviously metasomatic, involving movement of potassium into the host amphibolite and schist. In what appears to be a thermal contact effect, mica schist has been extensively recrystallized within 300 meters of the large granodiorite intrusive at Bass Rapids. This recrystallization has increased the grain size of the schist and has also obscured the earlier formed schistosity.
Figure 14.—Giant, post-tectonic muscovite porphyroblasts in Vishnu schist, near Hermit Creek.
Iron-rich chlorite is a common constituent of many rocks in the Canyon. The presence of this relatively low temperature mineral in a terrane of high-grade regional metamorphism is attributed to a retrogressive event or events. This is clearly evident in thin sections showing biotite and garnet in all stages of alteration to chlorite. In some rocks, biotite has been completely replaced by chlorite and potassium feldspar, an alteration that occurs only at temperatures below about 400°C (fig. 15). Degree of chlorite development is related to proximity to late shear zones. For example, chlorite is abundant in rocks near the Bright Angel and Vishnu faults, and the retrograde effects also extend beyond the zones of visible shearing. Thus, this low-grade metamorphism was probably associated with faulting. Almost certainly considerable water moved through the shear zones during and after faulting, and this facilitated recrystallization and also allowed the chlorite-producing hydration reactions to take place.

SUMMARY

Clastic sediments with minor carbonate and volcanic rocks were subjected to regional metamorphism and plutonism about 1,700 m.y. ago. The schists, gneisses, calc-silicate rocks, and amphibolites that were produced—the Vishnu Group—range in metamorphic grade from the greenschist to uppermost amphibolite facies. The predominant metamorphic structure is a schistosity, generally nearly vertical, which in a few localities can be seen to be cogenetic with nearly isoclinal folds. Later folding of the schistosity itself occurred in several relatively minor pulses of deformation. Two broad units of gneiss of uncertain origin show gradational contacts with the Vishnu Group in places, and share its metamorphism. They are termed the Trinity and Elves Chasm gneisses. Plutonism developed during, and outlasted, the regional metamorphism. It produced the Zoroaster plutonic complex—rocks ranging in composition from true granite to diorite and variable in fabric from directionless to gneissic. Most of these bodies are clearly of intrusive origin. Pegmatites and aplites, generally the last phase of this activity, are very abundant and were probably deposited by hydrothermal solutions. Metamorphism and plutonism were followed by development of shear zones, which offset metamorphic zones by several kilometers vertically and locally resulted in retrogressive metamorphism.
Figure 15.—Biotite porphyroblast replaced by chlorite (C) and K-feldspar (K), near the Vishnu Creek shear zone.
REFERENCES CITED


PRELIMINARY REPORT ON THE UNKAR GROUP (PRECAMBRIAN) IN GRAND CANYON, ARIZONA

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The Unkar Group is exposed extensively along the big bend of the Colorado River in eastern Grand Canyon and in smaller areas downstream around Bright Angel Creek and Muav Canyon 16 to 48 km (10 to 30 mi) to the west. The strata rest nonconformably upon, or are faulted down against, crystalline basement rocks and are overlain disconformably by the Precambrian Nankoweap Formation or with angular discordance by Cambrian strata. The Unkar Group includes 1,525 m (5,000 ft) of sedimentary rocks and the overlying 274 m (900 ft) of lavas. Sedimentary formations in ascending order are: Bass Limestone—57 to 100 m (187 to 327 ft) of dolomite, argillite and intraformational breccia or conglomerate including the basal Hotauta Conglomerate; Hakatai Shale—168 to 290 m (550 to 950 ft) of mudstone, sandstone, and shale; Shinumo Quartzite—345 to 412 m (1,130 to 1,350 ft) of quartz sandstone and subarkose; Dox Sandstone—951 m (3,120 ft) of quartz sandstone, siltstone, and mudstone. Four members are recognized in the Dox. The Unkar Group is interpreted as a very shallow water to locally subaerial sequence of lagoonal, tidal flat, and deltaic sediments with most clastic components coming from the east.

Precambrian sedimentary rocks in Grand Canyon were first reported by Powell (1874, 1875, 1876), who considered them to be Silurian in age. Walcott (1894, 1895) first named and described the Grand Canyon Series in detail and divided the sedimentary succession into two terranes: the Unkar (below) and the Chuar (above). He reported a thickness of some 3,660 m (12,000 ft) for the entire Grand Canyon Series, of which the lower 2,073 m (6,800 ft) was assigned to the Unkar Terrane.

Noble (1914) refined Walcott's descriptions, referred to the Unkar and Chuar as groups, and recognized five formations within the Unkar: 1) the Hotauta Conglomerate, 2) Bass Limestone, 3) Hakatai Shale, 4) Shinumo Quartzite, and 5) Dox Sandstone. The Unkar Group was subsequently studied by Van Gundy (1934, 1946, 1951) who assigned the topmost part of Walcott's Unkar succession to the Chuar and separated out the next unit below (Unkar division l b-e, Walcott, 1899, p. 216) as the Nankoweap Group, because it is bounded by unconformities above and below. Maxson (1961, 1967) mapped the Precambrian units in eastern Grand Canyon and referred to the Nankoweap as a formation but included it in the Unkar Group. We consider the Nankoweap to be a formation between the Unkar and Chuar Groups. The Chuar Group has recently been studied and divided into formations by Ford and Breed (1973).
Between the Unkar sediments and the Nankoweap Formation is a succession of basaltic lavas nearly 305 m (1,000 ft) thick. These rest conformably on the Unkar strata and have been called the Cardenas Lavas by Keyes (1938, p. 110) and Ford and others (1972) and the Rama Formation by Maxson (1961, 1967). The Cardenas Lavas (Hendricks and Lucchitta, this volume) are treated as the uppermost unit of the Unkar Group.

Thesis studies by several students from Northern Arizona University were recently completed on the following units of the Unkar: Cardenas Lavas (Hendricks, 1972), Bass Limestone (Dalton, 1972), Dox Sandstone (Stevenson, 1973). T. M. Daneker and V. S. Reed are currently working on the Shinumo Quartzite and Hakatai Shale, respectively. This paper is a progress report of the new data currently available on the Unkar sedimentary units. More complete papers presenting formal new names are in preparation. Names and age designations of the Precambrian strata in the Grand Canyon are summarized in figure 1.

GENERAL STRATIGRAPHY

The Unkar Group consists of about 1,525 m (5,000 ft) of mainly fine-grained red-brown sandstones, siltstones and shales, with minor dolomite and conglomerate, and is capped with 305 m (1,000 ft) of basaltic lavas. Five formations can be recognized in the group (fig. 1). A sixth unit, the Hotauta Conglomerate, is only locally distinguishable from the Bass Limestone and is here considered a member of the Bass.

The Unkar is exposed in three main areas of the Grand Canyon: 1) Near the mouth of Muav Canyon and adjacent smaller canyons tributary to the Colorado on both sides of the river within the northern Havasu Point and southern Powell Plateau quadrangles. This area is the Shinumo quadrangle of Noble (1914) in central Grand Canyon, where all the type sections for the sedimentary units of the Unkar are located (fig. 2). 2) Along and near Bright Angel Creek (fig. 2) in eastern Grand Canyon, where the Bass, Hakatai, and Shinumo formations are well exposed and readily accessible. 3) A large area extending from Hance Creek upstream to Lava Canyon within the Vishnu Temple quadrangle around the big bend in eastern Grand Canyon. In this last area, the Dox Sandstone crops out widely in the inner Canyon floor, the Cardenas Lavas are exposed in thin belts in the northeastern part, and the Bass, Hakatai, and Shinumo formations are exposed near river level in the west. Additional isolated small outcrops of the Unkar Group, mainly Dox Sandstone, occur at Nankoweap Valley and Clear Creek in eastern Grand Canyon, and at Tapeats Creek in central Grand Canyon.

The Unkar Group strata are everywhere gently tilted and are downfaulted against or rest nonconformably upon older Precambrian schist and granite. The Unkar is overlain disconformably by red-brown sandstones and siltstones of the Nankoweap Formation, or is bevelled by the erosion surface beneath the flat-lying Tapeats Sandstone of Cambrian age.

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| Figure 1.—History of Precambrian nomenclature in the Grand Canyon.
**BASS LIMESTONE**

The Bass Limestone, basal unit of the Unkar Group, was named by Noble (1914) for exposures in Bass Canyon in central Grand Canyon. Noble established the type section in Hotauta Canyon, where the Bass is 93 m (305 ft) thick. Van Gundy (1934) studied the Bass in the eastern part of the Grand Canyon and reported it to be a more complex lithology than that reported by Noble to the west. Dalton (1972) measured and described nine sections of the Bass, including almost all of the known outcrops in the Grand Canyon area.

The Bass lies nonconformably on the Vishnu Schist and is overlain conformably by the Hakatai Shale. The predominant lithology is dolomite with subordinate amounts of arkose, sandy dolomite with intercalated shale and argillite, and intraformational breccias or conglomerates that grade laterally into a dolomite and shale sequence to the northwest. Dalton (1972) found no limestone in any of the sections examined. He recommended that the Hotauta Conglomerate, which occurs locally at the base of the Bass, be included as a member of the Bass Limestone, and in this paper the Hotauta is treated as an informal member of the Bass.

The Bass ranges in thickness from 57 m (187 ft) at Crystal Creek to a maximum of 100 m (327 ft) at Phantom Creek and generally thickens to the northwest, except in the Crystal Creek section. Regional dip is to the northeast, an average of 11°. The distance between the easternmost outcrop (Hance Rapids) and westernmost section (Tapeats Creek) is approximately 61 km (38 mi).

The contact of the Bass with the crystalline basement rocks is distinct and can be located readily in the field. The upper contact with the Hakatai Shale is in places gradational but is generally marked by topographic and color differences. The color of the Bass is typically brown to red brown, in contrast to the brilliant orange-red color of the Hakatai. The Bass commonly forms a cliff, whereas the Hakatai forms a slope, except at Crystal Creek, where the Hakatai is also cliff forming and the Bass-Hakatai contact is gradational. Dalton (1972) placed the upper boundary at the highest dolomite and/or stromatolite-bearing bed beneath orange mudstone beds typical of the Hakatai.

**Petrology**

Dolomite is present throughout the known Bass outcrops. The beds are readily distinguished in the field by their resistant cliff- and ledge-forming character. The dolomite occurs as dolomicrite, clasts, and pore filling, but most is dolomicrite. The thick cliff-forming dolomite units become more cherty and stromatolitic to the west, with chert commonly replacing rock having high concentrations of stromatolites.
Argillite units, common throughout the Bass, are typically red or blue in color and thicken to the west. Dalton (1972) has interpreted the blue argillite, which is found only in the western outcrops, to have been deposited in deeper water than the red argillite.

Quartz grains occur in the dolomite units and locally are sufficiently abundant to form sandstone, particularly in the eastern sections where the quartz grains are of medium sand size, subangular to subrounded, and poorly sorted. Grains of detrital feldspar are present in small amounts and chert clasts are common in the sandstone beds in eastern Grand Canyon.

The basal conglomerate (Hotauta) typically consists of an assortment of gravel-size clasts of chert, composite quartz, zoned plagioclase, and micropegmatites, as well as sand-size clasts having abraded quartz overgrowths. The intraformational conglomerates and breccias within the Bass differ from the basal Hotauta conglomerate in containing a higher percentage of micritic lithoclasts.

Noble (1914) reported cubic crystal casts, presumably from salt, in a chert matrix of the Bass at Shinumo Creek. Dalton (1972) observed monoclinic rather than cubic casts in chert at Shinumo Creek. The casts appear to be polycrystalline dolomite that has replaced gypsum. No evaporites are now present in the Bass, but some of the breccia units appear to be the result of collapse, perhaps caused by removal of earlier formed gypsum.

Sedimentary structures

Symmetrical ripple marks are common throughout the Bass Limestone. Dalton (1972) concludes that they indicated oscillatory waves moving in a dominant west to east direction. Desiccation cracks of various sizes, abundant throughout the Bass, are commonly associated with intraformational conglomerates and are considered indicative of intertidal to supratidal environments. Small-scale normal and reverse graded bedding was observed in several thin sections. Many of the graded units are associated with the stromatolites, an association suggestive of a tidal flat environment.

Geologic history

The sea in which the Bass was deposited transgressed from the west across an eroded metamorphic and igneous terrane of the Vishnu Schist and Zoroaster Granite. This early transgression is recorded by the basal Hotauta conglomerate member. The unconformity at the top of the Hotauta and its locally restricted occurrence suggest that the marine transgression was interrupted by periods of regression and subaerial exposure.

As the marine transgression continued, carbonate deposition became predominant. A stromatolitic carbonate sequence having only minor contributions of land-derived sediments from the east was formed. Textural analyses of the Bass dolomites indicate a normal trend of deep-water mudstones grading eastward into more shallow deposits of
stromatolites and shallow-water mudstones. Transgression was eventually followed by a slower, regressive phase of sedimentation. This regressive depositional period is characterized by numerous ripple marks, mud cracks, and deposits of oxidized shaley materials indicating intervals of subaerial exposure and deposition of some land-derived sediments. Evaporite-forming conditions probably existed during this regression, as suggested by collapse breccia and pseudomorphs of dolomite after gypsum. Deposition of deltaic shales (Hakatai) followed the Bass sedimentation as the seas regressed farther to the west.

HAKATAI SHALE

The Hakatai Shale is the most colorful formation in the Grand Canyon, with colors ranging from bright orange to purple. It is easily eroded and commonly forms gentle to moderate slopes. Along with the Shinumo Quartzite, it has the most limited exposure of any Unkar Group formations in the Canyon. The contact between the Hakatai Shale and the underlying Bass Limestone is gradational in the east and sharp, though conformable, in the west. The upper contact with the Shinumo is conformable and gently undulating.

Four sections have been sampled and measured by V. S. Reed: Hance Rapids, Cheops Pyramid, Cottonwood in Bright Angel Canyon (a partial section), and the type section (Noble, 1914) in Hakatai Canyon (fig. 2). The Hakatai is divided into three distinct units in the west, with a possible additional unit in the east. The thickness almost doubles from east to west (table 1).

Table 1.—Thickness of Hakatai Shale

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<td>3</td>
<td>37 m (122 ft)</td>
<td>62.5 m (205 ft)</td>
<td>25.5 m (84 ft)</td>
</tr>
<tr>
<td>2</td>
<td>154.5 m (507 ft)</td>
<td>51 m (167 ft)</td>
<td>65 m (213 ft)</td>
</tr>
<tr>
<td>1</td>
<td>0</td>
<td>0</td>
<td>33.5 m (110 ft)</td>
</tr>
<tr>
<td>Total</td>
<td>289.0 m (949 ft)</td>
<td>171.5 m (563 ft)</td>
<td>170.0 m (558 ft)</td>
</tr>
</tbody>
</table>
Figure 2.—Sketch map of eastern and central Grand Canyon.
Stratigraphy

Unit 1

In the easternmost section at Hance Rapids, the contact with the Bass Limestone is difficult to define. It has been tentatively placed at the top of the highest prominent dolomite bed. Two of us (Dalton and Reed) have picked the boundary at different points some 33.5 m (110 ft) apart stratigraphically. Reed includes as unit 1 of the Hakatai a 33.5 m (110 ft) series of alternating sandstones and conglomerates that Dalton (1972) considers part of the Bass Limestone. This boundary problem will be resolved when both authors reexamine the evidence in the field.

The conglomerates in this unit consist of large clasts of metamorphic and volcanic rock fragments, up to 10 cm in size, and some chert, in a matrix of coarse quartz dolomite sand. The interbedded sands are arkosic and contain well-sorted and rounded grains of quartz, fresh feldspars, and rare dolomite lithoclasts. The sandstones commonly contain ripple marks and mud cracks.

Unit 2

Above the conglomerates and sandstones in the Hance Rapids section is a thick succession of highly fractured red mudstone. In exposures west of Hance Rapids this mudstone is the basal unit of the Hakatai Shale (fig. 3). The red mudstone is brittle, and pervasive fracturing that formed conchoidal surfaces has obscured whatever magascopic sedimentary structures might have been present. Approximately midway in the mudstone are two to four rather uniform sandstone beds 30 to 90 cm (1 to 3 ft) thick, with thin interbedded red shale. These sandstone beds are laterally extensive, and appear in the Hance section and the Cheeps and Hakatai Canyon sections to the west. The sandstone, notably free of clay, consists primarily of well-rounded and sorted quartz, with some fresh, rounded potassium feldspars and plagioclase. A few sandstone channels are present in the upper part of this unit in the western exposures.

Unit 3

Unit 3, consisting of shale and mudstone, is the most brilliantly colored of all the Hakatai units: deep red on fresh exposures, brilliant orange where weathered. Like unit 2 below, it is extremely brittle and has been intensively fractured into marble-size fragments; thus, most sedimentary structures have been obliterated.
Figure 3.—Composite columnar section of the Unkar Group, eastern Grand Canyon.
Unit 4

Unit 4, the uppermost sequence in the Hakatai Shale, is composed of a thick section of poorly consolidated, purple sandstones consisting predominantly of well-rounded and generally well-sorted medium-grained quartz, with minor amounts of both fresh and altered potassium feldspars, chert fragments, and polycrystalline quartz. Crossbedding is present throughout, with tabular-planar sets ranging from several centimeters (a few inches) to 60 or 90 cm (2 or 3 ft) in thickness. Ripple marks are common, mud cracks rare.

Preliminary petrographic examination of the sandstones indicates that multiple source rocks were available during deposition of the upper Hakatai. Sedimentary and metamorphic rock fragments are common, igneous rock fragments less common. Most of the quartz grains probably were derived from an older sedimentary source.

Environment of deposition

The Hakatai probably was deposited in a marginal marine environment with periods of emergence, as suggested by the red color and the presence of mud cracks and ripple marks. The presence of unaltered, rounded feldspar grains suggests an arid climate. The shale deposits probably formed in a broad, low-energy mud-flat environment, possibly deltaic. The sandstones in the upper part of the Hakatai Shale do not appear to be channel deposits but, rather, shallow-water marine sediments formed in an environment having sufficient energy to remove silt and clay size fractions. These sands may represent a slight deepening of the water after the shale and mudstone were deposited.

SHINUMO QUARTZITE

The Shinumo Quartzite, named by Noble (1914, p. 51) from exposures in Shinumo Creek in central Grand Canyon, consists of a thick succession of mainly red or purple quartz sandstones and subarkoses unconformably overlying the Hakatai Shale and conformably overlain by the Dox Sandstone. The sand grains are commonly coated with hematite and set in a siliceous cement. The Shinumo crops out intermittently through a 64-km (40-mi) distance within the Inner Gorge of the Grand Canyon between Papago Creek on the east and Tapeats Creek on the west. The total thickness increases slightly from 345 m (1,132 ft) at Papago Creek to 410 m (1,346 ft) at the westernmost exposure in Tapeats Creek (table 2).
Table 2.—Thickness of Shinumo Quartzite

<table>
<thead>
<tr>
<th>Units</th>
<th>Tapeats Creek -12 miles-</th>
<th>Shinumo Creek -14 miles-</th>
<th>Cheops Pyramid (part)</th>
<th>Papago Creek</th>
</tr>
</thead>
<tbody>
<tr>
<td>5</td>
<td>50.5 m (166 ft)</td>
<td>149 m (489 ft)</td>
<td>---</td>
<td>115 m (378 ft)</td>
</tr>
<tr>
<td>4</td>
<td>68 m (224 ft)</td>
<td>30 m (98 ft)</td>
<td>56 m (183 ft)</td>
<td>72 m (237 ft)</td>
</tr>
<tr>
<td>3</td>
<td>104 m (341 ft)</td>
<td>97 m (318 ft)</td>
<td>33.5 m (110 ft)</td>
<td>121 m (397 ft)</td>
</tr>
<tr>
<td>2</td>
<td>0</td>
<td>27 m (90 ft)</td>
<td></td>
<td>36 m (119 ft)</td>
</tr>
<tr>
<td>1</td>
<td>187.5 m (615 ft)</td>
<td>101 m (332 ft)</td>
<td>40.5 m (133 ft)</td>
<td>0.15 m (0.5 ft)</td>
</tr>
<tr>
<td>Total</td>
<td>410.0 m (1,346 ft)</td>
<td>404.5 m (1,327 ft)</td>
<td>---</td>
<td>345.00 m (1,132 ft)</td>
</tr>
</tbody>
</table>

Total: 1,346 ft (410.0 m) at Tapeats Creek; 1,327 ft (404.5 m) at Shinumo Creek; 1,132 ft (345.00 m) at Papago Creek.
Five mappable units within the Shinumo are recognized in four sections measured by T. M. Daneker. The lowest is an alternating succession of tan to purple coarse-grained to conglomeratic subarkoses and quartz sandstones. Sand grains are poor to moderately sorted and most commonly subrounded to rounded. Abundant small-scale tabular-planar crossbedding occurs throughout. This basal unit is more than 183 m (600 ft) thick at Tapeats Creek but thins consistently eastward to 15 cm (6 in) at Papago Creek.

Unit 2, the most resistant part of the Shinumo, consistently forms a vertical cliff in eastern Grand Canyon. It is a purple quartz sandstone and contains well-rounded well-sorted, fine to medium sand grains commonly having quartz overgrowths and set in a quartz cement. Individual beds are 5 to 10 cm (2 to 4 in) thick, where visible, and small-scale tabular-planar crossbeds occur locally. The unit has a maximum thickness of 36 m (119 ft) at Papago Creek but pinches out west of Shinumo Creek.

The middle unit is similar in lithology and color to unit 2 but is less well cemented and forms irregular, stepped ledges. The thickness ranges between 61 and 122 m (200 and 400 ft) but is more uniform than any of the other units. Ripple marks and tabular-planar and trough crossbeds are common throughout.

Unit 4, a deep-red to rusty brown quartz sandstone, is moderately well cemented and forms irregular ledges. Crossbeds, ripple marks, channels, and clay galls are common throughout; mud cracks are conspicuously abundant. Thickness ranges from 30 to 72 m (98 to 237 ft).

The uppermost unit is thickest (149 m) (489 ft) and best represented at Shinumo Creek. Consisting of fine-grained well-sorted and rounded quartz sand set in a siliceous cement, this unit generally forms a resistant series of ledges but locally forms massive single cliffs. The lower fourth of the unit is purple except for a distinctive 65 cm (2 ft) band of white in the middle. The upper part is light purple to white and has a "marble cake" appearance owing to the general presence of deformed and contorted bedding. Sedimentary structures include ripple marks, crossbeds, clay galls, abundant load casts and slurry slumped bedding, and rare mud cracks.

Preliminary analysis of the lithology and structures of the Shinumo suggests a nearshore, very shallow, marginal marine environment, probably in part fluvial or deltaic. The deformed beds and load casts in unit 5 indicate frequent slumping and deformation of unconsolidated water-saturated sediments.

DOX SANDSTONE

The Dox Sandstone, named by Noble (1914, p. 53) from Dox Castle (beneath which is located the type section in a tributary to Shinumo Creek), is predominantly interbedded quartz sandstone, siltstone, and mudstone and subordinant amounts of calcareous lithic and arkosic sandstone with thin shale interbeds. Only the lower third of the Dox is present at the type section in central Grand Canyon, but it is
extensively exposed on both sides of the Colorado River in eastern Grand Canyon between Seventy-five Mile and Escalante Creeks, where it has a maximum thickness of 952 m (3,122 ft). Stevenson (1973) has measured and described 13 sections of the Dox and recognized four members, which will be given formal status in a later publication but herein are referred to informally as the lower, lower middle, upper middle, and upper members (table 3).

**Lower member**

The lower member is 390 m (1,278 ft) thick in the Escalante-Unkar Creek area and consists of 244+ m (800+ ft) of light-tan to greenish-brown siliceous quartz sandstone and calcareous lithic and arkosic sandstone overlain by 122 m (400 ft) of dark-brown to green shale and mudstone. The tan to brownish color of this lower member is in marked contrast to the red or red-brown color characteristic of the rest of the Dox. Major outcrops are between Escalante and Seventy-five Mile Creeks on the southeast side of the Colorado and on the opposite right bank of the river. Contorted bedding, small-scale tabular-planar crossbeds, and graded beds, commonly having shale interclasts at the base, occur in the sandstone beds, and suggest deposition in a shallow subaqueous deltaic environment. Paleocurrent directions derived from crossbedded strata indicate a westerly source for the sediment.

**Lower middle member**

The lower middle member of the Dox is a cyclical succession of red mudstone, siltstone, and quartzose sandstone that is gradational with the members above and below. It is 280 m (920 ft) thick along Unkar Creek, where it is most completely exposed. The lower 213 m (700 ft) consists of a slope-forming repetition of red to brownish-red fine-grained quartz sandstone and red to maroon shaley siltstone and mudstone, with the latter two lithologies predominating. Load casts, scour marks, and mud cracks are common, and ripple marks rare. The upper part of the member is a thick and more resistant unit of red to maroon quartz sandstone that exhibits numerous channel features. The sandstones are arkosic, containing grains of quartz, orthoclase, plagioclase, and microcline set in a calcite or iron oxide cement. Low-angle tabular crossbed sets are common and, together with the channel features, indicate a flood-plain environment, with currents mainly from the west.

**Upper middle member**

The upper middle member, 155 m (508 ft) thick and occupying more than half the Dox outcrop area in eastern Grand Canyon, consists of red siltstone, mudstone and thin interbedded quartz sandstone and forms much of the extensive soft, smooth slopes that seem most typical of the Dox outcrops as viewed from the Canyon rim. Five distinct and continuous whitish marker bands 3 to 12 m (10 to 40 ft) thick in an
### Table 3.—Thickness of the Dox Sandstone members

<table>
<thead>
<tr>
<th></th>
<th>Shinumo Creek 26 miles-</th>
<th>Escalante-Unkar Creek 1 Composite-mile Creek 2 Cardenas 2 miles-</th>
<th>Tanner 2 Canyon -miles-</th>
<th>Espejo Creek</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Upper member</strong></td>
<td>---</td>
<td>---</td>
<td>92 m (303 ft)</td>
<td>92 m (303 ft)</td>
</tr>
<tr>
<td><strong>Upper middle member</strong></td>
<td>---</td>
<td>---</td>
<td>188 m (618 ft)</td>
<td>155 m (510 ft)</td>
</tr>
<tr>
<td><strong>Lower middle member</strong></td>
<td>Present but incomplete</td>
<td>280 m (920 ft)</td>
<td>58+ m (191+ ft)</td>
<td>51+ m (166+ ft)</td>
</tr>
<tr>
<td><strong>Lower member</strong></td>
<td>293 m (960 ft)</td>
<td>390 m (2178 ft)</td>
<td>---</td>
<td>---</td>
</tr>
</tbody>
</table>
otherwise uniformly red succession distinguish this member. These white marker zones are leached intervals within red beds. A stromatolite bed 15 to 45 cm (6 to 18 in) thick occurs just above the lowermost white marker band. The stromatolites are composed of dolomite with minor amounts of calcite and quartz silt and are in the form of laterally linked hemispheroids (Logan and others, 1964). Sedimentary structures concentrated in, though not restricted to, these light marker zones include salt crystal casts and wavy irregular bedding, with abundant ripple marks and mud cracks. The above features suggest a tidal flat or perhaps supratidal environment.

Upper member

The upper member of the Dox is 76 to 91 m (250 to 300 ft) thick, and exposures are limited to narrow bands well above river level and beneath basalt cliffs of the Cardenas Lavas, particularly along Cardenas Creek and Tanner Canyon. The member consists of slope-forming micaceous mudstone, which grades upward into more resistant red quartzose silty sandstone. Sedimentary structures include mud cracks and salt crystal casts in the mudstones and small-scale crossbeds and asymmetrical ripple marks in the silty sandstones. The contact with the overlying volcanics of the Cardenas Lavas is locally baked but appears conformable. The structures and lithology of the upper member suggest continuation of tidal flat conditions until burial by lava flows at the end of the Dox deposition.

PALEONTOLOGY

Fossils presently known from the Unkar Group are all algal stromatolites or other structures believed to be of algal origin. Dalton (1972) recognized three morphological types of stromatolites in the Bass Limestone: 1) biscuit-shaped mounds referable to Collenia undosa Walcott, and C. symmetrica Fenton and Fenton; 2) mat-like forms, resembling Collenia frequens Walcott; and 3) biohermic forms. The biscuit-shaped stromatolites are composed of dolomite and siliceous material and are commonly associated with mud cracks. The most common type of stromatolite in the Bass is the mat-like or encrusting form. They are generally restricted to the dolomitic rocks but locally occur in sandstones. The biohermic variety was found only near the top of the Shinumo Creek section. Domes are about 51 cm (20 in) in diameter and have 15 to 20 cm (6 to 8 in) of relief.

Insoluble residues of the Bass dolomites have yielded many flexible organic filaments up to 2 mm in length. Nearly all of these suspected life forms are threadlike and blue, green, or black in color. Calcispheres about 0.4 mm in diameter were observed in several thin sections. They are commonly associated with the stromatolites and may represent a stage in the reproductive cycle of algae as described by Stanton (1963).
Dalton (1972) found ovoid pellets of silty composition ranging in size from 0.1 to 3 mm. Their stromatolitic association, equant grain size, and similar composition suggest a boring organism that may have lived in symbiotic relationship with the algae communities.

Algal stromatolites in the form of spaced laterally linked hemispheroids (LLH-S) and stacked hemispheroids (SH-V) of Logan and others (1964) occur at one horizon in the upper middle member of the Dox Sandstone and are the only fossils known from the Dox.

Several medusoid structures reported from the Unkar Group and the Nankoweap Formation in eastern Grand Canyon have recently been reinterpreted as sedimentary structures. Brooksella canyonensis Bassler, described by Bassler (1941) and Van Gundy (1937) from the Bass Limestone, is considered by Cloud (1968, p. 27) to be compaction of sand around a gas blister, although interpreted by Glaessner (1969, p. 374-375) as a possible worm burrow.


Bivalve-like structures were reported from the lower Bass along the Bass Trail by Smith (1968). Specimens obtained from approximately the same locality by Trever Ford appear to the senior author to be oncolites or other algal structures.

AGE AND CORRELATION OF THE UNKAR GROUP

The Unkar Group rests upon crystalline basement rocks having a minimum age of 1.65 b.y. (Wasserburg and Lanphere, 1965, p. 735). Dikes and sills that cut the Unkar sedimentary rocks and the Cardenas Lavas which overlie them have been described by Hendricks and Lucchitta (this volume). The Cardenas Lavas at the top of the Unkar are Rb/Sr dated at 1,090 ±70 m.y. (McKee and Noble, this volume; Elston and Scott, in press). Thus the Unkar rocks are younger than 1.65 b.y. and older than about 1,100 m.y. The igneous rocks which formed the dikes within the Unkar and the basaltic flows of the Cardenas Lavas are probably broadly equivalent in time to similar diabase sills and basaltic lavas in the Apache Group of central Arizona. Radiometric dates from the Apache Group suggest a probable age of 1.2 b.y. (Silver, 1960).

Shride (1967, p. 80-83) suggested correlation of the lower Unkar strata with the Apache Group. The Mescal Limestone of the upper Apache was recognized as the lithologic equivalent and probable correlative of the Bass Limestone, and the Troy Quartzite, which overlies the Apache, as the correlative of the Shinumo Quartzite. If this correlation is correct, the Dox Sandstone is post-Troy in age and has no recognized equivalents elsewhere in Arizona.

Paleomagnetic data recently obtained by Elston and Scott (in press) suggest that, based upon pole positions, the Dox Sandstone is more nearly the correlative of the upper Apache Group and Troy Quartzite.
The Unkar Group may correlate with the Pahrump Group of the Death Valley region, California, where an age of 1.2 to 1.4 b.y. is reported by Cloud (1971). The Unkar is also probably equivalent in age to some part of the Belt Supergroup in the northern Cordilleran region, but detailed correlations are not possible at present.

REFERENCES CITED


THE LATE PRECAMBRIAN CHUAR GROUP OF THE EASTERN GRAND CANYON

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and

W. J. Breed
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ABSTRACT

The Chuar Group is restricted in exposure to the eastern part of Grand Canyon and tributary canyons within an area some 24 km long and 6 km wide. The group is 2,013 m thick and is divided into three formations and seven members. The lower two formations, Galeros (below) and Kwagunt (above), are predominantly argillaceous with subordinate thin limestone beds, while the highest, Sixtymile Formation, is mostly coarse breccia. Stromatolites are present at three horizons, one of them biohermal. The form-genera Inzeria, Baicalia, and Boxonia indicate an upper Riphean age. The megaplanktonic fossil Chuaria occurs near the top of the Kwagunt Formation. A pisolithic chert, also near the top of the Kwagunt, contains a microflora similar to the Bitter Springs chert of central Australia. The Chuar rocks may be contemporary with rocks below the Cambrian in eastern California, and with the Windermere Formation of the northern Cordillera and are probably younger than any other Precambrian rocks in Arizona.

INTRODUCTION

The Chuar Group was first described by Walcott (1894, 1895) and has recently been described in detail by Ford and Breed (1973a); only a summary account is presented here.

The group consists of dominantly argillaceous sediments with subordinate carbonates and a few sandstones. Totalling over 2,000 m in thickness, the group is divided into three formations and seven members, as listed in table 1. These crop out only in the eastern part of Grand Canyon, and there only in the upper parts of a number of right-bank tributary canyons to the Colorado. Exposures of the Chuar Group are in the upper parts of Nankoweap, Kwagunt, Awatubi, Sixtymile, Carbon, Chuar, and Basalt Canyons (fig. 1). These canyons transect a north-south syncline that has a very gentle northerly plunge, so that no one of them exposes a full succession. The upper members are exposed in Nankoweap and Kwagunt Canyons and on Nankoweap Butte; the lower members are best exposed in Basalt, Chuar, and Carbon Canyons. In all of these canyons, the Chuar sections are cut off on the east by the north-south Butte fault, which downthrows the Paleozoic sequence to river level about a mile to the west of the confluence of the Colorado and Little Colorado Rivers. Traces of three forms of Precambrian life, described below, have been found in the Chuar Group.

The Chuar Group rests disconformably on a mildly undulating surface of the Nankoweap Group originally described by Van Gundy (1951). The synclinally folded Chuar Group is covered unconformably by the flat-lying Tapeats Sandstone of Cambrian age.

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Figure 1.—Geologic map of the Chuar terrane.
Table 1.—Divisions of the Chuar Group.

<table>
<thead>
<tr>
<th>Formation</th>
<th>Member</th>
<th>Thickness</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sixtymile</td>
<td>---</td>
<td>36 m</td>
</tr>
<tr>
<td>Kwagunt</td>
<td>Walcott</td>
<td>255 m</td>
</tr>
<tr>
<td></td>
<td>Awatubi</td>
<td>344 m</td>
</tr>
<tr>
<td></td>
<td>Carbon Butte</td>
<td>76 m</td>
</tr>
<tr>
<td></td>
<td>Duppa</td>
<td>174 m</td>
</tr>
<tr>
<td>Galeros</td>
<td>Carbon Canyon</td>
<td>471 m</td>
</tr>
<tr>
<td></td>
<td>Jupiter</td>
<td>462 m</td>
</tr>
<tr>
<td></td>
<td>Tanner</td>
<td>195 m</td>
</tr>
</tbody>
</table>

GALEROS FORMATION

The Tanner Member, the lowest member of the Galeros Formation, consists of 12 to 24 m of massive, coarsely crystalline dolomite at the base, and 177 m of mostly shales above. The dolomite forms a massive ledge that caps the cliffs around Basalt Canyon, and an isolated outcrop on the south side of Chuar (=Lava) Canyon west of the Butte fault which was mapped by Maxson (1967) as basalt. A small faulted exposure also occurs in Nankoweap Canyon. The overlying shales crop out in much of Basalt and Chuar Canyons, and poorly preserved specimens of *Chuaria* occur near the top. The Tanner Dolomite was included in the Unkar Group by Walcott but was transferred to the Chuar by Van Gundy (1951).

The Jupiter Member of the Galeros Formation is about 460 m thick and consists of carbonates below and shales above. The basal division is 12 m of stromatolitic limestones, including undulating and broad domed forms of algal structure in a matrix of dolomitized tufa-like rock, with flat-pebble conglomerates at the base. In the upper part of the carbonate member are layers with abundant casts of gypsum crystals, and a few poorly defined solitary stromatolite columns, approaching the form *Inzaeria*, as well as undulating stromatolites of the form *Stratifera*. The remainder of the member is dominantly argillaceous, with numerous thin sandstones and siltstone beds. Rarely more than several centimeters thick, these show abundant ripple marks and mud cracks, with occasional rain prints and salt pseudomorphs. They commonly have hematite cement in patches, and locally weather to goethite box-stones. The shales are highly variable in color, from red-purple through ocherus yellow to pale green, with scattered blue-black micaceous bands.

The Carbon Canyon Member consists of interbedded limestone, shale, and sandstone beds, each several meters thick, with a total thickness of about 471 m. The limestones, commonly 1 to 2 m thick, consist of fine-grained dolomitic micrite, with scattered chert nodules, and common lenses of silt-size quartz grains. In places limestone grades
into calcareous siltstone. Most of the limestone beds show an irregular lamination, possibly in part being of algal origin. The tops of the beds are mostly ripple-marked and mud-cracked. The intervening shales vary from blue-black micaceous shale to red mudstones with scattered green bands. Sandstone beds are fewer in number and are rarely more than 0.6 m thick. They are generally green-grey, with subangular quartz grains set in a yellow-green carbonate matrix. Mud cracks are common in the sandstones, and some laminae show truncated incipient cracks that resemble worm tracks.

A single bed of limestone, some 70 m from the top of the member, has well-preserved stromatolite structures on weathered joint-faces. Up to 60 m high the stromatolites take the form of rapidly widening, irregularly branching columns, with strongly convex laminae. Dolomitization has destroyed internal detail, but they appear to fall within the form Baicalia, probably B. aff. rara Semikhatov.

The Duppa Member is mainly argillaceous; only a few thin scattered limestone beds occur through a thickness of 174 m. Siltstone beds up to 1 m thick are scattered throughout. Toward the top of the member are thin hematite-cemented sandstones with well-rounded grains. Above the sandstones are shales, generally micaceous, that grade into red mudstones.

KWAGUNT FORMATION

The Carbon Butte Member, 76 m thick, has at its base the only thick sandstone in the Chuar Group. This basal unit (24 m) is a red sandstone, which forms a scarp on the slopes of Carbon Butte. The unit provides a distinctive marker for the base of the Kwagunt Formation. A micaceous purple shale parting in the middle contains mud cracks, and current bedding occurs at the top. Red and purple mudstones make up most of the member; a bed of mottled weathered sandstone occurs in the middle of the section.

The Awatubi Member is 344 m thick and is dominantly argillaceous, with shales and mudstones of varying color interbedded with thin ferruginous siltstones. A massive stromatolite layer, 3 to 5 m thick, occurs at the base of the member. This basal unit consists of biothermal domes, each 2 to 3 m in height and width, and made up of a complex of columns 10 to 15 cm in diameter that alternate with confluent domes. The columns have nearly flat laminae and are generally parallel-sided with few branches. Most of the internal detail has been destroyed, but a clearly defined wall is present, indicative of the form Boxonia Koroljuk. The matrix between columns is generally crystalline dolomite, whereas that between bioherms is coarsely granular and porous dolomite. Flat-pebble conglomerate forms the nucleus of some of the bioherms.

About 9 m below the top of the member, black, finely fissile shales yield abundant Chuaria circularis from both eastern and western slopes of Nankoweap Butte. The former is thought to be Walcott's type locality. Other beds with Chuaria occur scattered through the upper part of the member.
The Walcott Member, the topmost subdivision of the Kwagunt Formation, is much more diverse in character than the other two. It is 255 m thick and with the Sixtymile Formation forms the cap of Nankoweap Butte. At the base is the remarkable "Flaky Dolomite" bed, which is about 2.5 m thick. Throughout the outcrop it consists of randomly oriented silicified flakes set in a fine-grained dolomite matrix. The flakes appear to be a disrupted algal-laminae, and occasional masses with domal shape are present. Scattered layers of dolomitized oolite also occur.

Above the Flaky Dolomite are black shales, with a few Chuaria interbedded with several silicified cherty pisolithic beds. The lowest is about 60 cm thick; the higher ones are only about 15 cm thick. The pisoliths are completely replaced by chert, with a matrix of iron-rich carbonate. The outer surface of the pisoliths contains a microflora described below. Occasional lenses of non-silicified oolite are entirely dolomitized. Toward the top of the member are two thick dolomitized limestones, each about 10 m thick, which contain oolites and sparse algal laminae.

SIXTYMILE FORMATION

The Sixtymile Formation (Breed and Ford, 1973) is in sharp lithologic contrast to the underlying rocks. It is primarily breccia and coarse pebbly sandstone, with subordinate cherty siltstone containing poorly developed slump rolls in the middle. The clasts in the breccias consist exclusively of Chuar detritus. Only 36 m thick, the Sixtymile Formation caps Nankoweap Butte. Both Van Gundy (1951) and Maxson (1967) mapped it as an outlier of Cambrian Tapeats Sandstone. The breccia is in sharp contact on the underlying shales with local erosion channels up to 3 m deep but otherwise concordant with the Chuar. Its stratigraphic relation to underlying beds is clear, however, in the recently discovered sections in Awatubi and Sixtymile Canyons. In both canyons the Sixtymile Formation is folded into the same syncline as is the underlying Chuar and is overlain unconformably by the Tapeats Sandstone of Cambrian age.

PALEONTOLOGY

The stromatolites of the Chuar Group noted above have been described in detail by Ford and Breed (1973a). While the digitate and columnar forms which are useful as stratigraphic index fossils in Russia and Australia are rare, the forms Inzeria, Baicalia aff. rara, and Boxonia are sufficient to indicate an upper Riphean age, i.e., late Precambrian.

The problematical fossil Chuaria circularis, named by Walcott (1899) from its occurrence in the Chuar terrane, has recently been studied in detail by Ford and Breed (1973b). The abundant black carbonaceous disks, found in black shales towards the top of the
Awatubi Member, are the remains of organic spheroidal bodies which generally are about 0.2 cm in diameter. They are not brachiopods as originally thought by Walcott but appear to be giant equivalents of primitive algae of the type known as Acritarchs. They have been placed in the family Leiosphaeridae, most of which are of nannoplankton proportions. Forms indistinguishable from Chuaria circularis have been found in Canada (in the Hector Formation of the Windermere Group), in the Vindhyan System of India (previously known as Fermoria), and in late Precambrian rocks of Iran and Australia. While accurate dating is still required in these areas, it is possible that Chuaria may be regarded as a late Precambrian index fossil. It probably represents a stage in the evolution of planktonic algae when they reached "giant" size for a restricted period of late Precambrian time.

The pisolithic chert of the Walcott Member has yielded a fossil microflora of at least eight types (Schopf and others, 1973). These fall into two broad categories, spheroids and filaments. The spheroids are suggestive of coccoid blue-green algae, which are well known from other late Precambrian cherts, including those in Australia. Of the filamentous thallophytes, one closely resembles Eomycetopsis, first described from the Bitter Springs chert of Central Australia (Schopf, 1968). This and the other filaments are present as closely tangled mats that form a rind on the pisoliths. Schopf and others (1973) have noted that this flora is the first to be described from Precambrian cherts in other than a stromatolite. Forms ecologically distinct from other assemblages may be expected to be discovered when the microflora is fully studied, even though the age of the Grand Canyon and Bitter Springs flora is much the same.

**AGE AND CORRELATION OF THE CHUAR GROUP**

Shride (1967) has suggested a correlation of the Unkar Group with the Apache Group of central and southern Arizona. The correlation of the Chuar with the Troy Quartzite of the Apache Group suggested by Wilson (1962) seems unlikely in view of the dominantly argillaceous character of the Chuar and the arenaceous character of the Troy. Furthermore the stromatolites in the Mescal Limestone of the Apache Group are unlike those of the Chuar. Thus it seems that the Chuar Group is post-Apache but older than the Cambrian.

Evidently the Chuar Group is also younger than the Cardenas Lavas, for which a minimum K/Ar age of 845 ±15 has been obtained (Ford and others, 1972). A date of 1090 ±70 m.y. has also been published for these lavas (Elston and others, 1973). The Chuar Group is also probably younger than the rocks that intrude the Unkar Group, although this has not been demonstrated.

No sedimentary group of late Precambrian age is known in Arizona that can be correlated with the Chuar. In the Death Valley region of California, however, an infra-Cambrian group hundreds of meters thick lies conformably below the Cambrian. At the base of this series is the Pahrump Group, reported by Cloud (1971) to be 1.2 to 1.4 b.y. old.
Because the Chuar is clearly younger than the Cardenas Lavas and diabases, it must be equivalent to some part of the infra-Cambrian of southeastern California. No single horizon can be correlated, but both the Chuar and the infra-Cambrian groups constitute a predominantly shallow-water sequence of sediments with stromatolitic limestones.

Farther north in the Cordillera, the Belt Supergroup appears to be, at least in part, similar in age to the Apache Group. The overlying Windermere Group was apparently deposited during the period from 850 million years ago to the start of the Cambrian at 570 million years (Harrison and Peterman, 1971). The Windermere Group thus falls into the same period of geologic time as the Chuar, although the chronological limits of both are uncertain in the absence of radiometric dates. Additional radiometric dating is clearly needed both in the Arizona Precambrian and elsewhere. The presence of Chuaria in both the Chuar Group and the Hector Formation of the Windermere Group provides paleontological support for correlating these groups.

The Chuar Group thus is older than the Cambrian and younger than the Cardenas Lavas. It is unique in Arizona but possibly is correlative with some part of the infra-Cambrian of southern California and with the Windermere Group of the northern Cordillera.

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UPPER PRECAMBRIAN IGNEOUS ROCKS OF THE GRAND CANYON, ARIZONA

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ABSTRACT

Basaltic igneous rocks occur as dikes, sills, and flows in the upper Precambrian Unkar Group of the Grand Canyon. Extrusive rocks, called the Cardenas Lavas, rest conformably on the Dox Sandstone; the dikes and sills intrude all sedimentary formations below these flows.

The extrusive sequence measured at Basalt Canyon is about 300 m thick and contains at least 14 lava flows and 10 sandstone interbeds. The lowest 93 m of section contains several sandstone lenses and abundant nodular or pillow-like structures. Before alteration, this unit was a medium-grained basalt similar in texture and mineralogy to the sills and dikes. Petrologic and chemical data suggest that this lower part of the section is a spilitic hyaloclastite that may be the altered effusive equivalent of the intrusive rocks. Igneous rocks above the 93-m level are fine-grained, with an intergranular to intersertal texture. These rocks change sequentially from basaltic andesite to basalt to basaltic andesite upward through the section. Characteristics of the lava and sandstone bed sequence suggest deposition in a low-energy shallow sea that became shallower with time and intermittently disappeared. A shallow-water environment probably was maintained by basin subsidence or rising water level during accumulation of the lava flows and sandstones.

Sills are confined to the lower shaly parts of the Unkar Group; related dikes are intruded above the sills along faults that predate or are contemporaneous with the sills. Sills crop out in seven locations in the Grand Canyon and range in thickness from 20 to more than 200 meters. They are composed chiefly of medium-grained diabase consisting of plagioclase of varied composition, subhedral to anhedral zoned olivine, and magnetite. Augite and minor biotite fill interstices. The dikes have a similar composition but are finer grained, as are the chilled margins of the sills. Early differentiation and crystal settling in the sills are shown by granophyre layers up to 6 m thick and felsite dikes, and by olivine-rich layers.

Radiometric dating of intrusive and extrusive rocks indicates a similar age of crystallization. The affinity in chemistry, combined with the relation in time and space of the flows, dikes, and sills, suggest that all were formed from a common parent magma. The sills and lower 93 m of the flow sequence may be contemporaneous; the remainder of the flows may represent a later phase of igneous activity that followed differentiation of the parent magma in a chamber below the level of the sills.

INTRODUCTION

Upper Precambrian rocks in the Grand Canyon region are divided in ascending order into the Unkar Group, the Nankoweap Group of Van Gundy (1951), and the Chuar Group. These together form the Grand Canyon Supergroup. As exposed in the eastern Grand Canyon, these rocks are approximately 3,650 m thick. The Grand Canyon Supergroup occupies a
stratigraphic position between the overlying Tapeats Sandstone of Cambrian age, and the underlying Precambrian crystalline basement complex. West of Hance Rapids (fig. 1) only the lower part of the Unkar Group is exposed in isolated and tilted fault blocks that generally dip from 10° to 15° northeast.

Igneous rocks intrude all sedimentary formations of the Unkar Group (fig. 2). Extrusive basaltic flows with interbedded sandstones of the Cardenas Lavas form the highest stratigraphic unit of the group (fig. 2). The name Cardenas Lavas is here informally used and is derived from the "Cardenas lava series" of Keyes (1938) and the "Cardenas Lavas" of Ford and others (1972). A geologic map of the eastern Grand Canyon by Maxson (1967) shows the approximate distribution of the lavas. This map, however, includes the Cardenas flows within the Rama Formation, which was the name informally given by Maxson (1961) to upper Precambrian diabase sills and dikes that occur further to the west.

Sills more than 20 m thick crop out in seven locations in the Grand Canyon (fig. 1). These sills are confined to the Bass Limestone and Hakatai Shale (fig. 2) — the lowermost formations of the Unkar Group. The sills range in thickness from 20 to more than 200 m and are composed predominantly of olivine-rich diabase that is uniform in texture and mineralogy throughout the area. Noble (1910, 1914) described diabase sills in the Bass Canyon-Shinumo Creek area that intrude the Unkar rocks at three stratigraphic levels. Maxson (1967, 1968) mapped the intrusive rocks of the Grand Canyon but did not describe the sills and dikes.

This report summarizes new data on the upper Precambrian igneous rocks of the Grand Canyon that (1) suggest the environment of lava deposition; (2) clarify the genetic relations of associated sills, dikes, and flows; and (3) provide critical information that may bear on interregional correlations of upper Precambrian rocks in the western United States.

Preliminary results of the current study, presented by Hendricks (1972a) and by Lucchitta and Hendricks (1972), are included in this report. Detailed reports on the intrusive rocks, the extrusive rocks, the ages of these rocks, and the spilitic alteration of the lower part of the Cardenas Lavas are being prepared, respectively, by Hendricks, Lucchitta and Hendricks, McKee, and Lucchitta and Hendricks.

ACKNOWLEDGMENTS

This paper in part summarizes a thesis presented by Hendricks (1972b) to the faculty of Northern Arizona University and titled "Younger Precambrian Basaltic Rocks of the Grand Canyon, Arizona," and in part observations and conclusions derived by Lucchitta and Hendricks from fieldwork begun in association with paleomagnetic investigations carried out by D. P. Elston.
Figure 1.—Location map showing outcrops of Unkar igneous rocks.
Figure 2.—Schematic cross-section showing stratigraphic and structural relations of Precambrian rocks in the eastern Grand Canyon.
Thanks are due for help of many kinds given us by Charles W. Barnes, Russell O. Dalton, Jr., Raymond L. Eastwood, Donald P. Elston, C. Sherman Grommé, Richard F. Holm, B. K. Lucchita, and Edwin H. McKee. We also thank the owners and staff of Canyoneers, Inc., for logistical support freely given; Allen Wilson for his expert services as river guide; and the National Park Service for permission to carry out research within the Grand Canyon National Park.

EFFUSIVE ROCKS - THE CARDENAS LAVAS

Thickness and lateral extent

The Cardenas Lavas crop out only in the eastern part of the Grand Canyon, where nearly continuous exposures occur on both sides of the Colorado River from the mouth of Unkar Creek in the southwest to just above the mouth of Chuar Creek to the northeast (fig. 1). Other outcrops occur along the northeast wall of Unkar Valley and at one isolated locality in Nankoweap Creek 4 km from its junction with the Colorado River. The most extensive exposures are northeast of the Colorado River from just north of the mouth of Unkar Creek to Chuar Lava Hill (fig. 3). This includes the Basalt Canyon section, which is the principal area for this investigation, and the Chuar Lava Hill section, which was the principal outcrop studies by Walcott (1894).

The regional dip of younger Precambrian rocks in the eastern Grand Canyon generally is to the northeast, so that the lavas and associated Algonkian rocks are truncated to the southwest by the ep-Algonkian unconformity. North-northwest of Chuar Creek the dip of the lavas carries them below the topographic surface. They only reappear at the surface in a remnant at Nankoweap Creek, about 21 km north of Basalt Creek.

Complete sections of the Cardenas Lavas range in thickness from 244 m at Unkar Creek (Walcott, 1894, p. 511) to 458 m at Chuar Lava Hill (Walcott, 1894, p. 515). The Basalt Canyon section is 292 m thick. The isolated remnant in Nankoweap Creek is approximately 60 m thick but may not represent a complete section because it crops out in an area cut by numerous faults.

Contacts

The upper and lower contacts of the Cardenas Lavas are well exposed in the vicinity of Basalt Creek. The basal contact has been traced for a distance of about 4 km southwest of Basalt Canyon. At Basalt Canyon the contact is sharp and smooth, with a relief of 10 cm at most; no lag material was observed at the contact. The uppermost 60 cm of Dox Sandstone appears to be mildly baked. Approximately 1 km south of the contact in Basalt Canyon, the Dox laminae immediately below the lavas exhibit folds and convolutions, 5 cm or less in amplitude, suggestive of soft-sediment deformation. At this location the base of the lavas locally incorporates small fragments of Dox.
Figure 3.—Basalt Cliffs as seen from the south. Width of view is approximately 10 km.
Four kilometers southwest of Basalt Canyon the nature of the Dox-Cardenas contact changes radically: a thin lava flow separated from the main mass of the lavas by about 3 m of reddish, well-bedded sandstone and siltstone similar to those of the underlying Dox, crops out for a distance of about 100 meters. A similar flow has been observed by Elston and Scott (1974) southeast of the Colorado River near Tanner Canyon, and by the authors near Cardenas Creek. All indications are that the lavas flowed over the Dox sediments before the latter were lithified and that the contact between the two formations is conformable and interfingering.

The contact between the Cardenas Lavas and overlying Nankoweap rocks is disconformable (Van Gundy, 1951). At Basalt Canyon, the contact has local relief of about 1 m, and the lavas below the contact are weathered. Along Tanner Canyon, a weathered ferruginous zone, locally about 10 m thick, is developed on top of the lavas (Elston and Scott, 1973). This zone is beneath a pre-Nankoweap erosional surface that, according to Elston and Scott, truncates the flows and initially sloped eastward.

Along the west wall of Basalt Canyon, the pre-Nankoweap erosion surface cuts down into the lavas in a southerly direction. One kilometer south of the contact as exposed along Basalt Creek, the Nankoweap rests directly on the lapillite unit shown in figure 4, indicating removal of about 60 m of section and a component of nearly 3° to the south-southeast for the initial slope of the erosion surface. As seen from a distance, the lapillite unit is cut out along the Basalt Cliffs 2 km southwest of Basalt Canyon, indicating that the initial southerly slope of the erosion surface is not a local feature. The easterly slope reported by Elston and Scott (1973) may represent a local modification of the southerly slope.

Field description - the Basalt Canyon Section

The dark-colored exposures of the Cardenas Lavas contrast markedly with the smooth reddish-brown and gray slopes formed by the Dox Formation below and the Nankoweap above. When viewed from a distance, outcrops of the Cardenas appear two-toned; the lowest 50–60 m of the formation forms a greenish slope, whereas rocks above this are dark gray to black and form cliffs. The massive appearance of the upper flow sequence is broken by boundaries between flow units and thin interbeds of sandstone. These sandstone beds typically are planar and of considerable lateral extent.

Field and petrographic descriptions of the Basalt Canyon section are given in figure 4. Only those features relating to interpretations of the environment of deposition are presented here. The lowermost 50 to 60 m of the green unit is a highly fractured basalt (?) weathering into granular rubble that covers the steep, smooth slope at the base of the lava cliffs. In places, the rubbly material grades laterally into less fractured flow rock. The upper part of the unit contains numerous
Sandy siltstone. Laminated to flaggy, Ts 2 moderately to well sorted. Laminations are lenticular. Sorting between layers much poorer than sorting within individual layers. Overall composition: quartz 75%; feldspar 25%; opaques 5%; muscovite 3%. Adjacent layers very markedly in composition. Quartz is angular and concentrated in the fine sand and silt. Feldspar, primarily plagioclase, is angular, prismatic, and in the fine sand size. Muscovite is in platy grains as much as 0.1mm in size arranged parallel to lamination. Opaque minerals, primarily magnetite, are widely distributed but especially abundant at the boundaries of laminations. Hematite coats most grains, giving the rock its dark red color. Authigenic quartz fills fractures and large pores. Volcanic-lastic glass are composed of clay, altered to talc, antigorite (?) and chlorite. Magnetite is present on boundaries of original mineral grains. Patches of devitrified glass are composed of clay, chlorite, epidote and zeolites filled amygdules. Top of unit is very amygdaloidal and rich in chlorite and epidote, disseminated and in amygduloid fillings. Spherical bodies, as much as 15cm, typically 8 to 11cm in diameter are common, especially near base of unit and near the 47m level. Unit is more resistant and less greenish than units below, forms a steeper slope.

Basalt, composed of plagioclase, olivine, Tb 1.0 pyroxene (?) and magnetite, with large prisms of altered glass. Plagioclase (An55-oligoclase) laths are up to 1cm in length. Mafic minerals are completely altered to talc, antigorite (?) and chlorite. Magnetite is present on boundaries of original mineral grains. Patches of devitrified glass are composed of clay, chlorite, epidote and zeolites (?). Original mineral percentages are: plagioclase 60%; glass 20-25%; mafic minerals 15-20%; minor magnetite.

Basalt, gray to gray-brown, f.g. Extensive chlorite-epidote alteration. Lower 46cm has abundant chlorite- and epidote-filled amygdules. Top of unit is very amygdaloidal and rich in chlorite and epidote, disseminated and in amygduloid fillings. Spherical bodies, as much as 15cm, typically 8 to 11cm in diameter are common, especially near base of unit and near the 47m level. Unit is more resistant and less greenish than units below, forms a steeper slope.

Basalt, similar to Unit 5, but less greenish, more tan and more cliff-forming. Common amygdule and fracture fillings of chlorite, epidote, locally quartz. Abundant spheroidal bodies as much as 60cm in diameter. Most probably are the product of spheroidal weathering, some may be pillows. Several flow units present, defined by chilled lower contacts as well as vesicles and amygdules near top. Wedging of flows and angular unconformities common within unit. Upper 4.5m of unit is relatively resistant and purplish. Uppermost 7 to 10cm is red and probably represents a weathering zone. Unit forms cliffs with minor slopes.

Sandstone, v.f.g., purplish, Discontinuous. Locally fills channels.

Basalt, similar to Unit 1, but less resistant and more green. Forms conspicuously greenish slope, typically covered by talus.

Basalt, dark greenish gray, fine-grained. Epidote common in groundmass, as blebs and in shear zones. Thin lenses of f.g. purplish sandstone near top of unit. More resistant than units below, forms subdued ledge. Weathers into coarse rubble than units below; some fragments are as much as 5cm in size. Composed of at least two flow units.

Basalt, gray to gray-brown, f.g. Extensive chlorite-epidote alteration. Lower 46cm has abundant chlorite- and epidote-filled amygdules. Top of unit is very amygdaloidal and rich in chlorite and epidote, disseminated and in amygduloid fillings. Spherical bodies, as much as 15cm, typically 8 to 11cm in diameter are common, especially near base of unit and near the 47m level. Unit is more resistant and less greenish than units below, forms a steeper slope.

Sandstone, brown to maroon, m. to f.g., laminated, jointed. Forms prominent sharp cliffs. Upper part moderately baked, hard. Lower contact is planar in gross aspect, but has local irregularities typically 30cm or less in relief, locally as much as 90cm. Some of the irregularities are wave-like in cross section and may represent fluting caused by currents.

Contact. About 60m exposed along strike near line of section. Contact is smooth, with relief of 1cm or less. No weathered zone, sand or pebble lag at top of Dox sandstone. Contact appears conformable. Baked zone.

Figure 4.—Measured section and petrographic descriptions of the Cardenas Lava at Basalt Canyon.
Basalt (?). Similar to Tb 4, but olivine content is less than 5%; chlorite is more abundant as an alteration product of groundmass and large plagioclase phenocrysts are altered to sericite, chlorite and epidote. Sphere (?) is present as an alteration product of augite.

Basalt, amygdaloidal. Intergranular to interstitial texture. Consists of randomly oriented plagioclase laths, altered olivine crystals and aggregates of pyroxene and chlorite-augite up to 0.5mm in size. Chlorite-augite may be an alteration product of glass. Plagioclase laths are up to 0.15mm in size. Magnetite-hematite fills radial cracks and coats grains. Pyroxene (augite) is present as aggregates of grains; individual grains are up to 0.1mm in size. Magnetite coats mafic minerals and forms large grains associated with these minerals. Chlorite and epidote occur as vesicle fillings and as granular to intersertal texture.

Aggregates of augite are composed of 5% magnetite, ilmenite and hematite. Andesite (?). Randomly oriented laths of plagioclase (An20–oligoclase-andesine) up to 0.5mm in length and a few completely altered pyroxene crystals, all in a matrix of chlorite and f. g. magnetite, ilmenite and hematite. Matrix may have been crystalline and/or glassy before alteration.

Andesite, strongly propylitized. Intergranular to interstitial texture. Randomly oriented laths of plagioclase (An20–oligoclase-andesine), phenocrysts of magnetite and hornblende and aggregates of clinopyroxene (augite), all set in a hyalocrystalline matrix of f. g. plagioclase, orthoclase, quartz and glass. Plagioclase laths are up to 1mm in length and show normal zoning. Aggregates of augite are composed of 5% of stubby crystals and colored by alteration products (green by epidote and chlorite, brown by hematite). Magnetite is associated with concentrations of mafic minerals. Hornblende crystals are euhedral-subhedral, pleochroic green-brown and poikilitic with quartz. Quartz occurs in irregular patches and as intergrowths with plagioclase filling interstices between plagioclase and pyroxene. Pockets of glass as much as 2mm in size are devitrified and now composed of brown-green pleochroic amphibole (?) and chlorite; some of the original spheroidite structures are preserved. Original mineral percentages are: plagioclase 55; augite 25; primary magnetite 5; quartz and orthoclase 5; hornblende 2. Alteration products of unknown origin and devitrified glass make up the remaining 8%.

Porphyritic andesite. Randomly oriented laths of plagioclase (An20–oligoclase-andesine) up to 0.5mm in length and minor phenocrysts of augite are set in a matrix of f. g. pyroxene and alteration products. Orthoclase and quartz occur as large irregular patches up to 0.5mm in size. Ilmenite and/or rutile needles are concentrated in the orthoclase and quartz patches. Chlorite and sericite are abundant. Overall composition: plagioclase 55%; augite 15%; magnetite-ilmenite 10%; quartz 10%; orthoclase <5%; altered aphanitic groundmass 15%.

CONTINUED ON NEXT PAGE
Andesite. Highly altered. Includes phenocrysts of plagioclase, pyroxene, hornblende, minor apatite, and large patches of orthoclase and quartz. Plagioclase laths, up to 0.3mm long, are highly altered to sericite (composition unknown). Clinopyroxene forms small (<0,1mm) grains that are highly altered to talc (?) and magnetite. Rhombohedral to anhedral pleochroic light green to brown hornblende forms <5% of rock. Concentrations of calcite and talc with rims of magnetite and hematite suggest olivine crystals that have been completely altered.

Basalt. Porphyritic, vesicular, amygdaloidal. Randomly oriented plagioclase laths, up to 3mm in length, and minor clinopyroxene (augite) phenocrysts are set in a brown (hematite) groundmass. Opaque material (magnetite-hematite) forms 20% of the rock. Amygdules are filled with quartz and serpentine (?). Alteration of plagioclase to sericite is nearly complete. Pyroxene is fresh. Vesicles originally occupied 10-20% of rock.

NANKOWEAP GROUP. Interbedded purplish sandstone, greenish shaly and conglomerate beds. Mostly thin-bedded. Unit forms a stair-like slope.

CONTACT. Unconformable. Local relief of 1-1.5m.


22. Amygdaloidal zone. Abundant amygdules are filled with quartz, epidote, chlorite and calcite. Unit weathers into crumbly slope.

23. Sandstone, f.g., baked. Thickness is very irregular; forms lenses.

24. Basalt. V.f.g. weathering brown to red-brown. Abundant vesicles and amygdules. Amygdules are filled with quartz, calcite, chlorite, epidote, serpentine, and talc. Chlorite-epidote alteration is common. Upper contact is very irregular, with relief of 3-6m. Unit forms a crumbly cliff.

25. Sandstone, f.g., purplish, baked. Thickness is very irregular; forms lenses.

26. Basalt, similar to that of unit 24, but with greenish outcrop color. Weathers into angular chips and forms a steep slope or subdued cliff. Units 24, 26, and 27 are distinctive from a distance, forming a triplet characterized by the following colors, in ascending order: tan, greenish, pinkish.

27. Lapillite. Scoriaceous lapilli, about 7.5cm in average size, and bombs as much as in in size, are set in a matrix of ash. Amygdules are filled with chlorite, epidote, quartz, and calcite. Forms rubbly slope. Bombs weather out in relief.

28. Basalt, f.g., vesicular, amygdaloidal. Vesicles are 1-10mm in size. Some are elongate parallel to bedding. Weathers into steep slope, the lower part of which is littered with nodules. Lowermost part of unit forms a cliff 1.5-3m high.

29. Sandstone, f.g., red to reddish purple on weathered surface well indurated. Thickness ranges from 3 to 6m. Flaggy bedding. Large-scale cross bedding.

30. Basalt, v.f.g., dark red to maroon in outcrop color, black to green-black on fresh surfaces. Composed of plagioclase, altered olivine and pyroxene, and magnetite-ilmenite. Chlorite-epidote alteration. Amygdules present near base and top of unit. Those near base are elongate parallel to bedding. Unit is massive, weathers into sharp crenules bounded by fractures.

31. Basalt, similar to that of unit 30, but with fewer amygdules. Forms pinkish cliff.
ovoid to spheroid bodies, the majority of which are produced by
spheroidal weathering. Some of the spheroidal shapes, however, are
suggestive of pillow lavas. Field and petrographic evidence indicates
that most of the greenish band consists of spilitic hyaloclastite.

Cliff-forming rocks above the lower unit appear uniformly layered
when viewed from a distance. Closer examination, however, reveals that
individual flows have primary dips and that some angular discordances
are common in the flow sequence. Planar erosion surfaces typically
overlain by thin resistant sandstone interbeds cut across both the
originally horizontal and dipping flows. It is these erosion surfaces
that give the impression of uniform layering. Flow boundaries are
marked both by sandstone beds and by vesicular zones; many of these
zones are inconspicuous. Detailed work will probably reveal more flow
units than are shown in figure 4.

Noteworthy structures in the section above the green unit include
an autoclastic breccia at the 100 m level, overlain by a unit with
pervasive platy jointing, well-preserved ropy structures at the 178 m
level, and a unit of lapillite with associated bombs in the interval
from 235 to 245 m above the basal contact. Structures higher in the
section suggest subaerial emplacement.

Sandstone beds generally range from less than 1 m in thickness to
as much as 6 m where they fill channels. Sandstone beds in the lower
part of the section are fine grained and relatively well sorted, and
tend to be platy or laminated. Higher in the section, the sandstone
beds are less well sorted, have a larger maximum grain size, rest on
surfaces with more irregularity, and appear to be more intensely baked.
Rubble derived from the underlying lava is incorporated into the bases
of these higher beds. The highest sandstone bed (at about the 263 m
level) locally has small-scale crossbedding.

Petrology and chemistry

Flows of the Cardenas Lavas can be divided into two groups on the
basis of petrography and chemistry: those below the sandstone bed at
93 m (fig. 4) and those above. Flows below this sandstone bed are
highly altered basalt which was originally hyalocrystalline with
ophitic, non-glassy parts. In rocks from the lower part of this unit,
mafic minerals are completely altered, with relict outlines preserved
by magnetite rims; plagioclase laths as much as 1 mm long are partially
altered to sericite and saussurite; unaltered parts of the plagioclase
have 0-5° extinction angles (albite twinning), indicating oligoclase.
PATCHES of devitrified glass consist of chlorite, epidote, and clay
minerals.

Ten meters below the base of the sandstone at the 93 m level, the
rock has a matrix of material similar to that described above but
containing patches of unaltered basaltic material that makes up
approximately 40 percent of the rock. The patches are roughly circular,
2-3 mm in diameter, and contain olivine, augite, and plagioclase (An45),
with minor magnetite. Both the altered and unaltered parts are similar
in their ophitic texture and grain size and tend to grade into each other.
Table 1.—Chemical analyses of intrusive and effusive rocks of the Inkar Group

<table>
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<th>Distance below top (metres)</th>
<th>BS5</th>
<th>BS6</th>
<th>BS7</th>
<th>BS8</th>
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Total in weight percent

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<th>Distance below top (metres)</th>
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<th>BS7</th>
<th>BS8</th>
<th>BS9</th>
<th>BS10</th>
<th>BS11</th>
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<td>100.19</td>
<td>100.45</td>
<td>100.24</td>
<td>78</td>
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</tbody>
</table>
All flows above the 93 m level are similar in texture but differ in vesicularity. These flows have an intergranular to intersertal texture and a smaller grain size than the rocks of lowermost 93 m of section. Notable mineralogic characteristics include presence of pleochroic green-brown poikilitic hornblende in samples Tb2 and Tb7 (fig. 4) but absence in other units; altered olivine in Tb4, Tb5, and Tb7 (?); and oligoclase in Tb2 but andesine in Tb4 and Tb5. The plagioclase type cannot be determined in samples Tb6 and Tb7 because of sericitization.

Chemical analyses for major oxides from nine samples of the Cardenas Lavas are shown in table 1. Comparison of these analyses shows two significant results: (1) Sample Tbl, from the lower 93 m, is chemically unique with respect to other units, and (2) lavas above 93 m change sequentially upward from relatively silicic to more mafic and then back to relatively silicic. Tbl differs chemically from the other lavas in having the highest silicon and sodium contents and the lowest total iron and potassium values. These data support the conclusion that this lower unit has undergone spilitic alteration. The more mafic lavas above the 93 m level are siliceous basalts; the relatively silicic ones are basaltic andesites.

INTRUSIVE ROCKS

Thicknesses and distribution

All major intrusive bodies of the Unkar Group are similar in texture, mineralogy, and chemistry throughout the area, and were probably emplaced during the same phase of igneous activity. The locations of these major intrusives are shown in figure 1. Sills range in thickness from 23 m at Hance Rapids and Clear Creek to a reported 300 m in Hakatai Canyon (Noble, 1914, p. 55). Thicknesses of other sills in the Grand Canyon include 140 m along Bright Angel Creek; 25-35 m along Phantom Creek; a minimum of 100 m along Crystal Creek; 100 to 150 m in the vicinity of Shinumo Creek; and 215 m along the Colorado River near Bedrock Canyon. Part of the Crystal Creek sill has been removed by ep-Algonkian erosion.

Petrology

All intrusions have fine-grained chilled margins, with the exception of the Phantom Creek sill, which has narrow shear zones at the contacts with the Shinumo Quartzite. Flow texture is indicated by aligned microlites. Chilled margins, about 30 cm thick, contain phenocrysts of olivine and plagioclase set in a groundmass of plagioclase microlites, clinopyroxene, and opaque material.

All sills of the Unkar Group have an interior composed of medium- to coarse-grained diabase containing olivine, plagioclase, clinopyroxene, magnetite-ilmenite and biotite with accessory apatite and sphene. Plagioclase (An$_{45-60}$) laths average 1.5 mm in length and are partially
to completely altered to sericite. Both normal and reversely zoned crystals are common in any thin section. The more calcic plagioclase is concentrated in the centers of the sills.

Anhedral subhedral olivine crystals as much as 1 mm across are partially altered to chlorite, talc, magnetite, iddingsite, and serpentine along borders and fractures. The fresh interiors of the grains have optic axial angles that indicate compositions of approximately Fa<sub>20</sub>. Interference colors suggest normal zoning.

Plagioclase laths and olivine grains are enclosed by large poikilitic clinopyroxenes. The pyroxenes have ZA C = 45°, 2V<sub>y</sub> = 50° and are brownish-pink, non-pleochroic, and fresh. Pyroxene with these optical properties occurs in the diabasic rock of all the sills. Large irregular grains of magnetite altered to hematite and biotite, as well as primary pleochroic brown biotite partially altered to chlorite, also occupy interstices between plagioclase and olivine grains.

Olivine tends to be concentrated in the center of the Grand Canyon sills, whereas clinopyroxene decreases in modal percent toward the center. In sills elsewhere, these relationships have been explained by a process of flow differentiation (Simkin, 1967; Drever and Johnston, 1967; Bhattacharji and Smith, 1964; Bhattacharji, 1967; Gibb, 1968; Komar, 1972a, 1972b) that involves movement of early formed particles (olivine and Ca-plagioclase) away from the margins of a sill or dike during flow of the magma. This movement, combined with gravitational settling, produces a gradual increase in olivine concentration from the lower contact upwards and an abrupt increase from the upper contact downward. Gravitational settling after emplacement may produce an olivine-rich layer near the base of the sill. In the sills of the Unkar Group the modal olivine changes abruptly from 0-5 percent near the upper chilled margins to 20-30 percent in the centers and then gradually decreases to about 10 percent near the bases. These results are in agreement with a flow differentiation model. The sill 800 m east of Shinumo Creek has an olivine-rich layer near the base. This layer contains about 50 percent olivine and probably represents gravitational settling following intrusion.

The thicker sills of the Unkar Group show textural variants from the normal diabase of the other sills. These variants include: (1) lumps or balls, about 10 cm in diameter, of ophitic intergrowths of clinopyroxene and plagioclase, and (2) pegmatite veins typically less than 1 m thick consisting of clinopyroxene and plagioclase. These variants were noted in the Hakatai Canyon-Shinumo Creek area by Noble (1914, p. 56-57). The relatively coarse grain size of these anomalous rocks suggests that they had a longer cooling history or were higher in water content than the remainder of the respective sills and may thus represent the last part of the sill to crystallize. Some of this late-crystallizing magma was concentrated in "lumps" and some in more tabular bodies, possibly along fractures.

Magmatic differentiation of the diabase is indicated by granophyre layers and by dikes of granophyre and white felsite. Although most of these differentiation products are confined to the sills, in the Hance area a felsite dike occurs stratigraphically above the sill and just
below a 6-m-thick diabase dike that probably is an offshoot of the sill. The felsite has pseudomorphs of completely altered plagioclase as much as 0.75 mm long, set in a matrix of plagioclase microlites and unidentified white isotropic material. The composition of the plagioclase phenocrysts is unknown because of the extensive alteration, but the microlites have extinction angles of nearly 0°, indicating An25-30.

Pink granophyre occurs both as dikes that cut the diabase sills and as layers that form the uppermost part of some of the sills. The Bright Angel Creek and Bedrock Creek sills are intruded by granophyre dikes that generally are subparallel to the attitude of the sills. These dikes range from 30 cm to 1 m in thickness and extend laterally for hundreds of meters. The sill 800 m east of Shinumo Creek is composed of 85 m of diabasic rock capped by 6 m of granophyre. The contact between granophyre and the overlying Hakatai Shale is traceable for 100 m. The contact is sharp, and no xenoliths of Hakatai were found in the igneous rock, suggesting that the granophyre was not produced by assimilation of the Hakatai. The transition from granophyre to diabase in this sill occurs in less than 1 m, suggesting that the residual magma "floated" to the top of the sill and crystallized with little late-stage mixing with the diabasic part of the sill. In the other thicker sills the residual material is more disseminated, and well-developed granophyre layers are not present. The upper parts of these sills are relatively silicic and may represent internal mixing during the final cooling of the sills. Table 1 gives chemical analyses of 17 samples collected from chilled margins and interiors of sills, and 2 analyses of dike rocks.

Contact metamorphism

Contact metamorphism and metasomatism resulting from the intrusion of sills produced chrysotile asbestos in the Bass Limestone. Asbestos fibers as long as 10 cm occur within 3 m of both the upper and lower contacts of these sills. Where intruded by sills, the Hakatai Shale has been baked into a knotted slate. Biotite porphyroblasts as much as 0.25 mm in size occur within 5 cm of the lower contact in Shinumo Creek. Below the Shinumo sill, andalusite and cordierite (?) porphyroblasts, replaced by muscovite and green chlorite, respectively, are present. These porphyroblasts become larger and less numerous away from the sill, reflecting a slower rate and lower density of nucleation. Approximately 5 m below the sill, no recrystallization has occurred. Noble (1914) noted that the contact metamorphism was greater below the Shinumo sill than above it.
RELATION OF INTRUSIVE TO EXTRUSIVE ROCKS

Studies by Walcott (1894), Keyes (1938) and Maxson (1961, 1966, 1967) discuss the intrusive and extrusive rocks of the Unkar Group without commenting on the relation between them. Maxson (1967) placed all igneous rocks of the Unkar Group into the Rama Formation, thus implying a direct association between intrusive and extrusive rocks.

A direct relation between intrusive and extrusive rocks should be reflected by similarities in absolute ages, in mineralogy, and in chemistry. Rb-Sr age determinations on the igneous rocks of the Unkar Group by McKee and Noble (this volume) and E. H. McKee (written commun., 1973) give ages of about 1,100 m.y.

Chemical analyses (table 1) indicate that most of the flows are much more silicic than the sills and therefore probably were not emplaced during the same phase of igneous activity. Chemical variation diagrams (fig. 5) suggest that the Cardenas Lavas and the intrusive rocks of the Unkar Group may have formed from a common parent magma. Despite some scatter, these graphs show a distinct trend. Analyses plotted on an AMF diagram (fig. 5B) suggest a trend showing a decrease in the relative magnesia without early crystallization of iron, assuming that intrusive and extrusive rocks have a common source. Kuno (1968) believes that a diagram of solidification index plotted against major oxides is a good indicator of progressive magmatic differentiation. The solidification index would decrease in value with progressive differentiation. Such a diagram (fig. 5C) also suggests a trend defined by analyses of both intrusive and extrusive rocks.

The mineralogy of the lavas above the 93-m level in Basalt Canyon shows that this part of the flow sequence is much more silicic than the intrusive rocks. Below the 93-m level, however, unaltered parts of the lavas are similar in mineral composition to the sills. Unfortunately, these unaltered parts are too small to be analyzed by conventional chemical techniques. Nevertheless, the similarities in mineralogy and texture suggest that the sills and the lower 93 m of lavas are comagmatic and probably coeval. The differences in chemistry probably reflect spilitization of the lower unit of lavas. Lavas above the 93 m level were extruded after differentiation of the parent magma.

REGIONAL CORRELATIONS

Generally unmetamorphosed sedimentary and igneous rocks of younger Precambrian age are locally exposed in the southwestern United States. Such rocks have been named the Grand Canyon Supergroup in northern Arizona and the Apache Group and Troy Quartzite (Ransome, 1915) in southern and central Arizona. In Nevada and California the Pahrump Series (Hewett, 1940) of younger Precambrian age crops out in a narrow belt about 40 km wide and 120 km long in the Death Valley region.
Figure 5.—Chemical variation diagrams for the Unkar igneous rocks. (A) Silica versus alkali diagram. (B) AMF diagram. (C) Solidification index (SI) versus major oxides.

\[
(SI) = \frac{\text{MgO} \times 100}{(\text{MgO} + \text{FeO} + \text{Fe}_2\text{O}_3 + \text{Na}_2\text{O} + \text{K}_2\text{O})}
\]
All these sequences of rocks contain diabasic intrusions that yield radiometric ages from 1050 to 1200 m.y. All the younger Precambrian rock sequences in Arizona contain lava flows; in southern and central Arizona the flows occur at two stratigraphic levels, both of which are below diabasic intrusions, whereas in northern Arizona flows occur only above the intrusions at one stratigraphic horizon. Correlation of all these rocks has been suggested by many authors: Darton (1910, 1925, 1932), Noble (1934), Stoyanow (1936), Shride (1967), McKee (1969), Nehru and Prinz (1970), Ford and others (1972), Ford and Breed (1973), Elston and Scott (1974). Correlation based on the absolute ages and stratigraphic position of igneous rocks only indicates that all these rocks are of approximately the same age. More refined correlations may be made possible through detailed studies of the structure, texture, mineralogy, and chemistry of sedimentary sequences. Refinement in the methods of paleomagnetism as well as increased knowledge of paleomagnetic pole positions also may provide a more definitive basis for correlation.

CONCLUSIONS

Results of this study enable us to derive the following conclusions:

1. The Cardenas Lavas were deposited in a shallow low-energy sea that became shallower in time and eventually disappeared. The rate of subsidence of the basin was roughly equivalent to the rate of accumulation of the lavas.
2. All intrusive rocks of the Unkar Group are similar in mineralogy, chemistry, and stratigraphic position, were probably emplaced during the same phase of igneous activity, and probably originated from a single parent magma.
3. The Cardenas Lavas, although more acidic than the intrusive rocks, probably originated from the same parent magma that produced the dikes and sills.
4. The lowermost 93 m of the flow sequence may be the altered equivalent of the intrusive rocks. Although the chemistry of this lower unit is different from that of the sills, the difference may be the result of spilitization of the flows.

REFERENCES CITED


Hewett, D. F., 1940, New formation names to be used in the Kingston Range, Ivanpah quadrangle, California: Washington Acad. Sci. Jour., v. 30, no. 6, p. 239-240.


RB-SR AGE OF THE CARDENAS LAVAS, GRAND CANYON, ARIZONA

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Publication authorized by the Director, U.S. Geological Survey
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ABSTRACT

Six whole-rock specimens of basalt from the Cardenas Lavas of the younger Precambrian Unkar Group yield a Rb-Sr isochron of 1.09 ±0.07 b.y. This age is believed to approximate the time of extrusion of the lavas. Potassium-argon age determinations on the lavas are considerably younger and may reflect a period of heating about 800 to 900 m.y. ago.

INTRODUCTION

Samples of basalt from the Cardenas Lavas and diabase from sills in the Unkar Group were collected for radiometric dating in conjunction with paleomagnetic studies of the younger Precambrian of the Grand Canyon (Elston and Grommé, this volume). Both Rb-Sr and K-Ar techniques were used to attempt to ascertain the age of emplacement of these igneous rocks. The Rb-Sr data plotted on a strontium evolution diagram (Rb$^{87}$/Sr$^{86}$ versus Sr$^{87}$/Sr$^{86}$) yield an isochron that is considered to represent the age of emplacement of the basalts.

ACKNOWLEDGMENTS

Noble was supported by National Science Foundation Grants GA-1546 and GA-38296 and by National Aeronautics and Space Administration Grant NGR-22-007-103. The advice of C. E. Hedge, R. W. Kistler, S. C. Grommé, D. P. Elston, and C. T. Wrucke, Jr., is particularly appreciated.

SAMPLES

Hand specimens of basalt from six flows spanning the 300-meter thickness of the Cardenas Lavas (Hendricks and Lucchitta, this volume) were selected for Rb-Sr analysis. They were collected in Basalt Canyon on the northwest side of the Colorado River, which exposes one of the thickest and most complete sections of the Cardenas Lavas in the Grand Canyon (fig. 1). The samples were obtained at fairly evenly spaced intervals from the stratigraphic section (fig. 2). No samples from the highly altered rock in the bottom 60 meters were analyzed. The sample of diabase included in this study was collected from a sill near Hance Rapids about 5 miles southwest of Basalt Canyon. Two of the samples from Basalt Canyon, and one from the southwest side of the Colorado River in an unnamed canyon west of Tanner Canyon, were dated by K-Ar method.
Figure 1.—A part of the eastern Grand Canyon with locations of sample-collecting sites used in this study.
Figure 2.—Stratigraphic column of the Cardenas Lavas at and near Basalt Canyon. Numbers correspond to samples used for radiometric dating.
ANALYTICAL METHODS

Strontium isotopic composition was determined on a 6-inch,
60°-sector, NBS mass spectrometer utilizing a triple-rhenium-filament
mode of ionization. Data were collected digitally (Stacey and others,
1971). All Sr isotopic data are normalized to a value of 0.1194 for
$^{86}\text{Sr}/^{88}\text{Sr}$. The mass spectrometer yields a mean value of 0.7080 for
the Eimer and Amend standard SrCO$_3$. The constants used in the Rb-Sr
calculations are:

$$\text{Rb}^{87} \lambda_\beta = 1.39 \times 10^{-11} \text{yr}$$
$$\text{Rb}^{87} = 0.283 \text{ g/g Rb}$$

Rubidium and strontium concentrations were determined by x-ray
fluorescence using an Mo tube operated at 75KV. The intensities of
the Compton-scattering Mo Kα peak was used to correct for matrix
effects, and an iterative computer procedure incorporating peak-shape
and background data determined on appropriately spiked quartz powders
(D. C. Noble and W. P. Doering, unpub. data, 1971) was used to correct
for peak interference and background curvature. U.S. Geological
Survey standard rocks AGV-1 (Rb = 67 ppm, Sr = 657 ppm) and BCR-1
(Rb = 48 ppm, Sr = 332 ppm) were used as standards. The precision
of the resultant Rb-Sr ratios is believed to be somewhat better than
that obtained by Doering (1968).

The values for all analyses are listed in table 1.

Argon analyses were made using standard isotope-dilution
techniques with a Nier-type 6-inch-radius 60°-sector mass spectrometer.
The potassium analyses were by flame photometer using a lithium
internal standard. The ± values of the K-Ar ages were assigned on the
basis of experience with duplicate analyses; they represent the
uncertainties in argon and potassium analyses, in the isotopic
composition and concentration of Ar$^{38}$ tracers, and in the concentrations
of the flame photometer standards. Constants used in the K-Ar age
calculations are:

$$\lambda_\beta = 4.72 \times 10^{-10} \text{yr}$$
$$\lambda_\varepsilon = 0.585 \times 10^{-10} \text{yr}$$
$$K^{40}/K_{\text{total}} = 1.19 \times 10^4 \text{ moles/mole}$$

The ages and analytical data are listed in table 2.
Table 1.—Rb-Sr analytical data for basalt flows in the Cardenas Lavas and one diabase sill in the younger Precambrian, Grand Canyon.

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<th>Sample number</th>
<th>Location lat. long.</th>
<th>Rb (ppm)</th>
<th>Sr (ppm)</th>
<th>Rb/Sr</th>
<th>Rb$^{87}$/Sr$^{86}$</th>
<th>Sr$^{87}$/Sr$^{86}$</th>
<th>±2σ</th>
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<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>b-8</td>
<td>36°07'00&quot;N. 111°51'15&quot;W.</td>
<td>106</td>
<td>193</td>
<td>0.551</td>
<td>1.60</td>
<td>0.73065</td>
<td>0.0003</td>
</tr>
<tr>
<td>Tb-6</td>
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<td>0.584</td>
<td>1.595</td>
<td>0.73030</td>
<td>0.0002</td>
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<tr>
<td>b-7</td>
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<td>0.72708</td>
<td>0.0002</td>
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</tr>
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<td>b-3a</td>
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<td>116</td>
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<td>0.74803</td>
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<td>0.793</td>
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<td>0.74274</td>
<td>0.0003</td>
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<td>(sill)</td>
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<td>0.70956</td>
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Table 2.—K-Ar analytical data and radiometric ages of basalts from Cardenas Lavas, Grand Canyon.

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<th>Sample number</th>
<th>Location</th>
<th>Location lat.</th>
<th>Location long.</th>
<th>K\textsubscript{2}O</th>
<th>Ar\textsuperscript{40}rad (mol/g)</th>
<th>Ar\textsuperscript{40}rad%</th>
<th>Age x 10\textsuperscript{6} yrs</th>
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<td>4.63x10\textsuperscript{-9}</td>
<td>98</td>
<td>810±20</td>
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<tr>
<td>Tb-6</td>
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<td>3.33</td>
<td>4.83x10\textsuperscript{-9}</td>
<td>95</td>
<td>790±20</td>
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<td>99</td>
<td>781±20</td>
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</table>

CONCLUSIONS

The data obtained from the six lava specimens plotted on a Rb\textsuperscript{87}/Sr\textsuperscript{86} versus Sr\textsuperscript{87}/Sr\textsuperscript{86} diagram are shown on figure 3. The best line through them determined by least-square fit using a modification of the procedure described by York (1969) gives an age of 1.09 ±0.07 b.y. and an initial Sr\textsuperscript{87}/Sr\textsuperscript{86} ratio of 0.7065 ±0.0015 (1σ). The points on the strontium evolution diagram define a good isochron (all but one of the points lie very near the line), suggesting that the Sr and Rb contents of the rocks have not been substantially disturbed since eruption. The single point from the diabase sill is plotted on the diagram, but was not used in the regression analysis, as it cannot be shown geologically to be related to the flows. Although this specimen has much lower Rb and higher Sr contents than do the lavas, other specimens from a differentiated sill at Shinumo Creek have Rb and Sr contents and Rb/Sr ratios which fall within the range found for the lavas (McKee, unpub. data). Nevertheless, the possibility cannot be ruled out that the sill at Hance Creek has a slightly different age and (or) initial Sr\textsuperscript{87}/Sr\textsuperscript{86} ratio than do the lavas. If the specimen from the sill is included along with the six specimens of lava, the slope of the isochron so defined gives an age of 1.18 ±0.04 b.y.

We consider the Rb-Sr age of about 1.1+ b.y. to be the age of extrusion of the Cardenas Lavas. This age is considerably older than radiometric ages determined by the K-Ar method. Ford and others (1972) report a K-Ar age of 845 ±15 m.y. for a lava from the formation, and three whole-rock K-Ar ages determined in this study are 810, 790, and 781 m.y. Continuous argon loss may have occurred as a result of weathering or heating from deep burial. If this was the case the consistency of these apparent ages is fortuitous. The consistency of the three K-Ar ages reported here suggests that the lower radiometric ages obtained by the K-Ar method may reflect an episode of heating about 800 m.y. ago.
Figure 3.—Rb-Sr whole rock isochron plotted on a strontium evolution diagram. The six points are from different basalt flows in the Cardenas Lavas. The Rb-Sr point for the diabase sill (Tis-1) is included but was not used in calculating the isochron.

Isochron includes b-3a, b-3, b-8, Tb-6, b-7, b-6

\[ t = 1.09 \pm 0.70 \text{ b.y.} \]

Initial \( \frac{Sr^{87}}{Sr^{86}} = 0.7065 \pm 0.0015 \)
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PRECAMBRIAN POLAR WANDERING FROM UNKAR GROUP AND
NANKOWEAP FORMATION, EASTERN GRAND CANYON, ARIZONA

by

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ABSTRACT

Paleomagnetic poles for the Unkar Group and overlying Nankoweap Formation of the Precambrian Grand Canyon Supergroup have been derived from detailed sampling across much of a 2,000-m-thick section of red beds and lava flows. The apparent polar wandering path plots in the present central and north-central Pacific Ocean, and its trace describes a complex double loop. The loops are elongate in a north-south direction and are superimposed on a general westerly trend. The age range of the Unkar Group and Nankoweap Formation, from isotopic and geologic considerations, and from paleomagnetic correlations with late Precambrian rocks of central Arizona, lies in the interval 1.4-1.0 b.y. The Grand Canyon pole path thus includes parts of magnetic intervals defined from other parts of the North American continent (Elsonian, MacKenzie, and Keweenawan). The complex character of the Grand Canyon pole path, derived from stratigraphically controlled data points, emphasizes the difficulty of constructing a polar wandering path solely by connecting poles from isotopically dated rocks. Nonetheless, there is a broad coincidence of pole positions in this age range, which indicates that the pole for North America did reside in the area of the present central Pacific Ocean during this part of Precambrian time, and that major oscillations of the pole path occurred in the present north-south direction. Characteristics in the pattern of the pole path hold promise for regional and perhaps intracontinental correlations of Precambrian rocks, and for establishing a magnetic chronology that may be linked with tectonic events in the North American craton.

INTRODUCTION

Correlation of Precambrian stratified rocks has long been hampered by an inherent lack of evidence for relative age of the kind that fossils provide for Phanerozoic rocks. A typical example exists in Arizona, where the correlation of late Precambrian rocks of central Arizona with rocks of the Grand Canyon Supergroup of northern Arizona has been the subject of considerable speculation. During the last decade, it has become recognized that stratified eruptive and sedimentary rocks contain a sequential record of changes of paleomagnetic directions and reversals of the earth's magnetic field (for example, DuBois, 1962; Helsley, 1969). Such a record should allow us to understand the history of apparent motion of the continental plate with respect to the geographical pole and also has potential as a geologic tool for correlating rock units. In order to test the usefulness of a paleomagnetic chronology, reference paleomagnetic sections are needed that are derived from sequences of rock which record, as completely as possible, long intervals of time. The Grand Canyon, recognized for its magnificent Paleozoic record, also exposes a thick section of well-preserved, gently dipping, essentially
unmetamorphosed Precambrian stratified rocks, the Grand Canyon Supergroup. We report here the initial results of paleomagnetic analysis across about 2,000 m of rocks in the lower and middle parts of the Supergroup.

STRATIGRAPHY AND STRUCTURE

Characteristics of the Grand Canyon Supergroup are summarized in table 1. The distribution of formations of the Supergroup in the eastern Grand Canyon is shown in figure 1, as are the locations of sections sampled for paleomagnetic analysis. The stratigraphic relationships of the Unkar Group, Nankoweap Formation, and lower part of the Chuar Group are summarized in figure 2. Current studies have been confined to the Unkar Group and Nankoweap Formation, which are dominantly red beds.

Basaltic lava flows, the Cardenas Lavas (Ford and others, 1972), occur at the top of the Unkar Group and are 1090 ±70 m.y. old based on Rb-Sr analysis (McKee and Noble, this volume). The characteristics of the Cardenas Lavas and the relationships between the Cardenas and the overlying Nankoweap Formation are described elsewhere (Elston and Scott, 1973, 197 ). A composite section of the Cardenas Lavas (fig. 3) shows the stratigraphic distribution of paleomagnetic samples from flows and from interstratified and bounding sandstone beds.

Beds of the Supergroup dip gently (about 8° or less) to the northeast. The regional dip is interrupted on the east, just east of Basalt Canyon, where the Cardenas Lavas are at the level of the Colorado River in a graben at Tanner Canyon Rapids. The structural simplicity and excellent exposures have allowed precise structural corrections to be applied to the paleomagnetic samples.

PALEOMAGNETIC STUDIES

Objectives of study

Sampling was undertaken at close stratigraphic spacings (1/3-3 m) to obtain average directions of remanent magnetization in stratigraphic units and to identify intervals that might record polarity transitions or anomalous behavior of the magnetic field. We report here only the mean directions of the remanent magnetization for stratigraphic units in which samples indicated generally consistent paleomagnetic directions. In some stratigraphic intervals, directions of magnetization were somewhat scattered. A major interval of this kind occurs at the level of sample groups D-2, D-3, and D-4, D-5, in the upper-middle part of the Dox Sandstone (fig. 2). Paleomagnetic results from these, and fine details of the magnetic record in the sampled units and in sills and dikes that intrude the Unkar Group, will be presented in a later report.
Table 1.—Generalized description of the Grand Canyon Supergroup,¹/ eastern Grand Canyon, northern Arizona.

<table>
<thead>
<tr>
<th>Thickness (meters)</th>
</tr>
</thead>
<tbody>
<tr>
<td>201</td>
</tr>
<tr>
<td>100</td>
</tr>
<tr>
<td>5</td>
</tr>
<tr>
<td>400+</td>
</tr>
<tr>
<td>10</td>
</tr>
</tbody>
</table>

Precambrian

Grand Canyon Supergroup (Grand Canyon Group of Powell, 1876; Grand Canyon Series of Walcott, 1894)


- Shale, gray and variegated; subordinate gray locally algal-bearing limestone and buff dolomite; red sandstone and breccia at top---------------- - - --- -------

Disconformity (Walcott, 1894; Van Gundy, 1951)

Nankoweap Formation (Nankoweap Group of Van Gundy, 1934, 1951; Nankoweap Formation of Maxson, 1961, 1967)

Upper member—(modified from Van Gundy, 1951)

- Sandstone, shaly and silty sandstone, and shale;
  mainly fine grained, red and purple; well stratified;
  thin- to medium-bedded, crossbeds, ripple marks, and
  mud cracks common; casts of wormlike trails locally abundant in shaly sandstone----------------- 100

Disconformity (?)

Ferruginous member¹/

- Sandstone, very fine to coarse grained, and siltstone,
  grayish-red to very dark red, finely laminated to
  massive, parallel-bedded, resistant; hematite laminae
  and hematite and locally magnesite cement abundant;
  altered detritus from adjacent lava flows abundant;
  unit rests on ferruginous weathered zone on lower,
  slope-forming part of Cardenas lavas west of Tanner
  Canyon----------------------------------------------- 5

Thickness of Nankoweap Formation from reconstruction------------------------------------------ 400+

Disconformity (Van Gundy, 1951)

Unkar Group (Unkar terrane of Walcott, 1894; Unkar Group of Noble, 1914)

Ferruginous weathered zone on Cardenas Lavas ¹/

- Deeply weathered and altered basaltic andesite;
  resistant; original textures preserved but groundmass
  pervasively stained and bonded by hematite, and
  locally altered to siderite and earthy hematite.
  Developed across truncated section of Cardenas
  Lavas west of Tanner Canyon----------------------- 10

¹/ As defined and described in Elston and Scott (in press)
²/ Ford and Breed, 1973
Table 1 (continued)

Cardenas Lavas (lava series of Walcott, 1894; Cardenas Lava Series of Keyes, 1938; Rama Formation of Maxson, 1961, 1967; Cardenas Lavas of Ford and others, 1972)

Basaltic andesite flows, dark-gray, greenish-gray and green. Lower part, about 76 m thick in the Basalt Canyon area, is dark to medium green, weathers to a crumbly slope, and contains several thin sandstone beds; overlain by 4.6 m of dark red-gray sandstone; upper part commonly forms cliffs and ledges, consists of six or more compact flows separated by thin, planar layers of sandstone 1 to 3 m thick. West of Tanner Canyon a ledge-forming flow, 4.6 m thick, underlies the lower rubbly weathering part and is separated from it by ~2 m of fine to very fine grained, thin-bedded red sandstone lithologically identical with the Dox Sandstone; this basal flow pinches out to the north and west. Contact with Dox may be unconformable in northeasternmost exposures of formation. At Tanner Canyon Rapids, section is capped by fine-grained parallel-bedded and laminated sandstone that is extremely well cemented with hematite; hematite alteration appears to locally extend downward ~15 m into underlying flow

Disconformity(?)

Dox Sandstone (Noble, 1914)
Lower third of formation, light brown and dark purple-red, crossbedded, cliff- and ledge-forming sandstone; upper two-thirds, red-brown, micaceous, slope and locally ledge forming silty sandstone and sandstone

Shinumo Quartzite (Noble, 1914)
Sandstone, quartzitic; crossbedded; white, brown, and locally pale purple to dark red; forms cliffs

Hakatai Shale (Noble, 1914)
Lower third, red-brown siltstone and silty sandstone; middle third, orange-red silty sandstone; upper third, fine- to medium-grained, crossbedded lavender sandstone; forms slopes and minor ledges

Bass Limestone and underlying Hotauta Conglomerate (Noble, 1914)
Dolomite and limestone, locally algae-bearing, and interbedded sandstone, conglomerate, and siltstone; purple-brown to dark red, and reddish brown; forms cliffs, ledges and steep slopes

Thickness of Unkar Group ~1700-2200

Thickness of Grand Canyon Supergroup ~3350-4600+
EXPLANATION

Phanerotoic Rocks

Unconformity

Tapeats Sandstone (Cambrian); locally includes Bright Angel Shale

Precambrian Rocks

Unconformity

Grand Canyon Supergroup

Chuar Group, undivided

Unconformity

Nankoweap Formation

Upper member (Nu)

Ferruginous member (NF)

Unconformity

Ferruginous weathered zone on Cardenas Lavas (fwz)

Unkar Group

Cardenas Lavas (C)

Unconformity

Dox Sandstone (D 1, 2, 3; 4, 5 & 6, 7 & 8, 9, 11 & 10)

Unconformity

Shinumo Quartzite (S 1, 2, 3)

Unconformity

Hakatai Shale (H)

Bass Limestone (B); includes Hotauta Conglomerate at base; silt near base not shown on map.

Crystalline Rocks

Unconformity

Vishnu Schist
Figure 1.—Generalized geologic map of Precambrian rocks, eastern Grand Canyon, Arizona, showing locations of sampled sections. Geology from Maxson (1967).
Figure 2.—Generalized geologic section of Unkar group and Nankoweap Formation, eastern Grand Canyon, showing intervals sampled for paleomagnetic analysis.
Figure 3.—Composite geologic section, Cardenas Lavas, showing units sampled for paleomagnetic analysis, from Basalt Canyon, graben at Tanner Canyon Rapids, and canyon west of Tanner Canyon.
Sampling and laboratory procedures

Samples were collected by means of a hand-held diamond-core drill, similar to that described by Doell and Cox (1965). Cores, 2.5 cm in diameter and 5-10 cm long, were oriented before being detached from the outcrop; the magnetic compass and clinometer were read to the nearest half degree, with a probably accuracy of about ±1 degree.

The natural remanent magnetization (NRM) of all cores was measured in an air-driven spinner magnetometer that has a noise level of $2 \times 10^{-7}$ emu. Pilot samples for each group of cores then were subjected to progressive stepwise demagnetization analysis, both thermal (200°-630°C) and alternating field (25-5000 oe) for the determination of intervals of stable magnetization and of characteristic magnetization (Zijderveld, 1967). Criteria that were used to select stable directions of magnetization included: 1) the end points of demagnetization paths, which commonly showed directions that shifted slightly away from the present field direction; 2) minimum dispersions for groups of samples from individual stratigraphic units; and 3) general trends of the magnetization vectors toward the origins of orthogonal demagnetization diagrams. Following this, the remaining specimens were cleaned at single or multiple demagnetization steps, and stable directions of magnetization were derived for groups of samples. Most of the red-bed samples were thermally demagnetized, whereas most of the lava-flow samples were demagnetized in an alternating field; samples from some of the lava flows responded to thermal as well as alternating field demagnetization. Details of the cleaning will be reported in a later paper.

Paleomagnetic directions

Concordance of directions, flows and associated sandstone of Cardenas Lavas

As shown by alternating field and thermal demagnetization analysis (Elston and Scott, 1973), and by strong-field thermomagnetic measurements, the magnetization in the Cardenas flows resides in magnetite and titanomagnetite, and locally ilmenohematite and (syngenetic or near-syngenetic) hematite. The remanence is principally a thermoremanent magnetization (TRM) acquired as the individual flows cooled below the Curie points of the magnetic minerals, which range from 560°C to 680°C. Some of the remanence may result from high-temperature chemical remanent magnetization (CRM). The acquisition of a TRM provides a geologically instantaneous recording of the direction of the earth's magnetic field and gives rise to what is called a virtual geomagnetic pole (VGP). VGP's, when averaged for Quaternary and late Tertiary rocks, have been shown to approximate closely the earth's geographic or spin pole (Irving, 1964, p. 108). A similar relationship is assumed to have existed in the geologic past.
In contrast with the flows, the remanence in red sandstone in the Unkar Group and Nankoweap Formation resides largely in hematite. Except for depositional hematite lamellae in the ferruginous member of the Nankoweap (Elston and Scott, 1973, and in press) hematite of the red beds presumably was emplaced sometime after deposition of the sandstone. The key question concerns the time of emplacement. If the hematite were emplaced shortly (a few thousand years or less) after deposition, then a suite of stratigraphically sequential samples would be expected to record changes in direction of the earth's magnetic field with time. Furthermore, paleomagnetic directions in individual flows and interbedded sandstones would be similar, and the average direction of flows would be virtually identical with the average direction of sandstones. Lastly, a correspondence of stable directions from flows and sedimentary interbeds would be strong evidence of the absence of any residual secondary magnetization that may have been impressed on the rocks by the present magnetic field as a consequence of weathering, as well as evidence that the remanence in the sedimentary strata was acquired very shortly after deposition. In such a situation, the paleomagnetic directions and polarities of the flows and sedimentary beds would truly reflect the history of the magnetic field at the time of deposition.

Stable paleomagnetic directions of the flows and the sandstones are virtually identical (table 2; fig. 4). The lowest sandstone bed in the Cardenas (s-1, fig. 3) is a Dox-like sandstone that seemingly is traceable into the Dox. This sandstone and two flows (b-1 and b-4) were sampled south of the Colorado River in beds that locally dip 15°W. (Elston and Scott, 1973, and in press). Paleomagnetic directions of these three units, after correction for dip, coincide with directions from nearly level flows and sandstone beds sampled in Basalt Canyon, attesting both to the acquisition of remanence before folding and its stability in folded beds. The uppermost sandstone (s-6, fig. 3) has been thoroughly impregnated by massive hematite. The paleomagnetic data suggest that impregnation occurred very shortly after deposition of the uppermost flow of the Cardenas.

We conclude that the remanent magnetization in sandstone interbeds of the Cardenas Lavas was acquired bed by bed, and that the distribution and pattern of directions seen in the flows (which provide spot readings of the magnetic field) are a measure of Precambrian secular variation (fig. 4). The pattern is less clearly displayed by the individual sandstone beds. This might be expected if the remanence in the red sandstone was acquired slowly over intervals of a few centimeters or less in a manner that tended to average the secular variation of the direction of the magnetic field. A similar progressive acquisition of remanence at about the same rate as deposition seems to have occurred in some other red beds, for example, the Moenkopi Formation of Triassic age (Helsley, 1969; Helsley and Shoemaker, 1973).
Table 2.—Paleomagnetic directions, flows, and sandstone of Cardenas Lomas, eastern Grand Canyon, Arizona.

<table>
<thead>
<tr>
<th>Ident. No./</th>
<th>Vector</th>
<th>Declination</th>
<th>Inclination</th>
<th>Poles</th>
<th>Stable</th>
</tr>
</thead>
<tbody>
<tr>
<td>NRM</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>b 8</td>
<td>Stable</td>
<td>10.917</td>
<td>262.4</td>
<td>48.1</td>
<td>48.2</td>
</tr>
<tr>
<td>s 5</td>
<td>Stable</td>
<td>6.841</td>
<td>257.7</td>
<td>45.4</td>
<td>36.7</td>
</tr>
<tr>
<td>s 4</td>
<td>Stable</td>
<td>3.935</td>
<td>271.6</td>
<td>53.0</td>
<td>46.3</td>
</tr>
<tr>
<td>b 7</td>
<td>Stable</td>
<td>7.980</td>
<td>261.2</td>
<td>51.6</td>
<td>349.6</td>
</tr>
<tr>
<td>s 3</td>
<td>Stable</td>
<td>7.974</td>
<td>258.8</td>
<td>51.7</td>
<td>263.9</td>
</tr>
<tr>
<td>b 6</td>
<td>Stable</td>
<td>6.930</td>
<td>266.6</td>
<td>50.7</td>
<td>20.9</td>
</tr>
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<td>b 5</td>
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<td>26.3</td>
<td>217.4</td>
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<td>258.9</td>
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<td>7.887</td>
<td>272.2</td>
<td>48.8</td>
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<td>277.4</td>
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<td>23.7</td>
</tr>
<tr>
<td>b 2</td>
<td>Stable</td>
<td>6.955</td>
<td>266.3</td>
<td>32.8</td>
<td>134.1</td>
</tr>
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<td>s 1</td>
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<td>7.877</td>
<td>253.0</td>
<td>35.9</td>
<td>56.9</td>
</tr>
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<td>b 1</td>
<td>Stable</td>
<td>6.963</td>
<td>253.3</td>
<td>50.0</td>
<td>154.5</td>
</tr>
</tbody>
</table>

For stratigraphic relations, see figure 3.

1/ N = the number of individual cores obtained from stratigraphically different parts of individual sandstone beds and of individual lava flows. For sandstone beds, each core is considered to be a measurement of direction of magnetization for which secular directions have been largely averaged as a consequence of slow progressive acquisition of remanence; each sandstone core thus is classed as a specimen. For individual lava flows, each core is considered to provide a geologically instantaneous measurement of the direction of the magnetic field (a spot reading, or virtual geomagnetic pole (VGP)), and each core is classed as a specimen VGP's of specimens from individual flows are averaged to provide a better estimate of the magnetic pole at the time of cooling of the flow. The VGP's from several flows, in turn, are averaged to obtain a paleomagnetic pole (or simply, a pole). The averaged paleomagnetic direction for each flow, a VGP, thus is considered to be a single sample.

2/ k = (N-1)/N-R.

3/ Semi-angle of the cone of 95% confidence level.

4/ Semi-major and semi-minor axes of the oval of 95% confidence level.


6/ Used for calculation of pole.
Figure 4.—Plots of paleomagnetic directions; flows and sandstone, Cardenas Lavas, Unkar Group, eastern Grand Canyon, Arizona. Flows: small dots are individual samples from eight flows. See figure 3 for key to stratigraphic position of flows and sandstone beds.
Directions of magnetization reported here are not based on a centimeter-by-centimeter progressive magnetization of the red beds. Most sections used for calculating average directions across the Unkar Group and Nankoweap Formation (table 3) are several tens of meters thick. The magnetization in these rocks could as well have been impressed simultaneously on stratigraphic intervals measurable in meters, and a valid polar wandering path could still have been drawn. However we do not think that this has occurred because the magnetization directions of a number of successive individual cores, sampled on 1/3 to 3 m intervals, appear to describe non-random patterns. Furthermore, abrupt changes of magnetization direction within the Bass Limestone and Hakatai Shale (table 3 and fig. 5) are not marked by lithologic changes, and hence seem to represent times of relatively rapid apparent polar movement.

Poles

Stable paleomagnetic directions and derived paleomagnetic poles are summarized in table 3. Plots of the poles, enclosed by ovals at the 95 percent confidence level, are shown in figure 5. The poles are connected in stratigraphic order, producing an apparent polar wandering path. We must emphasize two points of uncertainty that exist in the path:

1. The pole of the Shinumo Quartzite was derived from the S 1 group of samples (fig. 2), and the path connecting the Shinumo with Hakatai and Dox drawn through this pole is a conservative estimate. Directions obtained from the underlying S 2 and S 3 groups of samples plot much farther north, and lie near and appear streaked toward the direction of the present magnetic field. Curiously, these latter samples are from the most thoroughly silicified section of the entire Grand Canyon Supergroup, one that we expected to be least affected by present weathering. Samples from the S 2 and S 3 intervals exhibit extremely stable magnetization when heated to 630°C. Thus, the pole of the lower to middle part of the Shinumo may have lain much farther north than the pole of S 1.

2. The path connecting the pole of the ferruginous weathered zone with poles of the Cardenas Lavas and the ferruginous member of the Nankoweap (fig. 5, poles 11-14) is not as closely controlled, stratigraphically, as the path for the overlying and underlying units. Additionally, the paleomagnetic direction of the ferruginous weathered zone is based on a small number of samples. However, we have two reasons for using this data point: (a) the ferruginous weathered zone, on geologic grounds, clearly predates deposition of the Nankoweap; the zone was sampled where it is developed on flow b 4, its remanence resides not in magnetite but
Table 3.—Paleomagnetic directions, Unkar Group and Nankoweap Formation, eastern Grand Canyon, Arizona.

| Ident No. | Stratigraphic Unit | NRM | N | R | M | D | T | I | K | \(a_{95}\) | Pole Latitude | Pole Longitude | Pole Sm | Pole Sp |
|-----------|-------------------|-----|---|---|---|---|---|---|---|---|-----------|----------------|----------------|---------|---------|
| 15 | Nankoweap Fm., upper mbr; Nu | Stable | 25 | 23.186 | 265.7 | -9.7 | 13.2 | 8.3 | 0.6 S | 164.6 E | 8.4 | 4.2 |
| 14 | Nankoweap Fm., ferrug. mbr.; NF | Stable | 6 | 5.959 | 276.9 | 38.9 | 122.3 | 6.1 | 18.1 N | 172.6 E | 7.3 | 4.3 |
| 13 | Ferrug. weath. zone on Cardenas Lava; fzw | Stable | 6 | 5.911 | 323.3 | 65.8 | 56.5 | 9.0 | 60.6 N | 166.3 W | 14.7 | 12.0 |
| 12 | Cardenas Lava, sandstone; C | Stable | 46 | 44.394 | 263.7 | 44.6 | 28.0 | 4.0 | 10.4 N | 176.8 W | 5.1 | 3.2 |
| 11 | Cardenas Lava, flows; C | Stable | 45 | 43.502 | 261.7 | 43.8 | 29.4 | 4.0 | 8.6 N | 176.3 W | 5.0 | 3.1 |
| 10 | Dox Sandstone, upper; D 1 | Stable | 14 | 13.094 | 264.1 | 29.6 | 14.4 | 10.9 | 4.6 N | 174.4 E | 12.0 | 6.7 |
| 9 | Dox Sandstone; upper middle; D 6 | Stable | 11 | 10.877 | 292.2 | 55.8 | 81.1 | 5.1 | 36.5 N | 179.9 W | 7.3 | 5.2 |
| 8 | Dox Sandstone; middle; D 7 & 8 | Stable | 64 | 62.257 | 306.1 | 57.9 | 42.8 | 2.7 | 47.7 N | 178.3 E | 4.0 | 3.0 |
| 7 | Dox Sandstone; lower middle; D 9 | Stable | 20 | 19.498 | 295.9 | 58.3 | 37.8 | 5.4 | 40.2 N | 178.1 W | 8.0 | 5.9 |
| 6 | Dox Sandstone; upper lower; D 11 & 10 | Stable | 78 | 75.706 | 294.4 | 33.7 | 33.6 | 2.8 | 37.4 N | 176.8 E | 3.9 | 2.7 |
| 5 | Shinumo Quartzite, middle; S 1 | Stable | 9 | 8.953 | 277.1 | 73.1 | 171.9 | 3.9 | 33.7 N | 150.1 W | 7.0 | 6.3 |
| 4 | Hakatai Shale, middle and upper; H | Stable | 14 | 13.341 | 305.4 | 67.8 | 19.7 | 9.2 | 48.8 N | 163.3 W | 15.4 | 12.9 |
| 3 | Hakatai Shale lower and middle; H | Stable | 11 | 10.584 | 299.4 | 60.3 | 24.0 | 9.5 | 36.2 N | 173.3 W | 14.4 | 11.0 |
| 2 | Bass Limestone, middle and upper; B | Stable | 19 | 18.711 | 233.3 | 38.8 | 62.3 | 4.3 | 13.2 S | 161.8 W | 5.1 | 3.0 |
| 1 | Bass Limestone, lower and Hotauta Cg.; B | Stable | 23 | 22.421 | 271.4 | 44.3 | 38.0 | 5.0 | 16.0 N | 178.9 E | 6.2 | 3.9 |

1/ Poles are plotted on figure 5; for stratigraphic relationships and key to symbols, see figures 1 and 2. For definitions of N, k, \(a_{95}\), Sm, and Sp, see table 2.
Figure 5.—Poles, and apparent polar wandering path, Precambrian Unkar Group and Nankoweap Formation, eastern Grand Canyon, Arizona. Ovals enclose areas containing poles at the 95 percent confidence level. See table 3 for key to numbered poles.
in hematite which replaced magnetite of the flow, and its stable direction of magnetization is markedly different from that of the parent lava flow; and (b) this data point appears to be supported by a pole for a weathered site in 1140-1150 m.y. old diabase in central Arizona (Helsley and Spall, 1974; Elston and Scott, 1973, 197 ). The pole of the weathered diabase lies well north of poles of unweathered central Arizona diabase. It is a few degrees north of and about 20 arc degrees west of the pole of the ferruginous weathered zone (fig. 5, pole 13). Helsley and Spall considered the weathered diabase pole to be a stable Precambrian direction and included this data point in calculating an average pole for the diabase. The ferruginous weathered zone in the Grand Canyon and deeply weathered diabase of central Arizona thus may well reflect a time of erosion and weathering across Arizona that followed emplacement of sills and extrusion of the Cardenas Lavas. The polar wandering path during the erosion interval that separates the Cardenas and Nankoweap probably cannot be adequately documented from Arizona data alone.

DISCUSSION

The Cardenas Lavas, which are about 1,100 m.y. old (McKee and Noble, this volume), appear on paleomagnetic grounds to be somewhat younger than 1,140-1,150 m.y.-old central Arizona diabase (Elston and Scott, 1973, and in press). Poles derived from the upper part of the Apache Group and the lower part of the overlying Troy Quartzite lie about 10° north of the pole for central Arizona diabase (Elston and Scott, 1973, and in press), and the foregoing all plot in or near the group of poles reported here for the Dox Sandstone. This correspondence has led to a provisional correlation of stratified Precambrian rocks of central and northern Arizona, from which it is concluded that deposition of the upper part of the Apache Group and Troy Quartzite, and subsequent intrusion of diabase, occurred during deposition of the Dox Sandstone. This correlation also implies that the lower and middle parts of the Apache Group are equivalent to the lower and middle parts of the Unkar Group. If the base of the Unkar is broadly equivalent to the base of the Apache, it postdates the Ruin Granite of central Arizona (1.42 b.y. old; Livingston and Damon, 1968, p. 769) and thus is about 1.4 b.y. old.

The apparent polar wandering path derived from Grand Canyon rocks is complicated, even if smoothed to remove the smaller oscillations or reversals in path directions. It is much more complicated than other polar wandering paths that have been proposed for North America, all of which have been constructed principally by connecting poles derived from groups of isotopically dated rocks (DuBois, 1962; Spall, 1971; Robertson and Fahrig, 1971; Irving and Park, 1972; Larson and others, 1973).
The Grand Canyon Precambrian polar wandering path is not congruent in detail with any path previously published for North America. This is not surprising because the pole paths controlled by isotopic dating lack sufficient temporal resolution to unambiguously resolve a path that may oscillate about a narrow track. Because of this resolution problem, paths have been constructed on the assumption that polar wandering proceeds from point A to point B in the simplest and most orderly manner without rapid loops or oscillations.

The Grand Canyon path (Unkar Group and Nankoweap Formation), from isotopic age and geologic considerations, apparently represents the period ~1.4-1.0 b.y. ago. It encompasses, in time, much of the Elsonian (1.3-1.475 b.y.), and all of the MacKenzie (1.2 or 1.25 b.y.) and Keweenawan (1.0-1.2 b.y.) magnetic trends of Spall (1971). The Elsonian trend lies some 20° east of the pole of the upper Hakatai Shale (fig. 5, pole 4); the Mackenzie trend intersects the path for the Bass Limestone and lower Hakatai Shale (fig. 5, poles 2 and 3), and the younger end of the Mackenzie trend coincides with the Cardenas pole (fig. 5, poles 11 and 12); lastly, the Keweenawan trend coincides with the Dox and Nankoweap pole paths. Nevertheless, there is a broad agreement, indicating that the pole for the North American continent did reside in the area of the present central Pacific Ocean during this part of Precambrian time, and moreover, that the pole path did oscillate principally in a (present) north-south direction.

The Grand Canyon path shows some unexpected and marked regularities. There is a double loop, trending in a counterclockwise manner, that is offset to the west with decreasing age. The path is not only elongate in a north-south direction but in part appears rectangular. This is in contrast to the mostly broad, meandering open loops of previously proposed paths. The somewhat ordered character of the Grand Canyon path, and in particular its abrupt changes in direction, could, in the context of plate tectonics, reflect an interaction of the North American continent with other continental plates, and the pattern of looping might ultimately allow intercontinental correlations to be made. For example, if Africa is placed off the southwest coast of North America, as shown by Piper and others (1973, fig. 5), the looping Grand Canyon path broadly coincides with a path for Africa that also is complexly looped and that in part represents the same interval of time.

Lastly, current paleomagnetic investigations in central and northern Arizona (Elston and Scott, 1973, and in press) suggest that with adequate data, correlations can be made on the basis of paleomagnetic directions. If this is the case, it may be possible to correlate the Precambrian of Arizona with stratified and other Precambrian rocks elsewhere in North America.
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by
Edwin D. McKee

The Toroweap Formation: A new look
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PALEOZOIC ROCKS OF GRAND CANYON

By

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INTRODUCTION

In many sections of Grand Canyon, the Paleozoic rocks, which form a major part of the total thickness of its walls, begin high above the river level. Thus, for a party descending the river, they are seen only from a distance throughout long stretches of the canyon. On the other hand, if Marble Canyon is included with Grand Canyon, every one of these Paleozoic formations is exposed along the riverbank in one or more places. This situation exists because the Colorado River begins its dissection of the virtually flat-lying Paleozoic beds in the youngest rocks (Kaibab Limestone) at the head of Marble Canyon, near Lees Ferry; downstream it cuts through progressively older strata all the way to the Precambrian, which it first reaches in the vicinity of Nankoweap Creek in Grand Canyon. Farther downstream the earlier Paleozoic formations reappear near river level as a result of displacement along major faults. These faults, as recognized by John Wesley Powell (1875, p. 182), outline the principal plateau blocks (fig. 1).

Study of the Paleozoic strata dates back even before Powell's work, for two pioneer expeditions, each including a geologist of great ability and considerable discernment, traversed this area along the plateau to the south and each left a report clearly stating his contributions to stratigraphy. The first of these early geologists was Marcou (1856); the second was Newberry (1861). Powell's famous Colorado River trips of 1869 and 1871-72 resulted in the next scientific reports on this area, but for some unknown reason his reports give little specific data on the Paleozoic strata among which he had spent so much time and concerning which he undoubtedly had great interest. Perhaps he was too busy with the mere fact of survival; perhaps he was diverted by the large number of other new and strange phenomena that surrounded him.

John Wesley Powell's greatest contribution to our knowledge of Grand Canyon geology lies in his original genius, coupled with his recognition of projects worthy of investigation and his influence in stimulating colleagues in the conducting of such investigations. During the last several decades of the nineteenth century, virtually all of the Grand Canyon studies—the pioneer studies—were accomplished by this small group of scientists: G. K. Gilbert, C. E. Dutton, A. R. Marvine, and C. D. Walcott. Among other things, Gilbert was one of an intrepid band that actually forced a passage upstream along the Colorado River as far as Diamond Creek; Dutton became internationally known for his vivid word pictures, especially of the Toroweap area, following two seasons of fieldwork in a region that even now is just beginning to be well known. Walcott permanently associated his name with the geology of the region, chiefly because of two scientific expeditions: first, he descended Kanab Canyon and recorded new aspects of the stratigraphic section; second, and more important, he studied the late Precambrian and Cambrian geology in the isolated eastern end of the Grand Canyon, camping there much of one long winter and working under very adverse conditions.
Figure 1.—Section from west to east across the plateaus north of the Grand Canyon, with bird's-eye view of terraces and plateaus above. Horizontal scale, 16 miles to the inch; vertical scale, 4 miles to the inch. (From Powell, 1875, fig. 73.)

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Studies related to the stratigraphy of Grand Canyon that have been conducted since 1900 are far too numerous to mention here, but a fairly complete description of these studies and an analysis of their respective contributions have been made (McKee, 1969). In the present discussion a chronological approach is used. Each of the principal Paleozoic formations is briefly described and its genesis is evaluated in the light of recent studies; facies changes and other lateral variations encountered from one end of Grand Canyon to the other are discussed; unconformities of special significance are described and evaluated.

THE GREAT UNCONFORMITY AND PRE-TONTO ENVIRONMENT

The younger of the two great unconformities in Grand Canyon, originally made known through the writings of John Wesley Powell, is one of the most significant and most spectacular geological exhibits anywhere in the canyon walls. Its record is plainly seen from many vantage points on the canyon rims, but it is perhaps even more impressive where observed from closer sites along the Colorado. In places this unconformity is a remarkably flat, even surface for a distance of miles; it bevels the upturned ends of schists and other metamorphic rocks of early Precambrian time and is covered by flat-lying strata of the Cambrian. Elsewhere it is seen as cross sections of rugged hills or ridges, some hundreds of feet high, of late Precambrian quartzites and other resistant rocks, surrounded by and buried beneath sediments deposited in the Cambrian sea (fig. 2).

The time represented by this unconformity must be tremendous. It involves (1) the time required to develop a system of block fault mountains, outlined by a series of long northerly striking, normal faults (fig. 3), plus (2) the even greater time involved in eroding to a low hilly surface at least as much as 4,600 m (15,000 ft) of dipping late Precambrian strata. This great break in the geologic record was referred to by Walcott (1910) as the Lipalian Interval and he visualized it as a time "of unknown sedimentation between the adjustment of pelagic life to littoral conditions and the appearance of the Lower Cambrian fauna." Although an attempt was made at a later date (Hinds, 1939, p. 306) to minimize the magnitude of this unconformity and to dismiss the concept of a great land emergence with a major break in the record, calculations by Sharp (1940, p. 1260-61) of the time involved in lowering by erosion a land surface of the height represented by the Late Precambrian strata of Grand Canyon, indicated a period of roughly 100 million years. Thus, the time involved in the process of base leveling or the wearing away of mountains must have been very great according to any standards.

Details of the erosional record are so clearly preserved in numerous places along the canyon walls that some aspects of the history represented by this unconformity seem beyond question. In eastern Grand Canyon buried knolls and ridges projecting into the basal Cambrian strata were described by Walcott (1883) as illustrating
Figure 2.—Surface of erosion at base of Cambrian: (a) West side of Tapeats Creek. (b) South side of Colorado River opposite Bright Angel Creek. (c) North side of Colorado River opposite Horn Creek. (d) South side of Colorado River, 138 1/2 Mile Rapids. (e) West side of Tapeats Creek. (f) East side of river, north of 127 Mile Rapids. (g) East side of Pipe Creek. (From McKee and Resser, 1945, fig. 7).
Landscape of Grand Canyon Region at close of Algonkian Era

Figure 3.—Section along Colorado River. (From McKee, 1966b, p. 203).
"the sea breaking off and burying with drifting sand, fragments of the rocky islands." Farther west, on the south wall of Hotauta Canyon, an "island monadnock" is exposed (Noble, 1914, p. 62) "undercut by the waves of the sea... (preserving an) old sea cliff (at the base of which) huge angular blocks of Shinumo quartzite are incorporated in the Tapeats sandstone in places where they fell and lodged; farther out lie masses of boulders, worn and rounded by the pounding of the waves; and these boulders run into lenses of fine pebbly conglomerate, representing the shingle of the ancient beach, dragged out by the undertow."

The surface of weathering beneath the late Precambrian-Cambrian unconformity is well preserved in many places and furnishes additional clues concerning environment at the time of and immediately preceding transgression of the Cambrian seas. Although a dominance of mechanical disintegration was recorded by Hinds (1935, p. 50) as characteristic of this surface, later detailed studies by Sharp (1940, p. 1264) have indicated that weathering extended to a depth of at least 15 m (50 ft) below the surface. A verdict of "intense chemical weathering" was reached by Sharp on the basis of measurements of insoluble residue accumulation, evidence of iron-oxide residual concentrations, and progressive changes of certain minerals, especially biotite and feldspar. A further implication—that dominantly humid conditions prevailed throughout this region at the beginning of Paleozoic time—has likewise received considerable support from the results of these studies.

The clear, nearly continuous exposures in the walls of Grand Canyon of the great unconformity permit observation in detail of many features relative to the initial advance of the sea across the rocky Precambrian terrane and of the burial of this surface by marine sediment. As indicated by the relative age of various fossil assemblages, the sea advanced from the Cordilleran Geosyncline eastward across northern Arizona toward a landmass referred to as the Defiance positive element. This transgression began in the area of western Grand Canyon in Early Cambrian time as shown in figure 4 (note Olenellus-Antagmus horizon), but did not reach the eastern part of Grand Canyon until sometime in Middle Cambrian time (note Glossopleura-Alokistocare-Anoria horizon). Furthermore, the transgression was not accomplished as a single progressive and nearly uniform movement, but resulted from a series of marine advances caused by changes in relative sea level, each separated by a considerable period of nonadvance. Evidence supporting this concept of stagistep transgression is found in a series of measurements between certain key beds (time planes) and the surfaces of transgressive sandstones (fig. 4).
Figure 4.—Reconstruction of boundary between coarse sand and shale developed during transgression of the Cambrian sea. (A) Based on field data only. (B) Based on field data plus various theoretical considerations. (From McKee and Resser, 1945, fig. 12).
THE TONTO GROUP—TAPEATS SANDSTONE, BRIGHT ANGEL SHALE, AND MUAV LIMESTONE

Nomenclature

Among the first rocks in Grand Canyon to receive a formal name were those of the sandstone-mudstone-limestone sequence that forms the prominent Tonto Platform or bench in the lower canyon walls. They were called the Tonto Group by Gilbert (1874, 1875, p. 184). Nearly 40 years later these rocks were studied in the Shinumo quadrangle by Noble (1914, p. 61), who proposed but not without some misgivings because of gradational contacts, that they be divided on the basis of major lithologic stages into three formations. For these formations he adopted the names, in ascending order, Tapeats Sandstone, Bright Angel Shale, and Muav Limestone.

Subsequent to the detailed but localized studies of Noble, a stratigraphic investigation of the Tonto Group throughout Grand Canyon (McKee and Resser, 1945) has shown that these rocks grade vertically from one type into another, and involve major lateral changes through both lithologic gradation and intertonguing. Thus, even though the three formations can be recognized as rock units throughout the region, differences in age and facies are so considerable from place to place that a system of members (time-rock units) and tongues within the formations has been created in order to illustrate lateral relationships. This system, the units of which are separated largely by distinctive key beds or natural time planes, is illustrated in figures 5 and 6.

General description

The Tapeats Sandstone is a massive, cliff-forming unit with a thickness ranging from 30 to 91 m (100 to 300 ft) throughout the canyon area. In most places it is chocolate brown, but in some places it is gray or cream-colored and in others, a deep red brown. The sand is coarse to medium grained; coarse particles are dominant except in parts of the upper half, where medium-size grains are more common. Bedding is conspicuous because of contrasts in degree of cementation that cause layers to weather into alternating resistant ledges and shallow recesses. Flat, even beds up to a few inches thick are common, but by far the most abundant structural feature is crossbedding within layers ranging in thickness from 15 to 61 cm (1/2 to 2 ft). Most cross-strata are tabular planar or wedge planar but locally some are of trough type. Asymmetrical ripple marks, trilobite trails, and problematical worm borings are widely distributed and numerous at some localities. In many places the Tapeats grades upward into the Bright Angel through a zone in which coarse sandstone beds alternate with green shaly mudstones. This transition zone is arbitrarily assigned to the Tapeats Sandstone.

The Bright Angel Shale is a mixture of many lithologic types, mostly mudstones and fine-grained sandstones, but the dominant rock is a shaly, green mudstone. Some of this green rock is smooth, micaceous,
Figure 5.—Diagrammatic section of Cambrian deposits in Grand Canyon, showing stages in transgression and regression and distribution of facies from east to west. Time planes are horizontal; actual thickness ranges from 460 m (1,500 ft) in west to 245 m (800 ft) in east. (From McKee and Resser, 1945, fig. 1).
Figure 6.—Correlation chart of Cambrian formations and members in the Grand Canyon. (From McKee and Resser, 1945, fig. 2B).
and fissile; some is covered with fucoidal structures and splits along irregular, bumpy surfaces. Other rock types that occur at various horizons and localities are dark-magenta sandstones, gray platy siltstones with brown spots, pale-greenish-buff crumbly sandstone, and sandstones with concentrations of bright-green glauconite. Thus, in addition to the dominantly green color in the Bright Angel, magenta, yellowish to greenish buff, brown, and other colors are scattered throughout. Furthermore, the normal slope-forming topography of the formation is interrupted at many horizons with weak ledges or low cliffs formed by relatively resistant siltstones and sandstones. Fossils—especially trilobites, brachiopods, and Hyolithes—are locally common in the green shaly mudstones, and inarticulate brachiopods are numerous in some fine-grained sandstones. Trails of various animals are abundant at some horizons.

The Muav Limestone, uppermost formation of the Tonto Group, is composed chiefly of mottled gray limestone that forms a series of resistant cliffs. Although this limestone is thin bedded in eastern Grand Canyon, it is progressively thicker bedded and forms more massive cliffs toward the west. Between major cliff-forming units are parting beds that form benches or recesses, composed mostly of green, micaceous shaly mudstones, platy micaceous siltstones, platy, silty limestones and intraformational, flat-pebble conglomerates. Also included in the formation are facies of Girvanella limestone and of rusty-brown dolomite. Fossils are not common in the Muav, but numerous species of trilobites and some brachiopods are distributed throughout, especially in the parting beds. The entire rock sequence forms a gradational series with the underlying Bright Angel, and the lower boundary of the Muav is placed at the base of the lowest of the massive ledges of carbonate rock.

Key beds and time planes

Fortunately, from the standpoint of determining the nature of transgression and regression and also for establishing the time equivalency of facies within the Cambrian rocks of Grand Canyon, a number of distinctive and widespread marker beds or key beds occur within the Tonto Group (fig. 7). These key beds are very thin but persistent units that can be traced for distances of many miles and that are believed to represent approximate time planes. They are of three principal types: (1) zones of abundant but distinctive fossils with broad lateral extent but small vertical range, (2) thin but persistent units of intraformational, flat-pebble conglomerate, and (3) parting beds of platy siltstones or silty limestones that form ledges continuously exposed for great distances between major cliff units of massive limestone.
Figure 7.—Principal key beds and horizon markers considered to be approximate time planes in Grand Canyon Cambrian deposits:  
(a) Olenellus-Antagmus horizon.  (b) Tincanebits tongue, boundary beds.  
(c) Meriwitica tongue, boundary beds.  (d) Glossopleura-Alokistocare horizon.  
(e) Rampart Cave member, basal beds.  (f) Rampart Cave member top beds.  
(g) Sanup Plateau member, boundary beds.  (h) Spencer Canyon member, boundary beds.  
(i) Lower conglomerate beds.  (j) Peach Springs member, basal beds.  
(k) Kanab Canyon member, basal beds.  (l) Kanab Canyon member, middle beds.  
(m) Gateway Canyon member, basal beds and conglomerate.  
(n) Upper conglomerate bed.  (o) Solenopleurella horizon  
(p) Havasu member top beds.  (q) Top of Noble's "Muav C."  
(From McKee and Resser, 1945, fig. 3).
Transgression and regression

Cambrian strata of the Grand Canyon region are considered the result of a major transgression of the sea from west to east, interrupted by minor retreats or regressions, and terminated by a major regression. Some evidence for successive stages in transgression as furnished by the basal (Tapeats) sandstone has already been cited under the heading "The great unconformity and pre-Tonto environment." Additional evidence is found in the Muav Limestone where successively younger members extend farther eastward, except where temporary regressions result in short withdrawals of each facies. Thus, a series of gray limestone tongues projecting eastward into green shaly mudstone units of the Bright Angel, as shown in figure 8, was developed, presumably through alternate periods of rapid and slow basin sinking with accompanying sediment-filling to base level.

Distribution of facies

During times of transgression when the sea was advancing relatively rapidly toward the east and sediments were accumulating with little reworking, facies developed from the shore outward in a definite sand distinctive sequence. This sequence consists of conglomerate, coarse-grained sandstone, fine-grained sandstone, green shaly mudstone, glauconitic ferruginous beds, rusty-brown dolomite, Girvanella limestone and mottled aphanitic limestone. The same sequence was repeated in stairstep fashion at five different times as the sea advanced eastward, each step recording a stage in transgression (fig. 8).

Because regression represents a time of sea floor stability or of relatively slow sinking, when sedimentation fills available space up to base level and permanent deposits are few, a set of lithologic facies different from those of transgression was developed. These regressive deposits include silty, micaceous platy limestones, thin-bedded intraformational, edgewise conglomerates, and a distinctive type of glauconite bed as described by McKee (McKee and Resser, p. 55). Certain dolomite beds, probably formed in residual concentrated waters, likewise are characteristic of the regressive facies.

Faunal assemblages and zones

Extensive collections of fossils from the Tonto Group show that life was abundant and varied at many times and that it had wide distribution within favorable environments. The most common forms are trilobites and brachiopods, but sponges, cystids, gastropods, and Conchostraca are well represented. Most of the forms were described by Resser (in McKee and Resser, p. 185-220). They indicate that most species (unlike the genera) were very limited in range, with the result that well-defined zones are recognized. The most distinctive
Figure 8.—Reconstruction of stages in accumulation of Grand Canyon Cambrian deposits, based on interpretation of sedimentary record. (From McKee and Resser, 1945, fig. 11).
and widespread faunal zones in the Cambrian of Grand Canyon are, in ascending order, (1) *Olenellus-Antagmus*, (2) *Glossopleura-Alokistocare*, and (3) *Solenopleurella*. Other zones are more locally developed and some faunal assemblages seem to be strictly controlled by facies.

**Age of the Tonto Group**

Earlier attempts to date rocks of the Tonto Group consisted of correlations on the basis of lithology. These rocks were referred to the "Old Red" (Devonian) of England by Marcou (1856) and to the "Potsdam sandstone" (Cambrian) by Newberry (1861, p. 56). Later, they were considered Silurian by Gilbert (1874) and by Walcott (1880) and Carboniferous by Dutton (1882). Not until the middle 1880's, when its distinctive fauna had been collected and studied by Walcott and others, was the Cambrian age definitely determined. Even then, ideas concerning which part of the Cambrian was represented shifted at least 15 times during a few decades (McKee and Resser, 1945, table 1) until the currently accepted age range from Early Cambrian to the middle part of Middle Cambrian was determined.

**The Pre-Devonian Unconformity**

Erosion surfaces may be the time equivalents of great numbers of strata elsewhere in the geological column. Thus, between rocks of Cambrian and Devonian age in eastern Grand Canyon is an unconformity involving no recognizable discordance in stratification, yet marked by many irregularities or by erosion and representing a hiatus of considerable magnitude. The time involved in this hiatus probably spans all of Late Cambrian, Ordovician, Silurian, and Early to Middle Devonian time. In western Grand Canyon this time break was of somewhat shorter duration, for the Cambrian seas withdrew from east to west and, later, the transgressing Devonian deposits started there. However, the exact time represented by the break is not known. A thick sequence of dolomite beds above the Muav, described as "undifferentiated dolomites" (McKee and Resser, 1945, p. 77-79), may include Upper Cambrian and possibly Ordovician strata, both of which are recognized not far to the west in Nevada.

Irregularities of the surface that underlies Devonian strata in Grand Canyon consist mainly of channels of various sizes and shapes, cut into the resistant dolomite beds at the top of the Cambrian sequence. Many of the channels are relatively narrow and some are as much as 30 m (100 ft) deep. They were first noted in Kanab Canyon and at Temple Butte by Walcott (1880, p. 221; 1883, p. 438). Later the unconformity was studied in detail by Noble (1922, p. 49-51) between Garnet Canyon and Cottonwood Creek in eastern Grand Canyon. Channels of comparable dimensions have since been noted by the present writer below the thick Devonian sections in many parts of western Grand Canyon.
THE TEMPLE BUTTE LIMESTONE

Distribution

The Temple Butte Limestone of Devonian age is represented throughout eastern Grand Canyon by scattered remnants that fill channels and depressions in the eroded surface of the Muav Limestone. Nearly all of these remnant outcrops are less than 30 m (100 ft) thick and extend laterally for only a few hundred feet at the most. In the middle part of Grand Canyon the Temple Butte Limestone is several hundred feet thick and everywhere separates the Muav Limestone from the Redwall. Above the channel fills in that area, continuous beds of Devonian carbonate rock are relatively massive and evenly bedded. Farther west, the thickness of the formation becomes progressively greater and toward the lower end of Grand Canyon it attains a maximum of more than 300 m (1,000 ft) (McKee, 1937, p. 341).

As one follows the Colorado River through Marble Canyon, pockets of Temple Butte Limestone are first encountered near mile 38 and they are numerous along the canyon walls near water level for the next 16 km (10 mi). The type section is at Temple Butte on the west side of the Colorado River a few miles below its junction with the Little Colorado (Walcott, 1883, p. 438; Stoyanow, 1936, p. 503). Farther downstream, the Devonian rocks nearly everywhere are far above the river and most can be reached only with considerable difficulty.

Lithologic character

Devonian carbonate rocks that fill the channels eroded on the Muav surface are relatively easy to recognize in many places by their gnarly structure, saccharoidal texture, and purplish color. In places, the moderately thick, irregular beds are essentially horizontal but elsewhere they are roughly conformable to the concave profile of the channel; in still other places they are wavy or crenulated on a large scale.

In the more continuous strata of the Temple Butte, above the channel fills in western Grand Canyon, two principal types of rock compose the formation. One is a thin-bedded dolomite which has a smooth porcelainlike texture, but commonly weathers to a surface of even, ripplelike laminae that probably were formed by algae. This dolomite is tan to yellow or white, and it forms creamy white slopes or benches. The other type of dolomite forms resistant ledges of medium to thick beds. Mostly it is fine grained and steely gray, but commonly it weathers with a sandy texture and olive-gray surface (McKee, 1937, p. 341).
Fauna and age

Obscure molds and questionable imprints of fossils are present in many of the dolomites of the Temple Butte, but to date no clearly identifiable or diagnostic invertebrate fossils have been reported. This absence of fossils seems surprising in view of the excellent and abundant faunas that have been found in the Jerome Member of the Martin Formation, a probable correlative in central Arizona (Teichert, 1965).

The known fauna of the Grand Canyon Devonian consists of two discoveries—both fish. In 1879, "Placoganoid fishes" were discovered by Walcott (1880, p. 225) in the walls of Kanab Canyon near its junction with the Grand Canyon. In 1922, fish plates, identified as Bothriolepis, were discovered by Noble (1922, p. 51-52) in Sapphire Canyon. In a recent restudy of the Devonian fish from western United States, Denison (1951, p. 221, 230) concurred in the generic identification and the assignment of a Late Devonian age. The Temple Butte is currently assigned to the Frasnian stage or lower part of the Upper Devonian Series (McKee, in Poole and others, 1967, p. 887).

Environment of deposition

The Devonian rocks are the least studied and least understood of the Paleozoic strata in the Grand Canyon. The character and extent of the channels that the Temple Butte fills in the area between Ruby Canyon and Pipe Creek caused Noble (1922, p. 51) to suggest that "they look like cross sections of ancient stream valleys." The fish, Bothriolepis, from these deposits is recorded as "a characteristic element of Late Devonian fresh-water fish faunas throughout much of the world" (Denison, 1951). On the other hand, the widespread beds of pure dolomite that overlie the channel fills in western Grand Canyon probably represent a different environment. The common stromatolites or algal structures and other features indicate very shallow water; the dolomite suggests concentrated saline waters.

PRE-REDWALL UNCONFORMITY

Evidence that an unconformity occurs between rocks of Devonian age and those of Mississippian age in Grand Canyon is furnished both by the physical record and by faunal evidence of a hiatus. Nevertheless, the boundary between rocks of these two systems is in most places difficult to recognize. Nowhere has any angular discordance been recognized and in only a few places, mostly in western Grand Canyon, have conspicuous relief or local conglomerates been observed. Although no considerable uplift or appreciable amount of surface dissection took place during this interval, the faunal discontinuity shows a break in the rock record involving all of Kinderhook time in eastern Grand Canyon, much of that time farther west, and possibly the latter part of Late Devonian time throughout the region.
THE REDWALL LIMESTONE

Nomenclature

The Redwall Limestone is one of the most prominent formations in Grand Canyon forming in most areas a single sheer cliff 152 m (500 ft) or more in height. Because of the red color of its cliff face, stained by iron oxides washed down from the overlying Supai, this formation was given its name (Gilbert, 1875, p. 177). It characteristically develops giant overhangs and is cavernous in many places—features that at once distinguish it from other strata within the region.

Recent detailed studies of the Redwall (McKee and Gutschick, 1969) have demonstrated the advisability of dividing it into four members on the basis of well-defined differences in lithology. These distinctive units, which can be traced throughout the Grand Canyon and considerably beyond, are, in ascending order, the Whitmore Wash, Thunder Springs, Mooney Falls, and Horseshoe Mesa Members (fig. 9).

General description

The Whitmore Wash Member is a uniformly fine-grained dolomite in the eastern part of Grand Canyon, but farther west it is an even, fine-grained limestone. The member is thick bedded and nearly everywhere forms a resistant cliff about 30 m (100 ft) high, with a ledge at the top.

The Thunder Springs Member is distinctive and easy to recognize because it contains thin beds and elongate lenses of chert that alternate with layers of limestone or dolomite. The chert is white, opaque, and porcelainlike on fresh surfaces; it may remain white on weathering or it may take on a surface varnish of black or brown. The beds of carbonate rock between chert layers are mostly dolomite in eastern Grand Canyon, but limestone in the west. Like the chert, individual layers of carbonate rock are thin, commonly 2.5 to 5 cm (1 to 2 in) in some areas and 15 to 20 cm (6 to 8 in) in others.

The Mooney Falls Member is the thickest (51 to 107 m; 200 to 350 ft) and most massive part of the Redwall. Individual beds range from about 1 m (3 ft) to more than 6 m (20 ft) in thickness and even the very thick units have no apparent internal bedding planes or stratification surfaces. The rock is a pure limestone, except locally where it is dolomitic, and it is remarkably free of terrigenous material. Its texture ranges from aphanitic to very coarse grained, each size grade forming a zone that is repeated numerous times within rock units that externally appear massive. Large scale cross stratification of the tabular planar type occurs locally throughout the member.

The upper member of the Redwall, called the Horseshoe Mesa, is the thinnest, ranging from 11 to 38 m (35 to 125 ft) in the Grand Canyon. It is characteristically thin bedded and commonly weathers into a series of receding ledges in contrast to the underlying member.
Figure 9.—Section showing stratigraphic relations and thickness of the members of the Redwall Limestone along the Bright Angel trail, Grand Canyon. (From McKee, 1963, p. C22).
which forms massive cliffs. The Horseshoe Mesa Member is mostly aphanitic limestone, but it includes some fine-grained and oolitic limestones. In many localities a zone of thin-bedded chert lenses, like those in the Thunder Springs Member, occurs near the base.

**Lithology**

The Redwall Limestone is composed of numerous varieties of carbonate rock, but, except in a few places, it is remarkably free from insoluble residues and noncarbonate rocks other than bedded chert. Limestones of the Redwall, as classified by McKee and Gutschick (1969), are the following:

I. Granular limestone
   A. Peloidal limestone
   B. Skeletal limestone (exclusive of reefs)
      1. Bioclastic limestone
      2. Limestone dominantly of unbroken fossils
   C. Oolitic limestone
   D. Limestone conglomerate
II. Aphanitic limestone
III. Recrystallized limestone

Granular limestone, which is the dominant type, includes extensive beds of peloids (clastic carbonate grains of various sizes) and of fossil fragments, especially crinoidal; in some places, it contains abundant coals. These three varieties occur both as relatively pure types and as mixtures in different combinations. Also very common are the microcrystalline and cryptocrystalline limestones, usually referred to as aphanitic limestones, that probably formed from lime muds.

Many, but not all, of the dolomite beds in the Redwall probably formed as sedimentary dolomites. Their paleogeographic distribution in eastern or near-shore facies, their uniform, fine-grained texture, and other features suggest an early diagenetic origin. In contrast, some of the coarse-grained dolomites show spatial relationships to zones of faulting and are considered of secondary origin.

Chert beds and elongate lenses that are abundant at certain horizons in the Redwall are shown by the electron microscope to be composed of both microcrystalline quartz and chalcedonic quartz. Considerable evidence has been assembled (McKee and Gutschick, 1969) to show that this chert is an early replacement of carbonate mud on the sea floor. Delicate details of bryozoan, crinoid, and other fossil skeletons are preserved as molds in much of this chert.

**Transgression and regression**

Lithologic evidence indicates that the sea transgressed and regressed across the Grand Canyon region three times during Mississipian time. Evidence of the first and second transgressions is furnished by the thickness of carbonate beds in the Whitmore Wash
and Mooney Falls Members, respectively. The thick beds in these members must have formed when the sea floor was sinking relatively rapidly and base level allowed maximum accumulation of sediment as a result. Thin beds of the succeeding Thunder Springs and Horseshoe Mesa members suggest slow sinking of the sea floor and less sedimentation. Fossil evidence indicates that the first transgression began in late Kinderhook time in western Grand Canyon. The second transgression was in Osage time, followed by regression in early Meramec time. Evidence of a third transgression consists of isolated remnants of Chester-age rock preserved at the top of the Redwall along the Bright Angel Trail. Apparently the sea covered this region for the third time during the Late Mississippian, but most of the record was removed by pre-Supai erosion.

Cycles in sedimentation

A cyclic arrangement involving layers of carbonate rock of different grain sizes has been demonstrated to be characteristic of the upper two members of the Redwall throughout Grand Canyon. With limestone of aphanitic texture at one extreme, coarse-grained limestone at the other, and intermediate sizes between, a series of rock units each a few feet thick forms a sequence or cyclothem (fig. 10) that is repeated numerous times in each stratigraphic section of the members involved (McKee, 1960, p. 231). Although it is not believed that grain size of clastic carbonate particles bears any relation to transportation in terms of depth or distance from shore, available data suggest that the differences in grain size may be a function of systematic changes in water depth that effect differences in wave base.

Faunal zones and assemblages

The fauna of the Redwall is large and varied. The most common forms are foraminifera, brachiopods, corals, and crinoids; also common are bryozoans, gastropods, pelecypods, cephalopods, and blastoids. Fish, ostracodes, trilobites, holothurians, and algae occur in some parts of the formation. The larger fossil groups occur in distinctive associations, the most important of which are the coral-brachiopod-crinoid, coral-foraminifer-brachiopod, and brachiopod-bryozoan. The distribution of these associations apparently was controlled by environmental factors as indicated by the different facies of carbonate sediments.

Many of the Redwall faunas either are local in distribution or have long vertical ranges, and therefore have little stratigraphic value; however, the foraminifera, brachiopods, and corals are useful zone indicators. An almost continuous faunal succession through the formation, including six zones and four subzones, is represented by the foraminifers (Skipp, in McKee and Gutschick, 1969, chap. V). Among corals, the distinctive, widespread zones of Dorlodotia inconstans

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Figure 10.—Curves illustrating cycles in grain size of carbonate rocks forming upper part of Redwall Limestone at eight localities in Grand Canyon, Arizona. Numbers 1 through 5 beside each curve represent stages characterized by aphanitic or very fine grained texture. Numbers at top refer to positions on locality map. Dolomite which replaced limestone and thus locally obscures the original texture is shown by cross hatching. (From McKee, 1960, p. 232).
and Michelinia expansa are useful markers in many sections (Sando, in McKee and Gutschick, 1969, chap. VI). Many of the brachiopods may likewise be good zone indicators, but unfortunately have not yet been studied in enough detail.

THE PRE-SUPAI EROSION SURFACE

The contact between the Redwall Limestone and the Supai Formation as seen in the walls of Grand Canyon appears to be a flat, even surface between horizontal strata. Detailed examination shows that this is far from the fact. An unconformity, which represents a considerable amount of time and which formed karst topography, is represented (fig. 11).

Low but persistent relief, ranging from a meter or so to 12 m (40 ft), marks the upper surface of the Redwall. Channels and bowl-shape depressions penetrate its resistant upper-most beds and between these low areas are small mesas, flat-topped ridges and receding ledges. Some of the channels are partly filled with well-rounded gravels of jasper, chert, and other durable rocks that probably were transported from afar by streams. Other channels seem to be the result of solution processes for they contain large angular fragments of limestone, locally derived in a matrix of gnarly-bedded or structureless mudstone. Such products of weathering probably are residual materials accumulated in solution channels or sink holes. Abundant caverns in the upper parts of the Redwall may also be products of the karst development.

The magnitude of the hiatus represented by the pre-Supai unconformity can be estimated in a general way by faunal dating. In most places the rocks underlying the erosion surface are of early Meramec age, or possibly older in eastern Grand Canyon. Those overlying the break are of Morrow age. Thus, Mississippian rocks representing the upper Meramec and the Chester Series and those of Pennsylvanian age representing Springer time are absent. Although remnants of Chester-age rocks are known from near Bright Angel Trail and probably once covered the region, it seems probable that most of the pre-Supai erosion occurred during Early Pennsylvanian time and that latest Mississippian (Chester) rocks were largely removed then.

RED BEDS—THE SUPAI FORMATION AND HERMIT SHALE

Nomenclature

Red beds of Pennsylvanian and Permian age that give most of the bright color to the upper half of Grand Canyon's walls were originally referred to as the lower part of the Aubrey Group (Walcott, 1880). This sequence of beds was later named the Supai Formation (Darton, 1910, p. 25-27). Still later, it was subdivided and the upper 91 m (300 ft) of red shaly siltstone assigned to a new formation name—the Hermit Shale; the lower 183 to 213 m (600 to 700 ft) was redefined but retained the name of Supai (Noble, 1922, p. 59).
Figure 11.—Contact between Supai Formation and Redwall Limestone at various localities in Grand Canyon. (a-b) Hermit Trail. (c-d) Hance Trail. (e) Tanner Trail. (From McKee and Gutschick, 1969, fig. 31).
General description

In eastern Grand Canyon the Supai-Hermit red bed sequence consists, in general, of alternating deep-red shaly siltstone slopes separated by three major cliffs of fine-grained buff sandstone, mostly stained bright red on weathered surfaces. Along the Colorado River the Hermit Shale is first encountered about 8 km (5 mi) below Lees Ferry; the top of the Supai is exposed a few kilometers below near the junction of Soap Creek and Marble Canyons, and the Supai Formation composing the lower walls of Marble Canyon for the next 18 km (11 mi) downstream. In contrast, within Grand Canyon the Supai-Hermit strata are everywhere high above river level, but in some parts where they form the upper walls of the inner canyon, such as near Toroweap Valley, details can plainly be seen from below.

In western Grand Canyon the lithologic and topographic character of the Supai is very different from that farther east. Although total thickness is not much different, the percentage of resistant rock increases westward with the result that cliff units are more prominent there and ledges more narrow. Furthermore, limestone and dolomite form appreciable parts of the sequence as a consequence of a general westward gradation into limestones and of intertonguing of limestone and dolomite from the west.

A study of the Supai Formation, begun in 1958 by the author, indicates that this unit is readily divisible into four vertical subdivisions on the bases of contained fossils and differences in lithologic character. Further, each part is separated from subjacent and superjacent rocks by an unconformity marked by a surface of erosion, channeled to depths of as much as 14 m (45 ft), and by one or more layers of conglomerate that serve as key beds throughout the region.

Marine fossils—brachiopods, corals, crinoids, gastropods, pelecypods, and others—are common along the western margins of the Supai and are present eastward in the lower part as far as Havasu Canyon. Fusulinids are found in calcareous sandstones of the middle two units virtually throughout Grand Canyon. Land plants and reptile tracks occur at various horizons and at many places within the region.

Detailed studies of the many lithofacies of the Supai Formation are currently being completed and furnish much information on the environments of deposition. Statistical studies of cross-bedding dip directions and of grain-size distribution indicate a sand source to the north and sediment transport in a southerly direction.

The Hermit Shale, unlike the Supai, greatly increases in thickness from east to west, ranging from several tens of meters to more than 300 m (1,000 ft) in the length of Grand Canyon. The Hermit is mainly composed of nonresistant, shaly siltstone, and characteristically forms a bench or shelf. In western Grand Canyon, where this shelf is very wide and flat, it is referred to as the Esplanade.
The hiatus problem

As with many red bed sequences, the dating and the subdivision of the Supai-Hermit strata are made difficult by a general scarcity of fossils. The only diagnostic fossils within these rocks in eastern Grand Canyon are seed ferns and associated plants in the Hermit Shale. On the basis of this flora, the Hermit has been assigned an "upper Lower Permian" age (White, 1929, p. 38).

Marine fossils, mostly brachiopods and foraminifera, occur in thin limestone beds of the lower slope unit in the Supai, from western Grand Canyon east as far as Havasu Canyon. These fossils indicate an Early Pennsylvanian (Morrow) age for the basal strata of the Supai. Between the Hermit flora and the basal Supai fauna is a considerable thickness of red beds, representing a long span of time, from which no datable fossils have been found in eastern or central Grand Canyon.

In extreme western Grand Canyon, limestone tongues and lenses interbedded with Supai red beds have yielded marine fossils at numerous horizons in recent years. These fossils have shown that the rock sequence contains strata not only of Permian (Wolfcamp) and Pennsylvanian (Morrow) age, but that between these occur beds of Virgil and of Des Moines age. Presumably, all of these units are represented farther east by nonfossiliferous red beds of corresponding age.

A distinctive conglomerate that contains rounded pebbles of gray limestone and red siltstone and attains a thickness of as much as 14 m (45 ft) occurs throughout Grand Canyon at the base of the Wolfcamp rocks. It indicates the presence of an unconformity and marks a clearly visible break in the record. A somewhat similar conglomerate occurs at various places in western Grand Canyon, slightly above the highest Morrow-age fossils, and probably indicates an unconformity between strata of Morrow and Des Moines ages. This erosion surface may mark a hiatus involving Atoka time. Much more obscure, but probably also present, is another hiatus, between strata of Des Moines and Virgil age. To date, neither unquestioned physical evidence nor faunal record has been discovered to demonstrate this inferred hiatus, yet its presence seems probable from the faunal sequence to the west.

Current direction trends

Recent sedimentological studies of the Pennsylvanian-Permian red bed sequence, conducted in both field and laboratory, include grain-size analyses, insoluble residue determinations, calcium-magnesium studies, and statistical measurements of crossbedding types and directional trends. These investigations were made for each of the four principal time units within the sequence, and results have been plotted on base maps to show regional distributions.

The most conspicuous primary structural feature in the Supai and Hermit formations is cross-stratification. It is represented by large-scale, tabular planar crossbeds; by medium-scale, wedge planar crossbeds; and by trough-type structures, mostly with shallow festoon
forms. Cross strata are present both in sandstones and siltstones and also are in sandy limestones. Current-vector maps based on average directions of cross-strata dips show a dominant southerly direction of sediment transport and strongly suggest a nearby source in southern Utah.

Corroboration of current directions as determined from the dips of foreset beds is furnished by regional trends in grain size distribution and by lithofacies patterns. Texture analyses show that coarse and very coarse grains are chiefly in sandstones to the north, whereas maximum grain sizes to the south are finer. Likewise, analyses show that the average grain sizes of northern sandstones are coarser than those from the south. Finally, a great increase of carbonate rock to the west in Grand Canyon and also southeast from Grand Canyon suggests that extensive marine environments were in those directions.

THE COCONINO SANDSTONE

General description

The Coconino Sandstone is a deposit of clean, well-sorted quartz sand that forms a great sedimentary wedge across much of northern Arizona. The formation thins progressively northward and westward from a maximum thickness greater than 152 m (500 ft) along the Mogollon Rim in central Arizona to a narrow tongue that wedges out near the Arizona-Utah boundary. The Coconino is 17 m (57 ft) thick in the walls of Badger Canyon 8 km (5 mi) below the upper end of Marble Canyon, 29 m (96 ft) thick at Toroweap Point in central Grand Canyon, and 20 m (65 ft) at Grand Wash Canyon near the west end of Grand Canyon. In the Bright Angel Creek area it is about 91 m (300 ft) thick.

The most distinctive structure in the Coconino is the large-scale, wedge planar cross-stratification that is prominently displayed in the white cliff faces of this sandstone throughout the region (McKee, 1933a). The inclined laminae, having dips of as much as 34 degrees, have gently curving surfaces that in places are 18 to 21 m (60 to 70 ft) long. The beveled upper edges of individual sets are formed by low-angle erosion surfaces that constitute the bases of higher sets of cross strata. Other structures typical of the Coconino are long, parallel ripple marks, with rounded crests and oriented with axes parallel to the dip slopes.

Environment of deposition

The Coconino Sandstone is generally interpreted as an eolian deposit. Analysis of the cross-stratification suggests that it was mostly deposited as transverse-type dunes, although locally the barchan type can also be recognized. Essentially all of the high-angle foreset beds dip in a southerly direction, indicating that the wind transported sand from the north (Reiche, 1938). In a few places low-angle beds dip northward; these beds are interpreted as windward-slope deposits such as occur on recent dunes where the load exceeds
the transporting power of the wind. Ripple marks, whose shape and orientation are typical of those formed on the steep lee sides of modern dunes, probably formed where crosswinds were active.

**Fauna of the Coconino**

Well-preserved tracks and trails of both vertebrate and invertebrate animals are common on the steep foreset beds in various parts of the Coconino and constitute the only fossils in this formation. The tracks of vertebrates, believed to have been reptiles, include a wide variety of forms. Many of the animals had the size and proportions of small lizards, but others were wide-bodied with short limbs, and still others walked with long strides and had large feet. More than 20 types of track have been described as distinct species by Gilmore (1926), who also noted the burrows of wormlike creatures and trails probably of insects. These footprints and trails have been found nearly everywhere that the Coconino occurs, but as yet no skeletal remains have been discovered.

**THE TOROWEAP AND KAIBAB FORMATIONS**

**General description**

The Toroweap and Kaibab, which are the two uppermost formations in the walls of Grand and Marble Canyons, are composed in large part of carbonate sediment deposited during two transgressions of the sea during the middle part of Permian time. In both formations three subdivisions or members can be recognized; these members have been referred to as (1) (Gamma) the time of transgressing sea, (2) (Beta) the time of maximum advance, and (3) (Alpha) the time of regressing sea. The formations are separated by an unconformity, apparently marking a brief period after the Toroweap sea had withdrawn to the west and before the Kaibab sea began to readvance across the region.

The middle (or Beta) member of both formations is composed dominantly of carbonate rock and each forms a major cliff in the walls of Grand Canyon. In most other respects, these rocks are different. The Toroweap unit is massive limestone in western Grand Canyon, more prominent than the Kaibab Cliff in that area, but it is progressively thinner and more magnesian eastward, and is relatively inconspicuous at the east end of the canyon. The middle member of the Kaibab, in contrast, maintains a relatively great thickness throughout the area but changes in composition from limestone to sandy limestone to sandstone from west to east.

The basal (or Gamma) members of both Toroweap and Kaibab are relatively thin, weak, slope-forming units, and show evidence of the reworking of underlying sediments. In contrast, the upper (or Alpha) members are relatively thick; they are formed of red beds, thin residual limestones and, locally, beds of gypsum. These are regressive deposits formed as the seas withdrew westward; because the strata are easily eroded, they form slopes of receding ledges above the massive cliffs of carbonate rock that form the middle (or Beta) members.
Facies distribution

Both the Toroweap Formation and the Kaibab Limestone consist of a number of distinct facies, lithologic and faunal, that result from contemporaneous but contrasting environments. Further, a different set of facies is recognized within each member of these formations because these members are records of particular stages in the advance or retreat of the sea. Some facies are marine, some continental, and others represent various intermediate coastal environments.

An excellent example of contrasts in facies is illustrated by the middle (or Beta) member of the Kaibab (fig. 12). In the westward two-thirds of Grand Canyon this rock consists of relatively pure clastic limestone with abundant chert concretions and a fauna of brachiopods, corals, and bryozoans. In the Bright Angel section of eastern Grand Canyon the concretionary clastic limestone grades into sandy limestone containing beds of chert but few fossils. In extreme eastern Grand Canyon the same member is represented by a noncalcareous sandstone, with no chert but containing a molluscan fauna. Fossils in this facies are totally different from those in the limestone to the west. Still other facies, composed of magnesian limestones and with still different faunal assemblages, occur both north and south of the sandstone facies in eastern Grand Canyon.

Combinations of facies, comparable with but different from those in the Kaibab middle (or Beta) member, occur in other members of the Kaibab and also in the Toroweap. Generalized patterns of distribution have been plotted for these facies (McKee, 1938), but much still remains to be learned about their significance. In addition to those facies considered to represent "normal marine conditions" with their brachiopod assemblages, and others composed of highly magnesian carbonate rock and containing faunas of pelecypods and gastropods, facies consisting of red beds and gypsum are widespread in the regressive deposits of both formations and probably represent lagoonal conditions.

Cyclic deposition

Several types of cyclic sedimentation are represented in various parts of the Toroweap and Kaibab formations (McKee, 1964, p. 283). These cycles result from repetition or alternation of environments of deposition, with certain salts being concentrated by natural processes and deposited periodically, and detrital material of noncyclic nature continuing accumulation during times of no chemical precipitation.

In the upper member of both the Toroweap and Kaibab a sequence consisting of, from bottom to top, red beds, aphanitic limestone, and gypsum is repeated numerous times. Each unit in these sequences is from .3 to 2.5 m (1 to 8 ft) thick; the red beds are, in general, thicker than the chemical deposits. The detritus probably was transported by streams from the east and deposited in shallow lagoons or playas; the associated calcium carbonate and calcium sulfate were accumulated as precipitates from concentrated saline waters.
Figure 12.—Generalized map showing facies distribution in middle member of Kaibab Limestone, Arizona and Utah. (From McKee, 1964, fig. 3).
In the middle (or Beta) member of the Kaibab Limestone another type of cycle is represented in the chemical sediments. In a facies of sandy limestone between areas of pure calcium carbonate and of noncalcareous sandstone occur beds of earthy chert. Each lithologic cycle, as typically displayed along the Bright Angel Trail, is a few feet thick and consists, from bottom up, of sandy limestone with few or no fossils, sandy limestone with abundant fossils, and earthy bedded chert. The interpretation suggested to explain these cyclic deposits is that the silica now preserved in the chert beds was introduced into the sea by rivers from the east, that the waters containing concentrated silica, both during chert precipitation and for a time following, was unfavorable for marine life, but that gradually as normal sea environment prevailed, a marine fauna moved in from the open sea to the west.

The Permian fauna

The composition of the Toroweap and, especially, the Kaibab faunas has been of interest to geologists for many years; first, because certain elements in it are of wide distribution and therefore are useful in correlation with Permian rocks in many other regions and second, because some of the fossils that are clearly time equivalents can be shown to be strictly facies controlled. The fauna includes a large variety of forms, but many of these are nowhere found associated with certain others; some are strictly marine, others are probably brackish-water types, and still others are typical of supersaline conditions. The following list (McKee and Breed, 1969) gives the number of genera thus far reported for each of the main fossil groups.

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THE TOROWEAP FORMATION: A NEW LOOK

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ABSTRACT

In northern Arizona, the Toroweap Formation of Permian age is divided into three members which record a major west to east transgression, followed by a regression. The basal member, the Seligman, represents supratidal and continental environments at the land margin of the advancing Toroweap sea. Sabkha deposition gave way to marine carbonate deposition represented by the Brady Canyon member. The Woods Ranch member represents a regression of the Toroweap sea, as indicated by a return to supratidal and continental deposition.

Facies in the Brady Canyon member include: grain-supported skeletal limestone; mud-supported skeletal micrite; pelleted micrite and dolomicrite; sandy dolomite; aphanitic limestone and dolomite. Three minor transgressions and regressions are recorded by shifts in the carbonate facies. No facies in the Brady Canyon member is micrite-free, suggesting that low-energy conditions prevailed throughout Toroweap deposition in northern Arizona, the higher energy environments being farther west.

Evidence for supratidal facies within the Seligman and Woods Ranch members includes: the presence of thick evaporite deposits; algal stromatolites; dolomitization of carbonate units; and the scarcity of fossils. Evidence for a continental sabkha/dune facies within the Seligman and Woods Ranch members includes: horizontally bedded sandstones; the presence of solution collapse breccias, contorted bedding, and minor diapiric structures suggesting the former presence of gypsum; bimodal sandstones and large-scale crossbeds indicating an eolian environment.

Continental sabkha and dune deposition continued to the south and east of marine deposition during Toroweap time. The entire Toroweap Formation, therefore, is a good example of marine carbonate-sabkha-eolian deposition.

INTRODUCTION

In a summary paper on the Toroweap Formation, McKee and Breed (1969, p. 12) made the following statement: "Despite major advances in the study of carbonate rocks since 1938, no significant contributions have been made in the sedimentology of the Toroweap and Kaibab." This paper relates recent concepts of carbonate deposition to the interpretation of Toroweap sedimentation and applies new concepts of sabkha (salt flat) deposition to the understanding of the gypsum-sandstone facies of the Toroweap. Twenty-seven sections were measured, described, and sampled for this study. Thin sections were analyzed for texture and composition, and the data were integrated with recent carbonate and evaporite concepts which results in a new interpretation of the Toroweap facies.
The Permian Toroweap Formation in Arizona covers an area of approximately 65,000 km² (25,000 sq mi) (pl. 1a). The formation pinches out to the east, is eroded to the south and southwest, and extends westward into Nevada and northward into Utah (fig. 1). It overlies the Coconino Sandstone and is overlain by the Kaibab Limestone.

ACKNOWLEDGMENTS

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The junior author thanks George Billingsley for suggesting Marble Canyon as a research area; Robert Packard, Matthew Turner, Randi Martinsen, Norm Lewis, Chuck Warner, Dale Gray, and Don Marshall for assistance in the field; Harvey Butchart for sharing his knowledge of access to various side canyons in Marble Canyon; Canyoneers, Inc. for a river trip down the Colorado River; Edwin D. McKee (U.S. Geological Survey) for arranging to have the fossil material identified; and Ellis Yochelson (U.S. Geological Survey) and John Pojeta (U.S. Geological Survey) for identifying the fossil material.

PREVIOUS WORK

The Toroweap Formation was named and defined by McKee (1938), who subdivided it from the Kaibab Limestone. The type section that he chose is in Brady Canyon, an eastern side canyon to Toroweap (Tuweep) Valley. McKee recognized three members: an upper gypsum and red-bed member, informally named the alpha member; a middle limestone, informally named the beta member; and a lower sandstone and gypsum member, informally named the gamma member.

Sorauf (1962) suggested names for McKee's informal members—the alpha to be named the Woods Ranch member; the beta, the Brady Canyon member; and the gamma, the Seligman member. Although these names have not been formally adopted according to the criteria of the Code of Stratigraphic Nomenclature, their usefulness has been recognized (Bissell, 1969, p. 145) and they are used in this paper on an informal basis.

McKee (1938) recognized the transgressive-regressive nature of the Toroweap and defined several facies based on the presence or absence of certain fossils. Several theses have been written on the Toroweap since McKee's original work: Belden (1954), Aubrey Cliffs area; Schleh (1966), Parashant Canyon; Mullens (1967), Ashfork area; Pfirman (1968), Grindstone Canyon-Sycamore Canyon area; Evans (1971), Sycamore Canyon-Oak Creek area; and Benfer (1971), Oak Creek-Walnut Canyon area. Kirkland (1962) compiled a list of Toroweap fauna. Two dissertations
Figure 1.—Index map.
on Paleozoic stratigraphy were completed by Fisher (1961) and Sorauf (1962) in Parashant-Andrus Canyon, and Whitmore Wash area, respectively. Bissell (1969) related the Toroweap basin facies in Nevada to the Grand Canyon shelf facies in Arizona.

STRATIGRAPHY

Seligman member

The type locality of the Seligman member is in the Aubrey Cliffs, northwest of Seligman (fig. 1); the lithology is red calcareous sandstone. Gypsum beds occur in Hack Canyon, Toroweap Valley, and westward into Nevada. The lower boundary of the Seligman member is defined, throughout the Grand Canyon and to the south, as the first horizontal sandstone unit which truncates the uppermost crossbeds of the Coconino Sandstone (pl. lc). The boundary is conformable in most locations but is not distinct in the area of Marble Canyon where crossbedded units of the Coconino commonly intertongue with the Seligman member. Fisher (1961) had difficulty locating the basal contact of the Toroweap on the Shivwits Plateau (fig. 2) and suggests that the Coconino in that area is marine. The upper contact of the Seligman member is defined by the first appearance of a vertically continuous carbonate unit above the horizontal sandstones. The member is less than 15 m (50 ft) thick in the grand Canyon region but is reported by Bissell (1969) to be 152 m (500 ft) thick in the North Muddy Mountains in Nevada.

Brady Canyon member

The type section for the Brady Canyon member is in Brady Canyon, on the east side of Toroweap Valley (fig. 1). In Arizona the member has its greatest development to the northwest, where McNair (1951) measured 93 m (305 ft) at Pakoon Ridge (fig. 1). The member thins uniformly in an easterly direction except in Marble Canyon, where the rate of thinning increases just before reaching its depositional edge (fig. 3). The base of the Brady Canyon member is placed where the first carbonate appears above the horizontally bedded sandstone or bedded gypsum of the Seligman member. The basal carbonate is generally an aphanitic dolomite in the west and gradually changes to a sandy dolomite towards the east, grading into a dolomitic sandstone, at which point the Brady Canyon member can no longer be recognized. The top of the member is placed at the last appearance of carbonate below the Woods Ranch member. It is generally an aphanitic limestone or dolomite, with the quartz sand content increasing towards the east.

In the west, the Brady Canyon member is highly fossiliferous, it becomes less so in an easterly direction. In the western fossiliferous section, the fossils consist of brachiopods, bryozoans, crinoids, and endothyrids (pl. 2h). In the Marble Canyon area, only a restricted fauna of gastropods and bivalves is found. Stromatolites occur at the
Figure 2.—Panel diagram, Toroweap Formation, northern Arizona.
Figure 3.—Isopach map, Brady Canyon member.
top of the Brady Canyon member at the Bass Trail section (fig. 1). Chert nodules and layers are more abundant in the west. Limestone predominates in the west, and dolomite is abundant in the eastern area. The quartz content of the dolomite increases towards the east.

Bedding thickness ranges from 5 cm (2 in) to more than 7 m (20 ft). The bedding is generally horizontal, with some wavy bedding restricted to the more thinly bedded units. Primary sedimentary structures are rare. The upper surface of the Brady Canyon member is marked by mudcracks in Pigeon Canyon and Kane Canyon (fig. 1).

A Leonard age for the Toroweap has been determined by fossils contained in the Brady Canyon member (King, 1930; McKee, 1938; McKee and Breed, 1969).

Woods Ranch member

The suggested type section of the Woods Ranch member (Sorauf, 1962) is located in Whitmore Wash (fig. 1). The lithologies present in this section are thick evaporites, red and white sandstones, and thinly bedded carbonate units. The lower boundary of the member is chosen at the top of the last appearance of aphanitic dolomite or limestone. A change from cliff to slope between the resistant Brady Canyon member and the nonresistant Woods Ranch member helps locate the contact in the field. The upper contact of the member is chosen at the first appearance of a massive cherty limestone in the western Grand Canyon region, and at the first appearance of a massive sandy limestone to the east and to the south. The contact with the Kaibab is disconformable but is commonly difficult to locate precisely because talus slopes form where the nonresistant gypsum and sandstone units have been eroded from beneath the resistant Kaibab Limestone.

The Woods Ranch member shows no consistent thickening or thinning trends throughout most of the study area. It is more than 61 m (200 ft) thick at some locations. The massive gypsum is restricted almost entirely to the area north and west of the Colorado River (fig. 4). The gypsum interfingers with and grades into sandstone towards the north, east, and south. In Marble Canyon this change in facies is abrupt and closely paralleled by a geomorphic change from slope to cliff (pl. ld). In Pigeon Canyon to the west, the member consists mostly of sandstone, but west of this area gypsum again becomes the dominant lithology.

The Woods Ranch member is unfossiliferous, with the exception of the Schizodus zone, a thin but persistent limestone unit about 7 m (20 ft) below the Kaibab-Toroweap contact. The sandstones in the Woods Ranch member are horizontally bedded, except for a few large-scale crossbeds in the eastern and southern part of the study area. The gypsum units are massive and commonly have contorted structures. Interbedded with the sandstones and gypsum are thin-bedded carbonate units. Intraformational breccias also are common in the Woods Ranch member and are generally restricted to the sandstone units. Contorted bedding and what appear to be small diapiric structures (pl. 1e) are common and often associated with the breccias.
Figure 4.—Massive gypsum, Woods Ranch member.
Transition and eastern phases

McKee (1938) divided the Toroweap into three lateral phases: the western phase, where the three members are recognizable; the transition phase, which consists of horizontally bedded red and light-colored sandstones; and the eastern phase, which consists entirely of crossbedded white sandstones. The limits of the western phase are determined by the extent of the Brady Canyon member; beyond the depositional edge of the carbonates, the two sandstone members (Seligman and Woods Ranch members) are indistinguishable. The area dominated by horizontally bedded red sandstones is known as the transition phase. The transition phase is particularly well developed in Sycamore Canyon. The sandstones become lighter in color and more extensively crossbedded to the east and south. Within Oak Creek and Walnut Canyons (fig. 1) the eastern phase Toroweap is similar to the Coconino Sandstone in lithology and crossbedding.

PALEONTOLOGY

McKee (1938) summarized the paleontology of the Toroweap Formation in his original work on the Kaibab and Toroweap Formations. The Brady Canyon member is the only fossiliferous member, with the exception of the Schizodus zone in the Woods Ranch member. McKee recognized two major faunal facies: an open marine fauna to the west, and a molluscan fauna to the east. The open marine fauna includes brachiopods, bryozoans, crinoids, and horn corals. The molluscan fauna includes bivalves and gastropods, with a few scattered scaphopods and cephalopods. Kirkland (1962) compiled a faunal listing of subsequent finds in the Toroweap Formation. Belden (1954), Mullens (1967), Miller and Breed (1964), and Beus and Breed (1968) found additional species in the Toroweap.

The authors found endothyrids in several sections, as well as a large bivalve, Posidonia, in the Brady Canyon member. To our knowledge, neither of these had been previously reported from the Toroweap Formation.

PETROLOGY

Seligman member

The Seligman member consists of gypsum and fine-grained red calcareous sandstones, with medium to coarse quartz and feldspar grains. McKee (1938) noted that the Coconino Sandstone and the Seligman member are petrologically similar, a result of the reworking of the unconsolidated Coconino by the advancing Toroweap sea. The sandstones are mostly quartz arenites with a few fresh, rounded feldspar grains (pl. 3a). The Coconino also contains quartz and scattered feldspar grains (McKee, 1933).
Sedimentary structures are rare in the Seligman member. The sandstones are horizontally bedded except in the eastern region, where the Coconino intertongues with the Toroweap. Belden (1954) found channels in the Seligman member in the Aubrey Cliffs area. These channels have current-formed ripple marks that indicate a current direction towards the northwest. McKee (1938) reports an intraformational breccia in the Seligman member in Grand Wash Canyon (fig. 1) near Lake Mead.

Brady Canyon member

Five petrographic textural types are recognized in the Brady Canyon member: 1) grain-supported skeletal limestone, 2) mud-supported skeletal micrite, 3) pelleted micrite and dolomicrite, 4) sandy dolomite, and 5) aphanitic limestone and dolomite.

Grain-supported skeletal limestone

Grain-supported skeletal limestone (pls. 2c, 2d) occurs in an area west of a north-south line through the Thunder River section. The matrix is micrite, which is common throughout the Toroweap in northern Arizona. No micrite-free limestones were observed in any of the sections sampled. The skeletal material consists of disarticulated crinoid columnals, bryozoans, brachiopods, ostracodes, gastropods, endothyrids, echinoid and brachiopod spines, and trilobite fragments. Occasional floating quartz grains occur in this texture. The quartz is fine silt size and increases in abundance from about 1 percent at Pigeon Canyon, to 5 percent at Toroweap Valley, and as much as 25 percent at the Thunder River section. The quartz is angular and well sorted.

Mud-supported skeletal micrite

The skeletal components of this texture (pls. 2g, 2h) are similar to the grain-supported limestone, except that they are floating in micrite and are smaller, perhaps as a result of transportation by weak bottom currents. The mud matrix makes up 60-90 percent of the rock. In general, the massive limestones are composed of skeletal micrite, whereas the grain-supported skeletal limestones occur in the thinner bedded receding cliff units (pl. 1b).

Pelleted micrite and dolomicrite

This texture appears to be composed of pellets (pls. 2b, 2e) because of the shape and sorting. The pellets are elliptical to spherical and range from 0.1 to 0.2 mm in diameter. No true oolites were observed in sections sampled, although a few of the recrystallized peloids in the Ashfork section may have had an oolite origin. McNair (1951) reported oolites in the Whitmore Wash area in the lower limestones of the Brady Canyon member; however, Sorauf (1962), who measured many sections in Whitmore Wash, did not report any oolites.
The pelleted micrite facies is commonly dolomitized to a dolomicroite and is included in this facies when the original texture can be recognized. The dolomite texture is prevalent in the eastern part of the study area and also appears in some western sections, especially near the top and base of the Brady Canyon member. In some units pelleted micrite has been completely replaced by dolomite; in others, micrite has only been partly dolomitized. The dolomite crystals range from 0.025 to 0.062 mm in diameter.

**Sandy dolomite**

A sandy dolomite texture (pl. 2f) prevails in the Marble Canyon area and is also found in the western sections near the base and the top of the member. The dolomite is silt size, with abundant fine- to coarse-sand-size quartz grains. The quartz grains are subangular to well rounded and are commonly bimodally distributed: the coarse sand well rounded and the fine sand and silt subangular.

**Aphanitic limestone and dolomite**

Aphanitic limestone and dolomite (pl. 2a) commonly occur at the base and top of the western sections of the Brady Canyon member. They are probably present in the east but are masked by abundant quartz grains. The carbonate is aphanitic (<.004 mm), and the dense rock it forms breaks with a conchoidal fracture. No fossils or fossil fragments have been found in this rock unit. Desiccation cracks occur in this unit at the top of some of the sections of the Brady Canyon member.

**Diagenesis**

Three diagenetic textures were observed in thin section: silicification of fossils, dolomitization, and recrystallization. Silicification tends to preserve original textures, whereas dolomitization destroys them. Much of the silicification is well developed only in weathered samples, and thus appears to be a surface weathering effect. The dolomite percentage map (fig. 5) reflects the importance of early dolomite as controlled by depositional environments of high evaporation rates in a supratidal environment.

**Woods Ranch member**

The Woods Ranch member consists of massive evaporites, sandstones, and minor carbonate units. The evaporites are mostly gypsum, but some in Hack Canyon are dark-grey anhydrite (pl. 1g). The horizontally bedded sandstones are fine-grained quartz arenites, commonly red (pl. 3e). Large-scale crossbeds are confined to the light-colored sandstones, which are also quartz arenites. Both the red and light-colored sandstones are bimodal in places (pl. 3d), with the fine fraction (2-3\(\phi\)) absent. The sandstones are locally friable but more
Figure 5.—Percent dolomite, Brady Canyon member.
commonly are cemented by calcite or quartz overgrowths. Petroliferous dolomite (pl. 3c) occurs in the type section, in Hack Canyon, and in Soap Creek. The Schizodus zone occurs in many sections. Algal stromatolites (pl. 3b) occur in a few limestone units; other carbonates in the Woods Ranch member are thin-bedded aphanitic dolomites. The intraformational breccias are commonly angular blocks of sandstone, several inches in diameter, within a sandstone matrix of a different color (red blocks in white matrix, or the reverse). In a few locations the breccia fragments are fine-grained limestones enclosed in a sandstone matrix. In Jackass Canyon (fig. 1) the Woods Ranch member contains abundant breccias and contorted bedding, with minor diapiric structures (pls. le, lf).

**FACIES ANALYSIS**

**Brady Canyon member**

The five textural types in the Brady Canyon member, as outlined in the petrology section, represent depositional facies. These facies are related to each other both vertically and horizontally (fig. 6).

The type section in Toroweap Valley clearly exhibits the transgressive-regressive nature of the Toroweap Formation. The Brady Canyon member represents three minor transgressions and regressions, as shown in a west-to-east correlation (fig. 7) of stratigraphic sections. These are labeled T-1, R-1, T-2, R-2, T-3, and R-3, in ascending order. The photograph of the Brady Canyon member in Toroweap Valley (pl. 1b) clearly exhibits three massive cliffs, which are the skeletal micrite facies, each representing a regression. Note that the lowermost sandy dolomite pinches out between Kane Canyon and Soap Creek (fig. 7, T-1) and that the middle sandy dolomite (T-2) pinches out between Soap Creek and Jackass Canyon. In the third transgression (T-3) the upper sandy dolomite extends into Jackass Canyon. Each succeeding transgression oversteps the preceding one. The facies map (fig. 8) represents the facies distribution during transgression T-2. The aphanitic facies is not shown because it is masked in the sandy dolomite facies. Each of these transgressions was followed by a regression, as evidenced by the sandstone facies (pl. 3f) shifting westward and displacing the carbonate facies to the west.

The entire area of Toroweap deposition in Arizona was free from wave action or high-energy currents. The fact that micrite occurs in every facies of the Brady Canyon member indicates a quiet depositional environment. Irwin (1965) introduced a carbonate model for deposition in an epeiric sea (fig. 9). The Toroweap Formation of northern Arizona falls into Irwin's sedimentation zones III, IV, V. The Brady Canyon member exhibits zones III and IV, and the Seligman and Woods Ranch members for the most part exhibit zone V.
Figure 6.—Transgressive-regressive facies, Toroweap Formation, northern Arizona (after Irwin, 1965).

Regression

W
Pigeon Canyon
Toroweap Valley
Thunder River Trail
Kane Canyon
Soap Creek
Jack Ass Canyon

APEANITIC
PELLETED
SKELETAL
SKELETAL
LIMESTONE
MICRITE
MICRITE
LIMESTONE

GYPSUM
SAND
DOLOMITE
DOLOMICRITE
SANDY

Transgression

E

T-3
T-2
T-1
Figure 7.—West to east correlation of the Toroweap Formation, northern Arizona.
Figure 8.—Facies map, Brady Canyon member, during transgression T-2.
**Figure 9.**—Sedimentation and energy zones, Toroweap Formation, northern Arizona (after Irwin, 1965).

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<td><strong>LOW ENERGY (marine currents)</strong></td>
<td><strong>LOW ENERGY</strong></td>
<td>Silt size skeletal fragments</td>
<td>Sand size skeletal fragments</td>
<td>Algal &amp; skeletal carbonates grading from sand----mud size or oolites--pellets</td>
<td>Penecontemporaneous chalky dolomite</td>
<td>Evaporites</td>
</tr>
<tr>
<td><strong>HIGH ENERGY (waves and tides)</strong></td>
<td><strong>IV</strong></td>
<td>Not present in Arizona; see Bissell &amp; Chilingar (1968) for Nevada</td>
<td>Not present in Arizona; see Bissell &amp; Chilingar (1968) for Nevada</td>
<td>Skeletal micrites Grain-support mud-support pelleted micrite</td>
<td>Aphanitic limestone &amp; sandy dolomite</td>
<td><strong>V</strong></td>
</tr>
<tr>
<td><strong>LOW ENERGY</strong></td>
<td><strong>V</strong></td>
<td>Not present in Arizona; see Bissell &amp; Chilingar (1968) for Nevada</td>
<td>Not present in Arizona; see Bissell &amp; Chilingar (1968) for Nevada</td>
<td>Skeletal micrites Grain-support mud-support pelleted micrite</td>
<td>Aphanitic limestone &amp; sandy dolomite</td>
<td>Gypsum</td>
</tr>
</tbody>
</table>
Supratidal and continental facies

The Seligman and Woods Ranch members, together with the transition and eastern phases of the Toroweap Formation, can be discussed in terms of a coastal-continental sabkha/dune model. The model will be outlined, and the members and phases will then be discussed in terms of their position in the model.

Kinsman (1969) reports the development of Holocene coastal and continental sabkhas along the Trucial Coast of the Persian Gulf. The evaporites associated with coastal sabkhas are precipitated from brines derived from seawater, whereas the evaporites formed in continental sabkhas are precipitated from brines formed by the evaporation of ground water. The two sabkhas grade laterally into one another. The important distinction between the two sabkhas is the dominant type of sediment in each; the sediments in a coastal sabkha are mostly carbonates and evaporites, whereas those in continental sabkhas are mostly detrital and were deposited by wind and ephemeral desert streams (wadis). Evaporite formation is extensive in both sabkhas. The continental sabkhas are commonly associated with deserts, and the sediments are distinguished on the basis of sedimentary structures; the sabkha sediments are horizontally bedded and the dune sediments crossbedded. When conditions permit the formation of continental sabkhas (arid climate, low hinterland relief, and small eolian supply) they may extend inland for great distances. Kinsman (1969) reports that the Sabkha Matti in the Persian Gulf extends 80 to 100 km inland from the coast.

Woods Ranch member
(coastal-continental sabkha facies)

Indications that an arid climate prevailed during Toroweap time include presence of extensive evaporite deposits, fresh, rounded feldspars in the sandstones, and development of dune crossbedding. Paleogeographic reconstructions indicate that the Defiance-Zuni area was not a significant positive area at the time of Toroweap deposition (McKee and Breed, 1969). Thus, paleoclimatic and paleogeographic conditions were favorable for the formation of a coastal-continental sabkha/dune complex at the edge of the Toroweap Sea (fig. 10).

Layering in the gypsum deposits of the Woods Ranch member resembles bedding. Kerr and Thompson (1963) suggest that changes after burial may obscure original textures in evaporites, and what may appear to be bedding may actually be coalesced nodules. If the layering in the gypsum of the Woods Ranch member is true bedding, the deposits were formed in a standing body of water, perhaps a brine pan, in a supratidal environment. If it is the result of coalesced nodules, the gypsum was formed interstitially in a sabkha environment. Either way, the environment of deposition is supratidal. This interpretation is supported by the presence of nearby dolomite in the Brady Canyon member, which was probably formed by seepage refluxion of brines with a high Mg/Ca ratio caused by precipitation of gypsum in the supratidal
Figure 10.—Facies analysis, Woods Ranch member.
environment. Algal stromatolites in the Woods Ranch member indicate either a high intertidal or low supratidal environment, which further substantiates a tidal flat origin for the evaporites. The general absence of fossils (except the Schizodus zone) also indicates a restricted environment.

The bimodal size distribution of some of the red and light-colored sandstones of the Woods Ranch member indicates an eolian phase in the formation of the sand. According to Folk (1968), the bimodality results from deflation in interdune flats where the fine fraction is selectively removed, and the finer and coarser fractions remain. The polycyclic history of the sandstones on the Colorado Plateau, however, precludes precise interpretations based on textures alone.

Sedimentary structures and deformational features of the Woods Ranch member are important in determining the environment of deposition. The intraformational breccias are probably solution collapse breccias formed by leaching of gypsum. The evaporites of continental sabkhas are more susceptible to solution than the evaporites formed in coastal sabkhas because of the proximity to fresh water. The small diapiric structures and contorted bedding can be attributed to deformation associated with evaporites. Most of the sandstones of the Woods Ranch member are horizontally bedded, which is the proper association for sabkha sedimentation. The occasional large-scale crossbeds probably represent ancient dunes that infrequently covered the sabkha.

McKee (1938) noted the cyclic nature of the Woods Ranch member of the Toroweap. The repeated interbedding of "chemical limestones," gypsum, and red beds is easily explained by minor fluctuations in sea level, which would shift the coastal-continental sabkha boundary and result in sequences of tidal flat carbonates and gypsum, with interbeds of continental sabkha sandstones.

Transition phase (continental sabkha)

The transition phase of the Toroweap, the area dominated by horizontally bedded sandstone, is interpreted to represent deposition in a continental sabkha. Evans (1971) and Pfirman (1968) noted breccias and contorted bedding in the Grindstone and Sycamore Canyon areas (fig. 10) similar to those described by the authors in the Woods Ranch member near Marble Canyon (fig. 1). The association of horizontal bedding, breccias, and contorted bedding in the transition phase suggests that the continental sabkhas extended from the south rim of the Grand Canyon to the area between Sycamore and Oak Creek Canyons (fig. 10).

Eastern phase (dune facies)

In Oak Creek Canyon (fig. 10), the Toroweap Formation consists entirely of southerly dipping crossbedded white sandstones, part of McKee's eastern phase. Benfer (1971, p. x) compared the Toroweap Formation and the Coconino Sandstone in Oak Creek Canyon and noted only "subtle variations" in grain-size distribution and crossbeds. The coincidence of the dip directions for both formations with the
direction of paleowinds during the Permian (Reiche, 1938) strongly suggests an eolian origin for both. Vertebrate tracks clearly establish such an origin for the Coconino. With comparable lithology, grain size, crossbedding, and direction of transport in the eastern phase Toroweap, it seems probable that desert conditions prevailed during both Coconino and Toroweap deposition in the Oak Creek area. Benfer (1971) noted climbing ripples in the eastern phase Toroweap. These require large amounts of sand and rapid deposition (McKee, 1965), conditions that are satisfied in ephemeral desert streams.

Seligman member
(coastal-continental sabkha facies)

The Seligman and Woods Ranch members have similar lithologies, sedimentary structures, and deformational features. The Seligman member, therefore, is also interpreted as representing a coastal and continental sabkha facies.

The development of coastal sabkhas was not as extensive in the Seligman member as in the Woods Ranch member (fig. 11). The transgression from the west began slowly, with extensive development of coastal sabkhas. When the sea reached Hack Canyon (fig. 1), the rate of transgression increased and time did not permit accumulation of thick gypsum deposits. The following regression was slow across northern Arizona, and allowed the thick evaporite deposits of the Woods Ranch member to form.

According to Schleh (1966), the contact between the Coconino and Toroweap in Parashant Canyon (fig. 1) is extremely difficult to establish. Carbonate units occur beneath thick units of Coconino-like sandstones, and large-scale crossbeds occur well into the Seligman member of the Toroweap. The two formations are not separated by an unconformity in this area. The nature of the contact suggests that the Parashant Canyon area was subject to oscillations in sea level while the Coconino was being deposited elsewhere in northern Arizona. This interpretation is supported by Fisher (1961), who postulated marine deposition for the Shivwits Plateau area (fig. 2) contemporaneous with eolian deposition to the east. During Toroweap time, the sea encroached the area of eolian deposition. Because of the low depositional slope, extensive coastal and continental sabkhas formed, in which gypsum and horizontally bedded sandstones of the Seligman member were deposited on top of the beveled Coconino sands (fig. 12).

PREVIOUS INTERPRETATIONS

Others have attempted to interpret local areas of the Toroweap Formation, particularly in the transition and eastern phases, without an overall model for Toroweap deposition. The original interpretations will be discussed, and the data will then be reinterpreted in light of the proposed coastal-continental sabkha/dune model.
Figure 11.—Facies analysis, Seligman member and Coconino Sandstone.
Figure 12.—Schematic diagram of Coconino-Toroweap deposition (after Fisher, 1961).
Pfirman (1968) and Evans (1971) explained the breccias and contorted bedding in the transition phase in terms of subaqueous deposition, with the breccia resulting from "rip-ups" caused by wave action or erosion during storms, and the contorted bedding resulting from gravity slumping or loading. The present authors found crossbedding preserved in breccia fragments in Marble Canyon (pl. 1f) which indicates brecciation after lithification.

Pfirman (1968) noted the lack of crossbedding in the transition phase, which he attributed to a high-flow regime. Evans (1971) noted the lack of typical beach crossbeds in the Sycamore and Oak Creek Canyon areas and suggested that the strandline must be farther east. On the basis of crossbedding and presence of climbing ripples, Benfer (1971) suggested that the eastern phase Toroweap represents both eolian and marine deposition. Evans (1971) proposes a submarine sand wave origin for the southerly dipping crossbeds in the eastern phase.

The difficulties engendered by assuming a subaqueous origin for the transition and eastern phases of the Toroweap are overcome by the proposed model. The breccias and contorted bedding are associated with the leaching of gypsum originally deposited in a continental sabkha, the horizontal bedding represents sabkha sedimentation, and the high-angle crossbedding in the eastern phase has an eolian rather than subaqueous origin.

GEOLOGIC HISTORY

Toroweap deposition in Arizona began in the west, as shallow marine waters slowly transgressed across arid land areas of sand dunes and continental sabkhas. Sea level gradually rose, without waves or tidal currents, these having been dissipated far to the west. The Seligman member was deposited during this initial transgression, incorporating the sands from the Coconino dunes. The continental and coastal sabkhas were then covered by aphanitic limestones precipitated from a supersaturated brine. Some of these supratidal carbonates were later dolomitized. As the sea transgressed farther to the east, broad, shallow-water mud flats formed, and the lime muds were thoroughly burrowed and pelleted. Much of this mud facies was later dolomitized by refluxion of high-magnesium brines through the sediment. The area of mud deposition was somewhat restricted, and only a bivalve and gastropod fauna flourished.

As sea level continued to rise, normal marine faunas moved into western Arizona, as indicated by the abundance of brachiopods, crinoids, and bryozoans. There were three pulses of transgression followed by three minor regressions during deposition of the Brady Canyon member. Evidence for the regressions is the shifting of the carbonate and sandstone facies westward. After the third and most extensive transgression in the upper Brady Canyon member, the sea slowly withdrew at least as far west as Nevada. The area that had been the site of relatively deep marine deposition became the depositional site of coastal sabkhas. Gradual regression westward resulted in the
accumulation of thick gypsum deposits of the Woods Ranch member. Continental sabkhas and areas of eolian deposition extended to the east and south. One transgressive pulse occurred during deposition of the upper Woods Ranch member, which resulted in the widespread Schizodus zone. This preceded the major transgression that followed and deposited the Kaibab Limestone. The Toroweap Formation proves to be a good example, not only of a transgressive-regressive sequence, but also for marine carbonate-sabkha-eolian deposition.
PLATE 1

1a. NASA photograph of Grand Canyon, taken from Apollo 9. Approximate area of this study.

1b. Near the type section of the Toroweap Formation, in Toroweap Valley, Arizona. Note three massive cliffs; R-1, R-2, R-3, which represent three regressions in the Brady Canyon member (BC). The contacts with the Coconino Sandstone (Pc), and the Kaibab Limestone (Pk) are outlined.

1c. Lower contact of the Toroweap Formation, Jackass Canyon. Horizontally bedded sandstones of the Seligman member truncate the uppermost crossbeds of the Coconino Sandstone.

1d. Facies change in the Woods Ranch member, Rider Canyon, near Marble Canyon. The west to east change from gypsum to sandstone is closely paralleled by a geomorphic change from slope to cliff.

1e. Minor diapiric structure in the Woods Ranch member, Jackass Canyon. Beds are domed and brecciated around the squeeze-up. Hammer on right for scale.

1f. Breccia in Woods Ranch member, Jackass Canyon. Crossbedding is preserved within breccia fragments.

1g. Dark anhydrite layers and breccia zones in Woods Ranch member, Hack Canyon.


2c. Grain-supported skeletal limestone; micrite matrix. Brady Canyon member, Pigeon Canyon. Sample PC-3.


2e. Dolomicrite - Brady Canyon member, Toroweap Valley. Sample T-6.


2g. Mud-supported skeletal micrite - Brady Canyon member, Thunder River Trail. Note abundant quartz grains. Sample TR-7+45.

2h. Mud-supported skeletal micrite - Brady Canyon member, Pigeon Canyon. Note endothyrids. Sample PC-6.

Note: Scale of all photomicrographs, X25.
PLATE 3


3e. Red, fine-grained, quartz arenite in Woods Ranch member, near Marble Canyon. Sample SW-5.

3f. Sandstone unit within the Brady Canyon member in Marble Canyon. Sample SH-B-3.

Note: Scale of all photomicrographs, X25.
Plate 3
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MESOZOIC GEOLOGY

Mesozoic stratigraphy of northeastern Arizona
by
Richard F. Wilson

Mesozoic vertebrates of northern Arizona
by
Edwin H. Colbert
MESOZOIC STRATIGRAPHY OF NORTHEASTERN ARIZONA

by

Richard F. Wilson
Department of Geosciences
University of Arizona, Tucson, Arizona
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ABSTRACT

Mesozoic rocks form most of the surface outcrops north and east of the Little Colorado River in Arizona. Triassic and Jurassic rocks form a sequence of "red beds" of continental and transitional marine origin that are noted for their vertical variability and lateral facies changes. Marine and transitional Cretaceous rocks occupy the center of the Black Mesa basin, and also crop out in the eastern part of the Mogollon Rim.

INTRODUCTION

This report summarizes the Mesozoic stratigraphy (table 1) of northeastern Arizona and stresses the more recent work in the area. No attempt is made to cite all of the extensive previous work in this region; such citation can be found in the detailed reports that are noted here.

TRIASSIC ROCKS

In northeastern Arizona, three units are of Triassic or probable Triassic age: the Moenkopi Formation (Lower and Middle(?)-Triassic), the Chinle Formation (Upper Triassic), and most or all of the Glen Canyon Group.

Moenkopi Formation

The Moenkopi Formation in northeastern Arizona is a reddish-brown siltstone and sandstone unit that crops out extensively in the Little Colorado River Valley from St. Johns on the east to Cameron in the west, and northward from Cameron along the Echo and Vermilion Cliffs to the Utah state line (fig. 1). Outcrops are also present along the Mogollon Rim and in Monument Valley. The formation is present in the southern part of the Defiance uplift but absent in the northern part of the uplift. Recent studies of the formation in northern Arizona include those of McKee (1954), Repenning and others (1969), Stewart and others (1972a), and Baldwin (1973).

The Moenkopi rests unconformably on Permian rocks throughout northeastern Arizona: Kaibab Limestone in the western and southwestern part of the region, Coconino Sandstone in the central part, DeChelly Sandstone in Monument Valley and the southern part of the Defiance uplift, and the San Andreas Limestone in the area around St. Johns.
Table 1.—Summary of Mesozoic rock units in northeastern Arizona.

<table>
<thead>
<tr>
<th>Age</th>
<th>Rock Units</th>
</tr>
</thead>
<tbody>
<tr>
<td>Cretaceous</td>
<td>Yale Point Sandstone</td>
</tr>
<tr>
<td></td>
<td>light-colored, cross-stratified, cliff-forming sandstone as much as 91 m (300 ft) thick.</td>
</tr>
<tr>
<td></td>
<td>Wepo Formation</td>
</tr>
<tr>
<td></td>
<td>gray siltstone, mudstone, sandstone, and coal that forms ledges and slopes; 90-230 m (300-750 ft) thick.</td>
</tr>
<tr>
<td></td>
<td>Toreva Formation</td>
</tr>
<tr>
<td></td>
<td>a lower and upper light-colored, cross-stratified, cliff-forming sandstone member separated by a slope-forming member of variegated siltstone, and dark carbonaceous mudstone and coal; 43-100 m (140-325 ft) thick.</td>
</tr>
<tr>
<td></td>
<td>Mancos Shale</td>
</tr>
<tr>
<td></td>
<td>slope-forming, grayish, fossiliferous mudstone and siltstone; 145-235 m (475-770 ft) thick.</td>
</tr>
<tr>
<td></td>
<td>Dakota Sandstone</td>
</tr>
<tr>
<td></td>
<td>a lower and an upper cliff-forming light-colored, cross-stratified sandstone member separated by a dark slope-forming, carbonaceous siltstone and coal member; 15-45 m (50-150 ft) thick.</td>
</tr>
<tr>
<td></td>
<td>Brushy Basin Member</td>
</tr>
<tr>
<td></td>
<td>slope-forming, variegated claystone and sandy claystone; 0-75 m (0-250 ft).</td>
</tr>
<tr>
<td>Jurassic</td>
<td>Westwater Canyon Member</td>
</tr>
<tr>
<td></td>
<td>cliff-forming, grayish to yellowish, cross-stratified sandstone and minor mudstone; 0-90 m (0-300 ft) thick.</td>
</tr>
<tr>
<td></td>
<td>Recapture Member</td>
</tr>
<tr>
<td></td>
<td>slope-forming, reddish sandstone and mudstone; 0-180 m (0-600 ft).</td>
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<tr>
<td></td>
<td>Salt Wash Member</td>
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<tr>
<td></td>
<td>interstratified cross-stratified sandstone and greenish to grayish mudstone; 0-120 m (0-400 ft) thick.</td>
</tr>
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<td></td>
<td>Cow Springs Sandstone</td>
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<td></td>
<td>slope- to cliff-forming, grayish, planar cross-stratified sandstone; 0-105 m (350 ft) thick.</td>
</tr>
<tr>
<td></td>
<td>Bluff Sandstone</td>
</tr>
<tr>
<td></td>
<td>cliff-forming grayish planar cross-stratified sandstone; 0-25 m (80 ft) thick.</td>
</tr>
<tr>
<td>Triassic</td>
<td>Summerville Formation</td>
</tr>
<tr>
<td></td>
<td>a lower silty member of reddish brown horizontally stratified to contorted silty sandstone, and an upper sandy member of cross-stratified sandstone; 0-60 m (200 ft) thick.</td>
</tr>
<tr>
<td></td>
<td>Tocilto Limestone</td>
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<tr>
<td></td>
<td>yellowish to purplish siltstone and grayish limestone; 0-7 m (25 ft) thick.</td>
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<tr>
<td></td>
<td>Entrada Sandstone</td>
</tr>
<tr>
<td></td>
<td>a lower sandy member of cliff-forming, orangish cross-stratified sandstone, 0-106 m (350 ft) thick; a medial silty member of slope-forming, reddish sandy siltstone, 0-45 m (150 ft) thick; and an upper sandy member of cliff-forming, orangish sandstone; 0-73 m (240 ft) thick.</td>
</tr>
<tr>
<td></td>
<td>Carmel Formation</td>
</tr>
<tr>
<td></td>
<td>mostly slope-forming, horizontally stratified, reddish siltstone and minor sandstone, 0-60 m (200 ft) thick.</td>
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<tr>
<td></td>
<td>Navajo Sandstone</td>
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<tr>
<td></td>
<td>cliff-forming, light brown, planar-cross-stratified sandstone; 0-490 m (1,600 ft) thick.</td>
</tr>
<tr>
<td></td>
<td>Kayenta Formation</td>
</tr>
<tr>
<td></td>
<td>a northeastern assemblage of reddish cross-stratified sandstone, and a southwestern assemblage of reddish siltstone and minor sandstone; 0-210 m (700 ft) thick.</td>
</tr>
<tr>
<td>Triassic?</td>
<td>Springdale Sandstone</td>
</tr>
<tr>
<td></td>
<td>ledge-forming, pale red, cross-stratified sandstone 0-85 m (230 ft) thick.</td>
</tr>
<tr>
<td></td>
<td>Dinosaur Canyon Member</td>
</tr>
<tr>
<td></td>
<td>slope- to cliff-forming, reddish orange to reddish brown siltstone and sandstone; 0-105 m (350 ft) thick.</td>
</tr>
<tr>
<td></td>
<td>Lukachukai Member</td>
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<tr>
<td></td>
<td>cliff-forming, light brown, cross-stratified sandstone; 0-150 m (500 ft) thick.</td>
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<tr>
<td></td>
<td>Owl Rock Member</td>
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<tr>
<td></td>
<td>slope-forming, reddish siltstone and ledge-forming, grayish limestone, 0-105 m (350 ft) thick.</td>
</tr>
<tr>
<td>Jurassic?</td>
<td>Petrified Forest Member</td>
</tr>
<tr>
<td></td>
<td>slope-forming, variegated grayish to reddish, bentonitic claystone to clayey sandstone; 90-360 m (300-1,200 ft) thick.</td>
</tr>
<tr>
<td></td>
<td>Sandstone and siltstone</td>
</tr>
<tr>
<td></td>
<td>ledge-forming, lenticular sandstone and slope-forming claystone and clayey sandstone; 12-85 m (40-275 ft) thick.</td>
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<tr>
<td></td>
<td>Shinarump Member</td>
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<td></td>
<td>cliff- or ledge-forming, cross-stratified sandstone to conglomeratic sandstone; 0-30 m (100 ft) thick.</td>
</tr>
<tr>
<td></td>
<td>Moenkopi Formation</td>
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<tr>
<td></td>
<td>Principally slope- and ledge-forming, reddish brown siltstone and lesser sandstone 0-150 m (500 ft) thick. In Little Colorado River valley, the formation is divisible into a lower Wupatki Member of horizontally stratified and ripple-laminated siltstone, a medial Moqui Member of horizontally stratified siltstone and gypsum, and an upper Holbrook Member of horizontally to ripple-laminated siltstone and cross-stratified sandstone.</td>
</tr>
</tbody>
</table>

Rocks of Permian age

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Figure 1.—Index map of northeastern Arizona.
The Moenkopi thickens westward from a wedge-edge along the western and southern flank of the Defiance uplift to more than 150 m (500 ft) in the northern part of the Echo Cliffs region.

The three members of the Moenkopi recognized in the Little Colorado River Valley are, in ascending order, the Wupatki Member, composed principally of reddish-brown horizontally stratified and ripple-laminated siltstone; the Moqui Member, composed principally of reddish-brown and grayish siltstone, and minor white gypsum; and the Holbrook Member, composed principally of reddish-brown horizontally stratified and ripple-laminated siltstone and cross-stratified sandstone. The Holbrook Member probably represents most or all of the Moenkopi Formation in the southeasternmost part of the region—near St. Johns and the southern part of the Defiance uplift. A short distance north of Cameron, the Moqui Member cannot be recognized; from this point northward and westward along the Echo and Vermilion Cliffs, east of the Kaibab Plateau, the Moenkopi consists of a sequence of reddish-brown siltstone and sandstone to which no formal member designation is applied.

The Wupatki Member includes a prominent ledge-forming sandstone, 3 to 18 m (10 to 60 ft) thick, termed the lower massive sandstone by McKee (1954). This sandstone is near the base of the member at Holbrook, rises stratigraphically to the top of the member near Cameron and extends northward and westward beyond the limits of the Wupatki Member, where the sandstone occupies a position high in the undifferentiated sequence of Moenkopi Formation along the Echo and eastern Vermilion Cliffs. It is a key marker for physical correlation of the Moenkopi Formation.

Recent studies on the paleoenvironment of the Moenkopi Formation in northeastern Arizona (Baldwin, 1973) indicate that the strata beneath the lower massive sandstone accumulated on tidal flats or in lagoons; that the lower massive sandstone represents a transgressive-regressive cycle from tidal flats to shallow marine to beach; and that the strata above the lower massive sandstone are largely fluvial, with the exception of the Moqui Member, which represents deposition on tidal flats.

Chinle Formation

The Chinle Formation is present in outcrop and in the subsurface throughout northeastern Arizona from the Little Colorado River Valley in the south and east to the Echo and Vermilion Cliffs in the west and northwest. Recent studies of the formation in Arizona include those of Repenning and others (1969), Stewart and others (1972b), O'Sullivan (1970), O'Sullivan and Green (1973), and Breed and Breed (1972).

The Chinle Formation rests unconformably on the Moenkopi Formation throughout northeastern Arizona, with the exception of the central and northern parts of the Defiance uplift, where it rests unconformably upon the Permian DeChelly Sandstone.
Rocks assigned to the Chinle Formation range from 320 to 366 m (1,050 to 1,200 ft) in thickness in the eastern part of the region to more than 425 m (1,400 ft) near Cameron. The formation thins northward along the Echo Cliffs from Cameron to Lees Ferry, where it is about 300 m (1,000 ft) thick. This decrease in thickness is due to progressive beveling of the Chinle along an unconformity at the base of the overlying Glen Canyon Group.

The stratigraphy of the Chinle Formation on the Colorado Plateau is complex, and the formation has been divided into many members. On a regional basis, the formation can be divided into a lower and an upper part (Steward and others, 1972b). In northeastern Arizona, the lower part comprises (1) the basal Shinarump Member, a somewhat discontinuous unit of sandstone to conglomeratic sandstone of variable thickness; (2) a medial group of laterally equivalent members, from 12 to 84 m (40 to 275 ft) thick, of lenticular ledge-forming sandstone and slope-forming claystone and clayey sandstone termed the sandstone and mudstone member in the Echo Cliffs area, the Monitor Butte Member in Monument Valley, the lower red member in the Defiance uplift, and the Mesa Redondo Member in the eastern part of the Little Colorado River Valley; and (3) the upper Petrified Forest Member, composed of a thick sequence of variegated grayish to reddish bentonitic claystone and clayey sandstone. Ledge-forming sandstone units are present in the Petrified Forest Member in the southeastern part of the region and in the Defiance uplift. The lower part of the Chinle thins generally northward from more than 300 m (1,000 ft) in the Little Colorado River Valley to less than 150 m (500 ft) at the northern end of the Defiance uplift. The entire unit is considered of fluvial origin, deposited by streams flowing northward and northwestward from a source area, the Mogollon Highland, to the south (Stewart and others, 1972b; Poole, 1961).

The upper part of the Chinle Formation in northeastern Arizona consists of the Owl Rock Member, a unit of reddish siltstone and minor amounts of interbedded fresh-water limestone from 0 to 110 m (350 ft) thick. The member is absent at St. Johns owing to pre-Cretaceous erosion but present elsewhere in the region.

**Glen Canyon Group**

In northeastern Arizona, the Glen Canyon Group is composed of, in ascending order, the Wingate Sandstone, the Moenave Formation, the Kayenta Formation, and the Navajo Sandstone. The Wingate Sandstone was, until recently, assigned to the Triassic, the Moenave and Kayenta Formations to the Triassic(?), and the Navajo Sandstone to the Triassic(?) and Jurassic (Lewis and others, 1961). However, all of the Navajo Sandstone in northeastern Arizona is now thought to be of Triassic age (Galton, 1971; O'Sullivan and Green, 1973). Recent studies including part or all of the Glen Canyon Group in the region include those of Harshbarger and others (1957), Johnson (1967), O'Sullivan (1970), Stewart and others (1972b), and O'Sullivan and Green (1973).
Wingate Sandstone

In northeastern Arizona, the Wingate Sandstone comprises two members: the Rock Point Member (the lower) and the Lukachukai Member (the upper). The Rock Point Member, reddish-brown siltstone with very thick horizontal bedding, contains minor lenses of fluvial sandstone with trough cross-stratification in outcrops near the Utah state line. In outcrops along the east flank of the Defiance uplift, sandstone of probable eolian origin occurs as lenses and tongues with planar-cross-stratification. The member conformably overlies the Owl Rock Member of the Chinle Formation and ranges from 0 to more than 150 m (500 ft) in thickness. It is thickest along a northeast-trending belt extending from north of Winslow to the northern Defiance uplift. It thins to the southeast owing to truncation along an unconformity at the base of the Lukachukai Member of the Wingate Sandstone.

The status of the member is in question. It originally was considered part of the Chinle Formation but was placed in the Wingate Sandstone by Harshbarger and others (1957). Stewart and others (1972b) stress the equivalency of this unit to the Church Rock Member of the Chinle Formation in southeastern Utah. O'Sullivan (1970), however, has questioned the exact equivalency of the Rock Point Member to most of the Church Rock Member in Utah. Further study is need to resolve this question.

The Lukachukai Member of the Wingate is equivalent to the entire Wingate Sandstone of Utah and Colorado. This cliff-forming unit of light-brown planar-cross-stratified sandstone ranging from 0 to almost 150 m (500 ft) in thickness, reaches an erosional pinchout along a northeast-trending line extending from the central part of the Defiance uplift southwest to Ganado. To the west the unit grades laterally into the overlying Dinosaur Canyon Member of the Moenave Formation and is absent along the Echo and Vermilion Cliffs, except in the immediate vicinity of Lees Ferry. The member intertongues with the Rock Point Member along the east flank of the Defiance uplift, but to the west appears to unconformably overlie and truncate the Rock Point Member. The Lukachukai Member unconformably overlies the Owl Rock Member of the Chinle Formation in western exposures where the Rock Point Member is absent. The Lukachukai Member is of eolian origin, deposited by winds blowing from the northwest (Poole, 1962).

Moenave Formation

The Moenave Formation in northeastern Arizona comprises two members: the Dinosaur Canyon Member (the lower) and the Springdale Sandstone Member (the upper). The Dinosaur Canyon Member, reddish-orange to pale-reddish-brown sandstone and pale reddish brown siltstone, is thickest, more than 105 m (350 ft), in the southwestern part of its area of outcrop and thins eastward to a pinchout along a line that extends slightly east of south from east of Kayenta on the
north to near Ganado on the south. The southern pinchout is due to truncation by Jurassic rocks; the northern is caused by gradation into the Lukachukai Member of the Wingate Sandstone. The member rests conformably on the Lukachukai Member and unconformably on beds of the Chinle Formation west of the pinchout of the Lukachukai Member. Two facies of the Dinosaur Canyon Member are recognized along its southern exposures: an eastern facies composed of planar-cross-stratified sandstone of probable eolian origin deposited by winds blowing from the northwest, and a western facies of siltstone and trough cross-stratified sandstone deposited by streams flowing northwest, as in Chinle time (Johnson, 1967).

The Springdale Sandstone Member, a ledge-forming pale-red trough cross-stratified sandstone that crops out along the Echo and Vermillion Cliffs in the western part of the region, thickens northward from a wedge-edge near Tuba City at the southern margin of the Echo Cliffs to 85 m (280 ft) at Lees Ferry. Eastward, the Springdale Sandstone appears to merge with—and forms most of—the type Kayenta Formation at Kayenta. The Springdale Sandstone is of fluvial origin and was deposited by streams flowing toward the southwest.

Kayenta Formation

The Kayenta Formation as presently defined in northeastern Arizona consists of two distinct lithologic assemblages. One, present in outcrops in the northeastern part of the area in Monument Valley, the northwestern Defiance uplift, and in areas near the Utah state line east of the Echo Cliffs, consists largely of ledge-forming pale-red trough-cross-stratified sandstone deposited by streams flowing toward the southwest. This assemblage ranges from 0 to 75 m (242 ft) in thickness, thickens westward, and represents the type Kayenta Formation. It is probably the equivalent of the Springdale Sandstone Member of the Moenave Formation in the Echo and Vermillion Cliffs.

The other assemblage consists of slope-forming units of siltstone and minor trough cross-stratified sandstone, locally interstratified in its upper part with tongues of planar cross-stratified Navajo Sandstone. This assemblage crops out along the Vermillion and Echo Cliffs, and along the northern fringe of the Little Colorado River Valley. It reaches a maximum thickness of 215 m (700 ft) at the southern end of the Echo Cliffs. It appears to be a lateral equivalent of the Navajo Sandstone further to the north and east, and intertongues with the Navajo in that direction.

Thus, the two recognized assemblages of the Kayenta probably are not lateral equivalents, and the application of the name Kayenta to both is questionable. The Kayenta Formation pinches out eastward, owing to erosional truncation at the base of Jurassic rocks, along a line slightly east of north extending from west of Ganado to the northeast flank of the Defiance uplift.
Navajo Sandstone

The Navajo Sandstone, a cliff-forming light-brown to orange planar cross-stratified sandstone that crops out extensively in the northern part of northeastern Arizona, is more than 490 m (1,600 ft) thick near the western edge of the region, at Lees Ferry, and thins southward and eastward to an erosional pinchout along a southwesterly trending line extending from Four Corners to west of the Hopi Buttes. It is of eolian origin, deposited by wind blowing from the north and northwest (Poole, 1962).

JURASSIC ROCKS

In northeastern Arizona, Jurassic rocks crop out around all except the southern margin of Black Mesa, in the Four Corners area, and along the east flank of the Defiance uplift. Elongate separate outcrops extend northwestward from the west edge of Black Mesa to the Utah state line. The Jurassic lies unconformably on Triassic rocks throughout the region.

San Rafael Group

The San Rafael Group consists of, in ascending order, the Carmel Formation, Entrada Sandstone, Todilto Limestone, Summerville Formation, and Bluff Sandstone, the Cow Springs Sandstone, and the Morrison Formation. The Carmel Formation is of Middle and Late Jurassic age; the rest of the formations are classed as Late Jurassic. Studies of Jurassic rocks in the region include those of Harshbarger and others (1957), Craig and others (1955), and O'Sullivan and Craig (1973).

Carmel Formation

Throughout most of its area of outcrop, the Carmel Formation consists principally of slope-forming, reddish-brown, horizontally stratified siltstone and lesser ledge-forming sandstone. The formation ranges from 0 to more than 60 m (200 ft) in thickness, and reaches an eastern wedge-edge along the western flank of the Defiance uplift. To the south, at the southern margin of Black Mesa, the reddish siltstone facies of the Carmel grades into a reddish and white planar cross-stratified to horizontally stratified sandstone that is indistinguishable from the overlying Entrada Sandstone. Similar cross-stratified sandstone near Lupton may also represent the Carmel Formation (O'Sullivan and Craig, 1973).
Entrada Sandstone

The Entrada Sandstone is divided into three members in northeastern Arizona: a lower and an upper sandy member composed of cliff-forming orangish cross-stratified sandstone, both wind and water deposited; and a medial silty member composed of slope-forming reddish-brown horizontally stratified to gnarly bedded sandy siltstone (Harshbarger and others, 1957). The lower sandy member, from 0 to 105 m (350 ft) thick, is recognized in the northwestern and north central part of the area; the medial silty member, from 0 to 46 m (150 ft) thick, is recognized in all except the northwestern and westernmost parts of the area; and the upper sandy member, from 0 to 73 m (240 ft) thick, is recognized in only the eastern part of the area, near Four Corners and along the Defiance uplift. The Entrada in the southwest part of the area consists of reddish and white cross-stratified sandstone indistinguishable from that of the underlying Carmel Formation. Recent studies suggest that some of the correlations and member terminology of the Entrada in northeastern Arizona are in need of revision (O'Sullivan and Craig, 1973).

Todilto Limestone

Only the westernmost fringe of the Todilto Limestone is present in northeastern Arizona, on the northern part of the east flank of the Defiance uplift. The formation consists of from 0 to 7 m (25 ft) of horizontally stratified yellowish to purplish siltstone and grayish limestone of lacustrine origin.

Summerville Formation

The Summerville Formation is present in the northeastern part of northeastern Arizona, and grades laterally southward into Cow Springs Sandstone along a south-southeast-trending line extending from the Navajo Mountain area on the northwest to a few kilometers north of Ft. Defiance on the east. The formation ranges from 0 to more than 60 m (200 ft) thick and is divisible into a lower silty member and an upper sandy member (Harshbarger and others, 1957). The lower silty member is present in the Four Corners and Monument Valley area and consists of horizontally stratified to contorted reddish-brown silty sandstone and sandstone. The member is overlain by and grades laterally southward into the upper sandy member, which comprises the entire Summerville south and west of Four Corners and Monument Valley. The upper sandy member is composed of horizontally stratified to cross-stratified sandstone. Tongues of Cow Springs Sandstone lie within the Summerville Formation in its more southerly occurrences.
Bluff Sandstone

The Bluff Sandstone, the uppermost unit of the San Rafael Group, is a cliff-forming planar cross-stratified grayish sandstone of eolian origin that is recognized in the northeastern part of northeastern Arizona. It ranges from 0 to 24 m (80 ft) in thickness in this part of Arizona but is much thicker to the north in Utah. It is believed to merge with and be inseparable from the Cow Springs Sandstone to the south and southwest and is considered a tongue of the Cow Springs Sandstone (Harshbarger and others, 1957).

Cow Springs Sandstone

The Cow Springs Sandstone, a slope-to cliff-forming unit of grayish planar cross-stratified sandstone, is best developed in the southwestern and southern part of the region. In the southwest, the formation occupies the entire interval between the Entrada Sandstone below and Cretaceous rocks above and is as much as 105 m (350 ft) thick. It intertongues northeastward and northward with the Summerville Formation and with various members of the Morrison Formation. As noted, the Bluff Sandstone is considered a tongue of this formation. The Cow Springs is of eolian origin, and was deposited by winds blowing largely from the north (Poole, 1962).

Morrison Formation

The Morrison Formation, the uppermost unit of Jurassic age, is divisible into four members in northeastern Arizona: in ascending order, they are the Salt Wash, Recapture, Westwater Canyon, and Brushy Basin Members (Craig and others, 1955; Harshbarger and others, 1957).

The Salt Wash Member consists of interstratified sets of lenticular trough cross-stratified sandstone and horizontally stratified greenish to reddish mudstone of fluvial origin. The member is present northeast of a line extending from the intersection of the Colorado River with the state line southeast to the northern Defiance uplift. The member is several tens of meters thick in the more northern occurrences, near Navajo Mountain, but things rapidly to the southwest, east, and south. The disappearance of the Salt Wash along the western third of its line of pinchout may be due to lateral gradation into Cow Springs Sandstone; along the eastern two-thirds it may be due to lateral gradation of Salt Wash into Recapture Member (Harshbarger and others, 1957; O'Sullivan and Craig, 1973). Source area of the Salt Wash was probably to the southwest and west.

The Recapture Member, a slope-forming unit of fluvial origin, consists of cross-stratified reddish earthy sandstone and interlayered sandy mudstone that occupy the northeast half of northeastern Arizona north of a line extending southeast from the central part of Black
Mesa to a point a few miles south of Lupton. The member is more than 180 m (600 ft) thick at the northern end of Black Mesa. It thins northward, partly because it intertongues with the Salt Wash Member, and southward to a pinchout because it intertongues with the Cow Springs Sandstone. Source areas for the Recapture probably lay to the Southeast.

The Westwater Canyon Member, a unit of cliff-forming grayish to yellowish cross-stratified sandstone and minor mudstone of fluvial origin, has approximately the same limits of distribution as the Recapture Member in northeastern Arizona. It is as much as 90 m (300 ft) thick in part of the Defiance uplift, thins northward owing to intertonguing with the overlying Brushy Basin Member, and thins rapidly southwest owing to pre-Cretaceous truncation and possible intertonguing with Cow Springs Sandstone. Source areas for the Westwater Canyon were probably the same as for the Recapture Member.

The Brushy Basin Member consists of slope-forming variegated bentonitic claystone and sandy claystone. It is about 75 m (250 ft) thick near Four Corners (O'Sullivan and Craig, 1973) but thins southwest to a pinchout north of Black Mesa and the central Defiance uplift. Thinning is due largely to pre-Cretaceous erosion.

**CRETACEOUS ROCKS**

Rocks of Cretaceous age in northeastern Arizona occupy the Black Mesa basin, crop out along the eastern margin of the Mogollon Rim, and are present in small scattered outcrops near Four Corners. With the exception of a thin wedge-edge of Lower Cretaceous Burro Canyon Formation at the very northeast corner of the state, all Cretaceous rocks in northeastern Arizona are assigned to the Dakota Sandstone of Early and Late Cretaceous age, or to the Mancos Shale and Mesaverde Group of Late Cretaceous age, or to their equivalents. Recent studies of Cretaceous rocks in northeastern Arizona include those of Hazenbush (1972), O'Sullivan and others (1972), Repenning and Page (1956), and Young (1973).

**Dakota Sandstone**

The Dakota Sandstone, or its equivalent, rests unconformably on older rocks in northeastern Arizona. In the Black Mesa basin, the Dakota overlies the Jurassic Cow Springs Sandstone in the south and west, and the Morrison Formation to the north and east. South of Black Mesa, the Dakota or its equivalent rests on progressively older rocks: Triassic Chinle Formation near St. Johns, and Permian rocks along the eastern edge of the Mogollon Rim. This evidence of progressive southward beveling indicates uplift and northward tilt of what is now the southern edge of the Colorado Plateau before Cretaceous deposition.
In the Black Mesa area, the Dakota Sandstone, from 15 to 45 m (50 to 150 ft) thick, has been divided into three members that are locally discontinuous. In ascending order these members are: (1) a lower sandstone member, typically composed of 10 to 20 m (30 to 60 ft) of cliff-forming, light-colored, cross-stratified sandstone; (2) a middle carbonaceous member, typically composed of 6 to 12 m (20 to 40 ft) of slope-forming, horizontally stratified, dark carbonaceous siltstone and coal of lagoonal origin; and (3) an upper sandstone member, present in the northern half of Black Mesa, of light-colored, low-angle, cross-stratified sandstone, of probable beach origin. Equivalents of the Dakota Sandstone are recognized in a sequence of undifferentiated Cretaceous rocks along the eastern part of the Mogollon Rim.

**Mancos Shale**

In the Black Mesa basin, the Mancos Shale ranges from about 145 m (475 ft) thick at the southwestern edge of the Mesa to more than 215 m (700 ft) in the north. The formation is composed principally of slope-forming fossiliferous grayish mudstone and siltstone of marine origin. Thin orange silty sandstone sets are common in the medial part of the formation. The formation grades into the Dakota Sandstone below and the Mesaverde Group above.

Three biostratigraphic units recently have been recognized in the Mancos in the Black Mesa area (Hazenbush, 1972): a lower unit containing planktonic and benthonic Foraminifera of shallow marine origin; a middle nearly barren unit, containing only plant megaspores, that probably accumulated on a nearly subaerial mudflat; and an upper unit, containing arenaceous Foraminifera, that accumulated on a partially flooded mudflat. The age of the formation, based on both microfossils and megafossils, ranges from Late Cenomanian to middle Turonian. The Mancos of Black Mesa is equivalent to only the lower part of the Mancos Shale in the northern San Juan basin in northwestern New Mexico. No shale similar to the Mancos has been recognized in the Cretaceous sandstone sequence along the Mogollon Rim, but rocks of the same age as the Mancos are believed present there.

**Mesaverde Group**

The Mesaverde Group, in the Black Mesa basin, is composed of three formations; in ascending order they are the Toreva Formation, the Wepo Formation, and the Yale Point Sandstone. The Mesaverde Group on Black Mesa is equivalent to much of the upper part of the Mancos Shale in the northern San Juan basin in northwestern New Mexico; the Toreva is considered of middle to late Carlile age, and the Wepo and Yale Point of Niobrara age (O'Sullivan and others, 1972).
The Toreva Formation, from 42 to 100 m (140 to 325 ft) thick, consists of three members: (1) a lower sandstone member composed of cliff-forming cross-stratified sandstone of littoral marine origin; (2) a middle carbonaceous member, composed of slope-forming variegated siltstone, cross-stratified sandstone, dark carbonaceous mudstone and coal, that pinches out in the northern part of the Mesa; and (3) an upper sandstone member composed dominantly of cliff-forming fluvial and littoral marine sandstone and intertongued marine shale.

The Wepo Formation, from about 90 to 230 m (300 to 750 ft) thick, consists of slope-forming and ledge-forming gray siltstone, mudstone, sandstone, and coal of fluvial and paludal origin. The formation thins to the northeast owing to intertonguing with the Toreva Formation and Yale Point Sandstone.

The Yale Point Sandstone is present in the northeast part of Black Mesa but has been eroded away to the south. It is as much as 90 m (300 ft) thick and consists of cliff-forming, cross-stratified, light-colored sandstone of littoral marine origin.

REFERENCES


MESOZOIC VERTEBRATES OF NORTHERN ARIZONA

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FIGURE

1. Left hand page—Mesozoic formations of northern Arizona and their correlations with certain Mesozoic horizons in other parts of North America, and in Europe, Asia, and Africa

2. Right hand page—Characteristic fossil vertebrates from the Mesozoic formations of northern Arizona
ABSTRACT

Four Triassic horizons of continental origin, the Moenkopi, Chinle, Moenave, and Kayenta Formations, contain amphibians and reptiles. A few fragmentary reptilian remains have been found in the continental Triassic and Jurassic Navajo and Jurassic Morrison Formations. The one Cretaceous formation containing fossil vertebrates is the Mancos Shale, with marine fishes and reptiles. The Mesozoic vertebrates of northern Arizona closely resemble forms in other parts of the world, indicating avenues for the wide dispersal of land-living animals during this phase of earth history.

INTRODUCTION

The Mesozoic rocks of northern Arizona contain a considerable fossil record of vertebrate life, particularly of the land-living back-boned animals, during a time span that began in the early phases of Triassic history and continued well into the Cretaceous. Of the 17 stratigraphic units representing the Mesozoic depositional record in the northern part of the state, about half contain fossil vertebrates. These particular formations are so distributed within the Triassic, Jurassic, and Cretaceous periods that in spite of gaps in the sequence the fossils do give some concept as to the succession of vertebrate life in this region. The beds containing vertebrate fossils, from earliest to latest, are the Triassic Moenkopi, Chinle, Moenave, and Kayenta Formations; the Triassic and Jurassic Navajo and Jurassic Morrison Formations; and the Cretaceous Mancos Formation and Mesa Verde Group.

Many of the fossils found in the rock units of northern Arizona are closely related to and commonly identical with fossils found in other parts of western and eastern North America, and with fossils found as far distant as Europe, Asia, and Africa. The land-living animals of Mesozoic time were not restricted by our imposed geographical boundaries, especially at a stage of earth history when continents now far separated from each other were intimately connected, offering broad avenues for intercontinental dispersals.

TRIASSIC VERTEBRATES

Moenkopi Formation

The lowest Triassic unit in northern Arizona, the widely distributed and colorful Moenkopi Formation, is noteworthy because it contains an abundance of ripple marks and tetrapod footprints—clear cut evidence of fresh-water sediments, deposited in shallow streams and lakes. The footprints represent various land-living vertebrates, but a majority of them belong to the genus Chirotherium, named more than a century ago from footprints and trackways found in Germany.
These tracks represent a rather large thecodont reptile, the thecodonts being Triassic reptiles that evolved along several lines of adaptive radiation and were particularly important in that some were the direct ancestors of the dinosaurs, the crocodilians, the flying reptiles or pterosaurs, and the birds. (A recently expounded theory would remove the birds from a direct thecodont ancestry and would derive them from certain early dinosaurs (Ostrom, 1973).) Chirotherium was a quadrupedal reptile, as shown by the distinct fore and hind footprints; it walked with the feet well under the body. It was probably carnivorous, if one accepts the evidence of a carnivore skeleton, Ticinosuchus, from the Triassic of Switzerland, in which the bony structure of the feet match very nicely the tracks of Chirotherium.

Trackways and footprints are fairly common in the Moenkopi, but bones are rare. This is commonly the case in the fossil record; tracks and bones do not seem to go together, which suggests that conditions favoring the preservation of one type of evidence were not conducive to the preservation of the other. A few fragmentary reptilian bones (as well as some fish spines and scales) have been found in the Moenkopi Formation; however, the fossils of real consequence are those of labyrinthodont amphibians. The labyrinthodonts, characteristic of the late Paleozoic and the Triassic, are solid-skulled amphibians in which commonly only the openings for the eyes and the nostrils, and a median pineal opening between the eyes, pierce the heavy skull roof. Moreover, although the skull roof in these amphibians is very rugose, the bony rugosities are traversed by smooth canals between the eyes and across the back of the skull. These are the so-called "slime-canals," remnants of the sensory lateral line system of the fishes from which the labyrinthodont amphibians were descended. The limbs in the labyrinthodonts are short and commonly weak. Evidently these amphibians spent most of their time in shallow waters, where they probably preyed upon fresh-water fishes.

The presence of Chirotherium tracks and labyrinthodont amphibian remains, so characteristic of the Moenkopi Formation, is also typical of the Buntsandstein, the type Triassic of central Europe. Parotosaurus, a long-skulled Moenkopi amphibian, is comparable to amphibians in the Buntsandstein. Hadrokkosaurus, a Moenkopi labyrinthodont with a short, broad skull, is closely related to similar forms in southern Africa, India, and Australia. Still other Moenkopi amphibians, Cyclotosaurus and Rhadalognathus, may be compared with Cyclotosaurus from the Middle and Late Triassic of Europe. Therefore it is reasonable to think that the Moenkopi beds, partly of Early Triassic age, extend upward into the lower part of the Middle Triassic.

Chinle Formation

Whereas fossil bones are difficult to find in the brownish-red sandstones of the Moenkopi, they are widely preserved in the non-resistant and many-hued siltstones, sandstones, and clays of the Chinle Formation. The Painted Desert, which crosses northern Arizona obliquely, from the Petrified Forest at the southeast to the borders
of the Grand Canyon, is made up largely of rocks of the Chinle. The associated fossils provide a record of a varied fauna, of an assemblage of animals living in and along the borders of rivers and lakes, and on the higher ground between the watercourses.

Various fishes are found in the Chinle assemblage, but most of them beyond the borders of Arizona, in Utah and Colorado, and in Texas. These fishes are represented by heavily armored chondrosteans and holosteans, related in a general way to the modern sturgeons and garpike, respectively, and by fresh-water lobe-finned coelacanths, of which the modern survivor is the famous *Latimeria*, inhabiting oceanic waters off the east coast of Africa. In Arizona are found teeth of *Ceratodus*, a lungfish closely related to the modern lungfish, *Neoceratodus*, found in the rivers of Queensland, Australia.

Chinle amphibians are represented by the large labyrinthodont, *Metoposaurus*, the adults of which have huge, massive skulls, some of them over 60 cm (2 ft) long. These amphibians, among the last of the labyrinthodonts, attained a length of about 2 m (6 to 8 ft). They had partially cartilaginous skeletons and must have spent most of their time in the water.

Numerous reptilian forms occur in the Chinle. There are several thecodonts: a small, lightly built, bipedal carnivore, *Hesperosuchus*, found near Cameron; large, even gigantic crocodile-like phytosaurs, represented by the genus *Phytosaurus*; and large, heavily armored thecodonts, represented by the genus *Desmatosuchus*. The last two also have been found near Cameron, and one particularly noteworthy phytosaur skull, of gigantic proportions, indicates a reptile that in life must have been more than about 9 m (30 ft) long. It should be emphasized that the phytosaurs were not crocodiles, nor even ancestors of the crocodiles, in spite of their similarities to the modern reptiles of the tropics.

To continue the roster of Chinle reptiles, *Coelophysis*, an early theropod dinosaur, is particularly well known, especially from skeletons found in New Mexico. This is a small, delicately constructed bipedal dinosaur, clearly carnivorous and obviously quite agile. Finally, there are very large, heavy dicynodont reptiles (one branch of the therapsids or mammal-like reptiles), represented by the genus *Placerias*, abundantly preserved in a highly fossiliferous zone near St. Johns, Arizona.

Such is the Chinle fauna, but it is not exclusive to northern Arizona nor even to the Chinle Formation. Many of these same fossils are found in the Chinle Formation of Utah and New Mexico, in the Dockum Formation of Texas, in the Popo Agie Formation of Wyoming, and in the Newark Group of the eastern seaboard. Moreover, the combination of metoposaurs, phytosaurs, and early dinosaurs characterizes the type Keuper of central Germany and extends on to the Maleri beds of peninsular India. It is a characteristic Keuper assemblage, and its occurrence across 200 degrees of modern longitude and perhaps 40 degrees of latitude is a good example of the fact that fossils and formations are not necessarily coextensive.
Moenave Formation

A few fossils are known from the Moenave Formation, notably the trackways of theropod dinosaurs and some skeletons of an early crocodilian, found near Dinosaur Canyon. This ancestral crocodilian, placed in the genus Protosuchus, was a relatively small reptile, no more than about 1 m (3 ft) long. Nevertheless it is characterized by undeniable crocodilian anatomical features, which clearly distinguish it from its phytosaurian cousins. It is relatively long-legged and would seem to have been perhaps more terrestrial in habits than were the later crocodilians. Related genera are known from the Upper Triassic of South Africa and South America.

Kayenta Formation

The Kayenta Formation is known to be fossiliferous but as yet is incompletely explored. Most of the fossils that have been found are undescribed. Twenty years ago S. P. Welles (University of California at Berkeley) unearthed a well-preserved skeleton of a good-sized carnivorous theropod dinosaur near Tuba City. He named this dinosaur Dilophosaurus (Welles, 1954, 1970) because of the two longitudinal crests that characterize the skull. Other known fossils show that there were small theropod dinosaurs as well in the Kayenta assemblage. A skeleton of an armored ornithischian dinosaur has recently been discovered, seemingly the first Triassic ornithischian to be found in North America, which shows strong resemblances to the genus Fabrosaurus from South Africa.

Small armored thecodont reptiles have been found in the Kayenta Formation, as well as fragments suggesting a large genus similar to Desmatosuchus of the Chinle. Perhaps one of the more significant discoveries in the Kayenta Formation is that of numerous tritylodont reptiles, found near Kayenta. These fossils, just below the contact between the Kayenta and the overlying Navajo Formation, are represented by a considerable number of skulls of varying ontogenetic ages, and of complete skeletons. The tritylodonts are among the most mammal-like of the mammal-like reptiles. The skeletons from Kayenta are about 1 m (3 ft) long and indicate animals of stocky build, with strong limbs and short tails. The skull is superficially rodent-like, with strong incisor teeth separated from the many-cusped molar teeth by a diastema or gap. These fossils are classified as of reptilian rather than of mammalian relationships because of the rather subtle mechanism of the joint between the skull and the lower jaw. The Kayenta tritylodonts in many respects closely resemble Tritylodon, from the Upper Triassic Red Beds of South Africa. Furthermore, these Arizona fossils are closely related to Bienotherium from the Upper Triassic Lufeng beds of western China. Recently, similar tritylodonts have been found in the Upper Triassic sediments of Argentina.
JURASSIC VERTEBRATES

Navajo Formation

The Navajo Formation is of both Triassic and Jurassic age, since it transcends the boundary between the two periods. For the purpose of simplicity in the present discussion, it will be considered to be of Jurassic age.

The Navajo is largely of dune sand origin, an environment in which fossil preservation is unlikely, and indeed, except for dinosaur tracks, only two vertebrate fossils are known. One consists of a partial skeleton of a small theropod dinosaur, described by Charles Camp (1936) as Segisaurus, which is a lightly built dinosaur, seemingly similar to its predecessor, Coelophysis, of the Chinle Formation. The other type consists of some partial hind feet and ribs of a much larger theropod, related to Ammosaurus, a genus from the Newark Group of Connecticut.

These fossils, inadequate as they are, give us a glimpse of life during Early Jurassic, when there undoubtedly was a variety of reptiles wandering across western landscapes. Unfortunately, few of the inhabitants of those ancient lands have been preserved.

Morrison Formation

The Morrison Formation is widely exposed throughout western North America. To the north, especially in Colorado, Utah, and Wyoming, the formation contains abundant remains of late Jurassic dinosaurs, together with turtles, lizards, and crocodilians, and the almost microscopic bones of early mammals. Much of the popular concept of dinosaurs is based upon Morrison forms such as Allosaurus, Brontosaurus, and Stegosaurus, the skeletons of which are to be seen in some of the larger museums of this country.

The Morrison Formation in northern Arizona has not been extensively explored. Isolated bones indicative of characteristic Morrison dinosaurs have been found, and recently John Cosgriff (Wayne State University at Detroit) collected some associated dinosaurian remains, which have not yet been studied.

There is every reason to think that the Morrison fauna in northern Arizona, when more fully known, will be essentially similar to that in other parts of western North America.
CRETACEOUS VERTEBRATES

Mancos Formation

In contradistinction to the other formations that have been discussed, the Mancos is a marine sequence of black shales characterized by a large invertebrate fauna, but fortunately containing vertebrate fossils as well. The vertebrates are fishes, turtles, crocodilians, and plesiosaurs.

Among the fishes are sharks, particularly the batoid, Ptychodus, of which the heavy, grinding teeth occur as fossils. Also, remains of the large teleost fish, Xiphactinus (generally known as Portheus), so beautifully preserved in the Kansas chalk beds, are known.

Reptilian fossils such as turtles and crocodiles are known from fragmentary remains, and a recent discovery in northeastern Arizona has yielded the skeleton of a plesiosaur. It is a pliosaur, perhaps close to, if not identical with, Trinacromerum from the Cretaceous of Kansas. As in the case of other Mesozoic formations in Arizona, future exploration is almost certain to enlarge our knowledge of the vertebrates in the Mancos.

Mesa Verde Group

The Mesa Verde Group, in Arizona, is divided into three formations (from oldest to youngest): The Toreva, Wepo, and Yale Point. A large dinosaur vertebra is the only fossil vertebrate found to date in the Mesa Verde sequence in Arizona, but it is likely that more bones will be found in the future. The Mesa Verde to the north and east has yielded dinosaur bones of Upper Cretaceous aspect, similar to the bones of dinosaurs found along the Red Deer River in Alberta.

CONCLUSIONS

From the foregoing brief account it is evident that northern Arizona contains a record, in part well documented and in part fragmentary, of the vertebrates inhabiting this part of the world during the Mesozoic Era. Where the exploration of the fossil-bearing formations has been persistent and extensive, as in the case of the Triassic beds, the record is well documented and revealing. Where the search for fossil vertebrates has not been systematically pursued, particularly in the Jurassic and Cretaceous, the evidence is at best fragmentary. It is to be expected, however, that much more fossil material can be obtained in the future from the Mesozoic sediments of northern Arizona, which will expand greatly our knowledge of vertebrate life, particularly of life on the land, during the Mesozoic years.

Certainly the Triassic studies that have been carried on in northern Arizona provide a vision of abundant life on the land. Moreover, the fossils show that Arizona at that time was part of a larger world in which many land-living amphibians and reptiles were
Figure 1.—On next pages, left hand page—Mesozoic formations of northern Arizona and their correlations with certain Mesozoic horizons in other parts of North America, and in Europe, Asia, and Africa.

Right hand page—Characteristic fossil vertebrates from the Mesozoic formations of northern Arizona.

Moenkopi:  
M1 - Chirotherium, footprints made by a thecodont reptile;  
M2 - Parotosaurus;  
M3 - Hadrokkosaurus, labyrinthodont amphibians.

Chinle:  
C1 - Metoposaurus, a labyrinthodont amphibian;  
C2 - Phytosaurus, a thecodont reptile;  
C3 - Coelophysis, a theropod dinosaur;  
C4 - Placerias, a dicynodont therapsid reptile.

Moenave:  
MA - Prootosuchus, an ancestral crocodilian.

Kayenta:  
K1 - Dilophosaurus, a theropod dinosaur;  
K2 - a tritylodont reptile.

Navajo:  
N - a large Triassic dinosaur, similar to Ammosaurus.

Morrison:  
Mo - Brontosaurus, a characteristic Morrison dinosaur.

Mancos:  
Mn1 - teeth of Ptychodus, a batoid shark;  
Mn2 - Xiphacanthus (Portheus), a very large teleost fish;  
Mn3 - Trinacromerum, a plesiosaur.

M1 - from Peabody  
M2 - from Welles and Cosgriff  
M3 - from Welles and Estes  
C1 - from Sawin  
C2 - from McGregor  
C3 - from Colbert  
C4 - from Camp and Welles  
MA - from Colbert and Mook  
K1 - from Welles  
K2 - from Young  
N - from Marsh  
Mo - from Gilmore  
Mn1 - from Woodward  
Mn2 - from Osborn  
Mn3 - from Williston
distributed across wide ranges of latitude and longitude. Thus the fossil vertebrates of northern Arizona are of interest, not only because of the light they shed upon the Mesozoic inhabitants of this region, but also because of the evidence they provide concerning continental relationships at a time when Pangaea was breaking apart but yet was not so fragmented as to preclude the wide distributions of closely related terrestrial vertebrates across its vast surfaces.

SELECTED REFERENCES


CENOZOIC GEOLOGY, PALEOCLIMATE, AND ARCHAEOLOGY

K-Ar chronology for the San Francisco volcanic field and rate of erosion of the Little Colorado River
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Cenozoic volcanism and tectonism of the southern Colorado Plateau
by
Raymond L. Eastwood

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by
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by
Robert C. Euler and George J. Gumerman

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K-AR CHRONOLOGY FOR THE SAN FRANCISCO VOLCANIC FIELD AND
RATE OF EROSION OF THE LITTLE COLORADO RIVER

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FIGURE

Figure
1. Chronology of volcanism and terrace formation and sequence of geologic events within the San Francisco volcanic field .................. 233

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ABSTRACT

K-Ar data and ages are given for 22 basaltic and 9 silicic volcanic rocks from the San Francisco volcanic field. These dates and 4 from other laboratories are used to calculate the rate of downcutting by the Little Colorado River and to develop a chronology of volcanism, terrace formation, and Cenozoic depositional events.

An average rate of downcutting of probably 100 m (315 ft) per million years, over the last 2.4 m.y., is derived for the Black Point and Wupatki erosion cycles. Most of the erosion of the valley of the Little Colorado River probably occurred before the Bidahochi lake beds were deposited and during the Hopi-Zuni erosion cycles of Hemphillian age.

The chronology presented in this paper is consistent with the geomorphic classification of volcanic stages derived by Cooley (1962).

INTRODUCTION

Robinson (1913) was the first to subdivide the rocks in the San Francisco volcanic field chronologically. He divided the rocks into three periods of eruption: 1) Basaltic volcanism, to which he assigned a probable age of late Pliocene. 2) Eruption of lavas, ranging in composition from andesites to rhyolites, associated with six large volcanic cones and a somewhat greater number of small cones and laccoliths. An early Quaternary age was assigned to these silicic volcanic rocks. 3) Basaltic volcanism during the latter part of the Quaternary period.

Colton (1936), while director of the Museum of Northern Arizona, constructed a temporal classification of the basaltic volcanism based on geomorphologic criteria, primarily extent of erosion and weathering. His classification consists of five stages. The earliest stage includes basaltic lava flows, such as the flow on Switzer Mesa, which could be demonstrated to be older than the rocks of the San Francisco Peaks. The fifth and youngest stage includes Sunset Crater, to which he assigned an age of ca. A.D. 1067 (Colton, 1945).

Cooley (1962), applying the concept of correlatable erosional terraces and surfaces to the distribution of volcanic rocks, redefined Colton's classification. Cooley placed rocks that predate the formation of the San Francisco Peaks in a separate category, called older basalts. Colton's remaining stage I basalts were divided by Cooley into an older Black Point sub-stage (IB) and a younger Woodhouse Mesa sub-stage (IA). Cooley's classification has the distinct advantage that some of the erosional terraces and associated volcanic rocks can be related to dated fossil localities. Thus, Cooley was able to assign a Hemphillian age to the older basalts, a Blancan (?) age to the stage I basalts, and a Quaternary age to the remaining four stages.
Damon (1965) applied the first available radiometric dates to the volcanic sequences within the southern margin of the Colorado Plateau. He assigned the older basalts to Hemphillian time and the stage I basalts to Blancan time in agreement with Cooley's age assignments. However, at that time, insufficient radiometric data were available to enable correct assignment of ages to the younger stages.

This paper presents additional data and suggests a more refined chronology for the San Francisco volcanic field. This chronology is based on 22 K-Ar dates for basaltic rocks and nine K-Ar dates for silicic volcanic rocks measured in this laboratory. With one exception, the dated volcanic rocks lie within a region bounded by the 35th and 36th parallels and the 111th and 112th meridians. The one exception is Red Butte, which is on the Coconino Plateau, 7 1/2 miles west of longitude 112°W.

ACKNOWLEDGMENTS

This work was supported by National Science Foundation Grant GA-31270, Atomic Energy Commission Contract AT(ll-1)-689, and the State of Arizona. We are grateful to Maurice Cooley, Everett Lindsay, Richard Moore, Eugene Shoemaker, George Ulrich, Robert Sutton, and Edward Wolfe for helpful discussion. The analytical work took place over a decade, and a number of people assisted in it. We are particularly grateful to Judith Jenney, Peter Kresan, William Laughlin, Kathleen Roe, Robert Scarborough, and Anne Sigleo for their assistance. We are also grateful to Raymond Eastwood, Esther Holm, Richard Moore, Bruno Sables, George Ulrich, and Edward Wolfe for their assistance in sample collection.

K-AR DATA AND ANALYTICAL TECHNIQUES

The radiometric dates reported here have been obtained over the last decade. During this time, improvements in the analytical techniques have enabled us to date geologically young samples with increased precision. The techniques for isotopic dilution analysis for argon and flame photometric determination of potassium which were used until 1969 are described by Livingston and others (1967).

The potassium content was determined by both atomic absorption spectrophotometry and flame photometry in 1971-72 and by atomic absorption spectrophotometry since July 1972. The atomic absorption analyses give superior results on standard samples. The same solution was used for both atomic absorption and flame photometric determination of potassium (0.1 g sample as sulphate in 200 ml of aqueous solution containing 800 ppm Li and 1000 ppm Na). The atomic absorption spectrophotometer used is a Perkin-Elmer 403, single-beam instrument; the flame photometer used was a dual-beam modified Perkin-Elmer model 146. For the atomic absorption measurements, a 30-60 acetylene-air flame was used with the wavelength dial at 3827 Å (3-slot burner head in parallel position, slit 4, filter in).
Argon isotopes have been analyzed by the Nier type 6-inch, 60° sector field, mass spectrometer using dynamic mode until October 1970, both static and dynamic modes on aliquots of the gas sample from a single fusion until May 1973, and by static mode (2 analyses per fusion) since then. The static mode is about 300 times more sensitive than the dynamic mode and is helpful in analyzing geologically young samples.

Alundum crucibles were used until early 1973 as heat shields and also to contain any melt oozing out of the molybdenum crucible at high temperatures. We noted that some batches of alundum crucibles contained more atmospheric argon than others. Prebaking at 1350° did not release all the atmospheric argon from some alundum crucibles. Use of these crucibles resulted in poor precision and, consequently, inaccurate ages.

It may be noted that the standard error of a K-Ar date is primarily the result of atmospheric argon contamination. For a single determination the standard error is 2.3 percent with negligible air argon, 15 percent with 90 percent air argon, and 100 percent with 98.4 percent air argon.

We now use larger pyrex fusion envelopes (90 mm instead of 58 mm) instead of alundum crucibles. We mount four samples in the fusion system and bake the system over the weekend at 260°C. We also bake the fusion system overnight after every fusion. These modifications reduce the atmospheric argon contamination significantly. In addition, the NRC model 214 appendage pump in the inlet system removes the active gases and ensures pure argon for mass spectrometric analyses. We repeated a number of previous analyses with high atmospheric contaminations. The recent values for these samples supersede the older values.

The analytical data for the basaltic and silicic rocks from the San Francisco volcanic field are presented in tables 1 and 2, respectively. The uncertainty reported with any K-Ar age (tables 1 and 2) is the standard error (one sigma). Each rock sample is also identified in terms of Cooley's (1962) basaltic stage, or silicic volcanic stage (Cooley, oral commun. 1973), volcanic age-group (Moore and others, this volume), and the terrace or surface on which it rests (Cooley and others, 1969). The presence of a flow on a particular surface implies only that it postdates the surface formation, and the two may not be exactly correlatable. The lava may have flowed onto the surface just after its formation or at a much later time.

In addition to the data in tables 1 and 2, McKee (1973) has dated the basalt exposed in the Rio de Flag Canyon, adjacent to the Museum of Northern Arizona, at 6.0 ±0.3 m.y. McKee and McKee (1972) also report an age of 8.0 ±0.3 m.y. for a basalt flow that covers a gravel deposit in a large channel exposed in the east wall of Oak Creek Canyon. Smiley (1958) dates a stage V eruption from Sunset Crater at A.D. 1064-1065 by dendrochronologic methods. Lastly, Baksi (oral commun.) has determined a K-Ar isochron age of 0.071 ±0.004 m.y. on 18 drill core samples from the SP Flow.
### Table 1.—K-Ar ages on basaltic rocks from the San Francisco volcanic field, Coconino County, Arizona.

<table>
<thead>
<tr>
<th>Serial No.</th>
<th>Sample No.</th>
<th>Description</th>
<th>Percent K</th>
<th>Radiogenic argon-40 in $10^{-12}$m/g</th>
<th>Percent atmospheric argon-40</th>
<th>Age in m.y.</th>
</tr>
</thead>
<tbody>
<tr>
<td>B-1</td>
<td>UAKA-71-37</td>
<td>Whole rock, basalt from 3 m above the base of the flow capping Red Butte, elev. 7200 ft. The flow is on the Valencia surface (Cooley, 1962) and is an older stage basalt. Red Butte quad., lat. 35°49'10&quot; N., long. 112°05'23&quot; W.</td>
<td>0.95₂</td>
<td>16.82</td>
<td>85.5</td>
<td>9.40±0.91</td>
</tr>
<tr>
<td>B-1a</td>
<td>UAKA-71-37</td>
<td>Whole rock, basalt, second sampling from the same site as B-1.</td>
<td>0.95₂</td>
<td>14.7</td>
<td>61.6</td>
<td>8.68±0.23</td>
</tr>
<tr>
<td>B-2</td>
<td>UAKA-71-38</td>
<td>Whole rock, basalt that caps a resistant bench of Red Butte; probably a slump block, elev. 6400 ft. Older basalt stage. Red Butte quad., lat. 35°48'51&quot; N., long. 112°05'32&quot; W.</td>
<td>1.02</td>
<td>16.18</td>
<td>95.8</td>
<td>8.60±0.22</td>
</tr>
<tr>
<td>B-3</td>
<td>UAKA-71-4</td>
<td>Whole rock, basalt, oldest of 8 lava flows in Volunteer Canyon; overlies Kaibab Limestone. Elev. 6150 ft, Sycamore Point quad., lat. 35°07' N., long. 111°56'30&quot; W.</td>
<td>0.39₇</td>
<td>6.57</td>
<td>89.4</td>
<td>8.68±0.98</td>
</tr>
<tr>
<td>B-4</td>
<td>PED-15-64</td>
<td>Whole rock, basalt collected 10 m below the Ohio State Univ. observatory on Anderson Mesa. Older basalt stage. Elev. 7170 ft, Lower Lake Mary quad., lat. 35°06'54&quot; N., long. 111°33'20&quot; W.</td>
<td>0.89₀</td>
<td>9.88</td>
<td>79.6</td>
<td>6.00±0.30</td>
</tr>
<tr>
<td>B-5</td>
<td>PED-54-66</td>
<td>Whole rock, basal basalt flow in Oak Creek Canyon at Highway 89, switchbacks. Older basalt stage. Elev. 5700 ft, Mountaineer quad., lat. 35°01'33&quot; N., long. 111°44'22&quot; W.</td>
<td>1.00₅</td>
<td>10.5</td>
<td>87.2</td>
<td>5.88±0.91</td>
</tr>
<tr>
<td>B-6</td>
<td>UAKA-72-26</td>
<td>Whole rock, Woody Mtn. olivine basalt. This flow is displaced by the Oak Creek fault. Older basalt stage. Elev. 7430 ft, Bellemont quad., lat. 35°09'11&quot; N., long. 111°45'08&quot; W.</td>
<td>0.62₇</td>
<td>6.54</td>
<td>83.8</td>
<td>5.86±0.50</td>
</tr>
<tr>
<td>B-7</td>
<td>UAKA-73-118</td>
<td>Whole rock, basalt from Middle Switzer Mesa. Older basalt stage. Elev. 7020 ft, Flagstaff West quad., lat. 35°12'03&quot; N., long. 111°37'34&quot; W.</td>
<td>0.43₉</td>
<td>4.65</td>
<td>78.1</td>
<td>5.80±0.34</td>
</tr>
<tr>
<td>B-8</td>
<td>UAKA-73-155</td>
<td>Whole rock, basalt from Cedar Ranch mesa. Older basalt stage. Elev. 6760 ft, Ebert Mtn. quad., lat. 35°32'24&quot; N., long. 111°45'10&quot; W.</td>
<td>0.51₈</td>
<td>5.10</td>
<td>67.9</td>
<td>5.49±0.19</td>
</tr>
<tr>
<td>B-9</td>
<td>UAKA-71-3</td>
<td>Whole rock, basalt; 4th flow from the bottom in a sequence of 8 flows in Volunteer Canyon. It underlies the 50 m bedded ash horizon. Older basalt stage. Elev. 6280 ft, Sycamore Point quad., lat. 35°07' N., long. 111°56'30&quot; W.</td>
<td>0.93₆</td>
<td>8.02</td>
<td>57.6</td>
<td>4.75±0.23</td>
</tr>
</tbody>
</table>

226
<table>
<thead>
<tr>
<th>Sample Code</th>
<th>Description</th>
<th>Dates</th>
<th>Other Information</th>
</tr>
</thead>
<tbody>
<tr>
<td>B-10 UAKA-71-1</td>
<td>Whole rock, basalt; top flow of a series of 8 flows in Volunteer Canyon. This flow is continuous with a dike that crosscuts the sequence of lava flows. Older basalt stage. Elev. 6680 ft, Sycamore Point quad., lat. 35°07’ N., long. 111°56’30” W.</td>
<td>0.55, 4.26</td>
<td>84.9, 4.30±0.45</td>
</tr>
<tr>
<td>B-11 PED-7-70</td>
<td>Whole rock, basalt; Amphitheatre dike that intrudes palagonite tuff. Older basalt stage, Woodhouse age group. Elev. 5520 ft, SP Mtn. quad., lat. 35°39’30” N., long. 111°34’51” W.</td>
<td>1.04, 5.65</td>
<td>85.1, 3.05±0.39</td>
</tr>
<tr>
<td>B-12 PED-6-65</td>
<td>Whole rock, basalt; plagioclase phenocrysts excluded. Black Point flow on IB surface (Cooley and others, 1969) near Little Colorado River. IB stage, Woodhouse age group. Elev. 4840 ft, Black Falls NE quad., lat. 35°40’20” N., long. 111°20’39” W.</td>
<td>0.85, 3.59</td>
<td>2.39±0.32 (Damon and others, 1967)</td>
</tr>
<tr>
<td>B-13 PED-7-67</td>
<td>Whole rock, basalt; Amphitheatre flow that overlies the palagonite tuff of B-12. Stage IB, Woodhouse age group. Elev. 5450 ft, SP Mtn. quad., lat. 35°39’32” N., long. 111°35’03” W.</td>
<td>1.05, 3.44</td>
<td>54.7, 1.83±0.07</td>
</tr>
<tr>
<td>B-14 UAKA-73-116</td>
<td>Whole rock, basalt from Lava Point flow. IA stage, Woodhouse age group. Elev. 5420 ft, SP Mtn. quad., lat. 35°39’52” N., long. 111°31’14” W.</td>
<td>0.54, 0.91</td>
<td>96.5, 0.90±0.30</td>
</tr>
<tr>
<td>B-15 UAKA-73-119</td>
<td>Whole rock, vesicular basalt from Wukoki flow on Wupatki surface. II stage, Woodhouse age group. Elev. 4934 ft, Wupatki grad., lat. 35°30’01” N., long. 111°19’08” W.</td>
<td>0.81, 1.23</td>
<td>92.3, 0.87±0.14</td>
</tr>
<tr>
<td>B-16 UAKA-73-154</td>
<td>Whole rock, basalt from Kellam Ranch flow. IA stage, Woodhouse age group. Elev. 5400 ft, Merriam Crater quad., lat. 35°21’35” N., long. 111°17’54” W.</td>
<td>0.93, 1.62</td>
<td>89.6, 0.84±0.13</td>
</tr>
<tr>
<td>B-17 UAKA-73-153</td>
<td>Whole rock, basalt from flow on Woodhouse Mesa, one mile west of road, near Wupatki National Monument. IA stage, Woodhouse age group. Elev. 5260 ft, Roden Crater quad., lat. 35°29’15” N., long. 111°22’01” W.</td>
<td>1.03, 1.45</td>
<td>81.7, 0.786±0.059</td>
</tr>
<tr>
<td>B-18 UAKA-73-123</td>
<td>Whole rock, basalt from Vent 171 flow. Stage II, Tappan age group. Sampled near the benchmark. Elev. 5540 ft, Roden Crater quad., lat. 35°23’03” N., long. 111°20’13” W.</td>
<td>0.82, 0.98</td>
<td>92.7, 0.67±0.10</td>
</tr>
<tr>
<td>B-19 UAKA-73-23</td>
<td>Whole rock, basalt. One lobe of this flow of Shadow Mountain has been cut by a graben. Stage II, Tappan age group. Elev. 4300 ft, Cameron NW quad., lat. 35°59’46” N., long. 111°26’50” W.</td>
<td>0.36, 0.43</td>
<td>97.6, 0.62±0.23</td>
</tr>
<tr>
<td>Sample</td>
<td>Location Details</td>
<td>Description</td>
<td>Age (Ma) ± Error</td>
</tr>
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<td>------------------</td>
<td>-------------</td>
<td>------------------</td>
</tr>
<tr>
<td>B-20 UAKA-72-62</td>
<td>Whole rock, basalt from Tappan flow at the Little Colorado River. This flow has been cut by a graben trending NE at Moenkopi Wash and occupies channel of the ancestral Little Colorado River. Stage II, Tappan age group. Elev. 4300 ft, Cameron NW quad., lat. 35°53'14&quot; N., long. 111°27'18&quot; W.</td>
<td>1.20 ± 0.079</td>
<td>91.3</td>
</tr>
<tr>
<td>Serial No.</td>
<td>Sample No.</td>
<td>Description</td>
<td>Percent K</td>
</tr>
<tr>
<td>-----------</td>
<td>------------</td>
<td>-------------</td>
<td>-----------</td>
</tr>
<tr>
<td>S-1</td>
<td>PED-15-70</td>
<td>Sanidine from rhyodacite with sanidine and quartz phenocrysts from North Sugarloaf Mtn. Stage I, Woodhouse age group. Elev. 8260 ft, O'Leary Peak quad., lat. 35°22'42&quot; N., long. 111°36'42&quot; W.</td>
<td>6.74</td>
</tr>
<tr>
<td>S-2</td>
<td>BES-58-111</td>
<td>Biotite from porphyritic rhyodacite from North Sugarloaf Mtn. Stage I, Woodhouse age group. O'Leary Peak quad., lat. 35°22'42&quot; N., long. 111°36'42&quot; W.</td>
<td>6.04</td>
</tr>
<tr>
<td>S-3</td>
<td>BES-58-315</td>
<td>Biotite from tuffaceous unit, Sitgreaves Peak. Stage I, Woodhouse age group. Parks quad., lat. 35°21' N., long. 111°59' W.</td>
<td>5.51</td>
</tr>
<tr>
<td>S-4</td>
<td>UAKA-72-27</td>
<td>Whole rock, aphanitic latite near summit of Woody Mtn. Stage I, Woodhouse age group. Elev. 7940 ft, Flagstaff W. quad., lat. 35°08'35&quot; N., long. 111°44'57&quot; W.</td>
<td>3.22</td>
</tr>
<tr>
<td>S-5</td>
<td>UAKA-73-115</td>
<td>Whole rock, rhyolite from near top of Slate Mtn. Stage I, Woodhouse age group. Elev. 8130 ft, Kendrick Peak quad., lat. 35°52'09&quot; N., long. 111°50'34&quot; W.</td>
<td>3.75</td>
</tr>
<tr>
<td>S-6</td>
<td>UAKA-72-74</td>
<td>Sanidine from riebeckite rhyolite of the San Francisco peaks. Stage I, Woodhouse age group. Elev. 10860 ft, Humphrey Peak quad., lat. 35°19'40&quot; N., long. 111°40'00&quot; W.</td>
<td>4.16</td>
</tr>
<tr>
<td>S-7</td>
<td>UAKA-73-114</td>
<td>Biotite, White Horse Hills rhyolite. Stage I, Woodhouse age group. Elev. 9060 ft, White Horse Hills quad., lat. 35°23'24&quot; N., long. 110°42'12&quot; W.</td>
<td>6.50</td>
</tr>
<tr>
<td>S-8</td>
<td>UAKA-71-8</td>
<td>Sanidine phenocrysts from O'Leary rhyodacite collected from Forest Service road cut on elevation contour 8300 ft. Stage 3, Tappan age group. O'Leary Peak quad., lat. 35°23'50&quot; N., long. 111°31'36&quot; W.</td>
<td>8.94</td>
</tr>
<tr>
<td>S-9</td>
<td>PED-14-70</td>
<td>Sanidine, rhyolite from South Sugarloaf Mtn. Stage 3, Tappan age group. Elev. 8900 ft, Sunset Crater West quad., lat. 35°23'50&quot; N., long. 111°31'36&quot; W.</td>
<td>9.07</td>
</tr>
</tbody>
</table>
RATE OF EROSION AND TERRACE FORMATION

Two of the dated basaltic lavas flowed into the bed of the Little Colorado River, diverting or damming the river. According to Colton (1936), the Tappan lava flow entered the bed of the Little Colorado River, flowed upstream nearly a mile to the neighborhood of Cameron, and downstream at least 9 miles. The Little Colorado River was diverted from its old bed by this lava flow. Dr. Raymond L. Eastwood, who collected the Tappan lava flow sample, estimates from field relations that there has been 50-55 m (170-180 ft) of downcutting since emplacement of the flow.

Colton also pointed out that river gravels are found on top of the eastern end of the Black Point flow. "At one time the flow must have dammed the Little Colorado River so that the river flowed over the lava flow before it cut a canyon around the end" (Colton, 1936, p. 15). The top of the Black Point flow is at an elevation of 4,930 ft. The bed of the Little Colorado River is now at an elevation of 4,230 ft. Consequently, the river has cut down about 215 m (700 ft) in the 2.4 m.y. since emplacement of this stage IB flow.

The data and the calculated erosion rates are given in table 3. The two erosion rates, 105 m (340 ft) per million years at Tappan Spring Canyon and 90 m (290 ft) per million years at Black Point, agree within the error of the measurements. Thus, as a first approximation, the rate of erosion during the last 2.4 m.y. is approximately 100 m (315 ft) per million years. Cooley and Wilson (1968) estimate that 175 m (575 ft) of downcutting occurred at the head of the gorge of the Little Colorado River near Cameron since the beginning of Pleistocene time. According to Evernden and Evernden (1970, p. 85), "if the base of the Pleistocene is to be equated with the base of the Calabrian or with marine events occurring concurrently with deposition of the base of the Calabrian, the base of the Pleistocene will be approximately 1.8 m.y." Based on Cooley's estimate of the amount of downcutting and the Everndens' estimate of the duration of the Pleistocene, the rate of downcutting has been 100 m (320 ft) per million years, in excellent agreement with our estimates.

Table 3.—Downcutting by Little Colorado River.

<table>
<thead>
<tr>
<th>Location</th>
<th>Amount of downcutting (m)</th>
<th>Time elapsed (K-Ar, m.y.)</th>
<th>Rate of erosion (m/m.y.)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1 mile N.W. on confluence of Tappan Spring Canyon with Little Colorado Gorge (see sample no. B-20, table 1).</td>
<td>50-55</td>
<td>0.51</td>
<td>105</td>
</tr>
<tr>
<td>Black Point (see sample no. B-12, table 1).</td>
<td>215</td>
<td>2.4</td>
<td>90</td>
</tr>
</tbody>
</table>
It is tempting to extrapolate this rate of erosion into the Pliocene. However, geologic relations in the upper Little Colorado River do not allow such an extrapolation. For example, Repenning and others (1958), (fig. 3) have shown that the lower lacustrine member of the Bidahochi Formation crops out on both sides of the Little Colorado River between Holbrook and St. Johns. Within the Adamana 3 NE 7-1/2 minute quadrangle, the base of the Bidahochi is at an elevation of 5,580 ft. At the point nearest the outcrop, 3 miles away, the bed of the Little Colorado River is at an elevation of 5,310 ft. Thus, the net erosion since the deposition of the Bidahochi Formation is only 270 ft (80 m), and the river had cut down to an elevation of 5,580 ft in Hemphillian time. The gross amount of erosion is of course much greater. According to Dr. Everett Lindsay of the University of Arizona, near White Cone, the top of the lake beds extends to at least 6,300 ft and probably to 6,500 ft. The top of the lower member is at a similar elevation near Sanders. Thus, the gross amount of erosion, since deposition of the Bidahochi lake beds at these two localities, is about 1,200 ft (365 m).

The Little Colorado River must have already cut more than 1,500 ft (460 m) below the Anderson Mesa-Switzer Mesa surface (elevation about 7,100 ft) before the Bidahochi lake beds were deposited. Clearly, the Black Point and Wupatki erosion cycles postdate a period of deposition and must be considered a separate geologic episode.

CHRONOLOGY OF VOLCANISM AND SEQUENCE OF GEOLOGIC EVENTS

The K-Ar dates in tables 1 and 2, combined with erosion rates for the Black Point and Wupatki erosion cycles, have been used to evaluate the chronology of volcanism presented in figure 1. A new classification of basaltic volcanism (Moore and others, this volume) is presented along with Cooley's (1962) classification. Cooley (oral commun.) has also classified the stages of silicic volcanism. His stages, 1-3, correspond in time with the basaltic volcanic stages, I-III. We find no significant inconsistency in Cooley's classification of basalt stages. Cooley's stages IA and IB correspond closely to the Woodhouse age-group of Moore and others (this volume). The Tappan age-group includes Cooley's stage II and part of his stage III. The Merriam Crater age-group corresponds directly to Cooley's stage IV and the Sunset age-group to Cooley's stage V. There is no evidence for silicic volcanism during stages IV and V. Also, none of the silicic volcanic rocks which we dated fall into stage 2, although Cooley has placed the White Horse, Slate Mountain, Sitgreaves Peak and North Sugarloaf rhyolites in stage 2. Obviously, it is more difficult to place centers of silicic eruptions in a time sequence based on erosion surfaces than it is to classify the stratiform basaltic lavas that cap the erosion surfaces.

Cooley and others (1969) have estimated the height of the Black Point and Wupatki surfaces above the bed of the Little Colorado River in the Cameron-Winslow area. We have used their estimates and the
rates of erosion determined in this paper to place the formation of these terraces within a chronological sequence in figure 1 (column 5). Column 6 presents the reconstruction sequence of nonvolcanic geologic events within the San Francisco volcanic field modified from Cooley (1967).

According to Cooley and Akers, (1961), the ancestral Little Colorado River was flowing to the northwest, not far from its present course, during the Hopi-Zuni erosion cycles of Hemphillian age. We have presented evidence suggesting that the Little Colorado had already cut to more than 460 m (1,500 ft) below the Anderson Mesa-Switzer Mesa surface during the Hopi-Zuni erosion cycles. Subsequent to this time, deposition is indicated by a few exposures of gravel-bearing sediments associated with the older basalts. Judging from the state of erosion and height of the lake in the Hopi Buttes area, lacustrine deposition probably also took place within the valley of the Little Colorado River in the San Francisco volcanic field region. However, we know of no lake beds that have been positively identified.

Structural movements of presumed early Blancan age are indicated by normal faults that displace some of the older basalts. The Black Point erosion cycle occurred during mid-Blancan to early Irvingtonian time, leaving two prominent terraces between 120 to 150 m (400 to 500 ft) and 180 to 240 m (600 to 800 ft) above the present grade of the Little Colorado near Cameron (Cooley and others, 1969). The Black Point erosion cycle was followed by the deposition of gravels which locally are as much as 12 m (40 ft) thick. This episode was followed by the early Wupatki cycle, which left terraces at between 45 to 60 m (150 to 200 ft) and between 60 to 90 m (200 to 300 ft).

The occurrence of late-stage II structural movements is recorded by the graben, trending northeast-southwest, and cutting the Tappan lava flow at Moenkopi Wash (Rieche, 1937). Deposition of gravels with considerable sand and silt took place along tributary streams in mid-stage III time. Also, glacial outwash from the interior valley of the San Francisco Peaks (Updike and Pévé, this volume) postdates emplacement of the south Sugarloaf rhyolite, 210,000 years ago. Lastly, the late Wupatki erosion cycle that formed the modern valley of the Little Colorado River left erosion terraces at an altitude of 22 to 30 m (75 to 100 ft), 15 m (50 ft), and 9 m (30 ft) above the bed of the Little Colorado River in this region.

CONCLUSIONS

1. The chronology presented here, based on dated volcanic rocks and the rate of erosion during the Black Point and Wupatki erosion cycles, is consistent with the geomorphic classification of volcanic stages of Cooley (1962).
Figure 1.—Chronology of volcanism and terrace formation and sequence of geologic events within the San Francisco volcanic field. Darkened circles between the basalt stages represent K-Ar dates for basaltic lavas. Darkened circles on the right hand side of the silicic stages column represent K-Ar dates for silicic volcanoes (S.S.R. = South Sugarloaf Hill rhyolite, O.R. = O'Leary Peak rapakivi rhyodacite, M.H.G. = White Horse Hill rhyolite (formerly Marble Hill, R.R. = Riebeckite rhyolite, S.M.R. = Sitgreaves Peak rhyolite, N.S.R. = North Sugarloaf Hill rhyodacite). The logarithmic time scale is contracted by 20 percent for the stage IV and V basalt stages.
2. Most of the erosion in the valley of the Little Colorado River seemingly occurred during the Hopi-Zuni erosion cycles of Hemphillian age, which predated deposition of the Bidahochi lake beds. The grade of the Little Colorado River in the Black Point area immediately prior to the Black Point erosion cycle must have been at an elevation between 5,000 and 5,500 ft.

3. Erosion of the Little Colorado River bed during the Black Point to Wupatki erosion cycles occurred at an average rate of approximately 100 m (315 ft) per million years.

REFERENCES CITED


CENOZOIC VOLCANISM AND TECTONISM OF THE
SOUTHERN COLORADO PLATEAU

by

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ABSTRACT

Review of temporal, spatial, and petrological data for volcanic rocks of mid-Tertiary and Plio-Pleistocene age on or near the southern Colorado Plateau indicate that several systematic relations exist. The spatial occurrence of most volcanic fields define or coincide with structurally defined lineaments. Plio-Pleistocene volcanism occurred dominantly along a northeast-northwest lineament system. Geometric coincidence of these lineaments with axes and offsets of spreading centers of the East Pacific Rise in the Gulf of California suggests a tectonic causal relationship between Plio-Pleistocene volcanism of the southern Colorado Plateau and activity of these spreading centers.

Volcanic rocks of Plio-Pleistocene age of the southern Colorado Plateau are mainly alkaline, with subalkaline affinity. In contrast, volcanic rocks of mid-Tertiary age that rim the southern margin the Colorado Plateau are mainly calc-alkaline (or subalkaline) with alkaline affinity. The extent of subalkaline character of the volcanic rocks is correlative with higher strontium 87/86 ratios suggestive of admixture of crustal material.

The tensional tectonic environment for dominantly alkaline olivine basalt volcanism of Plio-Pleistocene age agrees with what has been learned about this volcanic episode in other areas of the western United States.

INTRODUCTION

The Cenozoic geology of the Colorado Plateau has been thoroughly reviewed by Hunt (1956), whose discussion of igneous activity noted that intrusive igneous rocks occur as stocks and laccoliths, and that volcanic rocks occur as large central volcanos and extensive sheets of basalt. Furthermore, laccoliths and large central volcanos are not found in areas adjacent to the Colorado Plateau. In reviewing the structural history, Hunt stated that upwarps, anticlines, and faults mainly trend northwestward and northward. These trends were apparently imposed by the pre-Cenozoic tectonic fabric of the region. Hunt also discusses the petrology of Cenozoic igneous rocks on the Colorado Plateau. The rocks are high in alkalis and aluminum, and Hunt (1956) subdivided them into four subprovinces on the basis of Na₂O/K₂O. Recent petrologic studies provide data relating to the petrogenesis of igneous rocks rimming the Colorado Plateau. Rock series of the southern Colorado Plateau have commonly been classed as alkaline, tholeiitic, calc-alkaline, or some combination thereof.

The tectonics of the southwestern United States and the Colorado Plateau have been discussed by Mayo (1958), Kelley and Clinton (1960), Damon and Mauger (1966), Atwater (1970), Damon (1971), and Scholz and others (1971). Damon and Mauger (1966), Atwater (1970), Damon (1971), Scholz and others (1971), and Lipman and others (1971, 1972) have emphasized the importance of relative motion of the North
American plate and East Pacific Rise to account for tectonism and volcanism in the western United States, particularly in the Basin and Range province. According to the plate-tectonic models, volcanism is related to subduction of oceanic crust and trench sediments along a Benioff zone. If the geometry of the East Pacific Rise and its offsets is correctly defined (Atwater, 1970), volcanism of the southern Colorado Plateau may not be consistent with these models (fig. 1); that is, the distance from the Colorado Plateau to the plate boundary seems excessive.

Lineament tectonics, independent of particular genetic models, provides a somewhat different perspective in that the occurrence of geologic features along linear trends may reflect deep-seated structural control of regional extent. Mayo (1958) and Kelley and Clinton (1960) emphasize the existence of lineaments defined by trends of structural elements such as faults, folds, or uplifts and/or location of centers of igneous activity. Four lineament directions can be classed into two systems: 1) a northwest-northeast-trending system, and 2) an east- and north-trending system (Mayo, 1958).

Recently, a greatly improved time scale for igneous activity in the southwestern United States has resulted from application of radiometric dating. Damon and Mauger (1966) and Damon (1971) have established the periodic nature of Late Cretaceous and Cenozoic magmatism in this region. These periods of magmatism are, 1) Laramide from 80 to 45 m. y., peaking at 63 m. y., and 2) mid-Tertiary from 40 to about 10 m. y., peaking at 25 m. y. (Damon, 1971). A third episode of volcanism has occurred during the last 10 m. y., but its characteristics are not yet well defined. Summaries of radiometric dates of igneous rocks in the southwestern United States by many authors document the existence of these three periods of magmatism (see for example, Armstrong, 1969; Damon, 1971; McKee and others, 1971; Silberman and McKee, 1972; Marvin and others, 1973; and Elston and others, 1973).

This paper examines Cenozoic igneous activity of the southern Colorado Plateau, mainly in Arizona and New Mexico, regarding its temporal, spatial, and petrologic characteristics. The spatial distribution is evaluated to test whether lineament tectonics adequately describes Cenozoic magmatism, and the temporal and spatial distribution is evaluated as to possible connections with the over-riding of the East Pacific Rise by the American plate. The petrologic characteristics of the volcanic rocks are compared with those of a calc-alkaline rock series which is presumably related to the subduction zones.
Figure 1.—Relationship of the Colorado Plateau to spreading centers of the East Pacific Rise (after Lipman and others, 1971).
ACKNOWLEDGMENTS

The author thanks C. W. Barnes and R. F. Holm of Northern Arizona University, and G. E. Ulrich and R. B. Moore of the U.S. Geological Survey for stimulating discussion and for comments which have led to improvements in the paper.

SPATIAL DISTRIBUTION OF CENOZOIC IGNEOUS ACTIVITY

Figure 2 shows the spatial distribution of intrusions and volcanic fields in the tectonic region of the southern Colorado Plateau. Generally, the areas delineated as volcanic fields are approximate outcrop patterns although in the Basin and Range province these areas represent the approximate areal distribution of the volcanic field prior to block faulting. Centers of Laramide igneous activity are not plotted in this paper for the Basin and Range (Livingston, 1973), nor are centers of mid-Tertiary volcanism plotted for southern Arizona.

It is clear that many centers of Cenozoic volcanism and plutonism occur along lineaments. The northeast-trending Jemez zone (I, fig. 2) and Front Range zone (II) and the northwest-trending southwestern Colorado zone (III) and central Colorado zone (IV) defined by Mayo (1958) are particularly obvious, as are the northwest-trending Henry (V) and La Sal (VI) lineaments of Kelley and Clinton (1960). Two other northwest-trending lineaments are definable: one by the centers of Datil volcanism in southwestern New Mexico and the Mount Baldy and San Francisco volcanic fields in Arizona (VII) and the other by volcanism in the Chiricahua, Superior-Superstition and Aquarius-Mohon Mountains of Arizona (VIII). The northwest lineaments have trends of about N. 45° W., and the northeast lineaments have trends of about N. 50° E.

It is not as clear that centers of volcanism define lineaments which are east and north trending. Mayo's (1958) north-trending Utah-Arizona belt (IX) and Pelonicillo belt (X) are suggested as are the east-trending Basaltic (XI) and Rico (XII) lineaments of Kelley and Clinton (1960).

Many of the Cenozoic volcanic fields occur on lineaments, which thereby validates the application of lineament tectonics. The northwest-northeast system of Mayo (1958) is dominant, as suggested by the orthogonal arrangement of volcanic fields that rim the southern Colorado Plateau. Hunt's (1956) suggestion that these trends are imposed by an ancient tectonic fabric is supported by: 1) northeast orientation of foliation of Precambrian rocks in Arizona (Gastil, 1960; Maxson, 1967; Boyce, 1972), and 2) general northeast and northwest orientation of strand lines and isopach lines for Paleozoic sedimentary rocks of this region. In the southern Colorado Plateau, the trend of the northwest lineaments as defined by the occurrence of volcanic fields (N. 45° W.) is different from that defined by Mayo.
Figure 2.—Distribution of Cenozoic igneous activity on and around the southern Colorado Plateau. Roman numerals are lineaments further identified in the text. Volcanic fields are identified as follows: A-M, Aquarius-Mohon Mountains; B, Bandera Lava Field; C, Chiricahua Mountains; D, Datil Volcanic Field; HB, Hopi Buttes; HM, Henry Mountains; L, LaSal Mountains; JM, Jemez Mountains; MB, Mount Baldy; MT, Mount Taylor; NB, Navajo Buttes; P, Pinacate Volcanic Field; RGD, Rio Grande Depression Lavas; S, Sentinel Volcanic Field; S-S, Superior-Superstition Mountains; SF, San Francisco Volcanic Field; SJ, San Juan Volcanic Field.

Location and age of Cenozoic igneous activity have been taken from the following sources: for Colorado, Armstrong (1969), Lipman and others (1970), Steven and Epis (1968) and Steven and others (1972); for New Mexico, Armstrong (1969), Bikerman (1972), Dane and Bachman (1965), Doell and others (1968), Elston and others (1968, 1973), Hunt (1938), Laughlin and others (1972), Lipman and Moench (1972), Ozima and others (1967), Seager (1973), Steven and Epis (1968), and Stormer (1972); for Arizona, Armstrong (1969), Damon and others (this volume), Evernden and others (1964), Marjaniemi (1968), McKee and others (1968), McKee and Anderson (1971), McKee and McKee (1972), Merrill (1970, 1973), Moore and others (this volume), Peterson (1968), Ratte and others (1969), Robinson (1913), Sheridan and others (1970), Watson (1967), Williams (1936), Wilson and others (1969) and Young and Brennan (1973); for Utah, Callaghan (1939), Gregory (1950), and Hunt (1956).
(1958) from Basin and Range structures (N. 30° W). The fact that the trend of lineament VII (fig. 2) coincides with a trend of recent earthquake activity along the Mogollon Rim (Simon, 1972 and Sturgul and Irwin, 1971) supports the idea that the N. 45° W. lineament is real. This difference of northwest trends may reflect the scale at which observations are made. The location of volcanic fields depends upon conditions in the lower crust or upper mantle where magma originates and upon zones of weakness in the deep crust, whereas structural features such as faults are controlled by conditions in the uppermost crust.

Among many local lineaments, there is another persistent trend defined by volcanic fields that has not been previously noted; namely, a set of three N. 20° E. trending lineaments. One includes the Pinacate volcanic field in Mexico, Sentinel and San Francisco volcanic fields in Arizona, and Navajo Mountain and the Henry Mountains in Utah (XIII). Another is suggested by the Uinkaret-Shivwitz Plateau volcanic fields in Arizona and those fields in southwestern Utah (XIV). A third may be defined by the Jemez Mountains and upper Rio Grande depression volcanics, which are also parallel to much of the Rio Grande River. This trend is within 10° of the trend of fold axes in the metamorphic crystalline rocks of the Grand Canyon (Boyce, 1972). It is also parallel to a trend in central Arizona along which crustal thickness abruptly changes (Warren, 1969). Therefore, an ancient deep-seated lineament direction is suggested.

TEMPORAL DISTRIBUTION OF IGNEOUS ACTIVITY

Figure 2 shows the spatial distribution of igneous activity during Laramide, mid-Tertiary (Oligocene-Miocene), and Plio-Pleistocene times. These temporal divisions are based on data reported in the literature cited above. The volume of Laramide or mid-Tertiary igneous rocks on the Colorado Plateau is insignificant compared with the volume of Plio-Pleistocene volcanic rocks. The boundary of the southern Colorado Plateau with the Basin and Range is marked by Plio-Pleistocene volcanic fields. Since mid-Tertiary volcanic rocks are abundant in the adjoining Basin and Range province, volcanism apparently progressed toward the Colorado Plateau. There are insufficient data at present to show a space-time distribution of Cenozoic volcanism as was done for the Great Basin by Armstrong and others (1969) and McKee (1971).

There is a definite temporal relation between the drift of the North American continent over the East Pacific Rise and mid-Tertiary and Plio-Pleistocene volcanism. Damon (1971) suggests that the westward drift of the North American plate relative to the East Pacific Rise began about 38 m.y. ago and ended about 10 m.y. ago. This is consistent with the data of Atwater (1970) and Yeats (1968), which suggest that the continental plate collided with the East Pacific Rise at the latitude of San Francisco about 30 m.y. ago and that the East Pacific Rise arrived beneath what is now the Gulf of
California about 9 to 12 m.y. ago. This is precisely the span of time during which mid-Tertiary volcanism occurred in the Basin and Range province. Clearly, the rifting of the continent and formation of the Gulf of California occurred during the time of Plio-Pleistocene volcanism on the Colorado Plateau and in adjacent areas.

The East Pacific Rise has been offset by transform faults; the trend of the axis of its spreading centers in the Gulf of California is northeast, whereas the trend of the transform faults is northwest (Larson and others, 1968; Lomnitz and others, 1970; Henyey and Bischoff, 1973). Faults of the San Andreas system such as the Imperial, the San Jacinto and Algodones faults, are parallel to the N. 45° W. lineament defined above. The axis of the East Pacific Rise is parallel to the N. 50° E. lineament (fig. 1). This suggests that activity of the East Pacific Rise served to preferentially open the northeast-northwest-trending system of lineaments, thus allowing Plio-Pleistocene volcanism to occur along these zones. This then may account for the predominant occurrence of volcanism along these trends rather than along those of the north and east system.

The temporal correlation of mid-Tertiary volcanism and westward drift of the North American continent relative to the East Pacific Rise strongly argues for a causative relationship (Damon, 1971). Evidence is presently too sparse to permit evaluation of the various volcanic-tectonic models to account for mid-Tertiary volcanism in southern Arizona and southwestern New Mexico.

PETROLOGIC CHARACTERISTICS OF LATE CENOZOIC VOLCANIC ROCKS

Igneous rocks can be classified into three major rock series, based on parent magmas and differentiation sequences: 1) the alkali olivine basalt series, 2) the tholeiite basalt series, and 3) the calc-alkaline basalt series. Criteria for distinguishing these rock series (Wilkinson, 1967) are applied to volcanic rocks of the southern Colorado Plateau with difficulty, partly because of the difficulty of uniquely defining members of a calc-alkaline rock series and partly because of the paucity of complete and detailed chemical and petrographic data. In the following discussion, conclusions have been taken from the literature with only cursory reevaluation. In some articles, the criteria used to place the rock series into one class or another are not clearly defined. Commonly, the distinction between alkaline and tholeiite affinities is based on the position of the rocks in the Na₂O + K₂O vs SiO₂ diagram defined for Hawaiian basalts by MacDonald and Katsura (1964).

As the volcanic rocks of the southern Colorado Plateau and environs are predominantly of Plio-Pleistocene and mid-Tertiary age, only these two periods of magmatism will be considered in the following discussion.

Generally, volcanic rocks of Oligocene-Miocene age that border the Colorado Plateau in New Mexico and Arizona have Peacock alkali-lime indices of 54 to 56 (Elston and others, 1968; Eastwood, 1970; Sheridan and others, 1970). These indices suggest that mid-Tertiary volcanism
tends to be more alkali-calcic than calc-alkalic. However, the observations of Elston and others (1968) and Eastwood (1970) indicate that the petrologic systems are more complex than the alkali-lime index suggests.

The petrography of these mid-Tertiary volcanic rocks tends to support the view that they are calc-alkaline or, in Wilkinson's terms (1967), at least subalkaline. Basalts, basaltic andesites, and andesites of this period contain or are associated with rocks that contain low-Ca pyroxenes (both hypersthene and pigeonite), olivine as phenocrysts and in the groundmass, and plagioclase in the labradorite and andesine ranges. Rarely olivine is found mantled by pyroxene. Most of the rocks have normative hypersthene and quartz, although the basalts are typically slightly nepheline normative (Eastwood, 1970; Sheridan and others, 1970). Merrill (1970) also found hypersthene and pigeonite in andesites of the Mount Baldy area. McKee and Anderson (1971) report hypersthene in tholeiitic basalts of the Hickey Formation (10 to 14 m. y.) in the Mingus Mountains of central Arizona.

Although no clearly alkaline volcanic rocks of mid-Tertiary age have been reported from the southern border of the Colorado Plateau the alkaline tendency of this calc-alkaline or subalkaline rock series has been noted. Sheridan and others (1970) recognize the calc-alkaline nature of ash flows of the Superior-Superstition volcanic field, and also the high alkali content of the basaltic rocks of that area and the fact that they are nepheline normative. Ratte and others (1969) identify a peralkaline ash flow of mid-Tertiary age in eastern Arizona.

Mid-Tertiary igneous rocks from the Four Corners area of the Colorado Plateau are distinctly alkaline. Williams (1936) documented the alkaline nature of the Navajo volcanic field, which consists mainly of minettes and trachybasalts. The potassium-rich nature of these rocks contrasts with the sodium-rich character of the Hopi Buttes volcanic field of Plio-Pleistocene age. Ratte and Steven (1967) and Lipman and others (1972) recognized that the rocks of the San Juan volcanic field are, or are related to, alkaline rhyolites.

From these data, it seems appropriate to characterize mid-Tertiary volcanism around the southern border of the Colorado Plateau as calc-alkaline (or subalkaline), with alkaline affinities in some of its members. This contrasts with the distinctly alkaline nature of mid-Tertiary volcanism in the Four Corners area of the Colorado Plateau.

The petrologic characteristics of the Plio-Pleistocene volcanic rocks of the southern Colorado Plateau are mainly alkaline, with some subalkaline affinities. The petrographic data of Robinson (1913) for rocks of the San Francisco volcanic field indicate the presence of alkalic amphiboles in rhyolites, hypersthene accompanied by augite or hornblende in dacites and latites, and olivine and augite in the basalts and andesites. Recent petrographic studies of siliceous rocks from the San Francisco Peaks verify the occurrence
of alkalic pyroxenes and amphiboles in these rocks (R. F. Holm, oral commun.). Petrographic and/or chemical analyses indicate that basalts from areas surrounding the central volcano are mainly alkali olivine basalts, although some have a high aluminum content (Hunt, 1956; Moore, 1973; Moore and others, this volume; McKee and Anderson, 1971; Condit, 1973; Murray, 1973). Ultramafic inclusions are common in many of the basalts (Cummings, 1972; Stoeser, 1973, this volume). Microprobe studies of the ferromagnesian minerals from dacites indicate the occurrence of fayalitic olivine and wide variation in pyroxene composition (R. B. Moore, oral commun.). As basalts are the most abundant rock type of the San Francisco volcanic field, the rock series could be described as alkaline. However, the complex nature of the ferromagnesian silicates in intermediate rocks indicates a subalkaline affinity or calc-alkaline affinity as Wenrich-Verbeek and Thornton (this volume) suggest.

The alkaline nature of the sodium-rich Hopi Buttes volcanic field has been determined by Williams (1936). These rocks include monchiquite, analcute basalt, and limburgite, each of which is characterized by augite, analcute, and minor amounts of olivine.

Results of petrologic studies have been published for several volcanic fields in northern New Mexico. Lipman and Moench (1972) found that Mount Taylor volcanism progresses from undersaturated alkali basalt, through silicic alkali basalt to late olivine tholeiite. Ultramafic inclusions occur in the early basalt flows. Ultramafic inclusions have also been described by Laughlin and others (1971) from basalt of Bandera Crater classified as olivine tholeiite by Laughlin and others (1972). According to Lipman and Moench (1972) the petrology of the Jemez Mountains volcanic field is similar to that of the Mount Taylor volcanic field.

Basalts related to the Rio Grande depression have been interpreted by Lipman (1969) and Lipman and others (1973) as being both alkalic and tholeiitic. Some of the alkali basalts seemingly are of mid-Tertiary age and are related to the San Juan volcanic field, which supports the previous characterization of that field. The alkali basalts are characterized by phenocrysts of plagioclase and diopsidic augite set in a groundmass of plagioclase, augite and olivine. Later basalts, such as the Servilleta flows, have smaller phenocrysts than the earlier alkali basalts, although their mineralogy is generally the same. Aoki (1967) suggested a subalkaline affinity for these later flows.

Stobbe (1949) has described the petrography of basaltic rocks in northeastern New Mexico that define an alkali olivine basalt series. The mafic members contain a typical alkali olivine basalt mineral assemblage in which the trachytes and phonolites are characterized by sodic pyroxene.

In summarizing the petrologic data of the Plio-Pleistocene volcanic rocks of the southern Colorado Plateau, it is clear that these rocks are dominantly alkaline. However, in every area studied there are some features, either petrographic or chemical, which suggest a subalkaline affinity. It is not clear whether this
subalkaline nature is calc-alkaline or tholeiitic or variable from one volcanic field to another. The alkaline nature of these Plio-Pleistocene volcanic rocks of the Colorado Plateau compares well with volcanic rocks of the same age from the Basin and Range province investigated by Leeman and Rogers (1970).

These observations support Christiansen and Lipman's (1972) view that late Cenozoic volcanism in the western United States is primarily basaltic and mainly alkali olivine basalt in character. A bimodal occurrence of basalt and silicic volcanic rocks has been noted by Lipman and Moench (1972) in the Mount Taylor volcanic field. The observations of Damon and others (this volume) and Moore and others (this volume) indicate that volcanism of the San Francisco volcanic field more recent than about 3 m.y. is also bimodal in the occurrence of basalts and rhyolites or rhyodacites. The silicic volcanism of the San Francisco Peaks (i.e., the riebeckite rhyolite and the Sugarloaf tephra) as well as of O'Leary Peak and the White Horse Hills occurs in this time span, as does basaltic volcanism in the northern and eastern part of the volcanic field. The first-period basalts of Robinson (1913) in the San Francisco volcanic field, which are between 6 and 10 m.y. old, may be an example of a dominantly basaltic field (Damon and others, this volume, and Christiansen and Lipman, 1972).

**STRONTIUM ISOTOPIC STUDIES OF CENOZOIC VOLCANISM**

Recent strontium isotopic studies provide data that contribute to understanding of the volcanism and tectonism of the southern Colorado Plateau. Bikerman (1967), Percious (1968), Eastwood (1970), and Stuckless and O'Neil (1973) reported initial strontium 87/86 ratios of mid-Tertiary volcanic rocks that range from 0.7063 to 0.7139. Regardless of the precise mechanism of magma derivation, the indicated heterogeneity of the derived rocks must reflect significant contribution of crustal material. A basaltic magma of mantle origin can be mixed with crustal material during its diapiric rise according to the model of Scholz and others (1971) or by the subduction model proposed by Armstrong (1968). Eastwood (1970) suggests that mid-Tertiary magmas were obtained by mixing of magma from a heterogeneous mantle with magma from the lower crust.

The above isotopic values contrast with those obtained on basaltic rocks of Plio-Pleistocene age from the southern Colorado Plateau and Basin and Range provinces in Arizona and New Mexico. Data from Hedge and Walthall (1963), Bikerman (1967), Damon and others (1969, 1970), Doe and others (1969), Kudo and others (1971), Leeman (1970), and Laughlin and others (1972) show that the initial strontium 87/86 ratios of Plio-Pleistocene volcanic rocks range from 0.7029 to about 0.7062. As Leeman (1970) has pointed out, this is the same range as that for oceanic basalts. Thus, the source regions of the mid-Tertiary magmas and Plio-Pleistocene magmas appear to have been fundamentally different, as suggested by Damon (1971).
Two general interpretations of these data are possible. The nearly continuous variation of strontium 87/86 ratios from those of primitive basalts, about 0.7030, to a value above that of average continental crust, 0.7095, could be emphasized. This would signify that mixing of crustal and mantle material occurred in both cases, but that the amount of crustal material admixed during Plio-Pleistocene volcanism was less than that admixed during mid-Tertiary volcanism. Evidence of crustal contamination has been found in lavas of Plio-Pleistocene age in northern New Mexico. Xenocrysts of quartz rimmed by pyroxene occur in many of the lavas of the Rio Grande depression (Lipman, 1969; Larsen and others, 1938). Lead isotopic studies of these xenocrystic alkali basalts also suggest effects of sialic crustal contamination (Doe and others, 1969). Xenocrysts of plagioclase that have higher strontium isotopic ratios than the host basalt occur in lavas from the Bandera Crater (Laughlin and others, 1971). Pushkar (in Damon and others, 1970) found a high strontium 87/86 ratio, 0.7186, in a rhyodacite from Mount Taylor volcanic field, apparently the result of assimilation of crustal rocks. Pushkar and Stoeser (in Damon and others, 1970) and Laughlin and others (1971) show that the strontium isotopic compositions of ultramafic inclusions and of host basalt are similar.

The second general interpretation is suggested by Leeman's (1970) observation above and also by the work of Kudo and others (1971): namely, that Plio-Pleistocene volcanism represents magma derived from a heterogeneous mantle, relatively uncontaminated by crustal material, whereas mid-Tertiary volcanism records significant admixing of crustal rocks by some means. Kudo and others (1971) found that hypersthene normative basalts of northern New Mexico had slightly higher initial strontium 87/86 ratios (0.7029 to 0.7078) than the nepheline normative basalts (0.7028 to 0.7044) and were interpreted as resulting from melting of a heterogeneous mantle. Lipman and others (1973) have determined potassium, thorium, and uranium contents for basalts of northern New Mexico. From these data, they also conclude that sialic crustal contamination was not significant. The similarity of these basaltic rocks to oceanic basalts of the Pacific was noted on the basis of potassium, thorium, and uranium contents.

SUMMARY AND CONCLUSIONS

Lineament tectonics provides a valuable, though at this point incomplete, means of understanding the spatial distribution of Cenozoic volcanic fields and intrusive rocks of the southern Colorado Plateau. The coincidence of the northeast-trending axes of spreading centers in the Gulf of California with northeast-trending lineaments accounts for the dominant northeast-northwest lineaments of Plio-Pleistocene volcanic fields of the southern Colorado Plateau. The tensional environment produced by Plio-Pleistocene activity of the East Pacific Rise provided access for mantle-derived alkali olivine basalt magmas having subalkaline affinities. The association of
alkali olivine basalt and tensional tectonics has been noted elsewhere in the western United States (Christiansen and Lipman, 1972).

The persistence of volcanism and plutonism along two lineament systems throughout Cenozoic time suggests that they are ancient and large-scale tectonic features. Correlation of these directions with structural elements of Precambrian age in Precambrian rocks attests to their aniquity. Because magmas of mantle origin occur along lineaments, these tectonic features therefore extend through the lithospheric crust and into the upper mantle, as suggested by Lipman and Moench (1972).

In contrast to Plio-Pleistocene volcanism, mid-Tertiary volcanism has been characterized as mainly calc-alkaline or subalkaline with alkaline affinities. This difference is also reflected in the relatively higher strontium 87/86 ratios of the mid-Tertiary rocks. These observations may indicate that significant amounts of crustal material were incorporated in the mid-Tertiary volcanic rocks, but not in the Plio-Pleistocene volcanic rocks.

Although it is clear that Plio-Pleistocene volcanism of the southern Colorado Plateau is related to activity of spreading centers along the East Pacific Rise, it seems equally likely that the volcanism is not related to a subductive mechanism. The occurrence of volcanic fields along ancient orthogonal lineament systems suggests a tensional environment. The diapir model of Scholz and others (1971) could account for this tensional environment, but the predominance of volcanism along the northeast-northwest lineament system is more easily accounted for by tension related to activity of the East Pacific Rise beneath the Gulf of California region. If this is true, magmas may have resulted in the upper mantle by reduction of pressure along these lineaments.

There are insufficient data to establish either the diapir model of Scholz and others (1971) or the subduction model of Armstrong (1968) for mid-Tertiary volcanism in Arizona and New Mexico. However, I believe trace element and strontium isotopic data favor the diapiric model.

Clearly, more petrographic and chemical data as well as careful field mapping in the Cenozoic volcanic fields of the southern Colorado Plateau region are needed for better understanding of the volcanic and tectonic processes involved.

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SOUTHWEST PALEOCLIMATE AND CONTINENTAL CORRELATIONS

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ABSTRACT

A continuous record of depositional and biological events is available from northern Arizona and the Southwest from which generally consistent climatic inferences can be derived for the past 50,000 years. Converging evidence indicates a major fluctuating pluvial cycle which culminated 20,000 to 30,000 years B.P. between pronounced nonpluvial intervals centered about 40,000 to 50,000 and 5,000 to 6,000 years B.P. The classic dating of the "altithermal" and "Little Ice Age" are broadly supported by the new time-stratigraphic data, and a systematic postglacial pattern of secondary climatic oscillations is reaffirmed.

Southwestern climates were mainly in phase with detailed midcontinental and Pacific Coastal reconstructions, but out of phase with continental glacier fluctuations as presently classified in Illinois. When continental glaciers were expanding, southwestern biotic communities were displaced to lower elevations and more southerly latitudes, and pluvial lakes expanded and alluvial deposition was augmented under conditions of more moist and cooler climates. As continental glaciers waned, the reverse biotic and geologic phenomena occurred.

Superposed on the major millenia-long trends of the late and post-Pleistocene climates of the Southwest were many secondary and tertiary climatic pulses of shorter durations and magnitudes. The major climatic impulses resulted in biotic migration of 1,000 m or more in elevation and latitudinal displacement of hundreds of kilometers. The shorter duration, lesser magnitude pulses moved biota only hundreds of meters and tens of kilometers from their present geographic distribution. The recorded lag between attainment of maximum pluvial conditions in the Southwest and those of maximum extents of continental ice presumably resulted from general retardation in continental ice accumulation along with changing ice dynamics and shifting nourishment centers. Time lags between the secondary glacial, pluvial, alluvial, and biotic oscillations appear to be less than present dating precision, which suggests almost immediate response to synchronous short term climatic trends. The pattern of southwest climatic change is consistent with other paleoclimatic data in suggesting cycles of 3,300 +years, 1,100 +years, and 550 +years, respectively of substage, stadial, and lesser (phase) rank. This pattern is defined in a proposed classification of late Quaternary time.

INTRODUCTION

Geologists, paleontologists, and paleoclimatologists have long been interested in the wealth of Quaternary data available in the American Southwest. The emerging data have prompted questions of synchronicity, magnitude, nature, and relationship of bioclimatic and geoclimatic events recorded in the Southwest as well as possible theoretical regulatory mechanisms. These problems have been the subject of discussion at several southwest workshops and seminars over the past decade which have revealed the need for a synthesis of the Southwest data and comparison of such data with those available from other regions of the continent. The following synthesis summarizes available data bearing directly on the character of past climatic patterns in the Southwest and in bordering continentally glaciated regions. From these comparative data we conclude that southwest climate shifted in phase with that causing the waxing and waning of continental ice. Furthermore the data suggest the existence of a multiple cyclic pattern of possible predictive climatic significance.

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MATERIALS AND METHODS

Numerous studies have demonstrated the occurrence of alpine glaciation in both Arizona and New Mexico (Antevs, 1954; Blagbrough, 1971; Breed and Blagbrough, 1967; Melton, 1961; Merrill, 1970; Sharp, 1942; Updike and Powe; this volume) and the concurrent displacement of biotic communities many hundreds of meters to lower elevations and hundreds of kilometers to the south (Malde, 1964; Martin and Mehringer, 1965). While dispelling any possible misconceptions of the magnitude of late Quaternary bioclimatic and geoclimatic events in the Southwest, the emerging data in some instances have been variously correlated and interpreted so as to leave unresolved basic questions of duration, amplitude, synchronicity and nature of the paleoclimatic phenomena (Aschmann, 1958; Bryan and Gruhn, 1964; Jennings, 1957; Maher, 1972; and Martin, 1963a). Were the episodes of erosion and alluviation, pluvial lake fluctuations, and shifting biotic boundaries coincident with major and secondary pulses of continental ice sheets? Was the magnitude of effect on southwestern
Figure 1.—Location of the bioclimatic and geoclimatic records cited or discussed in text.
appear to extend back beyond 30,000 years B.P. by assumption of uniform depositional rates based on C-14 dates; however, all but the Humptulips and San Augustin records lack the most recent 15,000 years or more due to erosion or pollen oxidation. These two best dated and longest records with uniform lithologies are compared in figure 2 with a number of other well-dated, overlapping but much shorter pollen records of the last 30,000 years from the Southwest.

The data exhibit a general parallelism of fluctuating relative abundance of conifer pollen achieving maximal values between 20,000 to 30,000 years ago. Preceding and following this period the relative abundance of conifer pollen is diminished, declining to pronounced minima not unlike the present, between 40,000 to 50,000 and between 5,000 to 6,000 years ago in most records. Superposed on this general trend are many secondary pulsations which are commonly replicated in records with comparable sampling intervals. The failure of all records to show each pulsation is most probably attributable to sampling interval differences, sample contamination, and differing site sensitivity to biologic response. The failure of fluctuations to be exactly coincident could result from slight regional climatic differences, but could also result from small errors in dating and slight variations in rates of deposition. However, when comparing portions of curves with commensurate detail, nearly all pulsations are coincident within the limits of C-14 dating resolution. We therefore conclude that if regional climatic differences exist between sites in western North America, they are not definable by the present data. Southwestern fossil pollen data obtained at different elevations also demonstrate broad synchronicity of bioclimatic events (fig. 3).

While climatic synchronicity is strongly suggested by southwestern pollen data, some comment must also be made regarding duration and magnitude of events. The major fluctuations which appear to be approximately three millenia in duration can be observed from unsmoothed data to consist of many secondary and tertiary pulses of lesser magnitudes whose durations are on the order of millenia, centuries, or decades (figs. 2 and 3). Martin and Mossiman (1965) have demonstrated that at least the more major of these in Lake Cochise are non-random fluctuations reflecting compositional changes of plant communities and has suggested that such fluctuations may be equivalent to continental ice margin metastability. Similar, seemingly synchronous palynological events are demonstrable at many other southwestern localities, such as Osgood Swamp in the Sierra Nevada and Adobe Valley, California (Adam, 1964; Batchelder, 1970); Meadow Valley Wash, Nevada (Madsen, 1972); Dead Man Lake in the Chuska Mountains of New Mexico (Bent and Wright, 1963); Lindenmeyer and Yellowstone, Wyoming (Hansen, 1951; Baker, 1970); Animas Valley, Colorado (Maher, 1976); various localities in Arizona (Hevly, 1964; Martin, 1963a, 1963b); and the Chihuahuan Desert areas of the Amistad Reservoir, Texas and Cuatro Ciengas, Coahuila (Bryant, 1966; Hevly, 1966; Meyer, 1973).
Figure 2.—Western United States bioclimatic records.

1. Redrock Lake curves (upper three curves, right corner): The uppermost right pollen curve was constructed from the pollen data of Pennak (1963) according to the bioclimatic equation: dominant higher-altitude species (Artemisia + spruce) minus dominant lower-altitude species (pine + grass). Increasing negative index values indicate increasing warmth and/or dryness. Time calibration assumes uniform depositional rates in homogeneous organic silts between 5 cm (946 +years) and 105 cm (5,560 +years), and between 105 cm and 185 cm (6,680 +years). The C-14 results of four of the 10 dated horizons in the section are clearly incompatible with stratigraphic position. As constructed by assuming uniform depositional rates, the curve is consistent with the results of all but one of the remaining six dated horizons. Its general accuracy is further supported by close agreements with comparably derived bioclimatic curves from Pennak's other C-14 dated records at nearby Albion Bog, Muskee Lake, and Silver Gate Lake Bog and from Maher's (1972) more closely dated record of Redrock Lake. Many of the apparent differences in trends among these curves can be directly attributed to differing sample sizes. The right curve of Maher (1972) reflect total arboreal pollen x 10^6/cm² (scale numbers 1 and 2) and percent organic matter. For comparative purposes, Maher's data are replotted according to the internationally accepted C-14 half-life of 5,570 years. The mean depths of the dated sections are as follows: 1590 +years-37.5 cm; 2450 +years-57.5 cm; 4290 +years-81.5 cm; 5150 +years-106.5 cm; 6870 +years-131.5 cm; 8460 +years-152.5 cm; and 9490 +years-167.5 cm.

2. Potato Lake (upper-central) curve: This pollen curve was constructed from data of Whiteside (1964, 1965) according to the bioclimatic equation: dominant higher altitude species (Artemisia + PAP) minus dominant lower altitude species (pine + juniper + oak + grass). Decreasing negative index values are indicative of cooler and more moist conditions and increasing negative values reflect the opposite. Boundaries between vegetation types were determined by composition of fossil pollen data compared to modern vegetation and pollen rain. Time calibration assumes uniform depositional rates in upper dark organic silts between 104 cm (9,900 +years) and surface; and in lower greenish silts based on the dated horizons at 104 cm (9,900 +years) and at 157 cm (14,400 +years).

3. San Augustin Plains curves: Pollen curve A constructed from pollen data of Clisby and others (1962) using the bioclimatic equation: dominant higher altitude species (2 x spruce) minus dominant lower altitude species (pine + 2 x desert herbs and shrubs). Trends of index values are as previously discussed. Boundaries between vegetation types established by pollen composition. Time calibration assumes uniform depositional rates in the homogeneous silty clays and clayey silts based on dated horizons at 4.2' (12,150 +years), 12.2' (17,650 +years), and surface; and in lower greenish silts based on dated horizons at 8 m (12,130 +years) is incompatible with stratigraphic position and is discarded. Curve B is further smoothed by using the average pollen sums in non-overlapping 300 year intervals; curve B is further smoothed to emphasize larger pollen trends by using average pollen sums in non-overlapping 1,200 year intervals. The dotted portion of curve B is inferred from dated strandlines in the San Augustin Plains area, and by cross-correlations with the other bioclimatic records.

4. Searles Lake curve: The pollen curve was constructed from pollen data of Roosma (1958) using the bioclimatic equation: dominant higher altitude species (pine + juniper) minus dominant lower altitude species (desert herbs and shrubs). Index trends and vegetation boundaries established as above. Time calibration assumes uniform depositional rates of 1,000 yrs/ft based on dated horizons in the Parting Mud section (Smith, 1968). Absolute pollen curves derived from the Parting Mud and Upper Salt Body sections by Leopold (1970) are generally consistent with Roosma's relative pollen frequency diagrams.

5. Swan Lake: Pollen curve was constructed from the data of Bright (1968). Index trends established as above. Time calibration assumes uniform depositional rates based on dated sections at 1.775 m (1,850 +years) at 5.9 m (10,190 +years) and at 8 m (12,090 +years).

6. Humptulips Bog: NAP, AP, and pine pollen curves were constructed from the data of Reusser (1964). Time calibration assumes uniform depositional rates based on dated horizons at 0.65 m (5,250 +years) and at 2.5 m (27,400 +years). Note the general agreement of this relatively complacent record with curve B of the San Augustin Plains.

7. Tule Springs and Murray Springs curves: Percent frequency pine pollen curves taken from Mehringer (1967) and from Mehringer and others (1967). Tule Springs profile #4 and Murray Springs pollen curves are positioned according to C-14 dated sections. Tule Springs profile #1, positioned by cross-correlations, is seemingly in accord with the infinite date of >37,000 years B.P.
Figure 3.—Comparison of bioclimatic records from various localities in the Southwest shown in figure 1.

**Sonoran and Great Basin Deserts (<1,500 m):** from various localities in southern Arizona (Martin, 1963a; Mehringer, 1967); Rampart Cave, Arizona (Martin and others, 1961); and Fishbone and Guano Caves, Nevada (Sears and Roosma, 1961).

**Chihuahuan Desert (<1,500 m):** from Bonfire Shelter and Devils Mouth, Texas (Bryant, 1966; and Hevly, 1966).

**Great Basin Desert (1,500 m):** from Hogup Cave, Utah (Aikens and others, 1968).

**Pinyon-Juniper, Woodland and Juniper Savana (2,100 m):** from San Augustin Plains, New Mexico (Clisby and Sears, 1956); and from Laguna Salada, Arizona (Hevly, 1964).

**Yellow Pine Forest (2,200 m):** from Potato Lake, Arizona (Whiteside, 1964).

**Spruce-Fir Forest (3,100 m):** from Redrock Lake, Colorado (Pennak, 1963). Higher altitude tree establishment in the White Mountains of California and the Snake Range of Nevada after Lamarche and Mooney (1967).
Comparison of fossil and modern pollen spectra indicate that the fossil spectra can be closely approximated by modern records of existing plant communities in the Southwest (Adam, 1964; Bent and Wright, 1963; Byers and Martin, 1965; Dixon, 1962; Hafsten, 1961; Hevly and others, 1965; Hevly, 1968; Maher, 1963, 1972; Martin, 1963a; Mehringer, 1965; Potter and Rowley, 1960). It is apparent from such comparisons that the southwestern biotic communities experienced considerable elevational migration during the late Wisconsin (Martin and Mehringer, 1965). Periods of increased relative abundance of arboreal pollen can be approximated by modern pollen spectra obtained from communities presently up to 1,220 m above the elevation of the fossil data (Hevly, 1964; Mehringer, 1966). Macroscopic plant data from fossil pack rat nests in modern desert environments of the Southwest, and even fossil vertebrate data from cave sites, indicate elevational depression of biotic communities up to 1,000 m below their modern occurrence (Harris and Findley, 1964; Mawby, 1967; Mehringer, 1966; Mehringer and Ferguson, 1969; Van Devender and King, 1971; Wells, 1966; Wells and Berger, 1967; Wells and Jorgensen, 1964). Invertebrate fossil data also indicate similar trends (Megard, 1964; Reger and Batchelder, 1971). Not only were biotic communities elevationally displaced (fig. 3), they or at least some elements were latitudinally displaced hundreds of kilometers to the south (Martin and Mehringer, 1965; Hevly, 1966).

The consistency of the above comparative data support the conclusion that in the Southwest, fossil spectra obtained well below timberline and characterized by increased arboreal pollen relative to the present can be interpreted as indicative of cooler and wetter climates. The reverse is true of fossil spectra characterized by diminished percentages of arboreal pollen which compare most favorably with modern pollen spectra obtained at or just below the elevation of the fossil locality. In the deserts, fossil pine pollen is generally indicative of mesic conditions, whereas at higher elevations recording past spruce-fir forests, pine is generally indicative of more xeric conditions; nevertheless, both low and high altitude records covering all or part of the last 15,000 years clearly exhibit the pluvial to post-pluvial transition occurring between 10,000 to 12,000 years B.P. (figs. 2 and 3).

The pollen records covering the post-pluvial period clearly record the classical three-fold subdivision proposed by Antevs (1948), including the "Little Ice Age" of Matthes (1942), which is lichenométrically documented in the Front Range of Colorado (Benedict, 1968). The major altithermal trend also is revealed by sediments, and by macroscopic and microscopic plant materials and bone accumulation rates in Hogup Cave (Aikens and others, 1968) (fig. 3). Paleontological evidence for the altithermal in northern Mexico, western Texas, southern New Mexico and southern Arizona remains undetected due to erosion (Bryant, 1966; Hevly, 1966; Mehringer and others, 1967; Meyer, 1973).

Broad general comparisons of southwestern paleoclimatic trends seem possible with climatic trends inferred from palynological data.
obtained in other areas of North America by Bryson and Wendlund (1967), Cushing (1965), Davis (1965), McAndrews (1966), Shay (1967), and Whitehead (1965).

Superposed on the broader trends are smaller secondary and tertiary pulsations which appear to be synchronous and of the same sign at different elevations in the Southwest (fig. 2). Warmer and dryer intervals, commonly characterized by erosion, seemingly bracket southwest cultural changes and possibly indicate climatic influence on prehistoric cultural evolution (Baumhoff and Heizer, 1965; Irwin-Williams and Haynes, 1970; Schoenwetter and Eddy, 1964; Schoenwetter and Dittert, 1968; Karlstrom and others, this volume, Euler and Gumerman, this volume) (fig. 6). Similar correlations between climatic events and migration of prehistoric populations have been suggested by Bryson and Wendlund (1967) for midcontinental areas, and by Husted (1970) for the northern plains and Rocky Mountain region.

Archaeological pollen data of the last 2,000 years obtained in Navajo Reservoir and at Picuris Pueblo, New Mexico as well as Hay Hollow Valley, Arizona reveal a number of minor bioclimatic events which are apparently correlative with decadal dendroclimatic events and local cut and fill sequences (Schoenwetter and Eddy, 1964; Schoenwetter and Dittert, 1968; Hill and Hevly, 1968; Robinson and Dean, 1969). However, vertebrate data of the last 2,000 years appear to indicate less biotic changes than do pollen data, which might suggest change in pollen production rather than movement of plant communities (Harris, 1963; Lawrence, 1951; Lincoln, 1962; Stein, 1963). Nevertheless, biotic changes are recorded both dendroclimatically and geologically and the most recent vegetation changes have been photographically documented (Fritts, 1965; Hastings and Turner, 1964; Robinson and Dean, 1969). Furthermore, chronologically synchronous climatic perturbations have also been documented in midcontinental areas by Bryson and Wendlund (1967). The magnitude of pluvial and post-pluvial displacement of biotic communities does not appear to be equal at all fossil localities. This probably is due to a variety of factors including mixing, sampling interval, and rates of deposition problems, as well as site sensitivity (ecotones more responsive than community interiors (Adam, 1964; Hevly, 1964).

The above data suggest synchronous movements of biotic communities and pollen production in the Southwest throughout the past 50,000 years; however, two of the presented records, those from Redrock Lake, Colorado, and Humptulips Bog, Washington, have in part been interpreted in a contrary manner. Because the interpretation of these records bears directly on the problem of synchronicity and variation of climatic trends in the west, we discuss the data of these records more fully and examine them for alternative ecological and climatic interpretations.
Redrock Lake, Colorado pollen curve

While Pennak's (1963) pollen record and reconstruction of paleoenvironments at Redrock Lake are interpreted to be in accord with other regional data, the pollen spectra and interpretation of Maher (1972) are strikingly different. As stated by Maher, the record undoubtedly indicates a general rise of vegetation zones near Redrock Lake during the postglacial period; however, Maher does not recognize the occurrence of a warm, dry, mid-postglacial period equivalent to the altithermal of Antevs (1948) or to the hypsithermal of Deevey and Flint (1957).

Pennak's record contains strong indication of the altithermal by abruptly increased values of Graminae and Cyperaceae but in Maher's pollen data no such increase is found. While total arboreal pollen does diminish during the appropriate time period, the drop may be accounted for by increased relative abundance of Nuphar and Potamogeton pollen. In Maher's pollen data no broad decrease of conifer pollen occurs; rather, it actually increases along with Alnus from c. 10,000 years ago up into the altithermal period. This increase of arboreal pollen and particularly the ratios of conifer pollen types suggested to Maher (1972) that the period may have been cooler and wet rather than warmer and dry.

Other elements in the pollen spectra suggest otherwise (fig. 4). The general increase of arboreal pollen from late pluvial time into the altithermal interval simply may be the result of a progressive rise in timberline from lower elevations to within and above the site during this time span, as is also suggested by the accompanying succession of upland sage to spruce to pine dominance. The pronounced decrease in arboreal pollen, in both frequency and absolute pollen amount, between 6,000 and 4,000 years B.P. is centered on the culmination of the altithermal as dated elsewhere (fig. 5). Further, because both Potamogeton and Nuphar are growing near their altitudinal limits, an increase in their abundance should indicate warmer temperatures. Pennak's recovery of increased grass and sedge pollen occurs in the same time interval, and is most reasonably interpreted as indicative of diminished water depth and warmer water. If so, in a near-timberline lake, increased temperature should result in increased productivity and deposition of organic matter. In fact, maximal organic matter content of sediment occurred during the culmination of the altithermal interval (fig. 4).

At other southwestern lakes Pediastrum tends to be more abundant during cooler intervals (Hevly, 1964; Whiteside, 1965). A similar relationship is suggested in Maher's pollen curve in which relative abundance of this algal relict is inversely proportional to the abundance of Potamogeton and Nuphar. Thus, at Maher's coring site an increase of Nuphar and Potamogeton pollen appear to be the altithermal indicators, but at Pennak's coring site within the same lake, grass and sedge pollen are the altithermal indicators. Such variations of non-arboreal pollen content within the same depositional basin
Figure 4.—Comparison of Redrock Lake biotic and sedimentation data: Curve 1, Pediastrum relative abundance; curve 2, Nuphar pollen relative abundance; curve 3, Pediastrum minus Nuphar index; curve 4, total arboreal pollen; curve 5, percent organic matter. The data clearly reveal the classic threefold division of the post-Pleistocene including the "Altithermal" indicated by diminished Pediastrum and total AP pollen and increased Nuphar pollen and organic matter production-accumulation. The classical "Little Ice Age" as well as the more recent "Little Ice Age" recognized by European workers are both indicated by increased Pediastrum and diminished Nuphar and arboreal pollen production as well as organic matter production-accumulation. See figure 2 for dating control.
have been previously noted by Davis (1963) and are to be expected in response to changing depth, water chemistry, temperature, and proximity to shifting shorelines (Hevly, 1964, 1974).

**Humptulips, Washington pollen curve**

Heusser (1964) utilizes the Humptulips data to reconstruct the probable mean temperature in that area during late Pleistocene and Holocene times. His derived temperature curve is not exactly parallel with events as reconstructed in the Southwest, nor as reconstructed more recently at nearby Kalalock, Washington (Heusser, 1972). The Humptulips pollen data probably are correlative with records from elsewhere in western North America, the apparent differences resulting from inherent difficulties of deriving valid quantitative temperature curves from complex pollen data. We also believe that the part of the Humptulips record exceeding 30,000 years B.P. can be alternatively interpreted in a relative climatic sense as a warm, dry, rather than a cold, wet interval. This interval is characterized by increased relative abundance of Sphagnum spores which elsewhere in the record correlate with warm intervals as interpreted by Heusser. Despite increased relative abundance of grass and sedge pollen, which influenced Heusser's interpretation of this period as cold and wet, the diminished relative abundance of high-elevation conifer pollen and slightly increased relative abundance of low elevation arboreal pollen are also indicative of a relatively warmer-dryer interval. This alternative interpretation agrees with the Heusser's interpretation of the Kalalock profile and, as will be shown, with glacial time-stratigraphy in nearby Puget Sound.

**SOUTHWEST GEOCLIMATIC DATA AND CONTINENTAL CORRELATIONS**

From the bioclimatic data discussed above we conclude that these data provide an internally consistent climatic record for the region over the past 50,000 years. The southwest geoclimatic data discussed below are compatible with this reconstruction. Comparative data further suggest that southwest paleoclimate was in phase with that of bordering regions, including the continentally glaciated parts of the continent. In figure 5, we directly compare the bioclimatic records with southwest pluvial records and with glacial records from the midcontinent, Puget Sound, and Alaska.

The pluvial chronology of Searles Lake is from Smith (1968, pers. commun., 1972). The Bonneville Basin reconstruction is modified from Karlstrom (1961). More recent data, particularly Bright's Swan Lake data obtained from a site in the outlet channel of the Provo level, support Morrison and Frye's (1965) conclusion that some of the strandline tufa dates are incompatible with the pluvial time-stratigraphy as dated by carbonaceous materials in deposits at intermediate and lower levels. The Swan Lake data essentially
Figure 5.—Bioclimatic records of the Southwest compared with published pluvial and glacial chronologies

Column 1. Proposed late Quaternary geologic (time-stratigraphic) classification with Cook Inlet, Alaska terminology (Karlstrom, 1966, 1970a, 1970b). Triple vertical lines positioned at estimated culminations in radiocarbon years B.P. of interglacial intervals separating or terminating deposits of proposed stage rank; double vertical lines at estimated culminations of intraglacial intervals separating deposits of proposed substage rank; single vertical lines at estimated culminations of subordinate nonglacial intervals separating deposits of proposed stadial rank.

Theoretical considerations indicate that dated reference boundaries placed at nonglacial culminations most rigorously satisfy the requirements for time-stratigraphic subdivision of time-transgressive depositional sequences such as​ tills and correlative deposits (Ray and Karlstrom, 1968).

Curves 2, 3, 4, 5, 6, 9, and 10. Bioclimatic records as described in figure 2, except that in 2, the middle curve is of spruce and pine rather than total AP.

7. Lake Bonneville Pluvial Record as modified from Broecker and Orr (1958) on the basis of new radiocarbon data (Karlstrom, 1961, this paper.)


require that the Red Rock Pass Provo level outlet was not reached by rising lake waters after 12,000 years B.P. At the same time the Swan Lake record seemingly confirms, by abrupt and parallel pollen changes, the pattern of late- and post-pluvial lake level fluctuations as earlier interpreted from published stratigraphic data from Danger Cave and lower strandlines (fig. 5).

The Port Talbot warm interval

The warm/dry interval between 40,000 and 50,000 years ago inferred from the San Augustin and Humptulips bioclimatic records (and also shown in nearby Kalalock (Heusser, 1972) ) coincides with the C-14 dated Knik/Naptowne interglacial stage of Cook Inlet (Karlstrom, 1955, 1961, 1964, 1968); the Port Talbot nonglacial interval in the Lake Erie Basin (Dreimanis and Karrow, 1972); the Plano Silt nonglacial interval in the Michigan Basin (Kempton and Hackett, 1968), and with the nonglacial deposits dated 34,900 to 50,000 years B.P. between tills in Puget Sound (Easterbrook 1974). That this nonglacial interval may have marked the culmination of a major warm interval in Puget Sound is further supported by the presented bioclimatic data and by the C-14 time-stratigraphic evidence that the plateaus of British Columbia (one of the major upland source areas for Puget Sound ice), remained ice-free from sometime before 43,800 to 19,000 years B.P. (Smith, 1970).

1/The Knik/Naptowne time-stratigraphic boundary is derived from boulder-count ratios calibrated by numerous C-14 analyses including 5 internally consistent dates ranging from >32,000 to >42,000 year B.P. (and including one finite date of 39,000 ±2000 years B.P.) obtained from insitu soil and peat and associated wood samples collected from the base of Naptowne drift at three localities. Correlation with the isolated marine sections of Woronzofian age, supported by provisional ionium/uranium dating of shells between 33,000 and 48,000 years B.P. (Karlstrom 1964; 1965) is incompatible with new dates (ca. 14,000 years B.P.) obtained from shells collected from the same horizon (Schmoll and others, 1972). If the new dating is valid, the Woronzofian marine horizon is probably glacioisostatic rather than glacioeustatic in origin, as noted by Schmoll and others (1972), and a new type section in Alaska would be required for the earlier marine incursion of Knik/Naptowne age locally recorded along the Alaska coast and elsewhere (Hopkins 1967; Karlstrom, 1968).
For years many Pleistocene researchers (see articles in Wright and Frey, eds., 1965) have assumed that the first major warm interval preceding the last major glaciation in the western cordillera was Farmdale in age and dated this interval between 22,000 and 28,000 years B.P. by direct correlation with the glacial classification proposed for Illinois and Wisconsin (fig. 5, column 13). A more recent synthesis of the glacial, pluvial, and alluvial stratigraphy of the western states (Birkeland and others, 1971) places the culmination of the Bull Lake/Pinedale interglacial soil between 40,000 and 50,000 years B.P. A similar age bracket is indicated for distinctive buried soils and nonalluvial deposits in reconstructed alluvial sequences for the San Joaquin Valley, California; the Piedmont area, Colorado; the La Sal Mountains, Utah; the Las Cruces—El Paso area, New Mexico and Texas; the Upper San Pedro Valley, Arizona; and Las Vegas, Nevada. Comparable time-brackets are also suggested for soils and nonglacial deposits between Tenaya and Tahoe tills in the Sierra Nevada, California; for buried soils and nonpluvial deposits between the Eetza and Wyemaha Formations of the Lake Bonneville Basin, Utah, and the Alpine and Bonneville Formations of the Lake Lahontan Region, Nevada. The seeming stratigraphically equivalent, but undated soil between Evans Creek drift and the Hayden Creek drift in the reconstructed Cascade Range, Washington, glacial sequence is placed somewhere between 20,000 and 40,000 years.

Most of the time-stratigraphic data from the western states as presently dated and interpreted support correlation of the major warm interval preceding the last glaciation, pluviation, and alluviation of the west with the Knik/Naptowne interglacial stage of Alaska and with the Port Talbot interstadial interval of the midcontinent. A recent reinterpretation of the glacial stratigraphy in the Yukon Territory, Canada (Porter, 1971) places the last major interglaciation of this region as Port Talbot, rather than Farmdale in age, which is also in agreement with the bioclimatic and geoclimatic data shown in figure 5.

Although an increasing number of researchers accept the evidence for the culmination of a major warm interval approximately 40,000 to 50,000 years ago, disagreements remain on the assignment of rank to this nonglacial interval. In the North American midcontinent it is assigned an interstadial rank, but in other regions of the world it is considered of interglacial rank. This disagreement results from definition differences, dating imprecisions and from insufficient, or conflicting, evidence relating to diagnostic vegetation types, and to positions of continental ice margins and sea level during this time interval. (Karlstrom, 1968; Dreimanis and Karrow, 1972; Grüger, 1972).
"Altithermal"

When Antevs (1948) introduced the paleoclimatic term "altithermal" for the North American equivalent of the European postglacial "climatic optimum" period of maximum warmth of Atlantic age (ca. 9,000 to 4,500 years B.P.), much data from southwestern United States had been proposed in support of such an event. Evidence from the western states as presented by Antevs (1955) includes the saline content of Abert and Sumner Lakes (Van Winkle, 1914), complete or near complete disappearance of alpine glaciers prior to the "Little Ice Age" (Matthes, 1939, 1942); prevalent wind deposition and arroyo cutting in the southwest (Hack, 1942; Evans and Meade, 1945; Judson, 1953); accumulation of caliche (Bryan and Albritton, 1943; Bretz and Horberg, 1949; Leopold and Snyder, 1951); the marked maximum of grass-chenopod-composite pollens in the Chewaucan Marsh and Warrior Lake sedimentary columns (Hansen, 1947; Flint and Deevey, 1951); and archaeologically dated stratigraphic data (Antevs, 1940; Cressman, 1951).

More recently the concept of the altithermal as the postglacial period of maximum warmth culminating between 9,000 and 4,500 has been seriously questioned. (See Deevey and Flint, 1957; Aschmann, 1958; Jennings, 1957, 1964; Martin, 1963a, and Bryan and Gruhn, 1964). However, archaeological and geologic data presented by others support the altithermal concept (see for example Husted, 1970; Haynes, 1968a; Baumhoff and Heizer, 1965; Malde, 1964; and Karlstrom, 1956). The concept now appears to be well documented by C-14 dated pluvial, alluvial, and glacial stratigraphy, which is consistent with the previously presented pollen data for many regions of the west.

The pluvial evidence, as interpreted by Smith (1968) in the Searles Lake basin, and by Morrison (1965, 1970) in the Bonneville and Lahontan basins, is of postpluvial lake desiccation and soil formation prior to regeneration of regional lakes during the "Little Ice Age" (fig. 5, curves 7 and 8). Such regeneration in the Searles Lake basin took place sometime between 3,500 and 7,000 years B.P. The more broadly C-14 bracketed Toyeh and Midvale soils of altithermal age in the Bonneville and Lahontan basins fall in the same time interval (Birkeland and others, 1971). An organic bed separating late pluvial lake deposits from overlying postpluvial alluvium in the San Joaquin Valley is dated at 5,480 ±300 years B.P. (W-1579: Deevey and others (eds.), Radiocarbon, v. 9, p. 515) and marks the culmination of the altithermal as dated elsewhere (Karlstrom, 1956; 1966).

Significantly, wood samples from lower Colorado River sediments considered to record the postglacial rise in sea level provide altithermal dates of 6,250 ±300 years B.P. and 5,380 ±300 years B.P. (W-1501 and W-1143: Deevey and others (eds.), Radiocarbon, v. 9,
p. 512-513) or contemporaneous with the Kasilofian transgressive culmination dated at ca. 5,500 years B.P. in the Cook Inlet region, Alaska (Karlstrom, 1965, 1968). Sea level fluctuations comparable to those inferred from bog stratigraphy in Cook Inlet (Karlstrom, 1961, 1964) are reported from the Netherlands (Jelgersma, 1961; Jelgersma and others, 1970) and Denmark (Jacobsen, 1964) (table 1).

Recent C-14 dated alluvial chronologies reported from many areas of the west support the earlier southwest investigators in the placement of a major break in the depositional record during the altithermal (fig. 6). Buried soils, nonalluvial deposits, and unconformities of altithermal age are placed between 7,000 and 4,500 years B.P. in the northern Arizona sequence (Hack, 1942) and in the Nebraska sequence (Schultz and others, 1951). In the Southwest, the same event is C-14 dated sometime between 2,300 and 7,200 years (Miller, 1958; Miller and Wendorf, 1958); and more closely C-14 dated between 4,500 and 7,100 years B.P. by Haynes (1968a). In the Rocky Mountain alluvial sequence (Scott, 1965), the center of the pre-Piney Creek sediments, recording conditions warmer than present, is dated at ca. 5,500 years B.P. In the southern Arizona San Pedro Valley sequence, the boundary between the Lehner Formation of late pluvial age, and the Escapule Formation of post-pluvial age is placed ca. 5,500 years B.P. (Haynes, 1968b). A deeply weathered pink sand unit in the Black Mesa area, correlated with Hack's buried weathered dune sand interval, dates 6,540 years B.P., a good altithermal date. This date, however, is provisional because it is derived from carbonate materials and requires confirmation by additional C-14 dating in the area.

The extensive glacial morainal sequences in the higher mountains of the Southwest recording Matthes' "Little Ice Age" are only in part C-14 dated. Based on a review of the regional evidence, Curry, D. (1970) places the boundary between late Pleistocene Pinedale moraines and Holocene moraines at 5,700 years B.P., and proposed the time-stratigraphic term "Sierran" for the current geologic-climate terms "Little Ice Age" and "Neoglacial." Curry's subdivision differs from recent geologic-climate classifications of the Neoglacial by Porter and Denton (1967) and by Denton and Karlen (1973), but it is consistent with the time-stratigraphic subdivision of Cook Inlet, Alaska, and with the late Pleistocene and Holocene morainal record of arctic Canada as dated by Andrews and others, (1971; 1972; 1973; and oral commun., 1974).

The work of Andrews and his students on the moraines bordering the Barnes ice-field provides for the first time a detailed and continuous continental glacial record for the time interval from Valders to the present. Thawed moraines of Cockburn (Cochrane) and other phases dated between ca. 9,000 and 6,000 years B.P., are contrasted with a complex of younger ice-cored moraines (Flint and King phases) dated between 5,000 years B.P. and the present. The boundary between the thawed morainal group and the ice-cored morainal group, bracketed between 6,000 and 5,000 years B.P. is considered to mark the culmination of a major period of warming, a
Figure 6.—Time-stratigraphy of northern Arizona compared with regional alluvial chronologies and with time-frequency curves of C-14 dated contacts in southwest alluvial and pluvial deposits. Left hand curve - data centered in overlapping 200 year intervals; right hand curve - data centered in overlapping 300 year intervals. N = 251 (unpublished compilation from Deevey and others, (eds.), Radiocarbon volume nos 1-15, Karlstrom, 1973).
Table 1.—Proposed classification boundaries compared with statistical time-stratigraphic data of southwest and midcontinent; dated sea level fluctuations (Netherlands, Jelgersma, 1961; Jelgersma and others, 1970; Denmark, Jacobsen, 1964); and maximum tidal force cycles (Stacy, 1967).

| Phase | Stage | Substage | Midcontinent n = 320 | Southwest n = 251 | Maximum tidal force peaks in anomalous years B.P. (before 1950) Average cycles ±2% | Sea level fluctuations Transgressive culminations in C-14 years B.P. Underlined figures = major transgressions
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Data centered in overlapping 200 year intervals
"thermal maximum," prior to renewed glacial activity, Curry's Pleistocene/Holocene subdivision as dated is consistent with this continental glacial evidence and coincides with the Naptowne/Alaskan interglacial stage and Pleistocene/Holocene boundary dated ca. 5,500 years ago in the Cook Inlet region (fig. 5).

The classic altithermal was dated by telecorrelation to the northern European postglacial chronology, and the boundaries were assumed, somewhat subjectively, to bracket the postglacial period when climate was warmer than present. The presented time-stratigraphic data indicate that this paleoclimatically defined and dated interval includes a series of important geoclimatic events, with maximum postglacial conditions culminating sometime between 5,000 and 6,000 years ago (Curry, R., 1970; Karlstrom, 1966). By definition (Ray and Karlstrom, 1968), this culmination marks the most appropriate time-stratigraphic boundary for subdivision of late pluvial deposition from the following "Little Ice Age" deposition.

Both "altithermal" and "Little Ice Age" have proven to be highly useful working concepts, in large part supported by the new time-stratigraphic data. As geologic-climate terms, however, they remain inappropriate for time-stratigraphic usage, and should continue to be used, if at all, in an informal sense. We find it convenient to continue the usage of altithermal in the restricted paleoclimate sense of a major warm interval centered between 5,000 and 6,000 years B.P.

CORRELATIONS WITH PROPOSED GLACIAL CHRONOLOGY OF THE LATE QUATERNARY

Based on an evaluation of C-14 dated Alaska and collated time-stratigraphic data, Karlstrom (1966, 1970a, 1970b) proposes a detailed classification for the last part of the Quaternary Period (fig. 5, column 1). The supporting data suggest a series of superposed and harmonically related paleoclimatic cycles with wavelengths of about 40,000 years, 3,000 to 3,500 years, 1,000 years, 550 years, and 280 years.

The spacing of the proposed stage, substage, and stadial time-stratigraphic boundaries in figure 5 reflects the inferred 40,000 year, the 3,300 year, and the 1,100 year climatic cycles. The inferred 550 year and 280 year cycles, although not formalized in the classification, are recorded locally in some of the most sensitive and closely dated paleoclimatic sequences available (Karlstrom, 1961, 1964; Karlstrom and others, this volume, this paper). Most of the data presented in this paper were either not published, or else not used, in deriving the proposed classification. These additional data therefore constitute a critical new test of the classification's validity, as will be assessed briefly below.
Stage cycle

The bioclimatic and geoclimatic data presented in figure 5 are compatible with the placement of the last stage in the proposed classification. The glacial cycle of ca. 40,000 years duration is bounded by the Port Talbot and the alithermal time-stratigraphic boundaries of interglacial rank. Based on available time-stratigraphic evidence, the stage deposits of Wisconsin age in the designated type locality of eastern Wisconsin are restricted to the same time interval (Karlstrom, 1969). If time-stratigraphic principles are to be consistently applied to Quaternary stratigraphy, then those previously unrecognized drifts of pre-Port Talbot age in Ohio, Indiana, and Illinois should perhaps be considered of pre-Wisconsin age rather than of early Wisconsin age, a current geologic-climate term historically inherited from a series of differing interregional correlations.

Substage cycle

The natural sampling intervals of Humptulips Bog (ca. 1,100 years) and the comparable smoothed interval of the San Augustin record (1,200 years) are such that these records should reflect parallel climatic trends, if such existed, with wavelengths of 3,000 or more years. The agreement between the broad trends of the two records and with the classification's substage time-stratigraphic boundaries of intraglacial rank is satisfactory. The combined data suggest a major secondary cycle of about 3,300 years duration, 12 of which occur between ca. 45,500 and 5,500 years B.P., and one and a part of another since 5,500 years B.P.

That the midcontinental record may reflect this same paleoclimatic cycle is indicated by important C-14 dated nonglacial intervals in type localities of the Michigan Lobe and contiguous Canadian morainal deposits. As shown in figure 5, columns 1 and 12, the intraglacial time-stratigraphic boundaries closely coincide with dated basal drift contacts (ca. 32,000 and ca. 29,000 years B.P.) of early Wisconsin tills of Rockian age (Black, 1962); with basal organics in the type silt section of Farmdale age in western Illinois (ca. 25,500 years B.P.); and with basal organics of the overlying Morton Loess section (ca. 22,500 years B.P.). Wood incorporated in basal Shelbyville till exposed in the same locality is dated at ca. 19,000 years B.P. marking the Gardena intraglacial interval of Leighton (1960).

Major ice readvances that deposited the Bloomington, Cary, and Mankato drifts are seemingly bracketed between ca. 15,500 years B.P. (Havana intraglacial: Karlstrom, 1961) and Two Creeks time (ca. 12,500 years B.P.: Frye and others, 1965). Contemporaneous with the Havana intraglacial is the major Lake Erie glacial retreat interval dated ca. 15,500 years B.P. in the Erie Basin (Dreimanis and Karrow, 1972).
Deposits of the next natural grouping of moraines of Valders age in Wisconsin are bracketed between Two Creeks and Timiskaming time (ca. 9,000 years B.P., Karlstrom, 1956), which just preceded the Cochrane (Cockburn) advances (Hughes, 1956; Falconer and others, 1965; Andrews and others, 1972). The last natural broad grouping of moraines in the continental glacial record are the post-5,500 years B.P. ice-cored moraines that border the Barnes ice-field. The ice-cored moraines are further divided into two main groups, the Flint phase and the King phase. The time-boundary separating the King and Flint phases based on lichen-dated moraines lies between 2,000 and 2,500 years B.P. (Andrews, pers. commun., 1974). As dated it is the correlative of the redefined Tustumena/Tunnel intraglacial boundary of Cook Inlet, Alaska, (Karlstrom, 1970b).

**Stadial and phase cycles**

The stadial cycle of ca. 1,100 years is strongly suggested by the secondary fluctuations of many of the southwest pollen and pluvial records (fig. 5, curves 2, 4, 5, 6, 7, 8, and 10), but only in those parts of the records with appropriate sample spacing and close stratigraphic definition (≤450 year intervals). The phase cycle of ca. 550 years is also strongly suggested but only in those relatively few records with local sample spacing and time-resolution of ≤200 years (fig. 5, curves 2, 6, and 10).

C-14 dated paleosols, unconformities, and basal contacts described from different areas in the Southwest (fig. 6) seem to cluster in certain restricted time intervals suggesting contemporaneous depositional impulses of regional significance. If so, C-14 dated alluvial and pluvial contacts throughout the southwest should cluster in these same time intervals. Wood, organic silt, charcoal, and peat samples associated with basal contacts of southwest pluvial and alluvial deposits as reported in the radiocarbon literature are plotted in the form of a time-frequency diagram (paleohydrologic curve, figure 6). The resulting pattern, best defined in those time intervals with most samples, shows pronounced clustering with most modal peaks coincident within ±100 years of the proposed interstadial time-stratigraphic boundaries, and subordinate clustering of approximately half this wavelength with most modal peaks coincident within ±100 years of the proposed phase time-stratigraphic boundaries (table 1).

At this level of analysis, correlation with events in the midcontinental glacial record is handicapped by the more generalized classification system used by most midcontinental researchers, and by the assumption that some of the recorded glacial advances such as the Valders and the Cochrane (Prest, 1970; Wright, 1971; Andrews, 1973) may record nonclimatic surges. However, the detailed reconstructions of the Lake Agassiz record (Prest, 1970), the preliminary morainal chronology of the Lake Erie Lobe (Mörner, 1970), and the C-14 and lichen-dated morainal sequence of arctic Canada (Andrews and others,
1972) reproduce many of the same short term stadial and phase
pulsations as recorded in the Southwest and in Alaska. The St.
Crois, Automba, Split Rock, and Nickerson glacial deposits as dated
in Minnesota (Wright, 1971) are the time-stratigraphic equivalents
of the Moosehorn III, and Killey I, II, and III stadial deposits.
The C-14 dated Bemis and Algona morainal deposits of the Des Moines
Lobe (Ruhe, 1969) are the time-stratigraphic equivalents of the
recessional Cary and Mankota deposits of the Lake Michigan lobe
region and of the Killey II and III stadial deposits in Alaska.
Based on C-14 dated boundaries, the type morainal deposits of Valders
age in southern Wisconsin correlate with the outermost Wisconsin
moraines of the James lobe in the Dakotas, and either with the
recessional Big Stone or else younger moraines of the Des Moines lobe
region. The combined morphologic and time-stratigraphic data
therefore suggest synchronous oscillation of ice margins, but also
record distinct regional differences in the timing of maximum advances;
probably attributable to changing topographic controls on frontal ice
dynamics and to shifting nourishment centers (Karlstrom, 1968,

In an analysis of published C-14 data, Karlstrom (1970a, 1970b)
notes that most of the dated glacial, fluvial, and lacustrine contact
deposits in the midcontinental drift area cluster around the proposed
interstadial time-stratigraphic boundaries (fig. 7). A more
comprehensive but yet preliminary analysis of the midcontinental time­
stratigraphic data is presented as a time-frequency diagram in
figure 5, column 11. The part of the curve that postdates ca. 14,000
years B.P. appears fairly well established, but refinements should
result from additional datings. The older part of the curve is
not as well controlled and includes dates with larger standard
deviations. The modal peaks of dated basal contacts generally
coincide with the midcontinental time-stratigraphic boundaries
derived from dated type and reference section data (fig. 5, column 12)
suggesting that these boundaries separate important contemporaneous
depositional events of regional significance. However these
boundaries differ in part from those defined in Frye and Willman's
(1973) recent revision of the chronology of the Lake Michigan
glacial lobe wherein different, and in some cases composite,
reference sections are used in naming and dating their subdivisions.

The midcontinental time-frequency curve also is similar to that
of the southwest record (table 1) strongly suggesting synchronous
depositional impulses, concurrently with continental glacial
advances, across the breadth of the North American continent.
Figure 7.—Late Quaternary classification and correlations (from Karlstrom, 1970b).
Theoretical considerations

C-14 dated stratigraphic evidence early suggested correlation of the Quaternary depositional events in upper latitudes with the broad solar insolation trends of Milankovitch (obliquity cycle = c. 40,000 years) and in some harmonic, or near-harmonic, fashion with Pettersson's (1914) shorter term tidal force cycles of ca. 1,700 and 90 years (Karlstrom, 1955, 1956, 1961; Fairbridge, 1961). More recent celestial mechanics calculations by Stacy (1967) reveal major tidal force cycles with average wavelengths of 556 years, 1,112 years, and 1,668 years which originate in Pettersson's calculated A.D. 1433 (517 years B.P.) interval (called the zero-check cycle by Stacy). These cycles are not precisely periodic because they vary ±2 percent from mean value depending on precessional movements of the Earth's geographic pole. The combined bioclimatic and geoclimatic data presented in this paper support direct correlation with these tidal force cycles (fig. 5, table 1).

The mechanisms by which shifts in solar insolation and tidal intensity affect climate remain somewhat enigmatic (Karlstrom, 1961; Fairbridge, 1967). The statistical relationship between worldwide precipitation and lunar phases (Brier, 1967), however, indicates that tidal effects may trigger changes in atmospheric circulation patterns directly affecting precipitation rates. Whatever the astroclimatic cause-effect relationship, direct or indirect, we consider that the time-stratigraphic data independently support the reality of the proposed stage, substage, stadial, and phase cycles. If, as suggested by our empirical data, the Milankovitch-Pettersson astroclimatic theory (Karlstrom, 1961) is correct, we have a powerful tool for both paleoclimatic classification and for predictive climatology.

In recent years serious attempts have been made by Quaternary researchers to predict probable future climate on the basis of past climatic patterns (Kukla and others, 1972). Because of differing chronological reconstructions and different assumptions used in interpretations, there is as yet little concensus on the probable course of future climatic trends. According to the proposed Milankovitch-Pettersson climatic theory (perhaps more appropriately named the Planetary-Solar Climate Working Hypothesis), the altithermal represents the culmination of increasing warmth and dryness in postpluvial time. The "Little Ice Age" represents the initial phases of generally deepening glacio-pluvial conditions. We are now within 25 years of the culmination of the current, generally recognized drought interval, marking the center of an interstadial interval. From that point forward into the future, climatic conditions in the southwest generally should become wetter and cooler, but with many superposed peneperridic intervals of differing amplitudes but progressively lessening "drought" conditions.
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A RESUME OF THE ARCHAEOLOGY OF NORTHERN ARIZONA

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ABSTRACT

A synthesis of the archaeology of northern Arizona is presented in terms of the Archaic Desert Culture and the post-A.D. 1 Anasazi and Hakataya cultural traditions of the prehistoric Indians of the Colorado Plateau. Specific reference is made to cultural dynamics and population movements that may have occurred because of changing paleoenvironmental conditions, especially around A.D. 1150 and following A.D. 1300.

INTRODUCTION

The archaeology and prehistory of northern Arizona and its environs on the Colorado Plateau have been studied for almost a century. The history of these studies has recently been summarized (Martin and Plog, 1973, p. 3-34). This paper summarizes prehistoric cultural development in the northern Southwest, especially northern Arizona, and the problems that archaeologists face in attempting to classify this development. It is intended, furthermore, as a prelude to the following paper (Karlstrom and others, this volume), in which paleoenvironmental data are related to archaeological data in a specific locale, Black Mesa, south of Kayenta, Arizona.

Until almost 10 or 15 years ago, archaeologists working in the Southwest struggled mainly with problems of time and space. Important strides toward the ordering of the data were made in the early decades of this century that provided for many locales the important chronologically and spatially documented culture-history upon which archaeologists today base their more anthropologically oriented behavioral studies.

One of the first, and frequently overlooked, attempts to order southwestern prehistoric cultures spatially was that of Hewett (1908). He attempted to classify prehistoric southwestern peoples in terms of physiographic units: those residing in the San Juan, the Little Colorado, the Rio Grande, the Gila, and the Chihuahua areas. It is apparent that Kidder (1924), in his classic introduction to the study of southwestern archaeology, was elaborating upon Hewett's culture areas; only two of Kidder's groupings, the "northern" and "eastern peripheral," are additions to those originally proposed by Hewett.

A relative chronologic framework, at least for specific localities, was developed in the second decade of this century when Nelson (1914) applied stratigraphic techniques to his excavations in the Galisteo Basin, south of Santa Fe, New Mexico. This was soon followed by the excellent, but again largely overlooked, studies of "horizontal" stratigraphy in the Zuni area by Kroeber (1916) and Spier (1917). These important pursuits were overshadowed by the initial developments of absolute tree-ring chronologies applied to southwestern prehistory by Douglass (1929) and by Haury and Hargrave (1931).
Meanwhile, localized differentiations of prehistoric cultural groups were being clarified. Fewkes (1909, 1911, 1916) described excavations in Mesa Verde. Kidder and Guernsey (1919) explored the Kayenta region of northeastern Arizona, and Judd (1930) followed by excavating Betatakin ruin. In northwestern New Mexico, Morris (1928) excavated Aztec ruin and, nearby to the south and east, Roberts (1929, 1930, 1931) reported his excavations at Shabik'eschee Village in Chaco Canyon, the Piedra district east of Durango, and the Kiatuthlanna region near the Arizona-New Mexico border. The prehistoric ruins in the area around Flagstaff were under intensive study (Colton, 1932), and Martin (1936) had begun his detailed investigations north of Mesa Verde in the Ackmen-Lowry district. These are but a few examples of the research that was being conducted in the Colorado Plateau. Most of this effort was oriented toward description of the ruins and their cultural contents in the several regions.

It was not until 1927, however, that archaeologists undertook a broad, almost pan-southwestern, culturally oriented scheme of classification. Kidder (1927) convened a conference of southwestern archaeologists which produced a system, the Pecos Classification that remains in use to this day although somewhat altered. The initial Pecos Conference was directed mainly at an attempt to arrive at "a basis for more precise definition of culture-stages" through "agreement as to diagnostic culture-traits" (Kidder, 1927, p. 490). Yet, temporal changes were not ignored as the conferees tabulated elements of architecture, village type, sandals, pictographs and ceramics. The resulting Pecos Classification consists of numbered stages (Basket Maker I, II, III, Pueblo I, II, III, IV, and V) of Pueblo development, and it remains in wide use today in spite of criticism. For example, Martin and Plog (1973, p. 30) remark that the Pecos Classification "has long since ceased to be scientifically valid" yet refer repeatedly to its designations. Regardless of the criticisms, archaeologists actively engaged in Pueblo archaeological research today utilize the Pecos system with good general understanding, realizing that the absolute chronology of the several stages varies somewhat from area to area.

Another system of classification is commonly used in an attempt to obtain more local and regional precision. This organization, from the general to the specific, was proposed by the Gladwins (1934) and later modified by Colton (1939). It defines temporally and spatially oriented phases (basic, temporally delineated units in the culture of an area) grouped into branches (geographically homogeneous but showing cultural variation through time), related branches joined into a stem, and related stems into a root. This, of course, was an attempt to also indicate cultural relationships.

The Gladwin-Colton classification also has been critized by Brew (1946, p. 44-66) and more cogently by Taylor (1948, p. 113-151), and other schemes have been proposed (Martin and Rinaldo, 1951; Daifuku, 1952).
AREA CULTURAL DEVELOPMENT

We shall confine our descriptions of cultural development to the several generally recognized prehistoric branches or traditions on the Colorado Plateau. There is general acceptance by archaeologists working in the Southwest that cultural dynamics prior to about A.D. 1 may best be described in terms of Paleo-Indian stage followed by a Desert Culture tradition and that cultures since the beginning of the Christian era may be discussed in terms of recognizable branches.

Paleo-Indian and Desert Culture

There is little definitive evidence for late Pleistocene big-game hunters in northern Arizona. A few fluted projectile points of a Clovis and Folsom affinity have been found near Kayenta, the Hopi villages, and Sanders (Danson, 1961; Olson, 1964; Ayres, 1966; Gumerman, 1966); but these have not been recovered from their original cultural contexts. Since mammoth, if not bison, remains have been reported from this region (Gumerman, 1966), the presence of at least Clovis hunters here may be postulated. Fortunately, other areas of the Southwest have yielded archaeological data correlated with paleoenvironmental conditions for the Anathermal and Altithermal (see for example Martin, 1963).

The Desert Culture, first described for the Great Basin (Jennings and Norbeck, 1955), pertains to a widespread and long-lived pattern of existence involving the hunting of small game and an increasing dependence upon wild vegetable resources for subsistence. Grinding stones, small stemmed indented-base and side-notched projectile points, and woven baskets, sandals and mats are "hallmarks" of the culture. While it may have developed from the earlier Paleo-Indian tradition, it is first recognizable in the northern Southwest some seven or eight thousand years ago.

North and east of Navajo Mountain, archaeologists from the Museum of Northern Arizona excavated two caves (Sand Dune and Dust Devil) that yielded artifacts designated as the Desha Complex. These included shallow basin milling stones, side-notched projectile points, sandals, and basketry. Radiocarbon dates of the sandals indicate an age of between 5,000 and 6,000 B.C. (Lindsay, and others, 1968).

Another Desert Culture Tradition, the Pinto Basin Complex, is also known from northern Arizona. Diagnostic stemmed indented-base points of this complex have been recovered on the surface of sites near Grand Canyon (McNutt and Euler, 1966) and near Window Rock on the Navajo Reservation (Ward, 1971). It has been suggested that the makers of 3,000- to 4,000 old split-willow-twig figurines recovered in stratigraphic context from sites in Grand Canyon, Walnut Canyon, and Sycamore Canyon in northern Arizona, as well as from caves in eastern California and Nevada, may have been Pinto Complex hunters (Euler and Olson, 1965). Radiocarbon dates for the figurines agree well with postulated dates for the Pinto Complex.
In 1972, a pre-ceramic archaic campsite was excavated by archaeologists from the Black Mesa Archaeological Project in Klethla Valley, immediately off the northern escarpment of Black Mesa. A blade, as yet unidentified as to raw material, was recovered from a charcoal hearth that has been radiocarbon dated at 785 B.C. Associated with the hearth were large quantities of split and wasted bone and many large, square-cut shell beads. We are unable to associate this site directly with known complexes of the Desert Culture; however it seems likely that it is related.

Some components of this southwestern Desert Culture are difficult to understand because material traits are scanty. However, some aspects of a Desert Culture tradition undoubtedly led to the development of some of the later cultural manifestations of the post-A.D. 1 northern Southwest.

Post-A.D. 1 cultures

The prehistoric cultures of northern Arizona and immediately neighboring regions dating from about A.D. 1 may be conveniently described in terms of several traditions (fig. 1). These will be discussed briefly with an emphasis upon cultural dynamics, especially those relating to widespread cultural changes, movements of people, and abandonments of areas. We hope in this way to bring archaeological data to bear upon possible paleoenvironmental changes that may have had a causal relationship to these dynamics. We shall confine our descriptions to those traditions or sub-cultures that have been included in two prehistoric cultural roots, the Hakataya (Schroeder, 1957) and the Anasazi (Kidder, 1936, p. 590; Colton, 1939, p. 9). Because of space limitations we have eliminated discussion of recent archaeological research by the Field Museum in the area of Vernon, Arizona, near Springerville.

The Hakataya Root

The Hakataya Root includes those peoples who lived in Arizona from the Colorado River east to the territory of the Anasazi and from the Gila River north to the Grand Canyon. They subsisted upon marginal agriculture, the hunting of small game, and the gathering of wild food plants. Food was roasted in subterranean stone-lined pits or cooked in undecorated ceramic vessels smoothed by a paddle-and-anvil technique, and fired in atmospheres that produced gray or brown surface colors. Mortars, milling stones, and "one-hand" manos were used to reduce uncooked foods. Other stone implements included percussion-flaked choppers and small, usually side-notched arrow points. Settlement patterns were typically scattered individual household units constructed of brush; rock-shelters were also in common use. The dead were usually, but not always, cremated.
Figure 1.—Approximate distribution of prehistoric cultural traditions in the northern Southwest at ca. A.D. 1150.
The Cerbat Tradition.—On the western edge of the Colorado Plateau were peoples of the Cerbat Tradition. From about A.D. 700 to 1150 they occupied a riverine and desert environment west of the plateau. Between A.D. 1150 and 1300 they moved eastward, ultimately inhabiting the territory from the Black Mountains bordering the Colorado River to the Little Colorado River and from near the Bill Williams Fork northward to the Grand Canyon. In this broad and varied region they maintained a stable culture, becoming known in historic times as the Pai, the Walapai and Havasupai, until they were disrupted by Anglo-American conquest in the 1860's. So unchanging was their way of life that they have been described as cultural conservatives in environmental diversity. They hunted deer, mountain sheep and small animals; farmed usually through spring-fed irrigation channels; and gathered a host of edible wild plants. They inhabited rock-shelters or brush wikiups, made undecorated brown pottery, and used, as diagnostic stone implements, basin milling stones with "one-hand" manos and small triangular side-notched projectile points. Their varied subsistence techniques enabled them to live unchangingly in a variety of environments and under what probably were varying climatic conditions for at least 1,100 years (Euler, 1958).

The Cohonina Tradition.—The Cohonina lived primarily but not exclusively to the east of the Cerbat and occupied the western plateau region as well as some areas west of the uplands, from about A.D. 700 (McGregor, 1951) until about A.D. 1150 (Euler, 1958). In contrast to the Cerbat, the people of the Cohonina Tradition, experienced a rather severe cultural change about A.D. 1150. Since they were first described (Hargrave, 1937), the Cohonina have remained somewhat enigmatic. Although there is an Anasazi cast to much of their material remains owing to interaction with their neighbors to the east, their basic way of life appears to have included very shallow surface dwellings as well as masonry "forts," troughed metates, gray pottery, and triangulary, unnotched projectile points. Undoubtedly, they engaged in some minimal hunting, farming, as well as the gathering of native plants. Both McGregor (1967, p. 136-137) and Schwartz (1959) have suggested that the Cohonina, around A.D. 1150, migrated into Grand Canyon to become the Havasupai. Euler (1958), on the other hand, believes that Cohonina culture, as described, disappeared at that date, to be supplanted by the Cerbat who, in turn, were the direct ancestors of both the Havasupai and Walapai. In many stratified sites, the diagnostic artifacts of Cohonina culture disappeared about A.D. 1150 (in terms of intrusive ceramic dating), to be replaced by the distinctively different Cerbat traits which persisted into historic times (Euler, 1958). Certainly, something occurred in northwestern Arizona in the twelfth century to cause these demonstrable cultural changes.
The Prescott Tradition.—To the south of the Cohonina is the territory of the Prescott Tradition. Little archaeological work has been done here, but from what is known it would appear that this was another prehistoric culture that varied little from about A.D. 900 to 1200, the time span for which we have evidence. These dates are based primarily upon intrusive ceramic types and may not be particularly accurate; only a few tree-ring dates, near A.D. 1200, have been obtained (Bannister, Gell, and Hannah, 1966, p. 15). The Prescott people, like the Cohonina, had a number of Anasazi-like features indicating interaction with those Pueblo people, first living in shallow pithouses, then building surface masonry pueblos and oval rock outlines. Maize agriculture, coupled with hunting and gathering, constituted their subsistence base. Their ceramics were made with a paddle-and-anvil technique, fired largely in an uncontrolled atmosphere, and crudely painted in black or red (Spicer and Caywood, 1936; Euler and Dobyns, 1962). Shortly after A.D. 1200 there is no further evidence of human occupation in the Prescott area until around A.D. 1600, when Spaniards found Yavapai Indians there; there is no evidence to relate these Yuman-speaking Indians to the prehistoric Prescott Tradition.

The Sinagua Tradition.—The archaeology of the Sinagua culture, in the vicinity of Flagstaff and to the south in the Verde Valley, is complicated. Detailed studies by Colton (1946) and his associates at the Museum of Northern Arizona as well as a later, brief synthesis by Schroder (1961) have, however, given us a general understanding of the Sinagua culture from about A.D. 500 until approximately A.D. 1200 in the Flagstaff area and about A.D. 1400 in the Verde Valley to the south. Because the eruption of Sunset Crater immediately east of Flagstaff about A.D. 1064 caused a major physiographic change in the Sinagua area, it has been convenient to discuss their cultural patterns in terms of pre-eruptive and post-eruptive periods (Schroeder, 1961). However, current researchers in the Sinagua area feel that other environmental and cultural factors, in addition to the eruption, were responsible for many of the post-eruptive changes (Peter Pilles, oral commun., 1974). The earliest Sinagua sites, consisting of shallow circular pithouses, are estimated to date from A.D. 500 to 700. This early culture may have been derived from the upper Little Colorado valley. By the beginning of the eighth century, however, the Sinagua, according to Schroeder (1961, p. 62), acquired traits from southern Arizona. Pithouses were square and timber-lined. Until the Sunset Crater eruption and ash fall, Sinagua culture changed very little. In addition to pithouse dwelling, the Sinagua farmed, hunted, and gathered wild vegetal foods. They produced quantities of undecorated red and brown pottery, finished with a paddle-and-anvil technique, and occasionally traded for Anasazi decorated pottery from the north and east.
After the eruption of Sunset Crater, what has sometimes been described as a prehistoric "land rush" occurred, as peoples from other cultures moved into the area covered by the Sunset ash fall to take advantage of its moisture-conserving qualities. A small group of Hohokam peoples settled at Winona Village about A.D. 1070 (McGregor, 1941), and Anasazi and Mogollon traits from the north and east were also present. The blending of these peoples and traits accounted for definite changes in the Sinagua cultural pattern. Pithouses were abandoned in favor of surface masonry pueblos, hundreds of which dotted the Sinagua region at this time.

Between A.D. 1125 and 1200 most Sinagua sites were abandoned and many of the people moved south into the Verde Valley, where Hohokam populations had preceded them (Schroeder, 1960). Here they built large pueblos such as Honanki, Tuzigoot, and Montezuma's Castle, at least the latter two of which were occupied until the early fifteenth century (Breternitz, 1960). From that time until Spaniards contacted Yavapai Indians there in 1582, the prehistory of the Verde Valley is not known.

The Anasazi Root

North and east of the general area that we have been describing lived the somewhat better known Anasazi peoples. These cultures, sometimes referred to as Basket Makers and Pueblos, are those to which the Pecos Classification applies. In addition, we recognize three major branches of sub-cultures—the Kayenta, Mesa Verde, and Chaco—and at least two minor variations—the Virgin and Winslow.

The Kayenta Sub-Culture.—The Kayenta region of northeastern Arizona has been the scene of long and extensive investigations, but until recently little synthesis has resulted from this work (Jennings, 1966).

From about the time of Christ until about A.D. 600, the Basket Maker II Kayenta lived in shallow pithouses and surface houses of brush and mud (Ward, in press). They used dry caves for storing food and burying their dead. Their subsistence was by hunting and gathering, with some reliance on maize and squash agriculture. The Basket Maker III period was marked by increasingly sedentary life, probably due to an increasing dependence on agriculture, and by the initial manufacture of ceramics. The decorated ceramics were black on a white or red background. An architectural combination of deep pithouses and slab-lined storage pits became the norm. From Basket Maker III to Pueblo I, about A.D. 600 to 850, there was a cultural continuum with little change that extended through the entire Pueblo I period. However, by the beginning of Pueblo II, about A.D. 1000, numerous changes took place. The population increased greatly and occupied uplands as well as the well-watered lowland areas to which earlier populations had been tied. In addition, there was an increasing uniformity in the village patterns among the thousands of Pueblo II sites. These villages were commonly composed of a line of surface
masonry habitation and storage rooms, a kiva (semi-ceremonial subsurface room), and a formalized midden. There also seems to have been a greater dependence on cultivated products and a decrease in the hunting of large game animals (Gumerman and others, 1972). By about A.D. 1150 there was a large-scale abandonment of much of the northern, western, and southern regions, as populations moved to lower and wetter areas. This movement was probably due to a climatic change that proved unfavorable to dry farming. The period from A.D. 1150 to 1250 appears to have been one of massive change and population shifts, although little is known about the cultural dynamics of that century (Dean, 1970, p. 150). By A.D. 1250 the population had moved into a number of large pueblos in the open or in cliff-sheltered localities. The spectacular cliff dwellings of Kiet Siel, Betatakin, and Inscription House were occupied at this time (Judd, 1930). By A.D. 1300, however, headward arroyo cutting prevented the Kayenta people from getting water from the streams to their fields (Dean, 1970) and partly because of this they abandoned the entire region and moved to the southern edge of Black Mesa, where their present-day descendants, the Hopi, still reside. The end of Pueblo III at about A.D. 1300 marked the end of the Kayenta as a definable cultural entity.

The Virgin Sub-Culture.—The Virgin Tradition is a peripheral and short-lived sub-cultural manifestation north of the Grand Canyon extending into southern Utah and southeastern Nevada (Aikens, 1966). Before about A.D. 900 the population was relatively small and although there were minor differences in architecture and artifacts, the general cast of the sub-culture was Kayenta. For example, although ceramic designs were similar to Kayenta designs, they were less skilfully executed and the construction of the vessels themselves was less refined. Between A.D. 900 and 1150 population increased, less-productive land was utilized, and the Virgin Sub-Culture became more distinctive. Village patterns, architecture, and artifact assemblages had a distinctive character. Nevertheless, enough similarities between Kayenta and Virgin remained to indicate a common heritage and continued social interaction between the two groups. The population was never as dense nor were the architectural or material remains as well fashioned as in the Kayenta region. By about A.D. 1130 the Virgin area was abandoned, and although evidence is scant, the people probably moved southeast to the Kayenta region. This movement of population was presumably due, at least in part, to a reaction to subsistence stress based on a change in climatic regime.
The Winslow Sub-Culture.—The Winslow Tradition occupied a part of the Little Colorado desert south of Black Mesa and north of the Little Colorado River. Like the Virgin Sub-Culture, it is a minor and little known culture that seems to have had its genesis in a Kayenta base. Also like the Virgin Tradition, it was not really distinctive until about A.D. 900. Before then the area was very sparsely occupied by Kayenta peoples. By A.D. 900, however, the population expanded, and developed a distinctive architectural and ceramic style (Gumerman and Skinner, 1968). The Little Colorado desert has no major streams except for the Little Colorado River itself. Consequently, the settlement pattern was one of many small dispersed communities situated to take advantage of the localized water sources and arable land. Unlike most other Anasazi Traditions, the Winslow people constructed several specialized sites as ceremonial centers, which apparently served as integrators for this dispersed group (Gumerman, 1968). Population continued to expand until approximately A.D. 1200 or 1250, when it was no longer possible to sustain inhabitants in this largely arid area, and the region was abandoned. Presumably, the population movement was again due to a climatic change which forced the Winslow people to areas adjacent to the Little Colorado River or to the Hopi Mesas, which had more dependable water supplies. The Winslow people were able to remain in their largely unfavorable environment slightly longer than other peoples in marginal farming areas such as Black Mesa and the Virgin area, probably because the subsistence stress was more easily coped with, or perhaps was less severe in the Little Colorado desert than elsewhere.

The Grand Canyon.—The Grand Canyon is geographically central to the Cerbat, Cohonina, Kayenta, and Virgin Cultures that we have briefly described. On its rim or in the depths, archaeologists have recorded occupation by peoples representing each of these cultures. This has been summarized most recently by Euler (1967).

Slightly before A.D. 700 and continuing until about A.D. 1000, Kayenta Anasazi made sporadic explorations and limited seasonal occupation of the inner recesses of the canyon. At about the same time, the Cohonina were settling selected locations near the South Rim. About A.D. 1000, a larger influx of Kayenta peoples took place, culminating a century later in a major occupation. By A.D. 1100, the canyon was occupied on both rims, and below, primarily by the Kayenta. A few Cohonina sites were occupied on the south, and in the northwestern sections of the canyon some Virgin occupation has been noted. Cerbat sites occur in the extreme western sections. The overwhelming majority of twelfth century sites, including hundreds of surface masonry pueblos with associated storage rooms, mescal pits, and occasional kivas, were built by the Kayenta, indicating a good adaptation of people to several sub-environmental niches. All of the above dates based on ceramic designs; there is only one small group of tree-ring dates from Grand Canyon, those from Tusayan ruin on the South Rim, occupied by Kayenta peoples around A.D. 1190 (Bannister, Dean, and Robinson, 1968, p. 11).
Between A.D. 1150 and 1200 there was a general abandonment of the Grand Canyon by all except the Cerbat, who, at that time were also expanding into the upland region to the south and east. From then until historic times Grand Canyon was sparsely utilized by the Walapai and Havasupai descendants of the Cerbat on the south side of the canyon, and by the Southern Paiute who followed the Virgin Anasazi into its North Rim tributaries, and rarely by Kayenta Anasazi and their Hopi descendants gathering salt from the "mines" near the mouth of the Little Colorado River.

To this point we have confined our remarks essentially to the prehistoric cultures of northern Arizona. However, since the dynamics of the three major Anasazi traditions are more or less interrelated, we shall describe briefly the development of the two that center in southwestern Colorado and northwestern New Mexico—the Mesa Verde and the Chaco.

The Mesa Verde Sub-Culture.—The archaeology of the Mesa Verde region has received considerable attention in recent years. Studies by the University of Colorado (Lister, 1966), the Wetherill Mesa Project (Hayes, 1964), and those of the Red Rock Plateau in the triangle formed by the Colorado and the San Juan Rivers (Lipe, 1970) may be cited as examples of published research.

From the first to the early fourth century A.D., Basket Maker people were established in the Animas River Valley north of Durango (Morris and Burgh, 1954, p. 85). These Indians, of a Basket Maker II stage, lived in houses of cribbed wall and roof construction, but otherwise were similar to the Kayenta Anasazi of this period. Hunting, gathering, and corn and squash agriculture constituted their subsistence patterns.

In the sixth century A.D. there was a geographical spread of the Mesa Verde people, and Mesa Verde itself was occupied for the first time in the last quarter of that century during the Basket Maker III stage (Breternitz, 1973, p. 9). The Basket Maker III stage continued until about the mid-700's, with occupation primarily in pit house villages. It was during this period that there appears to have been an influx of people onto Mesa Verde in the early and middle A.D. 600's.

Following this, from about A.D. 700 to 900, there was continued use of mesa-top villages (as opposed to later cliff dwellings) but with a transition from pit houses to substantial above-ground structures. From then until about A.D. 1150 (Hayes, 1964, p. 100), Mesa Verde culture developed, especially in the construction of coursed masonry pueblos and underground pilastered kivas, in elaborate black-on-white ceramic vessels, and in adaptation to the environment primarily through hunting and through farming by means of extensive water control systems on the mesa tops.
From A.D. 1150 to 1300, in the late Pueblo III stage of the Mesa Verde Anasazi, one primary change took place. Whereas some masonry structures had been built and occupied in rock shelters and caves in the cliffs at earlier times, the mid-twelfth century saw a general movement into cliff dwellings. These include well-known sites such as Cliff Palace and hundreds of lesser known structures. Off Mesa Verde proper, however, large open pueblos continued to be occupied. Furthermore, during this period, or perhaps earlier (Vivian, written commun., 1970), the Mesa Verde people expanded their range south to Aztec and Chaco Canyon. By the end of the thirteenth century the Mesa Verde region was abandoned.

The Chaco Sub-Culture.—While the boundary lines are blurred, the area lying generally to the south of Mesa Verde was occupied by peoples of the Chaco tradition. Although this was a broad region, extending from the San Juan River south to Zuni, our summary will primarily involve Chaco Canyon, the "urban" hearland of the Chaco people. Archaeological investigations here have a long history, but it has been only recently that holistic patterns of cultural dynamics were studied (Vivian, 1970). The establishment of the National Park Service's Chaco Project in 1972 provides a continuation of these concerns, which should lead to a greater understanding of the cultural processes that culminated in the building of large towns such as Pueblo Bonito in the eleventh century A.D.

The Anasazi began occupation of the Chaco Canyon region before A.D. 700. Basket Maker III pithouse villages were built on the mesa tops above the canyon (Roberts, 1929), but within approximately fifty years the Chaco Anasazi had moved to the canyon floor and had abandoned their pit houses in favor of above-ground structures.

By A.D. 850 a dichotomy between "town" and "village" architecture had begun (Vivian, 1970, p. 62), with the smaller villages usually located on the south side of Chaco Wash and the large towns to the north. About A.D. 1380 major construction took place in several of the towns, and they reached their maximum sizes; Pueblo Bonito, for example, contained over 600 rooms (Judd, 1964, p. 22). Numerous religious structures, as well as small and great kivas, were associated with these multi-story pueblos. Extensive water control systems, including canals carrying runoff from the cliffs above the towns, were built to irrigate crops. Broad roads not merely foot paths, connected several areas. Anasazi from the Mesa Verde moved into Chaco Canyon and constructed several sites, such as Kin Kletso. It is clear that some of the greatest development in Anasazi Culture took place in Chaco Canyon during the eleventh century.

If we have correctly assessed the tree-ring dates for this period, the last construction in both town and village sites was in the early part of the twelfth century. "By 1130 the canyon was being abandoned and it is likely that by 1150 few persons were residing in the Chaco. With the exception of a few minor occupations
Figure 2.—Approximate demographic shifts of prehistoric cultural traditions in the northern Southwest ca. A.D. 1150-1300.
in the canyon during the next two hundred years, pueblo occupation had ended" (Vivian, 1970, p. 68). This minor settlement, after A.D. 1250, was by Mesa Verde migrants who temporarily occupied Chaco Canyon. By A.D. 1300 they had moved to "fortified sites" on Chacra Mesa to the east, where they resided until not later than A.D. 1400 (Vivian, written commun., 1970).

**SUMMARY AND CONCLUSIONS**

So far, we have said little that is new to southwestern archaeology. We have been content to describe very briefly the prehistoric cultures of the Colorado Plateau (excluding two important regions—Canyon de Chelly because, in reality, so little is known, and the Navajo Reservoir district east of Mesa Verde and north of Chaco Canyon, because of space limitations) to emphasize that important cultural changes were going on in much of the area at approximately the same times.

As we review what is now known of the archaeology of the northern Southwest, we have seen three notable events:

1. The emergence of recognizable cultural variations from a probable Desert Culture theme shortly after the beginning of the Christian era. This is seen particularly among the three primary Anasazi Sub-Cultures, and it may be that only our paucity of field investigations obscures our vision for the Hakataya. Our earliest knowledge of the Kayenta and Mesa Verde Traditions is from the third and fourth centuries A.D., while that from the Chaco is somewhat later, the eighth century. The Virgin and Winslow Traditions become truly distinctive only about A.D. 900.

2. For the Hakataya, we recognize the existence of the Cerbat and Cohonina at about A.D. 700, the Prescott at ca. A.D. 900, and the Sinagua, earlier, at ca. A.D. 500. Coupled with an obvious spread of Anasazi populations about A.D. 700, that date becomes important in terms of the expansion of human societies on the Colorado Plateau. Also, A.D. 900 seems to mark a period of population growth throughout much of the area, especially in the uplands; this suggests a climatic change favoring dry farming.

3. Some time around A.D. 1150, the Cerbat peoples moved onto the Plateau, the Cohonina disappeared from the record, Grand Canyon was abandoned by the Anasazi, the Virgin people left their homeland on the Arizona Strip, the Prescott and Sinagua were beginning to depopulate their territories, and the Chaco was abandoned (fig. 2).
Only in the Kayenta and Mesa Verde regions do we see continuing populations of any size, and even here there were shifts in population distribution. Following A.D. 1300 there was a dramatic decrease in population throughout the region, and by A.D. 1450 most of the cultures discussed had disappeared from their former territories.

This paper sets the stage for the following study in which we (Karlstrom and others, this volume), note the potential importance of the phase system as a classificatory tool and the even more potentially valuable multidisciplinary study of paleoclimatology as a factor in prehistoric cultural dynamics.

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STRUCTURAL GEOLOGY

Synopsis of the Laramide and post-Laramide structural geology of the eastern Grand Canyon, Arizona
by
Peter W. Huntoon

Structural evolution of northwest Arizona and its relation to adjacent Basin and Range province structures
by
Ivo Lucchitta

The Bright Angel and Mesa Butte fault systems of northern Arizona
by
E. M. Shoemaker, R. L. Squires, and M. J. Abrams
SYNOPSIS OF LARAMIDE AND POST-LARAMIDE STRUCTURAL GEOLOGY OF THE EASTERN GRAND CANYON, ARIZONA

by

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ABSTRACT

Three systems of post-Paleozoic structures occur in the eastern Grand Canyon: north-trending monoclines, north-trending faults, and northeast-trending faults. These are tentatively dated as Laramide, Miocene to Recent, and Pliocene or possibly Miocene to Recent, respectively.

INTRODUCTION

The Grand Canyon is eroded into the tectonically stable Colorado Plateau, which is underlain by relatively flat-lying Mesozoic and Paleozoic sedimentary rocks. The tectonic fabric of the Colorado Plateau is dominated by generally north-trending structural zones that are spaced from 24 to 64 km (15 to 40 mi) apart (Gilbert, 1875a; Dutton, 1880; Davis, 1901).

A 1,220- to 2,440-m (4,000 to 8,000 ft) cover of Mesozoic rocks has been largely stripped from the vicinity of the Grand Canyon, and the canyon is eroded through 1,065 to 1,370 m (3,500 to 4,500 ft) of Paleozoic rocks. In the eastern Grand Canyon, approximately 3,960 m (13,000 ft) of Precambrian sedimentary rocks have been exhumed, as well as an igneous and metamorphic basement complex of undertermined thickness (Walcott, 1890).

The eastern Grand Canyon is carved through the Kaibab Plateau (fig. 1), a broad, gentle antclinal uplift that plunges toward the south across the Grand Canyon. Both the eastern and western margins of the Kaibab Plateau are abruptly terminated by north-trending structures. Elevations on the Kaibab Plateau to the north of the Grand Canyon exceed 2,840 m (9,300 ft), compared to the 610- to 855-m (2,000- to 2,800 ft) elevation of the Colorado River. The Marble Platform to the east of the Kaibab Plateau occurs between 1,200 and 1,950 m (4,000 and 6,400 ft); whereas the Kanab Plateau to the west occurs between 1,400 and 2,070 m (4,600 and 6,800 ft). The Coconino Plateau, south of the eastern Grand Canyon, reaches a peak elevation of 2,260 m (7,400 ft) along the Canyon rim. The surfaces of the Kaibab, Kanab, and Coconino Plateaus, as well as of the Marble Platform, are largely stripped surfaces developed on the Permian Kaibab Formation.

Three systems of deep-seated structures disrupt the Paleozoic rocks in the eastern Grand Canyon: 1) a system of north-trending monoclines, 2) a system of north-trending faults that are superposed on, or lie parallel to, the monoclines, and 3) a system of northeast-trending faults. As used here, a system is a series of structures that have parallel trends and identical structural characteristics. Presumably, members of a system are genetically related.

The absence of dating criteria handicaps interpretation of the sequence of tectonic events that disrupted the Paleozoic rocks in the eastern Grand Canyon. There are no Cenozoic volcanic rocks in the
Figure 1.—Locations of features in the eastern Grand Canyon region.
eastern Grand Canyon, and most of the Mesozoic and younger sedimentary rocks have been eroded away. Consequently, dates must be extrapolated into the region using information developed in nearby areas where stratigraphic and volcanic relationships are diagnostic. This is justified on the basis of the assumed common genetic relationship between members within a given system of structures. This notion is well rooted in the literature of the region and first appeared in a discussion by Gilbert (1875a, p. 57): "The force or forces that have produced them (the structures) are hence believed to be deep-seated, and uniform in kind and phase over large areas."

RELATIONSHIP OF CENOZOIC STRUCTURES TO PRECAMBRIAN STRUCTURES

Study of the outcrops of the Grand Canyon indicates that recurrent structural deformation tended to occur along the same zones of weakness. The record for this is particularly evident in the coincidence of Cenozoic structures with underlying Precambrian faults wherever Precambrian rocks are exposed. In many cases, more than one Precambrian or Cenozoic movement can be demonstrated along these zones. The eastern Grand Canyon literature is replete with documentation of this, (For examples, see: Walcott, 1890; Ransome 1908; Noble, 1914; Maxson and Campbell, 1933; Babenroth and Strahler, 1945; Van Gundy, 1946; Maxson, 1961; Huntoon, 1969, 1971.) The structural trends in the Paleozoic and younger rocks in the region were inherited from structural patterns that were established during Precambrian time (Hodgson, 1961).

LARGE-SCALE STRUCTURES

North-trending monoclines

Monoclinal folds are the largest structures that disrupt the Paleozoic rocks of the eastern Grand Canyon (fig. 2). All the monoclines in the region except the Supai dip toward the east. The displacements range up to 823 m (2,700 ft) across the structures in the vicinity of the Grand Canyon, and dips of the folded strata range from a few degrees to slightly overturned. In general, the monoclines trend northward; locally they branch and are sinuous in trend.

The early workers, including Powell (1873 and 1876), Gilbert (1875a), and Dutton (1880), regarded the monoclines and faults of the region to be homologous and to have resulted from vertical crustal movements. Walcott (1890) concluded that the East Kaibab monocline developed in response to renewed movement along an underlying Precambrian fault (Butte fault in Nankoweap Amphitheater), and that the movement was largely vertical.

Later workers, including Noble (1914), Maxson (1961), and Huntoon, (1971), recognized that the monoclines in the Paleozoic rocks
Figure 2.—North-trending monoclines in the eastern Grand Canyon region. Reverse faults underlying monoclines omitted.
of the Grand Canyon grade downward into reverse faults in the Precambrian rocks. This implies that these structures resulted from horizontal compression.

The monoclines are well exposed at several locations in the Grand Canyon. Characteristically, the Permian Kaibab Formation is bent into a sweeping fold with dips that exceed 25° in only a few places. However, the anticlinal and synclinal axial planes converge downward to a point that commonly lies near or in the basal Cambrian Tapeats Sandstone. Consequently, with depth, the width of the fold diminishes and the dip of the deformed strata increases. The lowermost Paleozoic rocks are commonly folded to the vertical or even overturned. The reverse faults associated with the monoclines commonly terminate upward in the Cambrian Bright Angel Shale, although they may extend as high as the Mississippian Redwall Limestone, or possibly into the Pennsylvanian-Permian Supai Formation. This is the case along the East Kaibab monocline in parts of the eastern Grand Canyon, where the total displacement across the structure is as great as 823 m (2,700 ft). In such areas, the Paleozoic rocks have undergone low-grade metamorphism (Walcott, 1890).

The transition from a reverse fault to a broad monoclinal fold is clearly exposed along the East Kaibab monocline in the walls of Palisades Canyon south of the Colorado River. Displacement on the underlying Butte fault there is 122 m (400 ft) at the level of the top of the Precambrian rocks. The Cambrian units are tightly folded and faulted. Dips in the Bright Angel Shale adjacent to the fault are vertical, but the fault dies out in the upper Cambrian rocks, where the fold becomes broader and smoother. Displacement across the monocline at the top of the 1,220-m (4,000 ft)-thick Paleozoic section is similar to that along the underlying reverse fault, but the fold is about 1.6 km (1 mi) wide and the dips do not exceed 15°.

The sinuosity and branching that are characteristic of the monoclines in the eastern Grand Canyon result from pre-existing trends of the Precambrian faults that were rejuvenated as the monoclines developed (Huntoon, 1971). Interestingly, the Precambrian faults were normal with the west side downthrown (Walcott, 1890; Noble, 1914; Van Gundy, 1946; and Maxson, 1961).

The monoclines in the eastern Grand Canyon are generally attributed to the Laramide orogenies. To date, the only evidence that has been advanced to support this interpretation comes from an unconformity between the Paleocene Pine Hollow Formation and the Paleocene-Eocene Wasatch Formation along a northward extension of the East Kaibab monocline at Canaan Peak, Utah (Gregory and Moore, 1931; Strahler, 1944b; and Babenroth and Strahler, 1945). The unconformity, however, has been attributed by Bowers (1972) to large-scale gravitational sliding, so the date may be invalid. A Laramide age for the eastern Grand Canyon monoclines is still attractive because other folds in the same system have been dated as Laramide elsewhere on the Colorado Plateau. Included are the Kanarra fold along the Hurricane fault (Gardner, 1941; and Averitt, 1964), Waterpocket monocline in Utah (Gilbert, 1877), and various monoclines to the east of the Grand Canyon (Kelley, 1955).
North-trending faults

A system of north-trending faults that are colinear with or parallel to the north-trending monoclines occurs throughout the region (fig. 3). In general, the west sides are downthrown and the faults are steeply dipping normal faults (fig. 4). Displacements range upward to about 400 m (1,300 ft) along the West Kaibab fault zone; however, displacement along the Central Kaibab faults is generally less than 120 m (400 ft). The West Kaibab faults are colinear with all segments of the West Kaibab monoclines and generally occur along the synclinal hinges of the folds. The en echelon components of the Central Kaibab faults southeast of Jacob Lake trend obliquely across the western branch of the East Kaibab monocline.

Powell (1873 and 1875), Gilbert (1875a and 1875b), and Dutton (1880 and 1882) described and traced the West Kaibab fault zone. Dutton (1882) observed that strata are downturned toward the fault plane on the west side of the structure and concluded that the faults postdate a monocline of the same trend. Noble (1914) observed the same phenomenon in Shinumo Amphitheater but failed to notice that beds east of the fault are upturned. Strahler (1948) concluded that no monoclines are associated with the West Kaibab faults and, in fact, that the downturned strata represent sag along the faults. Huntoon (1969) reexamined the question and found ample evidence for pre-fault monoclines by identifying upturned strata along the eastern sides of the faults in association with downturned strata on the west.

Downfaulting along many segments of the monoclines in the eastern Grand Canyon did not extend downward to the Precambrian rocks (fig. 4E). Colinear normal faults that exist only in the upper Paleozoic rocks occur along the Phantom-Grandview monocline, southern parts of the West Kaibab monoclines, and other localities. These faults are generally restricted to the synclinal hinges of the monoclines and terminate in the upper Redwall Limestone. The underlying Paleozoic rocks are commonly intensely jointed but not displaced. Seemingly, downfaulting began but did not progress enough to sever the entire section. In essence, the lower Paleozoic units unfolded slightly to accommodate the normal movement at depth, whereas the upper Paleozoic rocks yielded along faults.

The Central Kaibab faults are less well known, primarily because they are obscured by cover and vegetation. Strahler (1944a), as well as earlier workers, did not recognize the existence of continuous faults through the parks of the Central Kaibab Plateau; however, his later paper on the West Kaibab fault zone treats the major components in the Central Kaibab group (Strahler, 1948). Maxson (1967) mapped several of the Central Kaibab faults.

Two of the Central Kaibab faults are reverse faults, with the northeast side downthrown where they terminate in Bright Angel Canyon. Both dip steeply toward the southwest. One can be traced to a point 10 km (6 mi) north of the canyon rim where it scissors. Northward from the scissors point, the west block is downthrown and the structure is identical with other members of the north-trending system.
Figure 3.—Principal north-trending normal faults in the eastern Grand Canyon region.
Figure 4.—Recurrent deformation along the north-trending structures. A, Precambrian normal faulting. B, Erosion of Precambrian rocks and deposition of Paleozoic and Mesozoic strata. C, Monoclinal folding of Paleozoic and Mesozoic strata due to reverse faulting at depth. D, Normal faulting along the monoclines. E, Incomplete normal faulting along some monoclines due to slight normal movement along the underlying faults.
Davis (1901) recognized that the Kaibab Plateau and eastern Grand Canyon lie in a transition zone between great unbroken monoclines to the east and downfaulted monoclines to the west. The downfaulting occurred along the north-trending system of faults. Because many of these faults are colinear with the pre-existing monoclines, it is evident that recurrent post-Paleozoic deformation has occurred along the same basement structures. Dutton (1880) observed that the faults reversed the displacements along the monoclines. The north-trending faults are normal, which implies a tensional stress regime for their formation.

Dating of the north-trending faults in the eastern Grand Canyon has been, and continues to be, a subject of conjecture. No faults within the West or Central Kaibab groups have been dated by stratigraphic or radiometric means; however, insight into their age can be gained from datable faults in the north-trending system to the west. The renowned Hurricane fault of the western Grand Canyon is structurally similar to the Central and West Kaibab faults. Through the efforts of Dutton (1882), Davis (1903), Huntington and Goldthwait (1903), Gardner (1941), Averitt (1964), Kurie (1966), Hamblin (1970), Young (1970), and Lovejoy (1973), a partial chronology of events along this structure has been established. Large-scale normal faulting along the Hurricane fault appears to have commenced during the Miocene, before 18.3 million years ago. Faulting along the structure has been continuous since then. According to Hamblin (1970), as much as 75 percent of the displacement along the Hurricane fault in the vicinity of the Grand Canyon occurred during Pliocene time. A similar but less complete record of recurrent deformation exists along the parallel Toroweap-Sevier fault to the east, which indicated to Koons (1945) that early movement along this fault may have taken place during late Miocene or early Pliocene time. Recurrent movements have been recorded along the Toroweap fault in the Grand Canyon by Davis (1903), McKee and Schenk (1942), Koons (1964), and Hamblin (1970).

These observations as well as others along similar faults are summarized by McKee and others (1964) and indicate that the north-trending system of normal faults in the western part of the Colorado Plateau probably formed during Miocene time. According to the evidence along the Hurricane fault, the major movements appear to have taken place during Pliocene time. It is clear that faulting has continued to the present along the western Grand Canyon structures, and the same is probably true for the larger structures in the eastern Grand Canyon. However, evidence of recent faulting is presently unknown along the West and Central Kaibab faults.
Northeast-trending faults

The northeast-trending faults in the eastern Grand Canyon (fig. 5) are characteristically steeply dipping normal faults with no preferred downthrown side. The displacements along the faults in this system are generally less than 92 m (300 ft), but the faults are remarkable for their linearity and the influence they have on the topography. The Sinyala fault has a maximum displacement on the order of 8 m (25 ft), yet it can be traced for over 48 km (30 mi). The Eminence and Fence faults are boundaries of a large tensional structure that is shattered by discontinuous high-angle normal faults.

The best-known member in this system is the Bright Angel fault, which has a record of Precambrian, possible Paleozoic, and Cenozoic movements (Ransome, 1908; Darton, 1910; Noble, 1914; Maxson and Campbell, 1933; Van Gundy, 1946; and Maxson, 1961). Because criteria for positive-dating are missing, only one episode of Cenozoic movement, in which the relative displacement of the east block was downward 60 m (200 ft) at Grand Canyon Village, can be demonstrated along the fault. Recent work by Huntoon and Sears demonstrates a minimum of three periods of deep-seated Precambrian deformation along the structure at the Colorado River; relative net displacement of the west block was downward approximately 400 m (1,300 ft).

The Cenozoic northeast-trending faults are clearly tensional structures; however, they are virtually impossible to date in the vicinity of the Grand Canyon. The geometric relationship between the Bright Angel fault and the southern termini of the reverse portions of the Central Kaibab faults indicates that the Bright Angel fault postdates the major movements along the Central Kaibab group. The argument supporting this is circumstantial and presumes that the reverse portions of the Central Kaibab fault that terminate in Bright Angel Canyon were originally normal with the west blocks downthrown as along the other faults of the north-trending system. As the Bright Angel fault developed, the east block was downthrown. In the vicinity of the East Kaibab monocline, the trend of the Bright Angel fault also changed toward the north, indicating stress complications near the fold. Part of this stress may have been relieved by reverse movement on the southern portions of the Central Kaibab faults where they terminated against the Precambrian Bright Angel fault. This caused them to be downthrown to the east. If such a sequence is accepted, the northeast system of faults postdates the major displacements along the north-trending system.

The foregoing argument suggests that faulting along northeast trends began in Pliocene time, although some of this faulting could date back into the Miocene. Reiche (1937) and Akers and others (1962) record Quaternary and possibly recent motion along faults with northeast trends in the Cameron quadrangle, southeast of the eastern Grand Canyon, thus indicating that the northeast system of faults in the region has been active to the present.
Figure 5.—Principal northeast-trending normal faults in the eastern Grand Canyon region.
The previous discussions regarding the dating of tectonic events probably raise more questions than have been answered; however, a relative chronology emerges. The first structures that are considered here were the monoclines assigned to the Laramide orogenies. The north-trending faults formed next and appear to have been active since Miocene time. The northeast-trending faults are provisionally dated as Pliocene or possibly Miocene and probably have been active to the present. Dutton's (1882, p. 260) statement still holds for the eastern Grand Canyon:

"Whatever periods may have been assigned to the antiquity of past events have been assigned provisionally only, and the inferences are almost purely hypothetical. In the Plateau Country Nature has, in some respects, been more communicative than in other regions, and has answered many questions far more fully and graciously. But here, as elsewhere, whenever we interrogate her about time other than relative, her lips are sternly closed, and her face becomes as the face of the Sphinx."

**Small-Scale Structures**

**Reverse Drag**

Probably no phenomena associated with the normal faults in the Colorado Plateau have evoked as much commentary as reverse drag. Reverse drag is a downturning of strata in the downthrown block toward the fault plane antithetic to the motion along the fault. Powell (1875) observed reverse drag along some of the normal faults that cross the Colorado River. Dutton (1882) discussed examples of reverse drag along the Hurricane, Sevier, and West Kaibab faults and suggested that the anomalous dips resulted from monoclinal folding prior to faulting. This concept was accepted by Davis (1901), and by Johnson (1909). Noble (1914), in examining a branch of the West Kaibab faults in Shinumo Amphitheater, also attributed the reverse drag he observed to a faulted monocline. Koons (1945) studied the Hurricane fault south of the Colorado River and observed upturned strata east of the fault plane as well as the downturned strata to the west and felt this verified Dutton's faulted monocline concept.

Gardner (1941) attributed reverse drag along the Hurricane fault in Utah to sag along the fault that was related to the eruption of large volumes of lava. Strahler (1948) explained the eastward dips he observed along the downthrown side of the West Kaibab faults as sagging of the downthrown block contemporaneous with faulting. Mears (1950) examined reverse drag along the Oak Creek fault south of Flagstaff, Arizona, and concluded that the folds were due to
compression during a stage of deformation following the faulting. Other hypotheses invoked to explain reverse drag in the United States are reviewed by Hamblin (1965).

Hamblin (1965), in studying reverse drag along the normal faults in the western part of the Colorado Plateau, related the sag to curvature of the fault planes with depth. He convincingly argued that if the fault plane flattens with depth and tensional stress is applied, the rock in the downdropped block will sag into the void created along the fault. Huntoon and Sears have observed a similar relationship along the Bright Angel fault at Grand Canyon Village where the Paleozoic strata dip toward the west in the downthrown eastern block. As illustrated on figure 6, the reverse drag appears to be related to a change in dip of the fault plane between the Paleozoic and Precambrian rocks. The dip of the fault in the Paleozoic strata is between 76° and 87° E., compared with 45° to 80° E. in the Precambrian rocks. As the adjacent blocks displaced under tension, a void tended to develop in the Paleozoic section that filled through sagging of the downthrown block.

The concept of faulted monoclines is, however, supported by many of the reverse drag features observed in the Grand Canyon area. Huntoon (1969) documented the existence of a faulted monocline along the West Kaibab faults, in concurrence with Dutton's and Noble's earlier interpretations. The writer is, however, convinced that reverse drag as proposed by Hamblin or Huntoon and Sears can explain moderate to steep dips on the downthrown side of normal faults. In the case of faulted monoclines such as the West Kaibab structure, the dips on the downthrown block can be further accentuated by this process during later faulting.

**Attenuation of displacements**

The well-exposed outcrops in the walls of the Grand Canyon reveal minor structural phenomena that would go unnoticed elsewhere. One such phenomenon is the tendency for faults to die out with increasing elevation. Evidence for the transition from reverse faults of large displacement in the Precambrian rocks to monoclinal folds in the Paleozoic section has been presented.

A subtle expression of a similar phenomenon is well exposed along the northeast-trending Sinyala fault. This fault is clearly a minor high-angle normal fault, with a displacement of about 5 m (15 ft) at the level of the Cambrian rocks along the Colorado River. The displacement along the fault at the tops of the 610 m (2,000 ft) cliffs adjacent to the river is less than 1.5 m (5 ft). The difference is attributed to minor flowage within the intervening rocks. The fault completely disappears higher in the Paleozoic section.

Many minor faults that traverse the deeper parts of the canyons do not crop out on the adjacent plateaus. These represent incomplete faulting of the section analogous to that along the Sinyala fault.

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Figure 6.—Development of reverse drag along the Bright Angel fault. A, Geometry prior to Cenozoic faulting. B, Faulting without sag of hanging wall. C, Faulting with sag of hanging wall.
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STRUCTURAL EVOLUTION OF NORTHWEST ARIZONA AND ITS RELATION TO ADJACENT BASIN AND RANGE PROVINCE STRUCTURES

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ABSTRACT

The southwest part of the Colorado Plateau, in northwest Arizona, is composed of structural blocks dipping very gently northeast and delimited by great north-trending normal faults whose strike lengths are measured in tens to hundreds of kilometers and displacements in hundreds to thousands of meters. Lesser faults within the blocks generally parallel the major ones. Many of the faults are associated with folding that is contrary in sense to drag folding. Near the edge of the plateau, displacement resulting cumulatively from faults, monoclines, and an increase in dip produces a notable structural elevation of this edge.

On the Shivwits Plateau, westernmost of the structural blocks, an anomalous structural belt trending northeast is defined by northeast trends of normal faults, by two great bends in the Hurricane fault, by a set of closely spaced grabens trending north-northwest, by a general absence of volcanic centers, and possibly by local minor copper mineralization. This belt is aligned with a conspicuous zone of faults trending northeast in the Basin and Range province. This zone in part has transcurrent movement and may be part of the Las Vegas shear zone.

The southwestern Colorado Plateau and the adjacent Basin and Range province are parts of crustal blocks that differ markedly in physiography, structure, crustal properties, and geologic history. Many of these differences go back to at least Paleozoic time. The boundary between the two blocks is old, relatively sharp, and relatively fixed, and probably coincides rather well with the present physiographic boundary.

Parallelism between structural features of the plateau and those of the adjacent Basin and Range suggests that both have been subject to the same stress fields, but the much more subdued deformation on the plateau indicates that it has a more competent, buttress-like crust. In the plateau block, the basement has played a dominant role in controlling structural deformation. Zones of weakness, some very old, have focused deformation even under different stress fields. Many of these zones have been active repeatedly and commonly exhibit reversed movement. Thus, the compressional stress field associated with the Laramide orogeny to the west produced reverse faulting in the basement and sympathetic folding in the sedimentary cover. The later extensional stress field associated with Basin and Range deformation to the west resulted in normal faulting along many of the same faults.
Figure 1.—Location map.
INTRODUCTION

Geologic setting

For the purposes of this report, northwestern Arizona is defined as and limited to that part of the Colorado Plateau province that lies within Arizona and west of the Kaibab Plateau (fig. 1). This area is underlain by little-deformed sedimentary rocks of Paleozoic and Mesozoic age that have a regional dip of 2° NE or less. Near the west and southwest boundary, the dip steepens to as much as 4°. The most common rocks at the surface are the Kaibab Limestone of Permian age, the Moenkopi Formation of Triassic age, and mafic lava flows of late Cenozoic age.

The west boundary of the area is at the Grand Wash Cliffs (fig. 1), a consequent fault-line scarp along the Grand Wash fault. This fault is a regional feature that marks the boundary between the Colorado Plateaus and the Basin and Range provinces. The east boundary is at the Kaibab upwarp. In between are several blocks separated by north-trending faults. North of the Grand Canyon, these blocks are, from west to east, the Shivwits, Uinkaret, and Kanab Plateaus. South of the Canyon are the Hualapai and Coconino Plateaus (fig. 1).

The western Grand Canyon region is a key area for understanding the structural and tectonic transition between the Colorado Plateau and the Basin and Range. The location of the Shivwits Plateau directly at the transition makes this plateau of paramount importance in this context.

Previous work

The earliest report on the geology of the area is that of Newberry (1861) of the Ives Survey. Powell (1875) studied not only the Grand Canyon but also the area north of it. He was the first to describe many of its principal geologic features. Dutton (1882) concentrated on the Cenozoic geology of the area; Davis (1901, 1903) and Johnson (1909) discussed physiographic development as well as structural features; Darton (1910) summarized the geology of part of the area; Noble (1914) discussed and analyzed important structural features in part of the Grand Canyon. More recent structural and tectonic studies have been published by Averitt (1964), Babenroth and Strahler (1945), Cook (1957), Cook and Hardman (1967), Gardner (1941), Gilluly (1963), Hamblin (1965, 1970), Kelley (1955), Koons (1945, 1948), Lucchitta (1966, 1972), McKee and others (1967), McKee and McKee (1972), Moore (1969), Strahler (1948), Twenter (1962), and Young (1966). Shuey and others (1973) have just published a useful summary and analysis of geophysical properties at the transition between the Colorado Plateau and the Basin and Range.
ACKNOWLEDGMENTS

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DESCRIPTION

Northwest Arizona

The Colorado Plateau in northwest Arizona is composed of structural blocks elongate in a northerly direction, dipping gently northeast, and separated by faults trending roughly north. From west to east these faults are the Grand Wash, the Hurricane, and the Toroweap-Sevier (fig. 1). All are high-angle dip-slip and downthrown to the west. Strike lengths are measured in tens to hundreds of kilometers, displacements in hundreds to thousands of meters. The northerly trend of these faults is the result of alternating north-northwest and north-northeast to northeast segments. This pattern is repeated down to the outcrop scale, as pointed out by Hamblin (1970), and is reflected most conspicuously by the Hurricane fault (fig. 1,2). Displacement on the Hurricane and Toroweap faults decreases to the south, that of the Grand Wash fault decreases to the north.

Although some authors suggest that earliest movements on the faults may be as early as Laramide (Lovejoy, 1973 for the Hurricane fault), or post-Eocene and pre-Miocene (?) (Gardner, 1941 for the Hurricane fault), most agree that movement began in Miocene time (Hamblin, 1970; Koons, 1945; Averitt, 1964) and continued into the late Tertiary or, in the case of the Hurricane fault, the Holocene (Hamblin, 1970). These ages are based on the relations of the faults to sedimentary deposits and lavas in northernmost Arizona and southern Utah. Along much of its length, the Grand Wash fault has not moved since the upper 300 m of Muddy Creek Formation (an interior-basin deposit) was deposited, probably shortly after 10.6 m.y. ago (Damon, 1965; Lucchitta, 1966, 1972). Main movement occurred after eruption of the Peach Springs Tuff of Young, dated at 18.3 m.y. (Young, 1966). Near the northern border of Arizona, however, the Grand Wash fault has displaced lavas dated at 6 m.y. (Hamblin, 1970). Along the central part of the Grand Wash fault system, recent movements probably have occurred not along the main break but along the related Wheeler fault (fig. 1), which displaced deposits of Quaternary age (Lucchitta, 1966).
Figure 2.—Tectonic map of the Shivwits Plateau, Arizona.
Folds of tectonic origin are not common in northwest Arizona. The most prominent is the Kaibab upwarp, an anticlinal uplift trending approximately north to north-northwest and with structural relief of about 1,000 m. The Kaibab upwarp commonly is regarded as a Laramide feature, though recurring deformation may have continued during the Cenozoic. Other folding commonly is associated with the major faults of the area, and consists of downwarping toward the fault on the downthrown block commonly accompanied by upwarping toward the fault on the upthrown block. There has been no general agreement on interpretation of this feature, which has been called "reverse drag" by Hamblin (1965).

Gentle warps with relief of a few meters are common, especially in the northern part of the Shiwits Plateau. These warps are solution-collapse features resulting from solution in the gypsiferous alpha members of the Toroweap and Kaibab Formations.

Shiwits Plateau

Structural features on the Shiwits Plateau include faults and folds. Both generally parallel trends of the Grand Wash and Hurricane faults, the master faults of the area (fig. 2).

All Shiwits faults are high-angle dip-slip; some have scissors movement. Strike lengths are measured in kilometers to tens of kilometers, displacements in tens to hundreds of meters. Preferred orientations are northwest, north-northwest, and north-northeast (fig. 2). Northeast trends are relatively few and generally aligned with, as well as parallel to, the northeast segment of the southern great bend of the Hurricane fault. Most faults are concentrated in the eastern and western thirds of the Shiwits Plateau. The central part is relatively unfaulted. Grabens are common, especially in the eastern part of the plateau and in a belt trending about northeast and located generally north of the southernmost bend of the Hurricane fault. Although faults with either the east side or the west side down are equally common throughout most of the plateau, near the western boundary faults are consistently up to the west and associated with gentle monoclinal flexures also upwarped to the west. In many instances, these monoclines pass along strike into faults. Strata of the Shiwits Plateau dip very gently to the northeast. Near the western boundary, however, the dip increases to as much as 4°. This increase combined with the common elevation of the west side of faults and monoclines, results in a notable structural upwarping of the west edge of the Shiwits Plateau and of the Hualapai Plateau to the south of it.
Throughout most of the Shivwits Plateau, structural features can be determined only as being post-Triassic in age. In areas covered by lavas, faulting demonstrably occurred before, during, and after effusion of the lavas. However, the relative scarcity of faults that cut the lavas (fig. 2) suggests that most faults predate at least the youngest flows. Hamblin (1970), and Best and others (1969) report ages of about 6 m.y. for the oldest lavas (Stage I) on the Shivwits Plateau, and 2 m.y. or less for younger ones. If the dates apply to the lavas in the southern part of the Shivwits Plateau, which are largely Stage I, most faulting in that area probably occurred before 6 m.y. ago.

DISCUSSION

Problems and hypotheses

The area now occupied by the southwestern part of the Colorado Plateau has differed geologically from adjacent areas to the west throughout Phanerozoic time. Deformation has been slight; accumulation of Paleozoic sediments has been low; and topographic elevation has generally been the opposite of that of adjacent areas to the west: when these were high, the plateau was low, and vice versa. Although the present southern boundary of the plateau with the Basin and Range is relatively indistinct and marked by a transition zone about 100 km wide, the western structural and physiographic boundary is remarkably sharp and commonly less than 10 km wide. A person standing on top of the Grand Wash Cliffs, at the western edge of the Shivwits Plateau, sees to the west a seemingly endless succession of serrated ranges composed of tilted fault blocks and separated by deep structural basins that are in large part filled with sediments. The first of these basins west of the Cliffs, the Grand Wash trough (fig. 1), probably is 5,000 to 6,000 m deep (Lucchitta, 1966) and is bounded to the west by bedrock ridges whose strata commonly dip 45° or more toward the Grand Wash fault. East of the observer, by contrast, is a landscape of tablelands, scarps and stripped surfaces developed in subhorizontal strata that dip very gently northeast. This dip is not increased by rotation of fault blocks, in spite of the considerable displacement on some of the faults, and is disturbed only by monoclinal flexures that occur in the immediate vicinity of the faults. Deep, sediment-filled structural basins are absent.

As summarized recently by Shuey and others (1973), crustal properties of the western plateau differ markedly from those of the Basin and Range. On the plateau, the crust is thick, gravity low, heat flow low to normal, and aeromagnetic values are high; in the adjacent Basin and Range, the crust is thin, gravity high, heat flow high, and aeromagnetic values are low. Shuey and others (1973) emphasize that the locus of maximum change in these properties, as well as a seismic belt and a belt of anomalously high geomagnetic
variation, is a relatively wide zone, 50-80 km across, whose axis is somewhat east of the present physiographic boundary of the Plateau. Nevertheless, this zone, the present physiographic and structural boundary, the Wasatch line (eastern limit of Laramide thrusting and folding), and the Paleozoic hinge line all coincide remarkably well when viewed at a perspective commensurate with the crustal blocks involved, which measure about 600 km (plateau) and 400-800 km (Basin and Range) in an east-west direction.

A geologic discontinuity of one kind or another has existed at or near the present western boundary of the Colorado Plateau since Paleozoic time. This raises the questions of (1) what is the nature of the plateau crustal block, a block that apparently has remained internally coherent for 500 million years and that has continually behaved differently from its surroundings?; and (2) what is the nature of the tectonic boundary between the plateau block and the area to the west now occupied by the Basin and Range province? Answers to these questions are fundamental not only to understanding the plateau and Basin and Range provinces, but also crustal processes and plate tectonic theory in general.

The above problems are evaluated here in the context of Cretaceous and Cenozoic deformation of the southwestern boundary of the plateau, in the hope that this limited understanding will shed light on the general problems. Within this context, four hypotheses may be advanced to explain the structural and tectonic differences between plateau and Basin and Range.

(1) The plateau is being 'consumed' at its western edge by Basin and Range structures, in effect slumping at a grand scale into the Basin and Range province. This implies that the present boundary between the two provinces is ephemeral and has been moving eastward, and that the fault-bounded blocks of the western plateau represent a transition zone of incipient Basin and Range structure. The hypothesis also implies an eastward-migrating extensional stress field, a progressively younger age for plateau faults as one goes eastward, and a diffuse boundary between the plateau and Basin and Range provinces. A major change in crustal properties across the boundary is not required.

(2) The plateau and Basin and Range provinces are subject to and resulted from different crustal stress fields. This hypothesis does not require different crustal properties of the two blocks, nor a boundary between the two blocks that migrated with time, nor specifically a sharp or a diffuse boundary. It does require a decoupling between the two blocks and perhaps between the crustal blocks and the underlying mantle.

(3) The plateau and the Basin and Range provinces represent blocks that are subject to the same stress field but have different crustal properties, so that the stress field results in different styles of deformation in the two blocks. This hypothesis requires a relatively fixed and sharp boundary between the two blocks.
(4) The plateau and the Basin and Range were one geologic province before deformation and became subject to the same stress field, which decreased in intensity eastward. The present tectonic boundary between the two provinces marks the eastern boundary of the area where stress was sufficiently great to cause major deformation. This hypothesis does not require different crustal properties but does imply a boundary between the two provinces that is gradational and fixed in space. Consequently, Basin and Range structures would grade eastward into plateau structures, and both should be equivalent in age.

Interpretation

In order to discriminate between these hypotheses, it is necessary to discuss the tectonic style of the plateau in northwest Arizona, compare this style with that of the adjacent Basin and Range province, and analyze the nature of the boundary between the two provinces.

Structural style of the plateau in northwest Arizona

Monoclinal flexures associated with faults.—These flexures have been interpreted chiefly in two ways, as shown diagrammatically in figures 3a and 3b, respectively: (1) faulting of an earlier monocline whose displacement was opposite to that of the fault, an opinion held by Powell (1873), Dutton (1882), Noble (1914), Johnson (1909), Koons (1945), and Huntoon (1969); (2) downflexing and sagging of the downthrown block in response to extension and faulting, as proposed by Gardner (1941), Strahler (1948), and Hamblin (1965).

Neither Hamblin nor Gardner account for the upwarping of the upthrown block, which is common enough to require explanation, though not as common or conspicuous as the 'reverse drag' on the downthrown block. Hamblin (1965) explains some of the upwarping as resulting from pre-faulting folds whose axes form a low angle with the trace of the fault that cuts them, but this explanation cannot be applied in all cases. The faulted monocline hypothesis, on the other hand, commonly strains geometry when pre-faulting positions are restored. Neither hypothesis fully explains all observed facts, probably because of the respective assumptions that all folding predates faulting, and that the upwarping is unrelated genetically to the faulting and 'reverse drag' so that their spatial coincidence is fortuitous.

A combination of salient features of both interpretations accounts better for known relations and is shown in figure 3c. Old, relatively gentle and perhaps discontinuous monoclinal flexures were cut by faults of opposite displacement, and the initial monoclinal dip of the downthrown block was accentuated by the sagging envisioned by Hamblin (1965). This suggests that the sagging is much younger than
a) Faulted monocline hypothesis. Fault must always have same position with respect to monocline.

b) "Reverse drag" hypothesis of Hamblin (1965). No upwarping on upthrown block. Recurrent movement causes varying amount of flexing in units of different age, e.g. lava flows.

c) Hypothesis combining two hypotheses above. Old monocline, corresponding to reverse fault in basement at depth, is cut by normal movement along the same fault. Initial monoclinal warp on downthrown block is accentuated by flexing resulting from pull apart mechanism proposed by Hamblin (1965).

Figure 3.—Hypotheses on the origin of 'reverse drag'.

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the monoclinal folding and not genetically related to it. On the other hand, the association between monoclines and faults suggests that the monoclinal flexures are the surface expression of zones of weakness in the competent basement and were formed under compressive stress conditions probably related to the Laramide orogeny to the west. These zones later localized high-angle normal faults when the area was affected by the extensional stress field related to Basin and Range tectonics. The antiquity and repeated reversal of movement on faults, as well as the association of monoclines with faults at depth, are well documented for the Grand Canyon (Noble, 1914; Strahler, 1948; Maxson, 1961; Huntoon, 1969). Huntoon, in particular, recognized the importance of basement fractures in controlling deformation even under widely differing stress fields.

In summary, the monoclinal flexures and associated faults of the plateau block in northwestern Arizona suggest that (1) the block has been subjected to the same stress fields as the area to the west; (2) the resulting deformation has been different in degree (much more subdued) and perhaps in kind because of the greater competence and the buttress-like nature of the block, and (3) zones of weakness, some very ancient, in the basement of the plateau block have localized deformation even if produced by widely differing stress fields. These data support the third hypothesis but not the rest.

**Structural upwarping of the western edge of the plateau.**—Faults and monoclines upthrown to the west are associated with a marked increase in the dip of strata at the westernmost edge of the Colorado Plateau, where they form a band immediately east of and parallel to the Grand Wash fault. Many of the monoclines become faults along strike. Because most folds are of probable Laramide age on the plateau (or at least predate block faulting), the upwarp of the edge may be as old as Laramide.

The upwarp may reflect an ancient uplift southwest of the plateau (Lucchitta, 1966, 1972; Young, 1966) that is of probable Laramide age. This uplift, now reflected by the general northeast dip of strata in the southwest part of the plateau, was broken up by Basin and Range faulting in Miocene time. Alternatively, the upwarp may represent a monoclinal flexure of the kind discussed above and associated with the Grand Wash fault. In either case, the upwarp indicates that the western edge of the plateau has been involved in repeated deformation involving the basement and at least in part reflecting stress fields that probably are relatively old and encompassed both the plateau and the adjacent area to the west.
Sharpness and age of the Colorado Plateau-Basin and Range boundary

The present physiographic boundary in northwest Arizona is remarkably sharp (10 km wide). The boundary based on geophysical data is more diffuse but still narrow (80 km or less) compared to the dimensions of the crustal blocks involved (600 km). The boundary would be diffuse if the various plateaus bounded by high-angle normal faults were a transition zone between areas of typical plateau and typical Basin and Range structure. However, these plateaus, although exhibiting block faulting and extension tectonics, have neither the horsts (ranges) or grabens (basins), the great structural relief, nor the rotation of blocks typical of the adjacent Basin and Range terrane. Extension tectonics and block faulting are necessary but not sufficient to form Basin and Range structure. The plateaus could also be interpreted as a zone of incipient Basin and Range structure, but this is in poor accord with the ages of the associated faults, which are comparable to those of nearby faults in the Basin and Range. The plateaus probably represent neither Basin and Range structures that are just beginning to form nor structures that are older but have been arrested in their development. Consequently, the tectonic boundary—separating areas that differ markedly in structural style—is best placed at the western edge of the Shivwits Plateau, in coincidence with the physiographic boundary and with ancient tectonic boundaries such as the Paleozoic hinge line. In summary, the boundary is best viewed as old, relatively sharp, and relatively fixed. These properties are in poor accord with the hypothesis. The markedly different structural styles and the sharpness of the boundary do not support the fourth hypothesis.

Alinement and distribution of structural features on the plateau in comparison with the adjacent Basin and Range province

One of the most remarkable aspects of the structural pattern of northwestern Arizona is that both the faults on the plateau and the western boundary of the plateau closely parallel the structural grain of the adjacent Basin and Range province. This parallelism holds through the Lake Mead area, where the alinement changes from a predominantly northerly trend to a north-northwest to northwest trend. This can only reflect a common stress field, through-printing of basement fractures, or both. It would seem difficult to produce the parallelism without a common stress field.

In northwest Arizona, structural features trending northeast are not common. On the Shivwits Plateau, these features are concentrated in a zone trending northeast and aligned with the southernmost bend of the Hurricane fault (fig. 4). Other features with northeast trend in the same general area are (1) a straight segment of the Grand Canyon about 75 km long, (2) several diffuse but long lineaments visible on ERTS photographs, and (3) a northeast-trending segment of the Toroweap-Sevier fault.
Figure 4.—Northeast-trending features in northwest Arizona and southeast Nevada.
West of the plateau, a zone of conspicuous faults trending about northeast occurs north of Lake Mead, in alinement with the belt on the plateau (fig. 4). In the western part of Lake Mead, this zone has demonstrable transcurrent movement (R. E. Anderson, oral commun., 1973) and may be part of the Las Vegas shear zone. Volcanic centers are abundant south of the zone but absent north of it. Structural features trending northeast are uncommon elsewhere in southern Nevada.

A belt about 60 km wide that trends northeast is alined with the two bends of the Hurricane fault. This belt is characterized by a distinctive set of closely spaced grabens, by bends in the traces of faults in correspondence with bends of the Hurricane fault, and by virtual absence of volcanic centers. By contrast, volcanic centers are very abundant north and south of the belt.

The relations described above suggest that the colinear nature of the fault zone in the Lake Mead area and of the anomalous belt on the plateau reflects the same stress field and is not fortuitous. According to this interpretation, in the Basin and Range block this stress field resulted in faulting that was at least in part transcurrent. In the plateau block, by contrast, the stress field produced fractures with no demonstrable transcurrent displacement. Some of these fractures were then followed by normal faults generated by an extensional stress field. Examples are the faults trending northeast on the Shivwits Plateau and the northeast-trending segments of the Hurricane fault.

In summary, the parallelism between structural features on the plateau and in the adjacent Basin and Range province suggests that all these structures were formed in response to the same stress field. Furthermore, the lack of displacement of plateau structures as compared to correlative structures in the Basin and Range suggests that the plateau has responded as a more competent block than has the Basin and Range. These characteristics support the third hypothesis.

CONCLUSIONS

The southwestern part of the Colorado Plateau and the adjacent Basin and Range province differ markedly in physiography, structure, crustal properties, and geologic history. Most of these differences have existed through Phanerozoic time. The boundary between the two crustal blocks is best viewed as old, relatively sharp, and relatively fixed.

Parallelism between structural features of the plateau and the adjacent Basin and Range suggests that both have been subject to the same stress fields, but the much more subdued deformation on the plateau indicates that it has a more competent crust. This is also suggested by a northeast-trending structural belt that crosses the boundary between the two blocks. In the Basin and Range, the belt shows transcurrent movement; on the plateau, the belt shows no demonstrable transcurrent movement, but is reflected by anomalous northeast trends of structural features and by a set of grabens that may indicate a shear component of the stress field.
In the plateau block, the basement has played a dominant role in controlling structural deformation. Zones of weakness, some very old, have localized deformation even under different stress fields. Many of these zones have been repeatedly active, commonly with reversed movement. Thus, the compressional stress field associated with the Laramide orogeny to the west resulted in reverse faulting in the basement and corresponding folding in the sedimentary cover. The later extensional stress field associated with Basin and Range deformation to the west resulted in normal faulting along many of the same faults.

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THE BRIGHT ANGEL AND MESA BUTTE FAULT SYSTEMS
OF NORTHERN ARIZONA

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ABSTRACT

Regional geologic mapping using ERTS-1 pictures has led to the recognition of two parallel northeast-trending systems of normal faults, each of which can be traced more than 100 km. Many eruptive centers appear to be localized along these fault systems or along their extensions. The faults are chiefly observed in Phanerozoic rocks and have minor displacement but are interpreted by us to reflect fault zones of major displacement in the crystalline Precambrian basement.

The Bright Angel fault system extends as a continuous zone of normal faults from Cataract Creek on the southwest to the Echo Cliffs on the northeast. Beyond the Echo Cliffs, the system continues north-eastward to the vicinity of Monument Valley as a more diffuse, discontinuous zone of normal faults. The Bright Angel fault, Vishnu fault, and Eminence Break graben are among the larger individual members of the total system. The Navajo Mountain intrusive center lies along the discontinuous part of the system. Three major eruptive centers of the Mount Floyd volcanic field lie on the southwestern projection of the Bright Angel fault system. If the eruptive centers are included as part of the recognizable structural system, the Bright Angel system has a total known length slightly more than 300 km.

The Mesa Butte fault system, as now recognized, extends from Chino Valley on the southwest to Shadow Mountain on the northeast. Bill Williams Mountain, Sitgreaves Peak, and Kendrick Peak are principal silicic to intermediate eruptive centers of the San Francisco volcanic field that appear to be localized along the fault system. Red Mountain, Mesa Butte, and Shadow Mountain are prominent basaltic eruptive centers along the system; monchiquite diatremes at Tuba Butte and Wildcat Peak lie on the northeast projection of the fault system. The total distance from Chino Valley to Wildcat Peak is more than 200 km.

Comparison of the Bright Angel and Mesa Butte fault systems with a residual aeromagnetic map of Arizona reveals a close correspondence between the positions of the observed relatively minor normal faults and the margins of a series of large northeast-trending magnetic anomalies. Perhaps the most noteworthy feature of the aeromagnetic map is a 400-km-long northeast-trending belt of large positive aeromagnetic anomalies that extends from the vicinity of Congress to the northern border of Arizona. The Mesa Butte fault system lies along the southeast margin of this anomaly belt. Another large positive anomaly, bounded on the southeast by the Bright Angel fault, corresponds in the Grand Canyon to a belt of Precambrian amphibolite and schist. Most of the large positive aeromagnetic anomalies along the Bright Angel and Mesa Butte fault systems may correspond to similar bodies of mafic metavolcanic rocks, which have been offset along two major and perhaps several minor faults of Precambrian age. The normal faults that displace the overlying Phanerozoic rocks have been formed by renewed movement along these ancient fault zones, in response to dilation of the crust from late Tertiary time to the present.

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The ancient fault zones inferred to be present along the Bright Angel and Mesa Butte fault systems may be related in origin to the Shylock and Chaparral fault zones in central Arizona described by Anderson (1967). Both the Shylock fault zone and the Chaparral fault have right-lateral transcurrent displacement. As shown by Anderson, the Shylock zone has a probable minimum horizontal displacement of 8 km. A large contrast in the magnetic properties of the rocks on opposite sides of the fault zone, indicated by the aeromagnetic map, suggests the displacement may be several tens of kilometers or more. Comparably large right-lateral displacements may have occurred along the ancestral Bright Angel and Mesa Butte fault zones.

The location of epicenters of recent earthquakes and reports of earthquakes by residents in the region indicate that the Bright Angel and Mesa Butte fault systems are currently active.

INTRODUCTION

Among the structural features of northern Arizona, the Bright Angel fault has become perhaps one of the best known. It certainly must rank as the fault most often visited; thousands of tourists hike along part of the Bright Angel fault each year as they descend from the south rim of the Grand Canyon into the inner gorge along the Bright Angel trail, or from the north rim along the Kaibab trail. How many of these visitors are aware that a fault has controlled the route of easiest descent and, indeed, the course of Bright Angel Canyon, through which the Kaibab trail passes, is not known. It is clear, however, that the easy access into the Grand Canyon provided by these trails has led to close inspection of the fault by many geologists.

The Bright Angel fault was first described by F. L. Ransome (1908), who recognized that displacement on the fault had occurred during at least two widely separated periods of time, one Precambrian and one post-Paleozoic. The first map to portray the fault was a reconnaissance map of northern Arizona and New Mexico by N. H. Darton (1910), who also briefly described the fault and noted the two episodes of displacement. In 1914, a part of the fault lying south of the Colorado River was mapped in detail by L. F. Noble; although he contributed some important observations about the fault (Noble and Hunter, 1916, p. 101; Noble, 1918), the map was never published. Later, McKee described the Bright Angel fault in a short paper in 1929.

The problem of multiple episodes of faulting on the Bright Angel and related faults attracted the attention of Maxson and Campbell (1934), as they worked on the ancient crystalline rocks of the Grand Canyon. Much later, Maxson published a detailed map of the Bright Angel quadrangle and described a complex history of displacement on the Bright Angel fault and on intersecting faults (Maxson, 1961). Most recently, the Bright Angel fault was remapped, where it cuts exposed Precambrian rocks, by Sears (1973), who has reinterpreted the history of displacement.
Figure 1.—ERTS-1 picture of Coconino Plateau showing Bright Angel and Mesa Butte fault systems. NASA picture ERTS E-1104-17382-5, 4 Nov. 72.
Figure 2a.—Map of faults in northwestern Arizona. Most faults shown are normal faults; bar and dot on downthrown side.
Mapping utilizing ERTS-1 pictures

1. Lucchitta, Iva, this volume

Other sources of data

4. Holm, E.A., unpublished map
6. Twenter, 1962
7a. Pomeroy, 1959
7b. Marshall, 1957
7c. Marshall, 1956a
7d. Fillmore, 1956
7e. Morris, 1956
7f. McQueen, 1957
7g. Wells, 1960
7h. Marshall, 1956b
8. Shoemaker, 1960
11. Babenroth and Strahler, 1945
12. Cooley, M.E., unpublished map
13. Koons, 1945
14. Brillsewhack, 1952
15. Fuik, 1973
17. Moore, 1972
18. Huntmon, 1970
19. Phoenix, 1963
20. Cooley and others, 1969

Figure 2b.—Index to sources of information shown in figure 2a.
The present writers became interested in the Bright Angel fault in the course of a regional geologic investigation of the hydrographic basin of Cataract Creek, the principal drainage of the Coconino Plateau, which forms the south rim of the Grand Canyon. Starting from the southern boundary of the Bright Angel quadrangle, we traced the Bright Angel fault 30 km to the southwest. At this point the fault dies out, but other parallel faults continue to the southwest at least another 20 km. In this investigation we employed pictures taken from an altitude of 907 km by the first Earth Resources Technology Satellite (ERTS-1). While studying these pictures, we noted a remarkably long zone or system of faults of which the Bright Angel fault is a member. From the center of the Cataract Creek region the fault system can be traced northeastward for at least 270 km. A parallel system of faults that lies about 50 km southeast of the Bright Angel system can also be recognized in the ERTS-1 pictures (fig. 1). Some of the principal Cenozoic eruptive centers on the Coconino Plateau appear to be localized along these two fault systems.

At this stage in our investigation the evidence suggested that the two fault systems, which are chiefly observed in Phanerozoic rocks, might be related to major ancient structures in the underlying Precambrian crystalline complex. To test this hypothesis we compiled a map of the faults in northwestern Arizona (fig. 2) and compared this map with a residual aeromagnetic map of Arizona published by Sauck and Sumner (1971). About 60 percent of the fault map is based on previous published and unpublished maps. The remainder is based on new detailed and reconnaissance field mapping by us and by Ivo Lucchitta, using ERTS-1 pictures and aerial photographs. Comparison of the pattern of faults with the aeromagnetic map revealed many features of interest. In this paper we will discuss mainly the Bright Angel fault system and the parallel system of faults to the southeast, referred to here as the Mesa Butte fault system. Other systems of faults are briefly described.

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BRIGHT ANGEL FAULT SYSTEM

The Bright Angel fault system extends as a continuous zone of
northeast-trending normal faults from Cataract Creek on the southwest
to the Echo Cliffs on the northeast. Beyond the Echo Cliffs, the
system may be traced northeastward to the Monument upwarp in Utah
as a more diffuse discontinuous zone of normal faults.

South of the Grand Canyon, the Bright Angel system comprises
several distinct faults, each of which is some tens of kilometers
long. The Bright Angel and Vishnu faults are the principal members
of the system (figs. 3 and 4). Here the faults cross the Coconino
Plateau, which is capped by Kaibab Limestone of Permian age. Maximum
displacement of the exposed Permian rocks on each of the larger
faults is on the order of 100 m. On most faults, displacement is down
to the southeast. The downthrown side commonly exhibits reverse drag;
narrow prisms of beds belonging to the Moenkopi Formation of Triassic
age are preserved in a few places on the hanging wall. Stripping away
of the easily eroded Moenkopi from much of the rest of the plateau has
left relatively prominent fault-line scarps, formed by the upthrown
resistant beds of Kaibab Limestone.

A curious feature of the Bright Angel fault system, which is well
illustrated on the Coconino Plateau, is the tendency of individual
faults to die out as they approach northwest-trending monoclines
(fig. 4). Two small monoclines lie athwart the Bright Angel fault
system on the plateau. None of the individual faults in the system
cross either of these monoclines.

The Bright Angel fault is one of the longest members of the
entire fault system. It can be traced for a total distance of 65 km.
At its north end, the fault swings around from a northeasterly to a
northerly strike and then dies out as it approaches the East Kaibab
monocline (fig. 4). A series of en echelon faults continues farther
to the northeast along the main trend of the Bright Angel fault.

Displacement of Permian beds on the Bright Angel fault is down
to the southeast, both on the Coconino Plateau and on the Kaibab
Plateau. Where it crosses the Grand Canyon, however, a complex set
of relationships may be observed along the fault in older beds,
particularly in Precambrian rocks. The net displacement of beds
belonging to the Grand Canyon Supergroup, of late Precambrian age, is
up on the southeast. Their displacement is due primarily to one or
more episodes of reverse faulting in late Precambrian time (Maxson,
1961; Sears, 1973).
Figure 3a.—Oblique high-altitude aerial photograph of Mesa Butte fault system, looking southwest across Coconino Plateau: 1) Mesa Butte; 2) Cedar Ranch fault; 3) Mesa Butte fault; 4) Red Mtn.; 5) Slate Mtn.; 6) Source of Tappan Wash flow; 7) Kendrick Pk.; 8) Sitgreaves Mtn.; 9) Bill Williams Mtn. (U.S. Geological Survey high-altitude photograph)
Figure 3b.—Oblique high-altitude aerial photograph of Bright Angel fault system, looking southwest across Coconino Plateau: 1) Howard Hill; 2) Mt. Floyd. (U.S. Geological Survey high-altitude photograph)
Figure 4.—Map of part of Bright Angel and Mesa Butte fault systems, showing relation of faults to monoclines and Cenozoic volcanic rocks. Bar and dot are on downthrown side of faults; triangles indicate direction of dip of steep limb of monocline.
Northeast of the East Kaibab monocline, along the projected
trend of the main part of the Bright Angel fault, the narrow Eminence
Break graben, can be traced across the Marble Platform (fig. 4), a
stripped surface capped by the Kaibab Limestone. Beds southeast of
the graben lie at a higher elevation than the beds northwest of the
graben. Thus the displacement is greater on the fault bounding the
graben on the southeast side. A prominent northwest-facing fault-
line scarp, the Eminence Break, is developed along the southeast­
bounding fault. This fault is 40 km long and has a maximum displacement
of about 100 m. It dies out to the northeast at the Echo Cliffs
monocline. The fault bounding the northwest side of the graben is
only 30 km long and dies out before reaching the Echo Cliffs
monocline. To the southwest, toward the East Kaibab monocline, the graben bends
to the south and dies out precisely at the lower axis of the monoclinal
flexure. A few shorter faults that are parallel or subparallel with
the main part of the Eminence Break graben occur on nearby parts of the
Marble Platform.

A zone of en echelon grabens continues for a distance of 12 km
northeast of the Echo Cliffs along the projected trend of the Eminence
Break graben. Beyond this point, the Bright Angel fault system
becomes a broad diffuse zone of relatively short faults (fig. 2).
In general, faults become more and more widely spaced to the northeast,
and the northeastern limit of the system is not well defined. The
system extends at least as far as the Monument upwarp, Utah. One
branch of the system may be represented by a set of faults that extend
to the Comb monocline, just north of the San Juan River in Utah. The
easternmost fault of this set was described by Sears (1956). Between
the Echo Cliffs and Comb Ridge, the displacement on faults in the
Bright Angel system is generally less than 100 m, and none exceeds
20 km in length. At the surface the faults displace Triassic and
Jurassic rocks and, in the vicinity of Monument Valley, Permian
rocks.

MESA BUTTE FAULT SYSTEM

The Mesa Butte fault system, as now recognized, extends from
Chino Valley, near Paulden, Arizona, on the southwest, to Shadow
Mountain on the northeast. The known length of the fault system is
about 150 km, but detailed mapping of Precambrian terrane west and
southwest of Prescott, Arizona, may extend the system many tens of
kilometers farther to the southwest. About midway along its known
length, the fault system is concealed beneath Quaternary lava flows
of the San Francisco volcanic field. The recognized faults in the
system are thus grouped into a northeastern segment and a southwestern
segment.
The Mesa Butte fault is the principal member of the northeastern segment of the fault system. It was first noted by Johnson (1909) and was mapped and named by Babenroth and Strahler (1945). As may be seen in figure 3a, Mesa Butte, an elongate basaltic cinder cone, was formed by fissure eruption along the fault. Southwest of Mesa Butte, the fault is covered by Pleistocene lava flows that lap against the base of the fault scarp. The position of the fault and its displacement can be estimated with some precision, however, for a distance of about 10 km southwest of Mesa Butte. Farther to the southwest the fault scarp is completely concealed beneath younger lava flows and basaltic cinder cones (fig. 4). Northeast of Mesa Butte, the fault emerges from beneath the lavas and becomes the northwest-bounding fault of the spectacular long narrow Mesa Butte graben (fig. 3a). The graben is 300 to 400 m wide, 15 km long, and forms a trench in the surface of the Coconino Plateau 100 to 200 m deep. Formations exposed in the walls of the trench are the Coconino Sandstone, Toroweap Formation, and Kaibab Limestone, all of Permian age. The Mesa Butte fault can be recognized at the surface for a distance of about 35 km; maximum observed displacement is about 100 m.

A branch fault, the Cedar Ranch fault, joins the Mesa Butte fault near Mesa Butte. It can be traced southwestward about 15 km, where it disappears beneath younger lavas. The Cedar Ranch fault is covered by lava that laps against the fault scarp, but locally the lava is offset by faulting. As along the Mesa Butte fault, displacement is down to the southeast.

Northeast of the Mesa Butte graben and directly in line with it, is the sharply flexed, locally faulted southern salient of the Grandview monocline (fig. 4). Farther to the northeast the monocline swings around to the west and veers away from the Mesa Butte fault system. About 30 km beyond the Mesa Butte graben, a cluster of small northeast-trending faults near Shadow Mountain marks the northeastern limit of exposed faults in the Mesa Butte system. However, a series of monoclines along the northwest edge of Black Mesa and a segment of the Comb monocline along the southern margin of the Monument upwarp, near Kayenta, Arizona, may be controlled by the northeast extension of the ancestral Mesa Butte fault in the Precambrian rocks.

Faults belonging to the southwestern segment of the Mesa Butte fault system were mapped along Hell Canyon in the Paulden quadrangle and described by Krieger (1965). The principal fault in this segment was traced by Krieger 8 km northeastward from the Paulden quadrangle across the Tonto Rim, where it cuts Pennsylvanian and Permian rocks, and is shown on the geologic map of Arizona (Wilson and others, 1969). The continuation of this fault through the southern part of the San Francisco volcanic field can be recognized in the ERTS-1 pictures (fig. 1) and traced an additional 22 km. The total recognized length of the fault is 38 km. Displacement is down on the southeast; maximum
displacement of the exposed rocks is about 150 m. The fault disappears northeastward in a field of lava flows of Pleistocene age. Whether the displacement simply dies out or the fault is covered by younger lavas is not known, as the relations have not been studied in detail in the field.

Directly in line with the 38 km-long fault, 12 km southwest of its last recognized exposure in Hell Canyon, rocks correlated by Krieger (1965) with the Texas Gulch Formation of Precambrian age are in contact with the Mazatzal Quartzite of Precambrian age, the Tapeats Sandstone of Cambrian age, and the Martin Limestone of Devonian age, along close-spaced northeast trending faults. As mapped by Krieger (1965), the displacement that brought Mazatzal Quartzite in contact with the Texas Gulch Formation(?) at this locality is Precambrian (pre-Tapeats). The later displacement of Cambrian and Devonian beds may have been controlled by a Precambrian fault zone. It appears possible that the ancestral Mesa Butte fault is partly exposed in the Precambrian rocks at this place.

Still farther to the southwest, on the trend of the Mesa Butte fault system, rocks mapped as the Yavapai Series of Precambrian age are in contact with Precambrian granitoid rocks northwest of Granite Mountain (Wilson and others, 1969). This contact may be the southwestward continuation of the ancestral Mesa Butte fault. Work in the field is needed to test this hypothesis.

OTHER FAULT SYSTEMS

In addition to the Bright Angel and Mesa Butte fault systems, other swarms or systems of faults can be recognized from the pattern of faulting shown in figure 2. One of these, the Sinyala system, which comprises a swarm of northeasterly trending faults, is roughly parallel with the Bright Angel and Mesa Butte systems. Other systems trend northwest and approximately north (fig. 5). Each system is a relatively broad lane or zone of faults in which the individual faults tend to be parallel or subparallel with the overall trend of the lane. Where the lanes intersect, faults belonging to two or more systems are present. At these intersections, some individual faults follow the direction of one system for part of their length and then turn abruptly and follow the direction of another system.

The Sinyala fault system consists, in large part, of a set of en echelon faults. It is named for a fault 50 km long that crosses the Colorado River midway along the length of the system (fig. 2) and trends parallel with the total system. Although this fault is very long, its displacement generally is less than a few meters. Some other faults in the Sinyala system with much larger displacement are segments of long faults that follow one system for a distance and then turn and follow another. The faults are observed in rocks ranging in age from Precambrian to Permian.
Four northwest-trending systems of faults are recognized in northern Arizona (figs. 2 and 5): the Chino Valley system, the Cataract Creek system, the Kaibab system, and the Mormon Ridges system (fig. 5). The Chino Valley system of faults is observed mainly in Paleozoic rocks, but a set of fault scarps 20 km long is developed in alluvium in northwestern Chino Valley along the Big Chino fault of Krieger (1965 and 1967a).

The Cataract Creek system consists chiefly of a newly mapped swarm of faults that cut Permian, Triassic, and Tertiary rocks in the southern and western part of the Cataract Creek basin. Faults belonging to this system extend into the Shivwits Plateau on the northwest and across the southern part of the San Francisco volcanic field on the southeast.

The Kaibab system is a relatively broad lane of northwest-trending faults that extends from the southern Kaibab Plateau across the Grand Canyon onto the northern Coconino Plateau. Rocks ranging in age from Precambrian to Permian are cut by this system. As in the case of the northeast-trending faults of the Bright Angel system, many faults in the Kaibab system can be shown to have had a long history of displacement that began in the Precambrian.

The Mormon Ridges system is observed principally in Mesozoic rocks. It is named for ridges on the Kaibito Plateau formed along a close-spaced set of faults in the Navajo Sandstone of Triassic(?) and Jurassic age (Cooley and others, 1969).

Two roughly north-trending swarms of faults are present in the fault pattern of northern Arizona. They are referred to here as the Toroweap system and the Oak Creek Canyon system. Both of these systems cut rocks ranging in age from Cambrian to Tertiary. The Toroweap system includes a north-trending segment of the Toroweap fault (Koons, 1945, Twenter, 1962), a parallel segment of the Hurricane fault (Twenter, 1962), and a roughly north-trending segment of the Aubrey fault (Blissenbach, 1952), each of which has a displacement of several hundred meters. A number of smaller faults are also present in the system.

The Oak Creek Canyon system extends north from Oak Creek Canyon, through Flagstaff, Arizona, to the Kaibab Plateau. Just north of Flagstaff, the Oak Creek Canyon system is concealed by Quaternary and Pliocene(?) volcanic rocks of the San Francisco volcanic field; the large stratovolcano of San Francisco Mountain lies astride the system.

Northwest of the Sinyala system, faults are somewhat more uniformly distributed than in the region to the east and south (fig. 2). Major faults are present near this corner of the state, and northeast, north, and northwest-trending sets of faults are represented in this area, but the faults are not as clearly grouped into discrete lanes as they are in the adjacent region.
RELATION OF Eruptive CENTERS TO FAULTS

Inspection of the distribution of volcanoes in the San Francisco volcanic field reveals that about half of the silicic to intermediate volcanoes, described by Robinson (1913), lie along the Mesa Butte fault system. This relationship is easily recognized on the ERTS-1 pictures (fig. 1); these volcanoes form the most prominent peaks in the region. Closer study shows that Sitgreaves Mountain lies precisely on the projected trend of the Mesa Butte fault (fig. 7). Bill Williams Mountain lies a few kilometers northwest of this line, and Kendrick Peak lies a few kilometers southeast of the line, close to the projected trend of the Cedar Ranch fault. Slate Mountain, a relatively small silicic volcanic center, lies on the trend of the Cedar Ranch fault (figs. 3b and 4). The remaining silicic volcanoes in the San Francisco field occur either on the Oak Creek Canyon system of faults or on the Cataract Creek system.

Several hundred basaltic cinder cones are present in the San Francisco volcanic field, but only a few of these are related in an obvious way to the major fault systems. Red Mountain, one of the largest cinder cones, lies on the line of the Mesa Butte fault (figs. 3a and 4). Mesa Butte, formed by fissure eruption along the fault, is one of the northernmost eruptive centers in the San Francisco field. Well separated from the rest of the volcanic field, Shadow Mountain lies 45 km farther to the northeast along the Mesa Butte fault system (figs. 4 and 7). It is, however, a basaltic eruptive center of the San Francisco type (Condit, this volume).

Beyond Shadow Mountain lie the monchiquite diatremes and dikes at Tuba Butte and Wildcat Peak (fig. 7). These isolated volcanic centers are far removed from most other diatremes and alkalic basalts of the Navajo country (Gregory, 1917; Williams, 1936) but are close to the projected trend of the Mesa Butte fault system.

Eruptive centers are more widely spaced along the Bright Angel fault system. In the vicinity of Monument Valley, Utah, a broad swarm of minette dikes and two kimberlite pipes occur near the extreme northeastern end of the Bright Angel fault system. Navajo Mountain, a prominent structural dome in Utah, just north of the Arizona state line, occurs on the margin of the relatively diffuse part of the fault system. For many years the Navajo Mountain dome was thought to have been formed over a laccolith or stock (Gilbert, 1877; Gregory, 1917; Baker, 1936; Hunt, 1942) but without any definite evidence. The dome is distant from the known laccolithic mountain groups of the Colorado Plateau. A small syenite porphyry intrusion near the summit of the dome is reported by Condie (1964).

Near the southwestern end of the Bright Angel fault system, Howard Hill, a small structural dome that resembles the Navajo Mountain dome in shape but not in size, lies just beyond the end of the Vishnu fault (figs. 3b and 4). This dome may be located over a small stock. Still farther to the southwest, Trinity Mountain,
Round Mountain, and Mount Floyd, the principal eruptive centers of
the Mount Floyd volcanic field, occur on the trend of the Bright
Angel fault system. Mount Floyd, itself, lies almost precisely on
the projection of the Bright Angel fault (figs. 3b and 7).

PRECAMBRIAN ORIGIN OF FAULTS

The distribution of eruptive centers along the Bright Angel
and Mesa Butte fault systems and the relatively great length of these
systems suggest that the faults observed in Phanerozoic rocks may be
related to more profound, deep-seated structures in the crust. Where
the Bright Angel and Kaibab fault systems cross the Grand Canyon, the
relation of the faults to the deeper lying Precambrian rocks is
exposed. Here the principal displacement of Paleozoic beds along both
systems of faults was controlled by more ancient faults in the under­
lying Precambrian rocks.

Walcott (1889) was the first to demonstrate that displacement of
Phanerozoic rocks in the Grand Canyon had occurred along a Precambrian
fault. He found that the northwest-trending Butte fault, which occurs
along the East Kaibab monocline and which had reverse movement in
Tertiary time, had much larger displacement, but of the opposite
sense, in Precambrian time. Ford and Breed (1973) estimate as much
as 1.5 km of normal displacement on the Butte fault after deposition
of the Grand Canyon Supergroup and before deposition of the Tapeats
Sandstone of Cambrian age.

Noble (1914) recognized a number of other post-Paleozoic faults
that occur along Precambrian lines of displacement. He found that
both northwest-trending and northeast-trending faults are controlled
by Precambrian structure. On these faults, Phanerozoic displacement
generally is smaller and in the opposite sense from that which
occurred in late Precambrian time. The northwest-trending Muav fault
(fig. 6) was shown by Noble to have had a minimum normal displacement
of nearly 2 km in Precambrian time. A faulted monocline with much
smaller throw is developed in the Paleozoic rocks over the Muav fault.

Maxson and Campbell (1934) found that the crystalline rocks of
the Grand Canyon had been displaced by faults prior to deposition
of the Grand Canyon Supergroup. These faults trend in two different
directions: (1) N. 15-30° E., parallel with the schistosity of the
crystalline metamorphic complex (Vishnu Schist); and (2) N. 20°-30° W.,
parallel with a direction of master joints, which also were formed prior
to deposition of the Grand Canyon Supergroup. Some of these faults
are overlain by unbroken Precambrian strata; other faults of the
same network that displace overlying strata presumably were active
in post-Vishnu Schist pre-Grand Canyon Supergroup time (Maxson, 1961).

The Bright Angel fault belongs to the latter category. Maxson
(1961) inferred six episodes of displacement on the Bright Angel
fault:
1. Displacement of crystalline metamorphic rocks of early Precambrian age prior to intrusion of the Zoroaster Granite. Foliated migmatites in the metamorphic complex have been dated at 1695 ±15 m.y. and the Zoroaster Granite, at 1725 ±15 m.y., thus pointing to a major episode of Precambrian deformation at about 1700 m.y. (Pasteels and Silver, 1966).

2. Post-Zoroaster displacement of the crystalline complex prior to deposition of the Grand Canyon Supergroup.

3. Displacement after deposition of the Dox Sandstone of Precambrian age and prior to intrusion of diabase. Diabase intrusions in the Grand Canyon have been dated at 1150 to 1200 m.y. by L. T. Silver (pers. comm.).

4. Reverse faulting after deposition of the Grand Canyon Supergroup.

5. A second episode of reverse displacement prior to deposition of the Tapeats Sandstone of Cambrian age.


Sears (1973) infers a minimum of seven episodes of displacement on the Bright Angel fault:

1. Displacement of 60 m, up on the southeast, during deposition of the Shinumo Quartzite of Precambrian age.

2. Post-Dox Sandstone, pre-diabase displacement.

3. Local displacement during intrusion of diabase.

4. Post-diabase reverse displacement of 200 m.

5. Local post-Grand Canyon Supergroup scissors displacement.

6. Reverse displacement after deposition of Redwall Limestone of Mississippian age and prior to deposition of the lowermost beds of the Supai Formation, which are of Pennsylvanian age.

7. Post-Paleozoic normal displacement.

It appears that at least seven and possibly as many as nine distinct episodes of displacement can be documented at various places along the Bright Angel fault. Most of this displacement occurred during the Precambrian. Maxson believed that there had been a significant strike-slip component of displacement in post-Paleozoic time, but Sears has strongly challenged this interpretation.

The decipherable network of faults cutting the Grand Canyon Supergroup, which were active in Precambrian time, has been carefully re-examined by Sears (1973). His synthesis of this network is illustrated in figure 6. The more ancient northwest and northeast directions of faulting in the crystalline rocks are closely reflected in the late Precambrian pattern of displacement. This pattern is reflected, in turn, by Phanerozoic displacement along the Bright Angel and Kaibab fault systems. Most of the faults that were active in late Precambrian were reactivated in the Phanerozoic.
Figure 6.—Sub-Paleozoic geologic map of Grand Canyon National Park, from Sears (1973).
The early history of displacement on the Mesa Butte fault system is much less well known. An episode of Precambrian displacement is documented in the Paulden quadrangle (Krieger, 1965). Northeast of the Mesa Butte graben, minor displacement on the Grandview monocline in Permian time may be indicated by slump blocks in the Toroweap Formation of Permian age (Marshall, 1972).

AEROMAGNETIC AND GRAVITY ANOMALIES

Exposures of Precambrian rocks in northern Arizona are relatively limited, and it would be desirable to determine the relationship of fault displacements in the Phanerozoic to Precambrian structures over a broader area than the Grand Canyon. Aeromagnetic and gravity data provide powerful tools to examine Precambrian structure, especially in northwestern Arizona, where Phanerozoic rocks are generally less than 2 km thick and, except for Cenozoic volcanic rocks, very weakly magnetic. Local magnetic anomalies of limited extent and with steep gradients, associated chiefly with Cenozoic volcanic rocks in the San Francisco volcanic field, are readily distinguished from relatively broad anomalies with low gradients related to the Precambrian rocks. A residual aeromagnetic map of Arizona by Sauck and Sumner (1971) and a Bouguer gravity anomaly map of the state by West and Sumner (1973) reveal prominent structural trends in the Precambrian that are parallel with the observed systems of faults.

Comparison of the principal faults in the Bright Angel, Mesa Butte, and Sinyala fault systems with the residual aeromagnetic map reveals a close correspondence between the positions of these faults and the margins of a series of large northeast-trending magnetic anomalies. This correspondence has emboldened us to infer the traces of the concealed ancestral Bright Angel, Mesa Butte, and Sinyala faults shown in figure 7. Large displacement of the crystalline Precambrian rocks is postulated to have occurred on these faults in order to account for aligned linear margins of the anomalies. Significant displacement of the crystalline rocks along the Bright Angel fault is suggested by the observations of Noble (Noble and Hunter, 1916, p. 101).

The ancestral Bright Angel fault is exposed in the Grand Canyon for a distance of 14 km, but the ancestral Mesa Butte Fault is not exposed in the area shown in figure 7. Precambrian displacement on the Sinyala fault system has not been demonstrated on the outcrop; the ancestral Sinyala fault is inferred on the basis of the fault pattern in observed Phanerozoic rocks and on the relationship of the observed faults to the magnetic anomalies.

Perhaps the most noteworthy feature of the aeromagnetic map of Arizona is a northeast-trending belt of large positive magnetic anomalies 400 km long that extends from the vicinity of Congress to the northern border of the state. The amplitude of these anomalies ranges from about 300 to 700 gammas. Over most of its length in
Figure 7.—Simplified residual aeromagnetic map of northwestern Arizona showing ancestral Sinyala, Bright Angel, and Mesa Butte fault systems and related eruptive centers. Residual aeromagnetic intensity after Sauck and Sumner (1971). Eruptive centers along faults are shown by stars: 1) Trinity Mtn.; 2) Round Mtn.; 3) Mt. Floyd; 4) Howard Hill; 5) Bill Williams Mtn.; 6) Sitgreaves Mtn.; 7) Kendrick Pk.; 8) Slate Mtn.; 9) Mesa Butte; 10) Shadow Mtn.; 11) Tuba Butte; 12) Wildcat Pk.
northern Arizona the magnetic-anomaly belt corresponds to a belt of positive gravity anomalies with amplitudes of 10 to 30 milligals; the gravity-anomaly belt, however, is less well defined. The Mesa Butte fault system lies along the southeastern margin of the magnetic anomaly belt, and the Mesa Butte fault follows some of the steepest anomaly gradients on this margin. The monoclines along the northeast projection of the Mesa Butte fault system also follow the margin of the anomaly belt. This relationship suggests that the ancestral Mesa Butte fault continues northeastward beneath these monoclines (fig. 7).

Another large northeast-trending positive magnetic anomaly is bounded on the southeast by the Bright Angel fault. This anomaly, about 75 km long and 15 km wide, has a maximum amplitude of about 700 gammas. Where the anomaly is most pronounced, it corresponds to a well-defined positive gravity anomaly with an amplitude of 5 to 10 milligals. Where it crosses the Grand Canyon, the magnetic anomaly coincides approximately with a belt of amphibolite, migmatite, and schist referred to by Maxson (1961) as the Brahma Schist.

Maxson interpreted the Brahma Schist to be a sequence of metavolcanic and metasedimentary rocks overlying dominantly metasedimentary rocks of the Vishnu Schist. According to Maxson, the Brahma Schist is folded down into the Vishnu terrane in a large isoclinal syncline. Subsequent work has not supported this interpretation, however. The rocks called Brahma Schist by Maxson appear to be part of the Vishnu Schist sequence (Ragan and Sheridan, 1970); the name Brahma Schist has been abandoned. The positive magnetic and gravity anomalies along the prevailing strike of Maxson's Brahma Schist, on the other hand, indicate that the rocks in this part of the Vishnu terrane have higher mean magnetic susceptibility and higher mean density than adjacent parts of the Vishnu Schist, at least northwest of the Bright Angel fault. Probably these physical properties can be attributed to a greater than average abundance of amphibolite in this block of Vishnu terrane. By analogy, we suggest that the large positive magnetic anomalies associated with positive gravity anomalies adjacent to the Mesa Butte fault system are also related to belts of mafic metavolcanic rocks in the crystalline Precambrian complex.

Other large positive magnetic anomalies are bounded by inferred extensions of the ancestral Bright Angel fault (fig. 7), but these anomalies are more widely spaced than those along the Mesa Butte fault system. Some, but not all, of the positive magnetic anomalies along the Bright Angel system are associated with positive gravity anomalies.

The ancestral Sinyala fault has been drawn along the margins of a series of positive magnetic anomalies of relatively limited extent that are grouped into two broad aeromagnetic highs (fig. 7). The northeast trend of individual anomalies in these groups is not
as pronounced as along the Bright Angel and Mesa Butte fault systems. The positive magnetic anomalies along the Sinyala system correspond only very roughly with positive gravity anomalies.

Faults belonging to the northwest- and north-trending fault systems also follow linear margins of aeromagnetic anomalies, although these trends are not as obvious as the northeast trend on the aeromagnetic map. A good example is a north-trending magnetic-anomaly boundary that coincides with the north-striking Oak Creek Canyon fault. We conclude that the northwest- and north-trending fault systems are also controlled by faults with large displacement in the Precambrian crystalline basement.

**SHYLOCK AND CHAPARRAL FAULT ZONES**

The ancestral Bright Angel and Mesa Butte faults may be related in origin to the Shylock and Chaparral fault zones in the central Arizona mountain belt, which were described by Anderson (1967). Here Precambrian rocks are exposed over a wide area, and the detailed structure of the Precambrian has been worked out by Anderson and his colleagues (Anderson, 1959 and 1967; Anderson and Creasey, 1958, 1967; Anderson and others, 1971; Blacet, 1966; Krieger, 1965).

The Shylock and Chaparral fault zones are easily recognized in ERTS-1 pictures (fig. 8a), and the Shylock has been traced south of the area mapped by Anderson on the basis of the ERTS-1 data (extension shown by dashed lines in figure 8b).

The Shylock fault zone is a north-trending belt of roughly parallel, interlacing faults cutting Precambrian rocks, which ranges in width from about 1 to 3 km. As interpreted by Anderson (1967), the zone represents a major transcurrent fault with a minimum right-lateral displacement of 8 km. Estimates of this displacement are based on the offset of slices of quartz diorite in the fault zone. The total horizontal displacement may be much greater than 8 km.

A large contrast in the magnetic properties of the Precambrian rocks on opposite sides of the Shylock fault zone, indicated by aeromagnetic data (fig. 8b), suggests the displacement may be as great as several tens of kilometers. Positive magnetic anomalies on the west side of the fault are related, in a general way, to metavolcanic rocks of the Big Bug Group of Precambrian age. The anomalies do not correspond closely to mapped geologic units, however. A broad magnetic low on the east side of the fault corresponds, at least in part, to a cluster of plutonic rocks. In contrast to the relations observed along the Bright Angel and Mesa Butte fault systems, the positive magnetic anomalies along the Shylock zone are gravity lows, and the magnetic low coincides with a broad positive gravity anomaly with an amplitude of 25 milligals. The plutonic rocks east of the Shylock fault zone are evidently less strongly magnetized but have higher density than the metavolcanic rocks of the Big Bug Group.
Figure 8a.--ERTS-1 picture showing Shylock and Chaparral fault zones in central Arizona. NASA photo ERTS F-1104-17384-6, 4 Nov. 72.
Figure 8b.—Simplified residual aeromagnetic map showing Shylock and Chaparral fault zones. Traces of Shylock and Chaparral fault zones based on Anderson and others (1971) and on ERTS-1 pictures. Residual aeromagnetic intensity after Sauck and Sumner (1971).
In the Mingus Mountain quadrangle, along the southern border of the Black Hills, the Shylock fault zone is locally overlain by unbroken Tapeats Sandstone of Cambrian age (Anderson and Creasey, 1958). The large transcurrent displacement occurred before the Cambrian and after emplacement of the quartz diorite pluton that is offset by the fault. The quartz diorite has been dated at 1760 ±15 m.y. (Anderson and others, 1971). It is possible that the transcurrent displacement along the Shylock zone is related in time to faulting of the Vishnu terrane that occurred before deposition of the Grand Canyon Supergroup.

Near the Black Hills, the Coyote fault, which branches north-northwest from the Shylock fault zone, displaces Paleozoic and Tertiary rocks (Anderson and Creasey, 1958). North of the Black Hills, directly in line with the main Shylock zone, the Orchard fault also displaces Paleozoic and Tertiary rocks. At least two episodes of normal displacement have occurred on both faults (Lehner, 1958). The old transcurrent fault zone has clearly controlled the pattern of Phanerozoic normal displacement.

The Chaparral fault zone is a northeast-trending zone of distributive shear, up to a kilometer wide, that cuts Precambrian rocks (Anderson and Creasey, 1958; Krieger, 1965). Detailed structural features within the zone indicate right lateral slip (Krieger, 1965), and Anderson (1967) to account for separation of intrusive rocks common to both sides of the fault zone, suggests that appreciable right lateral displacement has taken place. The Chaparral fault zone trends toward the Shylock zone but does not offset it. It appears likely that the Chaparral joins the Shylock tangentially, but the critical area is concealed by Phanerozoic deposits.

We suggest that the northeast- and north-trending fault systems of northern Arizona are controlled by ancestral transcurrent faults in the Precambrian basement similar to the Shylock and Chaparral fault zones. The 1700 m.y. to 1800 m.y. old Precambrian terrane, represented by the Vishnu Schist in northern Arizona and the Yavapai Series (Anderson and others, 1971) in central Arizona and by associated plutonic rocks, probably was riven by right lateral faults prior to the deposition of the Grand Canyon Supergroup. Displacement on the ancestral Bright Angel and Mesa Butte faults may have been comparable to that on the Shylock fault zone. Major displacement probably occurred on a few main faults, which divide the crust into blocks tens of kilometers across. Minor shearing occurred within these blocks, particularly along their margins. Both the major and minor faults controlled later, dominantly vertical displacement in late Precambrian and Phanerozoic time.

Whether the ancestral northwest-trending faults are transcurrent faults or whether they are related tectonically to the northeast- and north-trending ancestral transcurrent faults is a problem that awaits further investigation. Evidence should be sought in the central Arizona mountain belt.
CENOZOIC HISTORY OF DISPLACEMENT

Cenozoic deformation in northern Arizona occurred in two widely separated periods. Folding on a monocline parallel with the northern end of the East Kaibab monocline took place after deposition of Upper Cretaceous strata and before deposition of beds now assigned to the Paleocene (Bowers, 1972). The principal folding of some monoclines on the eastern side of the Colorado Plateau occurred near the end of the Eocene (Shoemaker, 1956). Presumably, the monoclinal folding and broad regional warping of strata in northern Arizona took place during this episode of compressive deformation that lasted from latest Cretaceous to late Eocene, which commonly has been referred to as the "Laramide Revolution." Some reverse faulting along monoclines probably occurred during this early period of deformation. Subsequently, Paleozoic and Mesozoic strata were deeply eroded, and, in central Arizona, stripped entirely away. By mid-Tertiary time a widespread relatively mature erosion surface had developed (McKee and others, 1967), which is locally preserved beneath mid-Tertiary sediments and Oligocene and Miocene volcanic rocks (Krieger and others, 1971). Where the history of displacement on faults can be documented by stratigraphic evidence, most or all Cenozoic normal displacement has occurred after emplacement of these mid-Tertiary sediments and volcanics.

In northern Arizona large displacement took place on normal faults in Miocene time and has continued to the present. Displacement on the Grand Wash fault occurred mainly after emplacement of the 17 to 18 m.y. old Peach Springs Tuff (Young and Brennan, 1974) but before deposition of the Muddy Creek Formation of late Miocene and Pliocene (?) age (Lucchitta, 1972 and this volume). Some normal displacement along the Cottonwood Cliffs occurred prior to emplacement of the Peach Springs Tuff (Fuis, 1973). On faults farther to the east, most displacement postdates middle to late Miocene lavas but predates Pleistocene volcanic rocks. Faulting has continued into the Quaternary, however, as shown by relatively fresh fault scarps in alluvium in Chino Valley and near Cameron, Arizona (Akers and others, 1962), by minor offset of Pleistocene lava flows, and by the present seismicity of the region.

The Cenozoic history of displacement on the Mesa Butte fault system is fairly well known. On the southwestern segment of the system, most displacement postdates old basaltic lavas in the southern San Francisco volcanic field and predates lavas and sediments now assigned by Krieger and others (1971) to the Perkinsville Formation of Pliocene and Pleistocene age. In some places older units of the Perkinsville are displaced and younger units are not. Old basaltic lavas in the southern part of the San Francisco volcanic field have been dated at 11.1 ±0.5 m.y. and 14.4 ±0.6 m.y. from a locality near Sycamore Canyon by McKee and McKee (1972). Damon and others
(this volume) have dated a basalt from another locality near Sycamore Canyon at 8.68 ±0.98 m.y. These localities are 15 to 25 km east of the prominent fault scarp, where old basalts are displaced.

On the northeastern segment of the Mesa Butte fault system, most displacement occurred before the extrusion of lavas that follow the base of fault scarps and conceal the faults. Mesa Butte and the lava flows from this vent are clearly later than most of the displacement on the Mesa Butte fault system. Locally these flows are offset, however, both on the Mesa Butte fault and on the Cedar Ranch fault. On the basis of many dated basaltic lavas in nearby parts of the San Francisco volcanic field, normal displacement on the fault systems near Mesa Butte appears to have occurred between about 4.0 and 0.5 million years ago (Damon and others, this volume).

The youngest lavas displaced by faults of the Mesa Butte system are a flow from the Shadow Mountain vent, dated at 0.62 ±0.23 m.y. (Damon and others, this volume), and the Tappan Wash flow, dated at 0.510 ±0.079 m.y. (Damon and others, this volume). The Shadow Mountain lava is displaced about 13 m by a graben that cuts the flow (Condit, this volume). The Tappan Wash flow has been traced by M. Malin and E. M. Shoemaker from a vent southwest of Red Mountain, on the trend of the Mesa Butte fault (fig. 4), to previously recognized outcrops in Tappan Wash and along the Little Colorado River. Where it crosses the Cedar Ranch fault, it appears to be displaced several meters. Along the Little Colorado River (fig. 4) the Tappan Wash flow is displaced about 20 m by the Cameron graben (Reiche, 1937). According to Reiche, the graben is entirely younger than the flow. A channel was cut into the flow by the Little Colorado River, filled with alluvium, and then abandoned, all before development of the graben (Reiche, 1937).

Evidence of the time of displacement on the Bright Angel fault system is indirect. The principal displacement on some faults of the Cataract Creek system, which intersects the Bright Angel system near Cataract Creek, occurred after deposition of mid-Tertiary sediments and extrusion of basalt flows in the nearby Mount Floyd volcanic field. Basalt flows at Long Point, on the northern edge of this field, have been dated by McKee and McKee (1972) at 7.4 ±0.4 m.y. and 14.0 ±0.6 m.y. A well integrated dendritic drainage system in the Cataract Creek basin appears to be antecedent to the Bright Angel fault (fig. 3b). However, the prisms of Moenkopi Formation preserved along several faults in the system show that displacement began before the Moenkopi Formation was stripped from most of the Coconino Plateau. Several faults in the system have controlled the development of side canyons in the Grand Canyon. (See, for example, fig. 3b.) Probably the drainage pattern is antecedent (Hunt, 1956) to the fault system.

Several lines of evidence (e.g., Hamblin, 1970) indicate that normal faulting in northern Arizona has continued into the Holocene. Indeed, the fault systems are seismically active. Three moderately strong earthquakes (intensity VI to VIII on the Modified Mercalli Scale) occurred in 1906, 1910, and 1912 near or north of Flagstaff.
(Townley and Allen, 1939; Sturgul and Irwin, 1971; Coffman and von Hake, 1973). Newspaper accounts suggest the epicenters of the 1910 and 1912 earthquakes may have been near the intersection of the Mesa Butte, Oak Creek Canyon, and Kaibab fault systems. Sturgul and Irwin (1971) report 17 earthquakes of intensity IV or greater in northwestern Arizona during the period 1850 to 1966. Except for a few of these earthquakes, however, the epicenters are not known with sufficient precision for comparison with the fault systems.

As of January 11, 1974, 27 earthquakes in northwestern Arizona (fig. 9), for which the epicenters have been estimated to 0.1 degree latitude and longitude or better, were on record in the hypocenter data file of the National Oceanic and Atmospheric Administration. Only locations determined with five or more seismographic stations are included. The actual error in location of several of the epicenters plotted in figure 9 probably is greater than 0.1 degree. All focal depths are shallow, probably in the crust. The epicenter locations of relatively minor earthquakes that occurred after 1969 probably are most reliable.

All but one of the epicenters shown in figure 9 lie either within the previously defined fault systems, on the projection of these systems, or along a major fault. The earthquake plotted as No. 2 followed earthquake No. 1 by less than two hours. It is likely that the distance between the epicenters of these two earthquakes is less than indicated in figure 1; one or both of the epicenters may be significantly in error. Another pair of events (10a and 10b) is separated by 10 minutes in time and by about 20 km on the map; a third pair (14 and 15), along the Grand Wash fault, is separated by 2.5 minutes and 20 km.

The Toroweap, Sinyala, Bright Angel, Mesa Butte, Kaibab, and Oak Creek Canyon fault systems are demonstrably active. A magnitude 5.1 earthquake appears to have occurred along or near the Eminence Break graben in 1945. Five epicenters shown near the western border of the map are along or near the surface trace of the Grand Wash fault.
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<td>December 3, 1970</td>
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<td>December 16, 1970</td>
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<td>15</td>
<td>December 16, 1970</td>
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<td>23</td>
<td>April 12, 1973</td>
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</table>

Figure 9.—Distribution of earthquake epicenters in northwestern Arizona for the period 1938 to 1973. Epicenters shown with large circles generally are least accurately known; those shown with solid dots generally are most accurately known. Principal fault systems are shown with stipple.
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ECONOMIC GEOLOGY

Review of the development of oil and gas resources of northern Arizona

by

J. N. Conley
REVIEW OF THE DEVELOPMENT OF OIL & GAS RESOURCES
OF NORTHERN ARIZONA

by

J. N. Conley
State of Arizona Oil and Gas Conservation Commission
Phoenix, Arizona

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INTRODUCTION

Northern Arizona constitutes one of the few remaining unexplored and untested potentially hydrocarbon-productive onshore areas in the contiguous 48 states. The limited drilling exploration to date has proved the presence of oil and gas accumulations near the northeast corner of the state; it has indicated that other large areas have the basic geologic requirements for accumulations. For this review, "Northern Arizona" is considered to be that portion of the Colorado Plateaus province of the state north of the central mountainous region (fig. 1). For discussion purposes, this portion of the state has been subdivided into two regions, "northwest" and "northeast." Data have been taken from well and production records of the Oil and Gas Conservation Commission and from Peirce and others (1970), Peirce and Scurlock (1972), and Scurlock (1973).

HISTORY OF EXPLORATION

In the northwest region, the first recorded attempt to find oil was a hole drilled in 1905 to a reported depth of 611 m (2,003 ft) located in sec. 27, T. 18 N., R. 2 W., about 32 km (20 mi) north of Prescott, Yavapai County.

Only a few of the 48 test holes drilled in the northwest region since 1905 were situated on prospects defined by geological and(or) geophysical information. Some of the holes were drilled on known surface structural anomalies; others on weak geological information; and many were strictly "promotional" ventures.

In the northeast region, incomplete records indicate that two oil test holes were drilled in 1923-1924, one in northwestern Apache County and one in southwestern Navajo County. From then until late 1954 approximately 14 oil test holes were drilled. Most of these were drilled in southern Navajo County on surface anticlines; most of those on the Navajo Reservation were drilled on prospects based on geologic and(or) geophysical information. Early in 1954 Shell Oil Company attempted unsuccessfully to complete its Navajo 1 in sec. 6, T. 41 N., R. 29 E., Apache County, as a commercial oil well from reservoir rocks in the Pennsylvanian Hermosa Group. Shows of oil and gas were recorded also in rocks of Mississippian and Devonian age. In December of the same year Shell completed its Navajo 2, sec. 3, T. 41 N., R. 28 E., as the discovery well of the East Boundary Butte pool (fig. 2). The initial production gauge was 89,200 m³ (3,150,000 cubic ft) of gas and .06 m³ (3.6) barrels of oil per day through casing perforations at 1,384 m-1,398 m (4,540-4,585 ft) and 1,418 m-1,430 m (4,650-4,690 ft) in the Ismay and Desert Creek units, respectively, of the Pennsylvanian Hermosa group. This discovery, coupled with the late 1954 discovery of the Desert Creek pool about 23 km (14 mi) to the north, initiated an intensive exploratory program in the Four Corners area that resulted in the discovery of numerous oil pools in San Juan County, Utah.
Figure 1.—Map of State of Arizona, showing location of northwest and northeast regions.
Figure 2.—Map of northeast region of Arizona and adjacent portion of Utah, showing location of oil and gas pools. A one-well unnamed pool adjacent to Bita Peak is not shown.
Prior to 1957 the Navajo Indian Tribe, which controls approximately 10,000,000 acres of land in the northeast portion of Arizona, had put under lease several thousands of acres. During 1957 the tribe leased 198,842 acres, primarily to major oil companies. Subsequent geologic, geophysical, and drilling exploration during the 1955 to 1965 period resulted in the discovery of eight areally small oil and gas pools in the general East Boundary Butte area (fig. 2 and table 1).

Arizona's best oil pool, Dineh-bi-Keyah, was discovered in 1967. The discovery well was drilled by Kerr-McGee Corporation in sec. 32, T. 36 N., R. 30 E., Apache County. It was completed for an initial production gauge of 10 m$^3$ (634) barrels of oil per day through casing perforations at 872 m-880 m (2,860-2,885 ft) in a Tertiary igneous sill intruded into rocks of the Pennsylvanian Hermosa Group.

From early 1965 to October 1973, approximately 104 exploratory holes were drilled in the northeast region. The peak period of drilling activity occurred in 1967, 1968, and 1969. Most of the holes drilled were located in or near productive areas in the northeast corner of Apache County. Drilling near the Dry Mesa pool resulted in discovery of the Black Rock gas pool in 1971. Very little exploratory drilling for oil and gas has been done since 1969.

DENSITY OF DRILLING

Drilling activity generally tends to concentrate in and near productive areas. This has been the case for exploratory drilling in northern Arizona. There are vast areas in this part of the state devoid of holes drilled in search of hydrocarbons. Figure 3 presents graphically the few relatively densely drilled areas and the vast areas completely untested. Fortunately for the subsurface geologist, numerous holes drilled for stratigraphic, structural, mineral (potash and Halite), and helium information, plus hundreds of holes drilled for water, furnish stratigraphic and structural control in several areas lacking oil test holes (fig. 4). In the general Holbrook area of Apache and Navajo Counties, at least 126 holes have been drilled for stratigraphic information (predominantly for potash in the Permian Supai Formation). These holes had a depth range of 130 m (428 ft) to 1,030 m (3,380 ft) and an average depth of 448 m (1,471 ft). The total footage, drilled or cored, amounted to more than 56,523 m (185,391 ft). Most of the wells drilled for water reaching stratigraphic units useful in subsurface mapping are located in the southwestern half of northern Arizona.

Table 2 lists by counties the number of oil and helium test holes drilled, and pertinent data, in the northwest and northeast regions. The 18 holes in the northwest region bottoming in Precambrian rocks did not develop any regional stratigraphic information of value as most of them were drilled in Tps. 17 and 18 N., Rs. 4 and 5 E., in northern Yavapai County. Most of the 112 holes reaching Precambrian
Table 1.—Wells drilled for oil, gas, and helium in northwest and northeast regions of Arizona: table shows number of wells drilled, footage drilled, depth range, average depth, and number of wells penetrating pre-Pennsylvanian rocks.

<table>
<thead>
<tr>
<th>COUNTY</th>
<th>NUMBER OF WELLS</th>
<th>FOOTAGE DRILLED</th>
<th>DEPTH RANGE (Feet)</th>
<th>AVERAGE DEPTH (Feet)</th>
<th>NUMBER OF WELLS BOTTOMING IN PRE-PENNSYLVANIAN ROCKS</th>
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<tbody>
<tr>
<td></td>
<td>WELLS DRILLED</td>
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<td>215-3868</td>
<td>2607</td>
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<td>439</td>
<td>1,373,592</td>
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*Drilled for oil or gas - 88; drilled for helium - 120.
Figure 3.—Map of Colorado Plateau province (northwest and northeast regions), showing location of wells drilled for oil, gas, and helium.
Figure 4.—Map of Colorado Plateau province (northwest and northeast regions), showing location of wells drilled for oil, gas, helium, stratigraphic information (principally for potash deposits), and selected stratigraphically significant wells drilled for water.
Table 2.—Oil and gas pools in northeast region of Arizona: table shows location, discovery date, well status, and production data.

<table>
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<tr>
<th>POOL</th>
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<td>Black Rock</td>
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<td>Dry Mesa</td>
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<td>6</td>
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<td>North Toh-Atin</td>
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<td>1956</td>
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<tr>
<td>Teece Nos Pos</td>
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<td>1959</td>
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<td>Total</td>
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*Incomplete records for some wells
rocks in the northeast region were drilled in northeastern Apache County. As of October 1, 1973, a total of 439 oil or helium test holes had been drilled in northern Arizona; total feet drilled, 418,778 m (1,373,592 ft); and average depth, 954 m (3,129 ft). The deepest hole, drilled by Monsanto Company in sec. 1, T. 39 N., R. 29 E., was abandoned at a depth of 2,580 m (8,461 ft) in an igneous sill intruded into rocks of the Devonian Ouray Formation.

**STRATIGRAPHIC OCCURRENCE OF OIL AND GAS SHOWS**

In the northwest region, shows of oil have been noted in Permian, Pennsylvanian, and Devonian rocks (table 3). In the northeast region, shows of oil and(or) natural gas have been noted in Cretaceous, Triassic, Permian, Pennsylvanian, Mississippian, Devonian, and Cambrian rocks, and in Tertiary igneous sills intruded into these rocks. In this review a "show" has been considered to be any indication of hydrocarbons occurring naturally in rocks. The quality of shows noted varied from light oil staining visible on drill bit cuttings to measurable quantities of oil or gas recovered on drill stem tests. Most of the shows in the northeast part of the northeast region were evident in drill bit cuttings, in cores, and in the drill stem test recoveries of fluids and gas. In the south-central portions of Apache and Navajo Counties in the northeast region and all of the northwest region, most of the shows were noted in the drill bit cuttings; a few cores had oil shows.

**OIL AND GAS POOLS**

In seven of the oil and gas pools (productive, shut-in, or abandoned) the productive zones occur in the Hermosa Group of Pennsylvanian age (table 4). In the East Boundary Butte pool, for example, the Upper Hermosa, Lower Ismay, Upper, Middle, and Lower Desert Creek, and Akah units contain productive zones. With just two exceptions, all of the wells produce, or have produced, from zones in two or more units. The Lower Ismay and Upper Desert Creek units have the greatest number of productive zones. Limestone is the stratal lithology of all the Pennsylvanian productive zones in the East Boundary Butte and other pools nearest the extreme northeast corner of the state. Entrapment of hydrocarbons in these productive zones is controlled by a combination of favorable stratigraphic and structural conditions.

The Mississippian Redwall Formation is oil- and gas-productive in the Dry Mesa pool (table 4). An unnamed one-well pool in sec. 7, T. 40 N., R. 29 E., produced some oil from the Redwall prior to early abandonment. The stratal lithology of most of the productive zones in the Redwall Formation is dolomite.

The Devonian McCracken Sandstone is not productive currently in northeast Arizona. Two one-well pools however, Walker Creek and an
Table 3.—Stratigraphic occurrence of shows of oil and gas in the northwest and northeast regions (Not a correlation chart.)

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<th>BLACK ROCK</th>
<th>DINEH-BI-KEVAH</th>
<th>DRY MESA</th>
<th>E. BOUNDARY BUTTE</th>
<th>NORTH TOH-ATIN</th>
<th>TEC NS POS</th>
<th>TWIN FALLS CREEK</th>
<th>UNNAMED</th>
<th>UNNAMED</th>
<th>WALKER CREEK</th>
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**EXPLANATION**

- Geologic age of surface rock
  - Producing or Shut-in
    - Oil
    - Gas
    - Oil and Gas
    - Helium
- Deepest stratigraphic penetration and depth in feet
- Missing Section
  - Regional (Schematic)
  - Local
  - Unconformity

**Table 4.**—Oil and gas pools in northeast region of Arizona: table shows deepest stratigraphic unit penetrated, productive stratigraphic unit, productive status as of October 1, 1973, major unconformities, and missing geologic section.
unnamed one in sec. 36, T. 41 N., R. 30 E., produced some oil from this formation prior to abandonment.

In the Dineh-bi-Keyah pool, about 53 km (33 mi) south of the northeast corner of the state, the productive reservoir is a fractured syenite sill of Oligocene age. The sill intruded dense shale and limestone of the Lower Hermosa Group of Pennsylvanian age. The pool is near the north end of an anticline with more than 396 m (1,300 ft) of closure, but the oil accumulation is controlled primarily by stratigraphy (McKenny and Masters, 1968).

The 10 pools having a production history have produced a total of 21,964 m³ (13,925,291 barrels) of oil and 323,210 million m³ (11,405,313 million cubic ft) of gas. The gross value of hydrocarbons produced amounts to about $42 million as of October 1, 1973.

CONCLUSIONS

With the exception of a few relatively small areas, there have been no well-planned and efficiently executed exploratory programs in northern Arizona. This part of the state has many large unexplored or incompletely explored areas having the basic factors normally considered requisite for oil and gas accumulation. Numerous shows of oil noted in igneous sills intruded into rocks of Paleozoic age in the northeastern corner of the state suggest the possibility that there are other "non-normal" oil accumulations similar to the one at Dineh-bi-Keyah. Northern Arizona can be considered as an attractive unexplored onshore area offering potentially large accumulations of oil from a variety of traps in Paleozoic rocks. The possibility of accumulations existing in lower Triassic rocks should not be fully discounted, as minor shows of oil have been noted in a few wells.

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