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Stratigraphy and origin of the Chinle formation (Upper Triassic)

on the Colorado Plateau

By
John H. Stewart, 1928-

U. S. Geological Survey
OPEN FILE REPORT
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ABSTRACT

The Chinle formation of Late Triassic age consists of continental deposits of conglomerate, sandstone, siltstone, claystone, and limestone. It extends throughout most of the Colorado Plateau region, and ranges in thickness from 0 to 1700 feet.

Thin units or lenses of siltstone, and in some places sandstone and conglomerate, characterized by a peculiar mottling of red, purple, and grey occur directly below or in the basal few feet of the Chinle formation. The mottled strata are interpreted as remnants of a soil; the strata are widely distributed along an unconformity at the base of the Chinle formation and have a coloration similar to some present day soils.

The Chinle formation is divided on a lithologic basis into two parts. The lower part extends over the southern part of the Colorado Plateau and into adjacent regions. It contains four main stratigraphic units or groups of units, (1) Shinarump and related members, (2) Monitor Butte and related members, (3) Moss Back member and related units, and (4) Petrified Forest member. The Shinarump and Moss Back members are thin widespread cross-stratified sandstone and conglomerate units that weather to form ledges. The Monitor Butte and Petrified Forest members are composed of variegated bentonitic slope-forming claystone, clayey siltstone, and clayey sandstone, and locally contain ledge-forming sandstone units.

The upper part of the Chinle formation extends throughout most of northeastern Arizona, northwestern New Mexico, eastern Utah, and western Colorado. It is divided into two parts, (1) Owl Rock member and (2) Church Rock member and related units. The Owl Rock member is composed of reddish-brown siltstone and thin beds of pale-red and light greenish-gray limestone. The Church Rock member is composed of reddish-brown siltstone and minor amounts of sandstone. The sandstone is commonly cross-stratified and is abundant in a narrow belt extending from southwestern Colorado to central Utah.

The fossils and sedimentary structures in the lower part of the Chinle formation indicate that it is a vast alluvial plain deposit. The thin widespread sandstone and conglomerate units, such as the Shinarump and Moss Back members, are probably braided stream deposits, for they are similar to deposits of present day braided streams. The claystone, clayey siltstone, and clayey sandstone of the Monitor Butte and Petrified Forest members, on the other hand, probably are deposits of large meandering streams and of lakes. Stream directions, as indicated by the orientation of cross-strata, are north to northwest. The source area of most of the material in the lower part of the formation, as indicated largely by stream directions, was in southern Arizona and adjacent states—the Mogollon highland. The type of detrital material in the lower part of the formation indicates that the Mogollon highland was predominantly a volcanic terrain and that it also contained cherty limestone or dolomite, sandstone, metasedimentary rocks, and probably granitic rocks.

The fossils and sedimentary structures in the upper part of the Chinle formation indicate that it is a large deltaic deposit spread out into a lake. The deposits of cross-stratified sandstone that extend in a narrow belt from southwestern Colorado to central Utah are probably the deposits of a large river which formed the delta. Stream directions in and near this narrow belt of fluvial sandstone are dominantly northwest. High igneous and metamorphic terrains in western Colorado and adjacent regions--the Uncompahgre and Front Range highlands--were the main source areas during deposition of the upper part of the formation.

INTRODUCTION

The Chinle formation of Late Triassic age crops out widely on the Colorado Plateau (fig. 1), and consists of continental deposits of conglomerate, sandstone, siltstone, claystone, and limestone. The lower part of the formation contains brightly colored and variegated claystone and clayey sandstone and thin widespread conglomerate and sandstone layers. It contains the Shinarump, Monitor Butte, Petrified Forest, and related members. The upper part of the formation contrasts markedly with the lower part, and consists of reddish-brown, evenly bedded siltstone and minor amounts of limestone and very fine-grained sandstone. The upper part contains the Owl Rock and Church Rock members and related units. On most of the Colorado Plateau, the Chinle formation overlies the Moenkopi formation of Early and Middle (?) Triassic age and underlies the Wingate sandstone of Late Triassic age.

In most areas, the sedimentary rocks of the Colorado Plateau are flat-lying. The structural flatness is interrupted by long linear, or gently arcuate, monoclines that commonly extend for fifty or more miles and have a structural relief of one or two thousand feet. The Chinle formation is exposed in deep canyons cut into the flat-lying rocks, along and extending away from the base of extensive escarpments held up by flat-lying resistant formations, and along the monoclines. The climate in most parts of the Colorado Plateau is arid, vegetation is sparse, and exposures are excellent. In most areas, every foot of rock is exposed and can be studied in detail.

The study which is the basis of this report consisted of a regional stratigraphic analysis of the Chinle formation covering 190,000 square miles of Utah, Colorado, Arizona, New Mexico, and Nevada--the Colorado Plateau and some adjacent regions. The study was designed to obtain information regarding the areal distribution, local and regional differences in rock type, sources and character of constituents, and conditions of deposition of these strata. This report summarizes the stratigraphy of the Chinle formation, and is primarily concerned with the environment of deposition of the formation and the paleogeography of Late Triassic time.

The regional stratigraphic work consisted of detailed correlations of lithologic units throughout the Colorado Plateau and a few regions adjacent to the Plateau, with the purpose of establishing a firm background in the distribution, lithology, facies, and thickness of strata. About 100 stratigraphic sections were measured and described on outcrops, and many sections measured by other geologists were studied in detail. In general, stratigraphic sections were measured at a spacing of 15 to 30 miles along outcrops. Stratigraphic units were correlated between sections on the basis of lithologic characteristics and also by tracing of units along outcrops. A minor amount of time was spent in study of logs of drill holes.

The work was conducted by the U. S. Geological Survey and was a part of a large project to study the Triassic strata of the Colorado Plateau done partly on behalf of the U.S. Atomic Energy Commission. The field work was carried out during the field seasons of 1952 through 1956. Office compilation was during the winter months of these years, during 1957 and 1958, and on a part-time basis at Stanford University during 1959 and 1960. The writer also worked in close cooperation with other members of the U. S. Geological Survey who studied lithofacies relationships, sedimentary structures, clay mineralogy, sedimentary petrology, and pebble types in the Triassic rocks.

The writer was assisted in the field at various times by George A. Williams, Omer B. Raup, Forrest G. Poole, and Richard F. Wilson, and others, and the help of these geologists is gratefully acknowledged. The stratigraphic sections and correlations in northern Utah and northwestern Colorado were mostly made by F. G. Poole in close cooperation with the writer. The writer is grateful for the use of this information. The writer has also used both published and unpublished information collected by other geologists working with the U. S. Geological Survey in the study of the Triassic rocks. This information includes the work of R. F. Wilson on lithofacies relationships, of F. G. Poole, O. B. Raup, and G. A. Williams on sedimentary structures, of L. G. Schultz on clay mineralogy, of R. A. Cadigan on sedimentary petrology, and of William Thordarson and H. F. Albee on pebble types. The specific contributions of these geologists are noted in the text, but the writer is particularly grateful to these people for many valuable discussions concerning the Triassic rocks. J. W. Harshbarger, M. E. Cooley, and C. A. Repenning helped greatly in the study of the stratigraphy of the Chinle formation in the Navajo Indian Reservation in Arizona. L. C. Craig, who supervised much of the work, made valuble suggestions and provided continual encouragement throughout the course of the study. The writer is also grateful to G. A. Thompson, W. R. Dickinson, and other professors at Stanford University for their advice and help during the course of the study and during the preparation of the manuscript.

The writer has made extensive use of published and unpublished material on the Triassic stratigraphy of the Colorado Plateau. The stratigraphic correlations, except where noted, and the conclusions reached in the interpretative part of the report are, however, the contribution and responsibility of the writer.

PREVIOUS WORK AND HISTORY OF NOMENCLATURE

Geologic work on the Colorado Plateau was begun in 1853 and has continued at an expanding rate. The earliest work, in 1853, was by Jules Marcou (1856, 1858), who traversed central New Mexico and Arizona and recognized various stratigraphic units including the New Red sandstone corresponding to the present day Triassic rocks. J. S. Newberry (1861, 1876), first as a member of the Ives expedition in the years 1857 and 1858 and later of the Macomb expedition in the year 1859, briefly described the geology, including the Triassic strata, in northern Arizona, northwestern New Mexico, southwestern Colorado, and southwestern Utah.

During the 1860's, 1870's, and 1880's, the Colorado Plateau was the site of four great surveys that were sponsored largely by the U. S. government. These are the famous Wheeler, Hayden, King, and Powell Surveys. The Wheeler Survey (U. S. Geographical Surveys West of the One Hundredth Meridian) under the direction of the U. S. Army undertook geologic study of the Colorado Plateau, particularly the western border in southwestern Utah and northwestern Arizona. The Hayden Survey (Geological and Geographical Survey of the Territories) under the direction of the Department of the Interior covered most of Colorado and small adjacent parts of Utah, Arizona, and New Mexico. The King Survey (U. S. Geological Exploration of the Fortieth Parallel) under the direction of the U. S. Army described the geology of parts of northern Utah and Colorado. The work done under the direction of Major

John W. Powell, and often referred to as the Powell Survey, included study carried out with private funds, as well as under the Department of the Interior. This work was in parts of northern Arizona, eastern Utah, and northwestern Colorado. These Surveys recognized and named many stratigraphic units, some of which were correlated throughout a large part of the Colorado Plateau. Perhaps the most impressive attempt at regional correlation was that of Major Powell (1876) who recognized four major groups of strata of Jura and Trias age, as he called them. The divisions are in ascending order: (1) the Shinarump group, (2) the Vermillion Cliff group, (3) the White Cliff group, and (4) the Flaming Gorge group. Powell correlated these groups between the Uinta Mountains on the north and northwestern Arizona on the south. The Shinarump group included the Moenkopi and Chinle formations as recognized today.

Of the names proposed by Powell only the term Shinarump is used today, and that has been modified to be the Shinarump member of the Chinle formation. Powell used the term Shinarump in a dual sense; in one sense to describe a group and in the other sense to describe a thin sandstone and conglomerate unit (his Shinarump conglomerate) within that group. This latter usage is the one that has been retained. Powell (1873) first used the term Shinarump to describe one of the erosional cliffs (the Shinarump Cliffs) in southwestern Utah. Gilbert (1875a) and Howell (1875) used the term Shinarump conglomerate to describe the resistant unit that forms the Shinarump Cliffs, and Gilbert and Howell are often given credit for the name Shinarump conglomerate.

References to Powell in the works of Gilbert and Howell, however, leave little doubt that they considered that Powell was the originator of the term, although Powell did not formally use the term in a publication until 1876. The Shinarump conglomerate was considered a separate formation until it was redefined as the Shinarump member of the Chinle formation by Stewart (1957).

Whitman Cross and his associates extensively mapped and described the geology of the San Juan Mountains region (fig. 1) in southwestern Colorado and published their findings in a series of publications dating from 1899 to 1914. In this work, Cross and his associates defined the Dolores formation (Cross and Purington, 1899) and later modified this definition (Cross and Howe, 1905a). The term Dolores formation as used in Cross's modified definition is still generally used today in the San Juan Mountains region and is used in this report. The Dolores formation, however, is a unit of Late Triassic age and is entirely equivalent to the more extensively recognized Chinle formation. The only reason for retaining the term Dolores formation in the San Juan Mountains region in preference to the term Chinle formation is the factor of usage--the term Dolores formation is the accepted term for these rocks in the San Juan Mountains region.

In a series of articles about the Navajo Country, H. E. Gregory (1914, 1916, 1917) described many of the salient features of the Triassic stratigraphy in north-central and east-central Arizona. Gregory's paper (1917) on the Navajo Country is an outstanding contribution to the stratigraphy of the Colorado Plateau. Although much of Gregory's work has been revised, many of his names and stratigraphic divisions are still used today. Gregory (1917) recognized the Shinarump conglomerate and named the overlying Chinle formation. His Shinarump conglomerate is largely the same unit that is now called the Shinarump member of the Chinle formation. Gregory named the Chinle formation for exposures in Chinle Valley in east-central Arizona and recognized four divisions of the formation, which are in descending order the A, B, C, and D divisions. These divisions correspond to units now given formal names. The A division is the Church Rock member of the Chinle formation; the B division is the Owl Rock member; the C division is the Petrified Forest member; and the D division is the Monitor Butte and lower red members. These divisions, now recognized as formally named members, are the basis of much of the detailed stratigraphic work recently done on the Colorado Plateau.

In the 1920's, 1930's, and 1940's, many detailed geologic mapping projects and stratigraphic studies were carried out on the Colorado Plateau. These studies were mostly under the direction of the U. S. Geological Survey. The most important stratigraphic paper in this period was one by Baker, Dane, and Reeside (1936) on the correlation of Jurassic formations in part of Utah, Arizona, New Mexico, and Colorado. This paper, although mainly concerned with Jurassic formations, also contains considerable information on Triassic rocks. Other papers during this period that contain important stratigraphic information include Baker (1933, 1936, 1946), Dane (1935), McKnight (1940), Hunt (1953), Longwell and others (1923), and Gilluly and

Reeside (1928).

In the Zion Park region in southwestern Utah, Gregory (1950) recognized the Shinarump conglomerate and the overlying Chinle formation. He divided the Chinle formation into four members, which are in ascending order: (1) lower sandstones, (2) Petrified Forest member, (3) Springdale sandstone member, and (4) upper sandstones. The thin "lower sandstones" unit is difficult to recognize as a distinct member and is included in the Petrified Forest member in this report. Gregory's paper contains the original definition of the Petrified Forest and Springdale sandstone members. Later work by Harshbarger and others (1957) and Averitt and others (1955), however, has shown that the upper part of Gregory's original Petrified Forest member as well as his Springdale sandstone member and upper sandstones are part of the Glen Canyon group, which includes the Wingate sandstone and overlies the Chinle formation elsewhere on the Colorado Plateau. The Chinle formation in the Zion Park region, therefore, is only the lower part of what was originally considered to be the Chinle formation in that region. Similarly the Chinle formation as presently recognized in southern Nevada is only the lower part of what was originally considered to be Chinle formation (Wilson and Stewart, 1959). The nomenclature used in this report in southwestern Utah, northwestern Arizona, and southern Nevada is shown in figure 2.

In northeastern Arizona and southeastern Utah, members of the Chinle formation have been defined recently. Stewart (1957) proposed that the Shinarump conglomerate be redefined as the Shinarump member

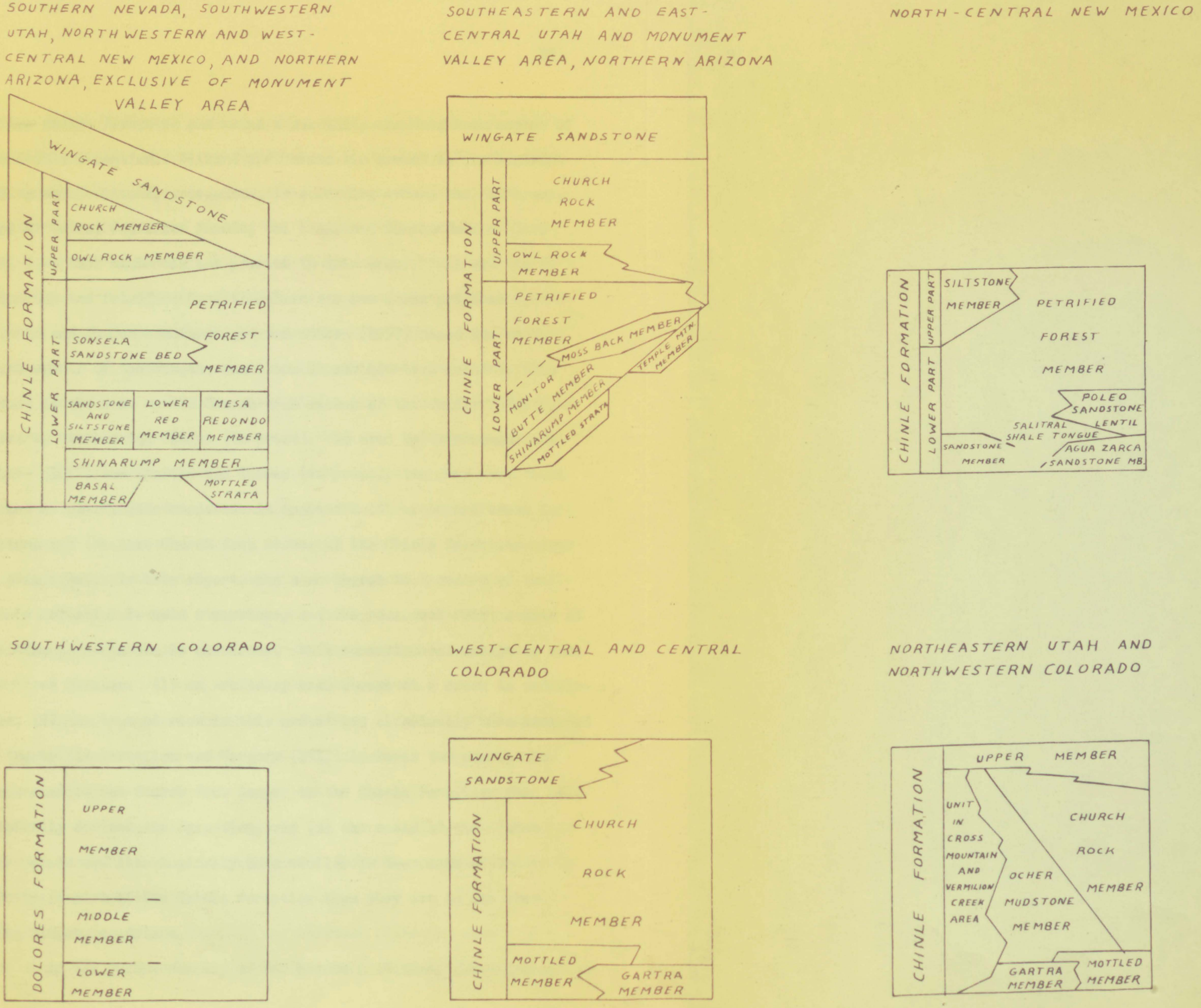


FIGURE 2.-- NOMENCLATURE OF CHINLE AND DOLORES FORMATIONS, AND WINGATE SANDSTONE ON THE COLORADO PLATEAU

of the Chinle formation and named a new unit, the Moss Back member of the Chinle formation. Witkind and Thaden (in press) in the Monument Valley area, Arizona, recognized, in ascending order, the Shinarump, Monitor Butte, Petrified Forest, Owl Rock, and Church Rock members (the Moss Back member is not present in this area). All but the Shinarump and Petrified Forest members are new names proposed by Witkind and Thaden. Harshbarger and others (1957) named the Rock Point member of the Wingate sandstone in northeastern Arizona; this member is the same as the Church Rock member of the Chinle formation named by Witkind and Thaden (in press). As used by Harshbarger and others (1957) and Witkind and Thaden (in press), the name Rock Point member of the Wingate sandstone is applied south of Laguna Creek in Arizona and the name Church Rock member of the Chinle formation north of this creek. In this report, the name Church Rock member of the Chinle formation is used everywhere, and the name Rock Point member of the Wingate sandstone is abandoned. This nomenclatural change is made for three reasons: (1) an arbitrary name change at a creek is undesirable, (2) the type of rock in this member has classically been included in the Chinle formation and Gregory (1917) included the strata now designated as the Church Rock member in the Chinle formation when he originally defined the formation, and (3) the rocks of the Church Rock member are lithologically more similar to the rocks of the underlying part of the Chinle formation than they are to the overlying Wingate sandstone.

In the Navajo country of northeastern Arizona, Akers and others

(1958) and Cooley (1959), recognized the following members of the Chinle formation: the Shinarump, lower red, Petrified Forest, and Owl Rock members. They considered the Church Rock member, their Rock Point member, to be part of the Wingate sandstone. The lower red member corresponds to the Monitor Butte member as recognized to the north. Akers and others (1958) also recognized a prominent sandstone unit in the Petrified Forest member--the Sonsela sandstone bed. Cooley (1958) proposed the name Mesa Redondo member in east-central Arizona for a unit somewhat similar to the lower red and Monitor Butte members. The Monitor Butte, lower red, and Mesa Redondo members all occupy approximately the same stratigraphic position, but occur in separate areas. All three names are used in this report because each has a use to describe a slightly different lithologic type. The terminology adopted in this report in northeastern Arizona and southeastern Utah is shown in figure 2, and consists essentially of the Shinarump, Monitor Butte (lower red and Mesa Redondo), Moss Back, Petrified Forest, Owl Rock, and Church Rock members.

In north-central New Mexico, Northrup and Wood (1946) defined the Agua Zarca sandstone member, the Salitral shale tongue, and the Poleo sandstone lentil in the lower part of the Chinle formation. The name Poleo was originally proposed by von Huene (1911), but the other two names were new. The nomenclature of Northrup and Wood is largely used in this report (fig. 2).

In northeastern Utah and northwestern Colorado, strata of Late Triassic age have been divided by many geologists (Thomas and others, 1945; Huddle and McCann, 1947; Kinney, 1951 and 1955; Brill, 1944; Donner, 1949; Sheridan, 1950) into the Shinarump conglomerate and Chinle formation. Because of regional work on the Colorado Plateau, the Shinarump conglomerate in this area can be shown to be a separate unit and is referred to as the Gartra member of the Chinle formation in this report. The name Gartra was originally proposed by Thomas and Krueger (1946) who used the term Gartra grit member of the Stanaker formation. The Stanaker formation is the same as the Chinle formation used in this report. Other members are also recognized in the Chinle formation in northeastern Utah and northwestern Colorado (fig. 2).

CLASSIFICATION OF DETRITAL SEDIMENTARY ROCKS

A new classification of detrital sedimentary rocks is used in this report. The classification is similar to those proposed by Gilbert (1955, fig. 96 and 97) and Pettijohn (1957, p. 290-293), but differs from both in important details. The classification was developed in order to express in a rock name two factors that are lacking in the other classification, namely (1) the mean grain size of the rock, i.e., whether the rock is a conglomerate, sandstone, siltstone, or claystone and (2) the dominant clay type, i.e., kaolinite, illite, or montmorillonite, in the rock, if the clay is quantitatively an important part of the rock. In addition, the classification expresses the factors of provenance, maturity, and fluidity (Pettijohn, 1957) that are important elements in other petrologic classifications. No complicated new names have been invented. In addition, many of the controversial or ambiguous terms of the other classifications have been avoided.

The fundamental part of each rock name is a textural term such as conglomerate, sandstone, siltstone, and claystone. The textural term is modified by mineralogic names that describe the composition of the rock. As examples, the dominant rock type of the Church Rock member is an illitic arkosic siltstone; the dominant rock types of the Petrified Forest member are illitic-montmorillonitic claystone and montmorillonitic volcanic sandstone; and the dominant rock types of the Shinarump member are kaolinitic quartz sandstone and quartz sandstone. These names, it is hoped, give an immediate impression of the type of rock without requiring knowledge of the quantitative basis of the classification.

The classification is based on separation of the detrital components of the rock into four categories, which are (1) quartz, chert, and quartzite, (2) feldspar, (3) unstable fine-grained rock fragments (lithic and volcanic), and (4) clay minerals. These are the same basic categories used by Pettijohn (1957) and Gilbert (1955). The rock types can be shown on a compositional tetrahedron expressed in terms of the four detrital components (fig. 3). Four major divisions are recognized on the basis of the amount of argillaceous material: (1) less than 10 percent clay minerals, (2) 10 to 50 percent clay minerals, (3) 50 to 75 percent clay minerals, and (4) more than 75 percent clay minerals. The various names used in each of these four divisions are shown on figure 4. The name of the dominant type of clay in the rock is used to indicate the amount of clay in the rock. Thus in a rock containing less than 10 percent clay, no clay name is added as a modifier to the textural name (e.g., quartz sandstone); in a rock containing 10 to 50 percent clay, the first modifier of the textural name is a clay name (e.g., illitic arkosic siltstone); in a rock containing 50 to 75 percent clay, the second modifier of the textural name is a clay name (e.g., arkosic illitic claystone); and in a rock containing 75 to 100 percent clay, the only modifier is a clay name (e.g., montmorillonitic claystone).

The advantages of the classification are that the mean grain

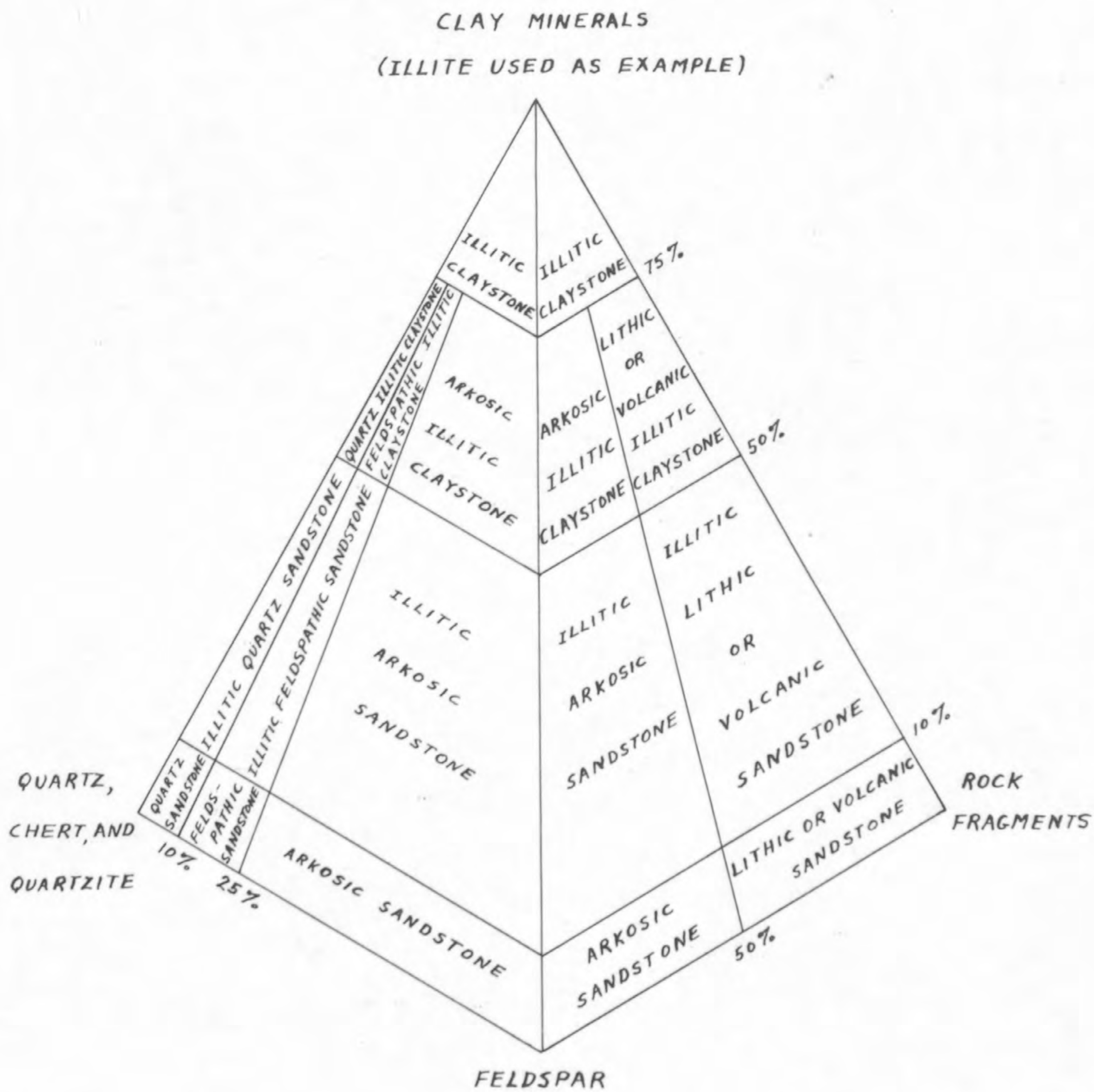
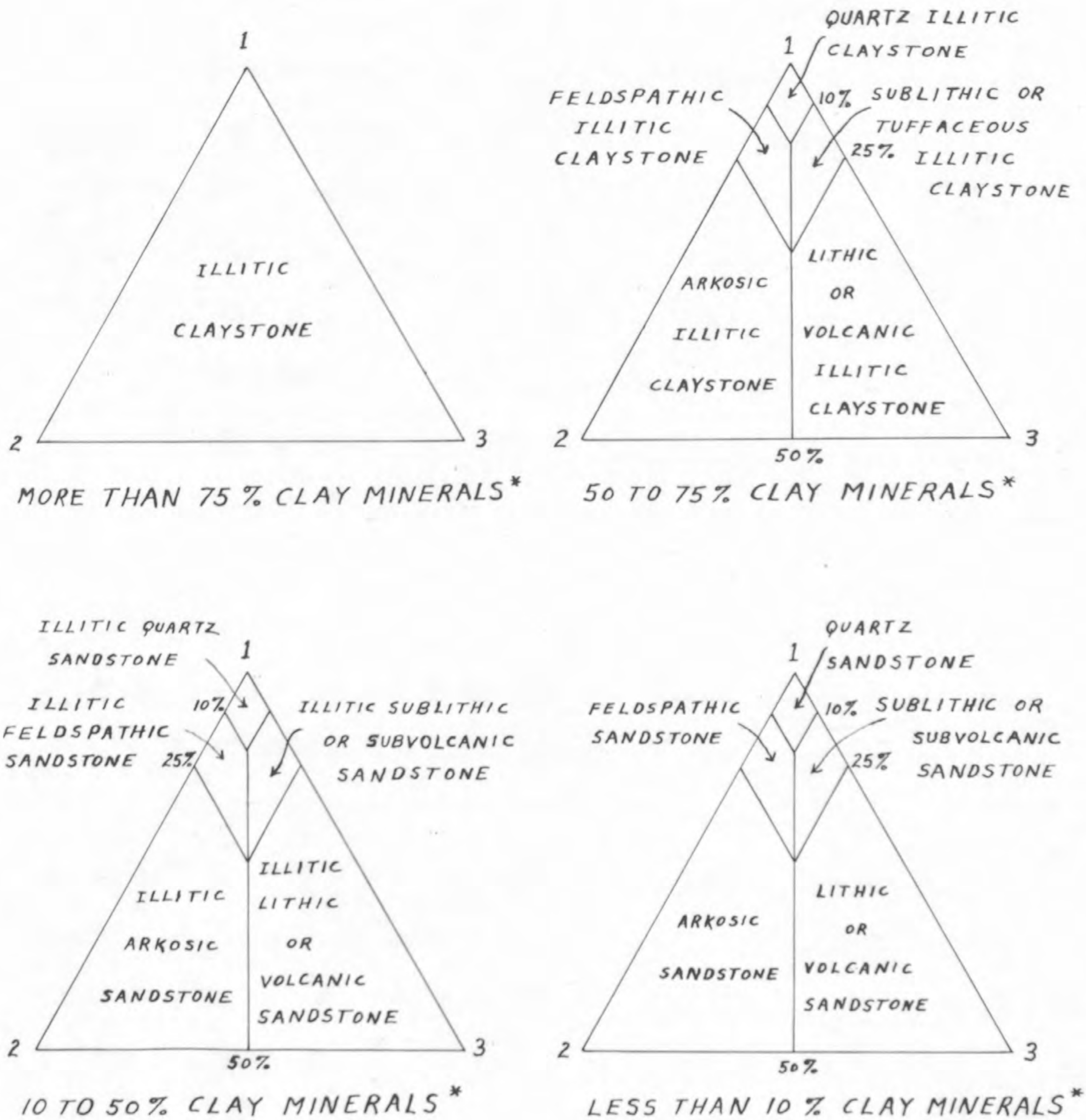


FIGURE 3.-- COMPOSITIONAL TETRAHEDRON SHOWING ROCK TYPES



COMPONENTS:

1. QUARTZ, CHERT, AND QUARTZITE
2. FELDSPAR
3. ROCK FRAGMENTS (LITHIC AND VOLCANIC)

* ILLITE USED AS EXAMPLE

FIGURE 4.-- CLASSIFICATION OF DETRITAL SEDIMENTARY ROCKS

size and dominant type of clay of the rock are shown. Few, if any, of the advantages of the other classifications are lost. In the rock of the Chinle formation, claystone and siltstone are the dominant rock types, and some compositional classification is needed for these rocks. Other compositional classifications make little or no distinction between rocks of different grain size. Also the different types and origins of the clays in the members of the Chinle formation point out the need for including the name of the clay in the rock name. Kaolinite is the dominant clay in the Shinarump and Moss Back members; montmorillonite and illite-montmorillonite in the Monitor Butte and Petrified Forest members; and illite and montmorillonite-illite in the Owl Rock and Church Rock members. The clays in each case indicate a different origin. The kaolinite probably formed by intense subareal weathering of feldspar, clay minerals, or other minerals. The montmorillonite indicates weathering of volcanic debris or devitrification of glassy volcanic material. The illite probably represents subaqueous diagenetic alteration of other clays in the basin of deposition. Because of the diverse origins of these clays, some method is needed to indicate the type of clay in the rock, and the proposed classification does this.

One difficulty in the use of the proposed classification is that the type of clay in the rock is often difficult to determine. Much of the clay identification in the Chinle formation is based on X-ray diffraction work (Schultz, in press; Cadigan, 1959a, p. 554-556), and thus a classification based on clay types is practical.

If the classification is used elsewhere and the clay cannot be identified, the word "argillaceous" or "clayey" could be substituted for the clay name where the clay name modifies the terms conglomerate, sandstone, and siltstone, and the clay name could be dropped entirely where it modifies the term claystone. Thus such terms as clayey arkosic sandstone, arkosic claystone, and claystone could be used.

The classification given here is designed to describe only the detrital components of the rock. The cement, however, can be indicated by use of modifiers such as calcareous or siliceous where these components constitute from 25 to 50 percent of the rock. Thus a quartz-rich rock, for example, with 30 percent carbonate cement would be called a calcareous quartz sandstone.

The classification makes a distinction between detrital volcanic material and pyroclastic volcanic material. Detrital volcanic material includes all volcanic debris that is abraded and transported into the basin of deposition by sedimentary processes. Pyroclastic volcanic material includes material that is unabraded and introduced into the basin of deposition aerially from a volcanic vent. The detrital volcanic material is an integral part of the classification and is indicated by the modifiers "volcanic" and "subvolcanic." Pyroclastic volcanic material is not included in the classification. A rock, however, that contains 25 to 50 percent pyroclastic material can be indicated by the modifier tuffaceous. Thus a quartz-rich rock with 30 percent pyroclastic material is called a tuffaceous quartz sandstone. Rock with more than 50 percent pyroclastic material is called tuff. Such a distinction between detrital volcanic debris and pyroclastic material is generally difficult to make. The distinction is necessary, however, to indicate the different means of transportation of the material in the rock, and is in conformity with the definition of tuffaceous and tuff given by Wentworth and Williams (1932) and discussed by Hay (1952).

Table 1.--Dominant Rock Types of the Chinle Formation Using Proposed Classification*

(Rock names in parentheses indicate subordinate rock type)

Member	Dominant Rock Type
Church Rock	Illitic arkosic siltstone Montmorillonitic-illitic arkosic siltstone (Illitic arkosic sandstone) (Illitic feldspathic sandstone)
Owl Rock	Illitic arkosic siltstone Montmorillonitic-illitic arkosic siltstone (Calcareous arkosic siltstone) (Limestone)
Petrified Forest	Illitic-montmorillonitic claystone Montmorillonitic volcanic sandstone Montmorillonitic subvolcanic sandstone
Moss Back	Quartz sandstone Kaolinitic quartz sandstone Kaolinitic feldspathic sandstone
Monitor Butte	Illitic-montmorillonitic volcanic sandstone Illitic-montmorillonitic feldspathic sandstone Illitic-montmorillonitic claystone (Kaolinitic feldspathic sandstone) (Feldspathic sandstone) (Quartz sandstone) (Calcareous feldspathic sandstone)
Shinarump	Kaolinitic quartz sandstone Quartz sandstone

*Data mostly from Cadigan, 1959a, and L. G. Schultz, in press.

DESCRIPTION AND CLASSIFICATION OF SEDIMENTARY STRUCTURES

The Chinle formation contains a wide variety of sedimentary structures including current lineation, ripple marks, uniform bedding, cross-stratification, contorted structures, channels, and swales. These structures are mainly the product of aqueous currents. In addition, the upper part of the Chinle formation and the overlying Wingate sandstone contain eolian dune deposits in which large scale cross-strata occur. The types and varieties of sedimentary structures occurring in the Chinle formation, and some closely related types in the Wingate sandstone, are briefly described below starting with the smaller features such as current lineation and ending in large scale features such as channels and swales.

Current lineations

Current lineations (Stokes, 1947) are linear streaks of sand grains occurring along flat and smooth bedding plains (fig. 5). They form parallel to the direction of current flow and represent a stream-lining of the flat sand bottom in response to the passing current. These structures can commonly be observed in modern streams, and are fairly common in the fluvial deposits of the Shinarump and Moss Back members of the Chinle formation.

Figure 5.--Current lineation in the Moss Back member of Chinle formation in the White Canyon area, Utah

Aqueous ripple marks

Two basic types of aqueous ripple marks occur in the Chinle formation--parallel and cusp. A structure that is interpreted to be formed by the deposition of layer upon layer of cusp ripple marks is abundant in the Chinle formation, and is referred to as rib-and-furrow structure. Rib-and-furrow structure, if it is truly composed of ripple marks, is the most common type of ripple marked structure in the Chinle formation. The ripple marks, including rib-and-furrow structures, are most common at the tops of fluvial sandstone units such as the Shinarump and Moss Back members, and in sandstone units in the Monitor Butte member, although ripple marks may occur anywhere in the formation. The ripple marks usually occur in very fine-grained sandstone or in coarse siltstone.

Parallel ripple marks consist of linear ridges and troughs (fig. 6, 7, and 8). In cross-section, the ridges may be symmetrical (symmetrical or oscillation ripple marks) or asymmetrical (asymmetrical or current ripple marks). The short steep slope is on the down current side in asymmetrical ripple marks. In the Chinle formation, both symmetrical (fig. 7) and asymmetrical (fig. 8) ripple marks occur. The wave lengths of the ripples in the Chinle formation are generally about 1 inch, although they may be as long as 3 inches. The ripple index (ratio of the amplitude of the wave from trough to crest to the wave length) is generally between 1:9 and 1:12, which is close to the ripple index of 1:4 to 1:10 listed by Kindle and Bucher (1926) as being characteristic of water-formed ripples.

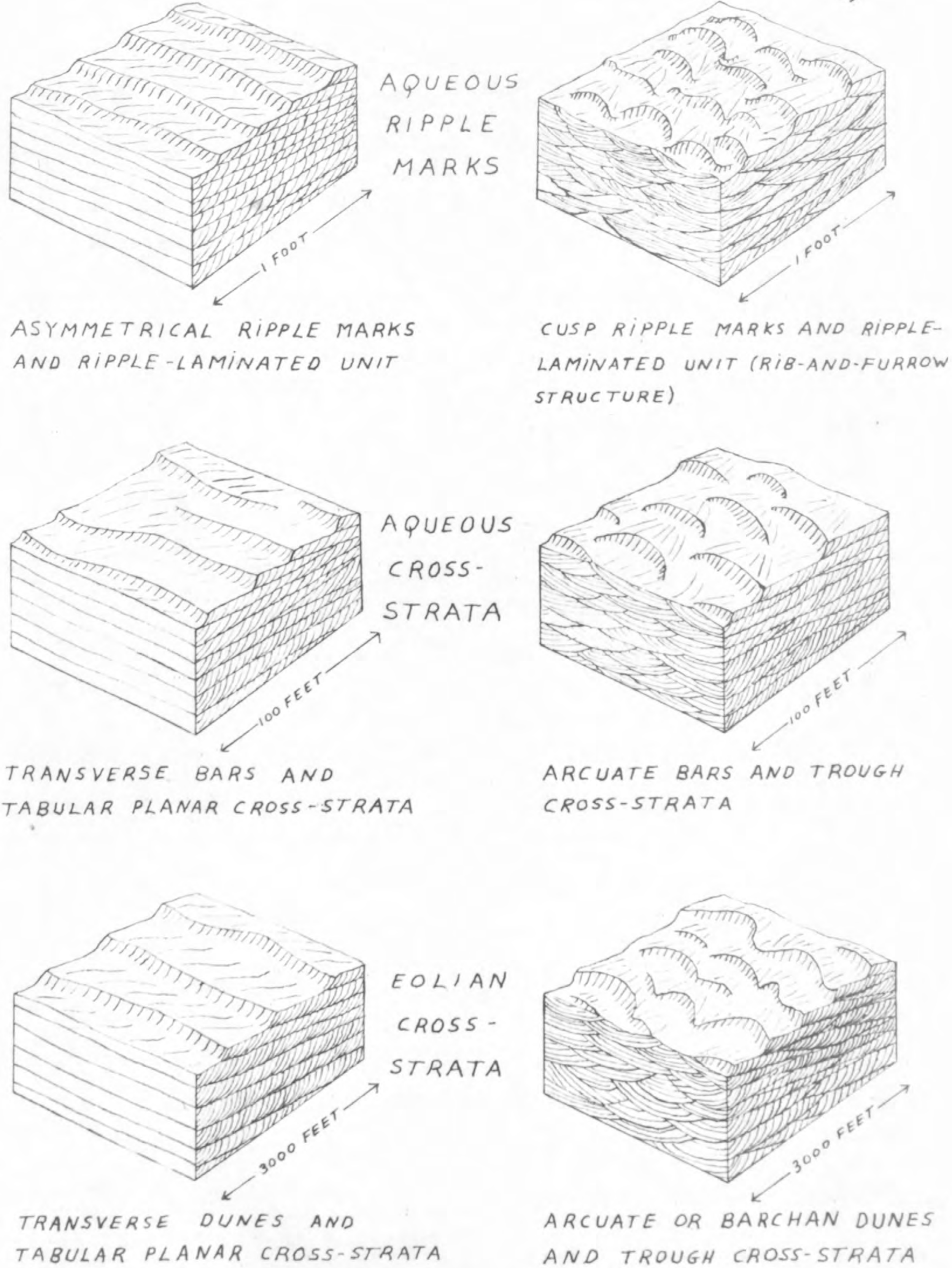


FIGURE 6.--DIAGRAM OF TYPES OF RIPPLE MARKS AND AQUEOUS AND EOLIAN CROSS-STRATA

Figure 7.--Symmetrical parallel ripple marks (fluvial) at top of Shinarump member of Chinle form... area, Arizona

Figure 8.--Asymmetrical parallel ripple marks (fluvial) at top of Shinarump member of Chinle formation in Lees Ferry area, Arizona

Layer upon layer of parallel ripple marks can be deposited, forming a thick unit of superimposed ripple laminae (fig. 6). Such units, composed of parallel ripple laminae, are probably rare in the Chinle formation, although they are common in some recent stream deposits such as those on the delta of the Colorado River described by McKee (1939).

Cusp ripple marks consist of arcuate ridges with the concave side downstream (fig. 6 and 9). They commonly appear similar, except in size, to barchan dunes. The arcuate ridges are generally 3 to 4 inches across and are distributed on a bedding plane in an irregular pattern, or with a slight tendency for the cusp marks to be alined downstream. The forms and patterns made by cusp ripple marks are quite varied, however, and many variations from ideal cusp marks to irregular marks with asymmetrical forms and patterns can be seen in the field.

A structure thought to be composed of layer upon layer of cusp ripple marks is particularly abundant in the Chinle formation. This structure has been called "rib-and-furrow" by Stokes (1953) who first described the structure, and this term is used in this report. Rib-and-furrow structures consist of parallel ridges separated by shallow troughs. The troughs are generally 3 or 4 inches wide and are filled with layer upon layer of miniature arcuate cross-laminae. As determined from other sedimentary structures, the troughs trend downstream. The troughs and ridges are discontinuous in a downstream direction. They generally extend a half foot to a foot, although some may be traced for several feet. When viewed from above (fig. 10), the structures

Figure 9.--Cusp ripple marks (fluvial) at top of Shinarump member of Chinle formation in Lees Ferry area

Figure 10.--Plan view of rib-and-furrow structure at top of Shinarump member of Chinle formation in Lees Ferry area, Arizona

Figure 11.--Side view of rib-and-furrow structure from lower red member of Chinle formation in east-central Arizona (rock about 8 inches across)

appear as indistinct discontinuous troughs filled with arcuate cross-laminae that trend across the troughs and abut against the ridges at an acute angle. In plan view, the cross-laminae are concave down-stream. When viewed from the side (fig. 11), the structures appear as miniature scour and fill cross-strata, with cross-laminae dipping gently downstream and filling shallow troughs.

Eolian ripple marks

Eolian ripple marks in ancient rocks are a rare occurrence. An unusual type has been described by McKee (1945) from the Coconino sandstone of Arizona; in addition, Poole (1957) and the writer have noted eolian ripple marks in the Triassic and Jurassic rocks of the Colorado Plateau.

The eolian ripple marks noted by the writer (fig. 12 and 13) are in very fine- to fine-grained sandstone of the Wingate sandstone at Ft. Wingate, New Mexico. They occur on the gently dipping parts of cross-strata that are interpreted to be the foreset strata of a dune. The ripples have essentially parallel crests. The crests are about 6 inches apart, and the amplitude of the waves is about 0.13 inch. The ripple index is 1:45 which is within the range of ripple indices of 1:20 to 1:50 (or more) given by Kindle and Bucher (1926) as characteristic of wind-formed ripples and is much different from the ripple indices for aqueous ripple marks.

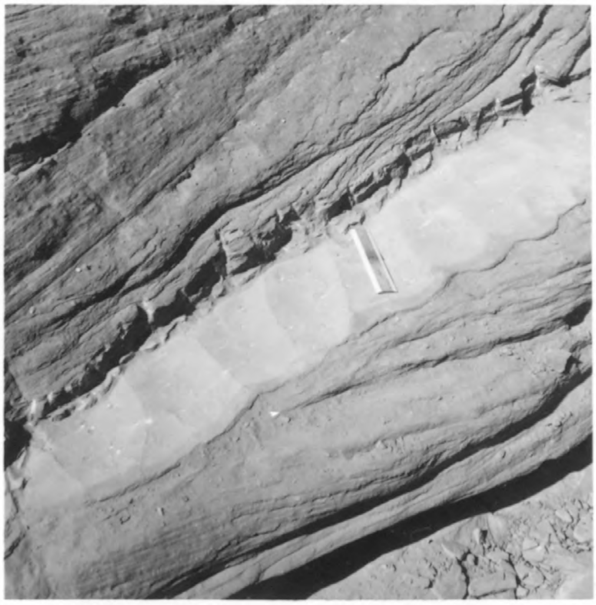


Figure 12.--Eolian ripple marks in Wingate sandstone near Ft. Wingate, New Mexico (note 6 inch scale)

Figure 13.--Trough sets of eolian cross-strata in Wingate sandstone near Ft. Wingate, New Mexico, showing location of eolian ripple marks shown in figure 12. Circle marks location of hammer resting on ripple-marked surface.

Aqueous cross-strata

Cross-strata are the dominant sedimentary structure in fluvial units such as the Shinarump and Moss Back members and the Sonsela sandstone bed of the Chinle formation. In addition, much of the Monitor Butte and Petrified Forest members and in some areas much of the Church Rock member contain cross-strata. An understanding of the types and origin of cross-strata is essential, therefore, in interpretation of the environment of deposition of the Chinle formation. A brief description is given here of the types of cross-strata; the origin of cross-strata is discussed later (see interpretations).

The cross-strata in the Chinle formation can be divided into two main groups on the basis of the shape of the sets in which the cross-strata occur. A set is defined by McKee and Weir (1953) as "a group of essentially conformable strata or cross-strata, separated from other sedimentary units by surfaces of erosion, non-deposition, or abrupt changes in character." One type of cross-strata in the Chinle formation occurs in tabular planar sets in which the upper and lower boundaries of the sets are parallel, or essentially parallel, flat surfaces of erosion. The other type of cross-strata occurs in trough sets in which the lower boundary, and in most cases the upper boundary as well, are curved surfaces of erosion. Of the two types of cross-strata, the tabular planar type is by far the most abundant in the Shinarump and Moss Back members and the Sonsela sandstone bed. Trough sets, however, may be the most abundant type in the Monitor Butte and Petrified Forest members, although cross-strata are difficult to see in detail in these units.

The tabular planar type (fig. 6, 14 and 15) of cross-strata occur in sets which generally range in thickness from one-half foot to two feet. Individual sets can be traced laterally along exposures for at least 200 feet. In plan view, the cross-strata appear as laminae dipping and striking uniformly. The cross-laminae do not curve in plan view as is characteristic of the cross-laminae in trough sets. In cross-section, the cross-laminae are concave upward and become tangential downwards with the lower bounding surface of the set. The maximum dip of the cross-strata is generally about 20 to 25 degrees.

The trough type of cross-strata (fig. 6 and 16-22) occurs in a variety of shapes and sizes, and the varieties are difficult to classify. The trough type of cross-strata in the Shinarump and Moss Back members occurs in sets that generally range in thickness from one-half foot to two feet. In plan view, the sets are narrow elongated features commonly 5 to 20 feet long and 2 to 5 feet wide, with blunt rounded terminations upstream (fig. 16, 17, and 18). Downstream the sets are cut off by the development of other sets. The cross-strata, in plan view, are curved and are convex upstream (fig. 16). In a cross-section cut along the length of the trough, the sets either are lens shaped (fig. 19) or are tabular layers that resemble tabular planar sets in cross section (fig. 6). In a cross section cut across the trough, the lower boundary of the set is U-shaped (fig. 20) and the upper boundary is a surface of erosion marked by the U-shaped boundaries of overlying sets. The dips of the individual cross-strata are

Figure 14.--Tabular planar cross-strata (fluvial) in Shinarump member of Chinle formation near St. George, Utah

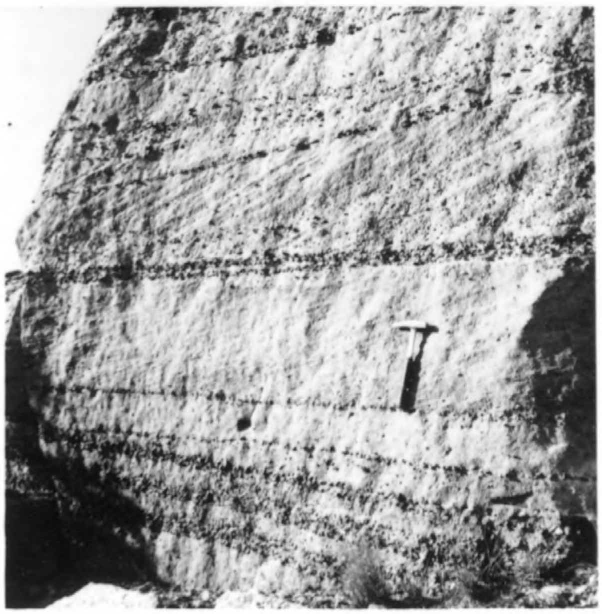


Figure 15.--Tabular planar cross-strata (fluvial) in Shinarump member of Chinle formation near Kanab, Utah

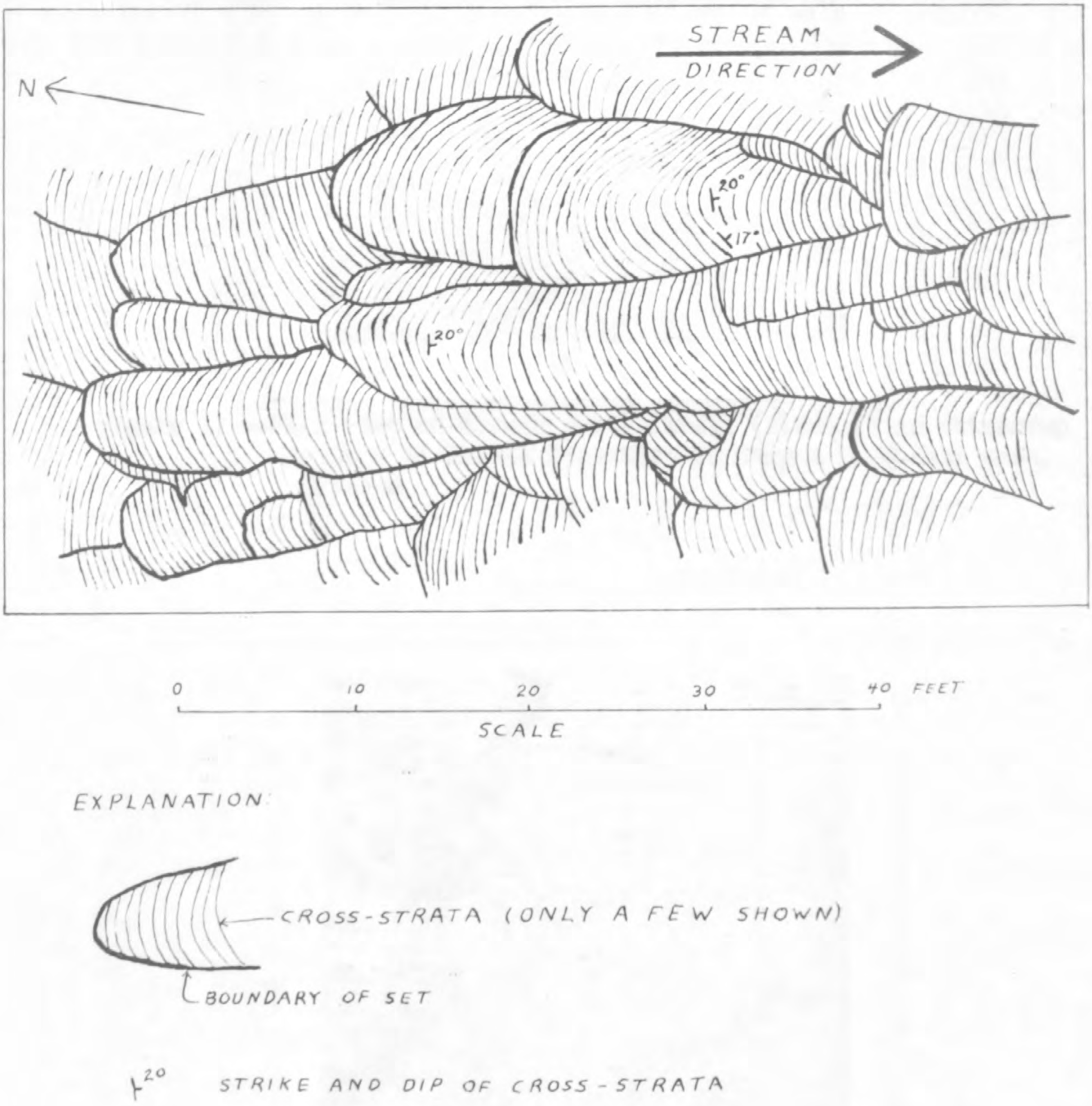


FIGURE 16.-- SKETCH OF PLAN VIEW OF A GROUP OF TROUGH CROSS-STRATA IN SHINARUMP MEMBER OF CHINLE FORMATION AT CANYON DE CHELLY, ARIZONA

Figure 17.—Plan view of trough cross-strata (fluvial) in Shinarump member of Chinle formation in Canyon De Chelly area, Arizona

Figure 18.—Plan view of trough cross-strata (fluvial) in Shinarump member of Chinle formation in Canyon De Chelly area, Arizona

Figure 19.--Side view of trough cross-strata (fluvial) in Shinarump member of Chinle formation near Cameron, Arizona

Figure 20.--End view (looking downstream) of large trough set of cross-strata in Shinarump member of Chinle formation near Cameron, Arizona

Figure 21.--Shallow trough sets of low angle cross-strata (fluvial) in Petrified Forest member of Chinle formation near Cameron, Arizona

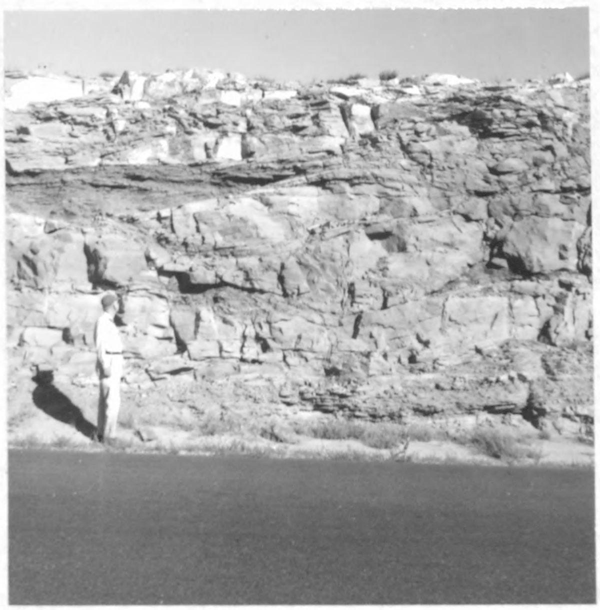


Figure 22.--Side view of deep trough sets of cross-strata (fluvial) in sandstone bed in Petrified Forest member of Chinle formation in east-central Arizona

generally about 20 to 25 degrees.

The cross-strata in the clayey sandstone of the Petrified Forest member occur generally in trough sets (fig. 21), but the scour at the base of the sets is shallower than in the sets of the Shinarump and Moss Back members. In addition, the dip of the cross-strata is rarely more than 5 or 10 degrees. As will be discussed later, these cross-strata in the Petrified Forest member probably formed under different conditions than those of the Shinarump and Moss Back members.

Other varieties of trough cross-strata occur. For example, in some sandstone units cross-strata occupy fairly deep, well-defined troughs (fig. 22), and the cross-strata appear to have filled in "holes" that developed on the bottoms of streams. In addition, extremely large and irregular shallow troughs characterize the cross-strata occurring in sediments that fill broad channels (ancient river courses). Trough cross-strata, therefore, include a variety of types that probably formed under different physical conditions. Their classification is unsatisfactory and generalizations about them are difficult to make.

Eolian cross-strata

Cross-strata are the dominant sedimentary structure in the eolian Wingate sandstone, that overlies the Chinle formation, and in lenses or tongues of sandstone, similar to the Wingate sandstone, that occur in the upper part of the Chinle formation in northeastern Arizona and northwestern New Mexico. These cross-strata, as will be discussed later, were formed in sand dunes. These eolian cross-strata resemble those formed by aqueous processes, and the two types cannot always be told apart. The eolian cross-strata, however, are in most cases on a much larger scale; they are commonly 10 or more feet long, whereas aqueous cross-strata are generally less than 10 feet long. In addition, the dip of the eolian cross-strata is commonly somewhat higher than that of aqueous cross-strata; eolian cross-strata generally dip about 30 or more degrees whereas aqueous cross-strata dip about 20 to 25 degrees.

Two main types of eolian cross-strata occur--tabular planar and trough cross-strata. The tabular planar eolian cross-strata (fig. 6 and 23) are similar to those of aqueous origin except that they are generally in thicker sets, commonly 5 or 10 feet thick; some sets are 40 or more feet thick. Most of the trough cross-strata of eolian origin (fig. 6 and 24) are different from those of aqueous origin. The bounding surfaces of the set are surfaces of erosion and dip generally about 10 degrees, but the curvature of the surface is very slight in contrast to the aqueous cross-strata in which the curvature of the bounding surfaces is marked. Some of these cross-strata approach the shape of planar cross-strata in which the planar surfaces between sets dip about 10 degrees. In plan view, all of the eolian trough cross-strata curve gently but perceptibly.

Contorted strata

Contorted strata, both on a small and a large scale, occur in the Chinle formation. A reversal of dip or slumping in the upper part of tabular planar cross-strata is the most common type of small scale contortion (fig. 25 and 26). In cross section, the cross-strata and

Figure 23.—Tabular planar cross-strata (eolian) in Wingate sandstone near Gateway, Colorado. Set is about 15 feet thick.

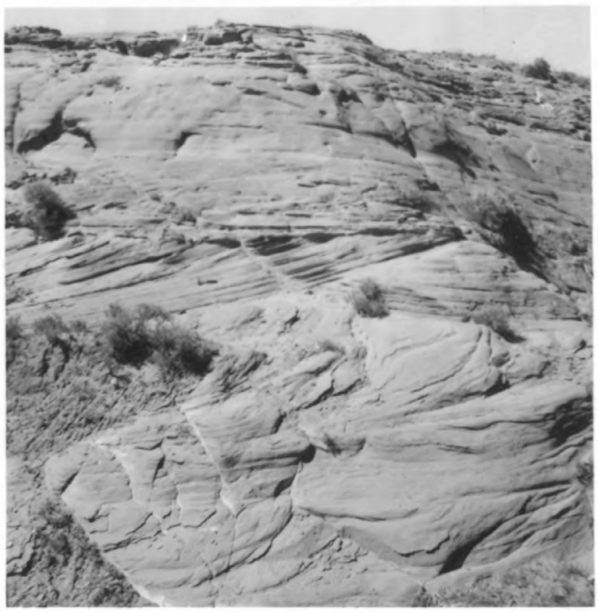


Figure 24.—Trough cross-strata (eolian) in Wingate sandstone near Ft. Wingate, New Mexico

Figure 25.--Contorted cross-strata in Moss Back member of Chinle formation in White Canyon area, Utah

Figure 26.--Contorted cross-strata and uncontorted tabular planar cross-strata in Moss Back member of Chinle formation in White Canyon area, Utah

the contorted upper part of the cross-strata appear like a recumbent fold; individual laminae have the shape of a U laid on its side (fig. 25). In some sets, the U-shaped pattern is not well developed, and the upper part of the set is contorted in an irregular pattern (fig. 26). The contorted cross-strata are common in fluvial sandstone units such as the Shinarump and Moss Back members.

Contorted strata involving entire beds or groups of beds are common in the Monitor Butte and related members of the Chinle formation and also occur rarely in other parts of the formation. Most commonly the contorted strata consist of well-cemented, ripple-marked sandstone beds interbedded with claystone, siltstone, and clayey sandstone. The sandstone is intricately folded (fig. 27 and 28) or occurs in irregular blocks lying with almost any conceivable strike and dip. Such contorted strata may be small features covering only a hundred square feet and involving 5 feet of strata or they may cover many acres and involve 50 feet or more of strata. In some places, the contorted strata are truncated and overlain horizontally by undisturbed strata, indicating that the contortion was caused by slumping shortly after deposition of the strata.

Channels and swales

Channels are large U-shaped scours or valleys cut into a stratum and filled with sediment of the overlying unit (see illustrations in Finch, 1959, Pl. 7 and 9). Most commonly the channels are a few hundred feet wide and 50 to 100 feet deep. Some channels can be traced over a sinuous course for many miles, whereas others can be

Figure 27.--Contorted strata in Monitor Butte member of Chinle formation in Capitol Reef area, Utah

Figure 28.--Contorted strata in Monitor Butte member of Chinle formation in Capitol Reef area, Utah. Note flat-lying strata truncating folded strata.

traced only a part of a mile. Most of the channels are cut into the Moenkopi formation and filled with sediment of the Shinarump member, but channels also occur at the base of the Moss Back member and rarely elsewhere in the formation. As will be discussed later, some of the channels are considered to be the scour channels of rivers; others probably are filled river valleys formed during an older period of erosion.

Swales (Witkind, 1956) or valleys are broader relief features. They commonly range in width from 1 to 3 miles and have a relief of about 40 feet (Witkind, 1956). These swales are considered to be broad river valleys which were later filled with sediment. Channels commonly occur in the deeper parts of swales, and channels are probably more common in swales than elsewhere.

STRATIGRAPHY

The Chinle formation is composed of a variety of lithologic types including thin widespread light-colored sandstone and conglomerate layers, variegated bentonitic claystone and clayey sandstone, pale-red or greenish-gray limestone, and red nonbentonitic siltstone and sandstone. The formation extends over most of the Colorado Plateau and beyond the Plateau in several directions. On the north, the name Chinle formation is generally used in Colorado and Utah, but rocks of equivalent age and similar lithologic type in Wyoming are designated by the name Jehm formation or the Popo Agie member of the Chugwater formation. On the east, in north-central and central Colorado the Chinle formation pinches out on the flanks of the ancestral Uncompahgre and Front Range highlands. On the southeast, rocks of Late Triassic age extend into southeastern Colorado, eastern New Mexico, the Panhandle of Oklahoma, and western Texas (McKee and others, 1959). In these areas, the term Dockum group, consisting of the Santa Rosa sandstone and the Chinle formation, is mostly used. On the south, the Chinle formation probably originally reached a limit along the north flank of the Mogollon highland, an ancient highland which occupied southern Arizona and adjacent parts of California and New Mexico. The Chinle formation is well defined in the Spring Mountains near Las Vegas, Nevada, but correlation of the formation west of this area is uncertain. Upper Triassic rocks, however, occur extensively in western Nevada and at several localities in southeastern California

(Reeside and others, 1957). The Chinle formation represents only a part of the extensive Upper Triassic deposits of the western United States (McKee and others, 1959).

On the Colorado Plateau, the Chinle formation ranges in thickness from 0 to slightly over 1700 feet. The formation is thickest in northeastern Arizona and northwestern New Mexico.

The Chinle formation unconformably overlies the Moenkopi formation of Early and Middle (?) Triassic age on most of the Colorado Plateau. In the eastern part of the Colorado Plateau, the Moenkopi formation is absent, and the Chinle formation rests unconformably on rocks of Permian age. On the flanks of the Uncompahgre and Front Range highlands in Colorado and New Mexico, the Chinle formation locally rests unconformably on Paleozoic rocks older than Permian, and in areas where the Chinle formation covers part of these highlands, it rests on igneous and metamorphic rocks of Precambrian age.

The lower boundary of the formation is everywhere an unconformity. The unconformity is a remarkably flat plane surface, only interrupted in some areas by swales, channels, or scours cut into the underlying rocks and filled with the strata of the Chinle formation. In only a few areas, mostly along the flanks of the Uncompahgre and Front Range highlands, can an angularity be noted between the Chinle formation and the underlying sedimentary rocks. The most notable angularity is in the Ouray area in southwestern Colorado where the Dolores formation (a lateral equivalent of the Chinle formation) rests with an angular discordance, which is locally as great as 6 or 7 degrees, on formations of Paleozoic age (Cross and Howe, 1905b). The Ouray area is on the west flank of the ancestral Uncompahgre highland.

The Chinle formation is disconformably, or in some areas conformably, overlain by formations of the Glen Canyon group in most of the Colorado Plateau region. In the central part of the Colorado Plateau, the Wingate sandstone (Upper Triassic) of the Glen Canyon group overlies the Chinle formation. The contact of the Chinle formation and Wingate sandstone is a flat smooth plane and is considered to be a disconformity in most places. In northeastern Arizona and possibly in other areas also, the Chinle formation and the Wingate sandstone interfinger and intergrade. In northwestern Arizona, southwestern Utah, and southern Nevada, the Moenave formation (Triassic ?) or the Moenave and Kayenta formations undifferentiated (Triassic ? and Jurassic ?) disconformably overly the Chinle formation. In northeastern Utah and northwesternmost Colorado, the Navajo (Nugget) sandstone (Jurassic ? and Jurassic) overlies the Chinle formation. Along the eastern margin of the Colorado Plateau, the Entrada sandstone (Upper Jurassic) truncates older formations eastward and unconformably rests on the Chinle formation in north-central New Mexico and in south-central, central, and parts of north-central Colorado. Along the southern margin of the Colorado Plateau, the Dakota sandstone (Lower and Upper Cretaceous) truncates older formations southward and on outcrops in east-central Arizona and in west-central New Mexico unconformably overlies the Chinle formation. In east-central Arizona, the Dakota sandstone truncates the entire Chinle formation and rests directly on the Moenkopi formation that underlies the Chinle formation elsewhere. Such a southward beveling and total truncation of the Chinle formation by erosion prior to the deposition of the Dakota sandstone can only be demonstrated in east-central Arizona, but probably occurred elsewhere or everywhere along the southern margin of the Colorado Plateau.

The Chinle formation is divided into a lower and an upper part. The lower part of the formation consists of variegated bentonitic claystone and clayey sandstone of the Monitor Butte, Petrified Forest, and related members and of thin widespread ledge-forming sandstone and conglomerate units such as the Shinarump and Moss Back members. The upper part of the formation consists of reddish-brown horizontally bedded siltstone and minor amounts of limestone, ripple-laminated siltstone and sandstone, limestone pebble conglomerate, and cross-stratified sandstone. The upper part of the Chinle formation consists of the Owl Rock member below and the Church Rock member above, and locally of strata equivalent to these members but called by different names. The contact between these two parts of the Chinle formation is gradational and intertonguing, and although the lithologic differences between the two parts of the formation are marked, the exact boundary is difficult to locate precisely. In some areas, the contact is gradational over 100 to 200 feet. In spite of the difficulties in locating the exact boundary, these two parts of the formation differ from one another not only in lithologic type but also in environment of deposition and source of constituting material. The stratigraphy and origin of the Chinle formation is discussed, therefore, in terms of these two parts.

The separation of the Chinle formation into the lower and upper parts is useful in all of the Colorado Plateau, except in northeastern Utah and northwestern Colorado. In these areas, red beds (Church Rock member) belonging to the upper part of the formation are recognized, but the other strata of the formation, including the Gartra, mottled, ocher mudstone, and upper members, cannot with certainty be assigned to either the upper or lower part of the Chinle formation. These latter members are not described in this report, but are included, however, in some of the diagrams. A possible basal member of the Chinle formation in southern Nevada probably belongs with the lower part of the Chinle formation, but is not discussed.

In addition to the major parts of the Chinle formation described above, peculiar mottled strata occur near the basal contact of the formation. Because they are unique, these mottled strata are described separately.

Mottled strata

Mottled strata are thin units or lenses characterized by a peculiar mottling of reddish purple, pale reddish brown, and light greenish gray and occur directly below, near, or at the base of the Chinle formation. The mottled coloration is believed to have formed by some process of alteration of preexisting rocks, probably during the formation of a soil. Locally the mottled strata are continuous and well-defined, and are given member names (Temple Mountain member of the Chinle formation in the San Rafael Swell in central Utah, and the mottled member of the Chinle formation in northwestern Colorado and northeastern Utah, fig. 29). Elsewhere, these strata are given the informal name "mottled strata."

The mottled strata have a widespread but spotty distribution on the Colorado Plateau. They probably occur on less than 10 percent of the outcrops of the Chinle formation, but have been found in almost every area of the Plateau and adjacent regions. The most characteristic occurrence is in lenses 5 to 10 feet thick that extend along the outcrop for 100 feet to several thousand feet. In some areas, the mottled strata are continuous for many miles. The Temple Mountain member of the Chinle formation (the name applied to the mottled strata in the San Rafael Swell) extends for many miles without interruption and occurs on 85 percent of the outcrops of the Chinle formation in the San Rafael Swell (Robeck, 1956).

The mottled strata occur at the base or in the basal part of the Chinle formation, or at the top of the formation that directly underlies the Chinle formation. The rocks directly below the Chinle formation on which the mottled strata are developed include the Moenkopi formation in much of the Colorado Plateau, the Cutler formation (Permian) and Precambrian igneous and metamorphic rocks in southwestern Colorado, the De Chelly sandstone (Permian) in east-central Arizona, and other Permian strata in north-central New Mexico. Locally the strata occur both in the basal few feet of the Chinle formation and in the top few feet of the underlying formation, and the basal contact of the Chinle formation lies within the mottled strata.

The mottled strata may be siltstone, sandy siltstone, sandstone, or conglomerate, or even granitic or metamorphic rock. The distinctive character of the rock is its mottled coloration which, in its typical form, is striking and unmistakable (fig. 30). Reddish purple, pale reddish brown, and light greenish gray are intricately mottled; irregular blotches generally one or two inches across on one color are intricately interwoven with blotches of the other colors. The red and purple parts are colored mainly by hematite (Schultz, in press; W. D. Keller, oral communication). In a few areas, irregular vertical gray bands, a few inches across, produce conspicuous vertical stripes.

The most abundant lithologic type of the mottled strata is siltstone, even though the mottled strata may occur in almost any rock type. The mottled siltstone in the Chinle formation commonly contains scattered fine to very coarse rounded grains of quartz. Such grains of quartz have not been noted in the mottled strata of the Moenkopi formation, and the presence or absence of these grains, therefore, is one basis for distinguishing the mottled strata of the Chinle formation from that of the Moenkopi formation.

The mottled strata in the basal part of the Chinle formation, in addition to siltstone, also contain local lenses of conglomerate and sandstone. The sandstone and the matrix of the conglomerate is medium to very coarse grained; the conglomerate contains granules, pebbles, and cobbles of quartz. Locally the sandstone and conglomerate

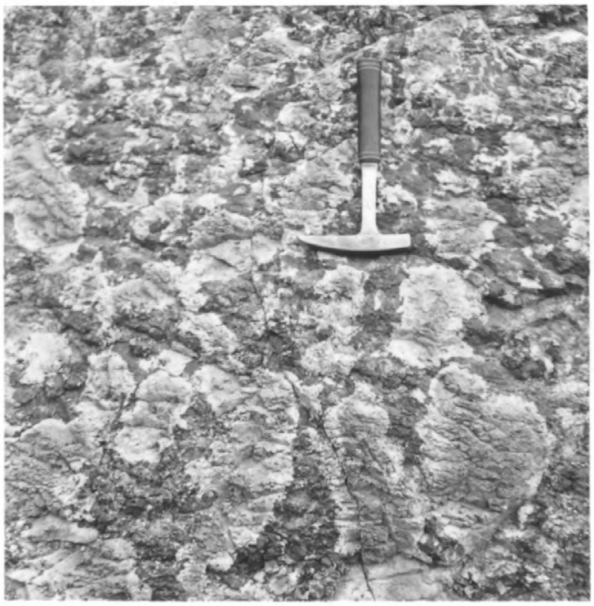


Figure 30.--Mottled strata about 7 miles up the Colorado River from Moab, Utah

are cross-stratified. In some areas, the mottled strata in the Chinle formation contain lenses of jasper and carbonaceous material (Hobeck, 1956).

The clay types in the mottled rocks are particularly important in interpreting the origin of the mottled strata. Schultz (in press) has studied these clays and his results are summarized below.

The clays in the mottled strata in the top few feet of the Moenkopi formation are generally (1) illite, (2) about an equal amount of mixed layer illite-montmorillonite in which the illite layers are only slightly more abundant than, or equally abundant to, the montmorillonite layers, and (3) an equal or slightly less amount of poorly crystallized kaolinite. The mottled strata differ from the underlying red rocks of the Moenkopi formation in that they commonly contain more kaolinite, more mixed layer illite-montmorillonite in which the illite layers are only slightly more abundant than, or equally abundant to, the montmorillonite layers, and less illite than the strata in the unaltered Moenkopi formation. In addition, the kaolinite is a more poorly crystallized type in the mottled strata, and the mottled strata rarely contain chlorite and never contain feldspar, although both are common in some parts of the Moenkopi formation.

The dominant clay in altered Precambrian igneous and metamorphic rocks in southwestern Colorado, lying below the Chinle formation, is mixed layer illite-montmorillonite in which the illite layers are only slightly more abundant than, or about equally abundant to, the montmorillonite layers. Illite, poorly crystallized kaolinite, and chlorite occur as minor clay minerals. The same clays are generally present regardless of the composition of the underlying unaltered crystalline rock which may be granite, diorite, amphibolite, or chlorite schist.

The clay mineralogy of the mottled strata in the Chinle formation shows more variety than in the mottled rocks in the Moenkopi formation or Precambrian. In many places, the mottled strata contain poorly crystallized or well crystallized kaolinite as the dominant clay. In other places, the dominant clay is mixed-layer illite-montmorillonite in which the illite layers are only slightly more abundant than, or about equally abundant to, the montmorillonite layers. In still other places, mixed-layer illite-montmorillonite, in which the montmorillonite layers are dominant, is the chief mineral. Chlorite occurs in minor amounts in some samples.

The mottled strata are most commonly 5 to 10 feet thick. The Temple Mountain member of the Chinle formation (mottled strata in the San Rafael Swell) is generally about 20 to 30 feet thick and attains a maximum thickness of 101 feet in a channel fill (Robeck, 1956). The mottled strata are unusually thick--at least 200 feet--at a locality along the Colorado River about 7 miles northeast of Moab, Utah.

The contacts of the mottled strata with the underlying and overlying strata are, in most places, poorly defined and transitional. The mottled coloration commonly gives way downward into the uniform color of the underlying rocks. The upper contact is commonly sharp, but it may also be gradational.

Lower part of the Chinle formation

The lower part of the Chinle formation consists of variegated bentonitic claystone and clayey sandstone and thin widespread sandstone and conglomerate layers. It extends throughout southernmost Nevada, northern Arizona, southeastern Utah, southwesternmost Colorado, and northwestern New Mexico (fig. 31). Probably much of the Dockum group of eastern New Mexico, the Panhandle of Oklahoma, and western Texas correlates with this part of the Chinle formation. The lower part of the Chinle formation is over 1000 feet thick along its southern margin and thins gradually to the northeast. It thins and grades out into the upper part of the Chinle formation along a northwest line extending through northernmost New Mexico, southwesternmost Colorado, and east-central Utah. For the purposes of description, the lower part of the Chinle formation is conveniently divided into four stratigraphic units, (1) Shinarump and related members, (2) Monitor Butte and related members, (3) Moss Back member and related units, and (4) Petrified Forest member.

Shinarump and related members

The Shinarump member is the basal member of the Chinle formation (fig. 29), except locally where it is underlain by mottled strata of the Chinle formation. The Aqua Zarca sandstone member and an unnamed sandstone member in north-central New Mexico and the Gartra member in northeastern Utah and northwestern Colorado are lithologically similar to and occupy the same stratigraphic position as the Shinarump member. These members, however, lie outside of the depositional area of the Shinarump member.

The Shinarump member occurs in a region of about 140,000 square miles in the southern part of the Colorado Plateau and westward into Nevada, although it is absent in several large areas and many small areas within this region (fig. 32). Figure 32 is an interpretation of the depositional pattern of the Shinarump and related members based on the distribution of the members along outcrops. Away from outcrops the distribution, as shown on the figure, is hypothetical.

The Shinarump member is composed typically of yellowish-gray and pale yellowish-orange, fine- to coarse-grained sandstone. Lenses of conglomeratic sandstone and conglomerate containing granules and pebbles predominantly of quartz, quartzite, and chert are common. Silicified and carbonized wood are also common in the member. The member is almost entirely cross-stratified. Tabular planar sets, generally 0.5 to 2 feet thick, are the most common type of cross-strata; trough sets of about the same thickness also occur, and in some areas is the dominant type of cross-strata. The Shinarump member is a resistant unit (fig. 33) that forms vertical cliffs and in some areas underlies broad benches.

The sandstone of the Shinarump member is composed of subround to subangular grains of quartz, a small amount of potassium feldspar (generally 5 percent or less), and an even smaller amount of altered volcanic material (Cadigan, 1959a, p. 543-544). These grains are set in a matrix of kaolinitic mud which averages about 14 percent of the rock (Cadigan, 1959a, p. 543-544). The sandstone is generally

EXPLANATION

INFERRED LIMITS OF MAJOR DEPOSITIONAL AREAS OF SHINARUMP, AGUA ZARCA SANDSTONE, SANDSTONE, AND GARTRA MEMBERS OF CHINLE FORMATION. LIMITS ARE BASED ON DISTRIBUTION ON OUTCROPS, WHERE POSSIBLE, BUT ARE HYPOTHETICAL BETWEEN OUTCROPS

FIGURE 32.-- AN INTERPRETATION OF THE DEPOSITIONAL PATTERN OF THE SHINARUMP, AGUA ZARCA SANDSTONE, SANDSTONE, AND GARTRA MEMBERS OF THE CHINLE FORMATION

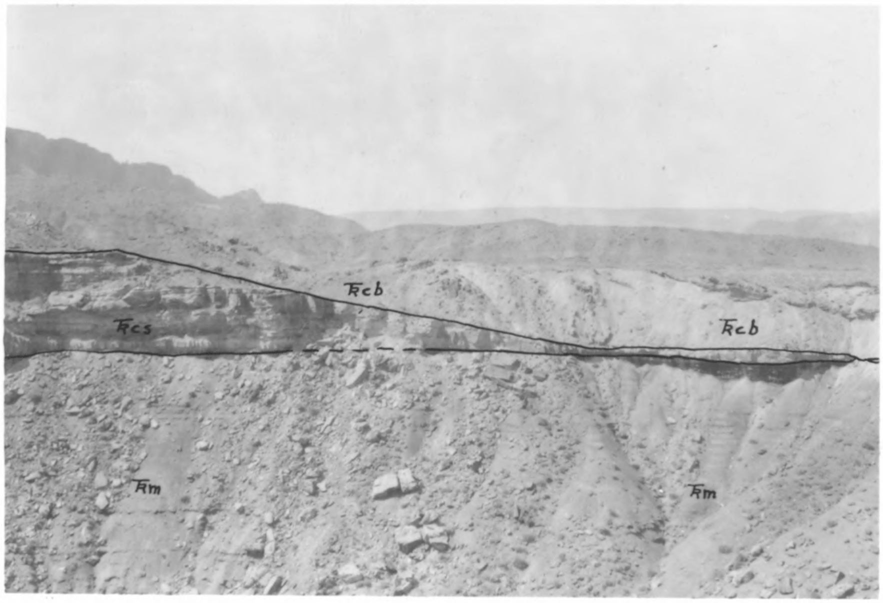


Figure 33.--Shinarump and Monitor Butte members of the Chinle formation on Monitor Butte in northern part of Monument Valley area, Utah. A channel is cut into the Shinarump member and filled with strata of the Monitor Butte member. The sediment-filled channel may be a "clay plug"--an abandoned channel or slough deposit. (Trm, Moenkopi formation; Trcs, Shinarump member of the Chinle formation; Trcb, Monitor Butte member of the Chinle formation)

very weakly cemented by isolated patches of carbonate minerals (dominantly calcite) and iron oxide cement. The rock can be classified in most areas as a kaolinitic quartz sandstone or as a quartz sandstone, although some of the rocks are feldspathic sandstone or kaolinitic feldspathic sandstone.

Quartz, quartzite, and chert are the dominant types of granules, pebbles, and cobbles occurring in the Shinarump member (William Thordarson and H. F. Albee, written communication [redacted]). The average amount of quartz varies from 8 percent in the St. Johns area to 83 percent in the White Canyon-Elk Ridge area; the average amount of quartzite varies from 15 percent in the White Canyon-Elk Ridge area to 49 percent in the Kanab area; and the average amount of chert varies from 2 percent in the White Canyon-Elk Ridge area to 49 percent in the southern Defiance Uplift area (William Thordarson and H. F. Albee, written communication [redacted]). The percentage of quartz pebbles is unusually high (83 percent) in the White Canyon-Elk Ridge area; elsewhere it is generally near or less than 50 percent.

The mean and maximum sizes of granules, pebbles, and cobbles decreases gradually northward in the member (Thordarson and Albee, written communication [redacted]). The mean sizes of gravel in the St. Johns, Holbrook, and southern Defiance Uplift areas in east-central Arizona is 21 mm and the maximum sizes range up to 284 mm. In the Circle Cliffs area, 200 miles to the north, the mean size is 15 mm and the maximum sizes range up to only 63 mm. The Shinarump member in the White Canyon-Elk Ridge area is unusual, however; here the mean sizes of pebbles decrease from 25 mm in the eastern part of the area to 12 mm in the western part of the area.

Many of the chert pebbles in the Shinarump member contain fossils, clearly indicating that they were derived from older sedimentary rocks, which were probably cherty limestone. The fossils include fusulinids, brachiopods, bryozoa, and, to a lesser extent, gastropods, pelecypods, corals, algae, crinoidal material, sponges, ostracods, and echinoid spines (William Thordarson and H. F. Albee, written communication [redacted]). As will be discussed in more detail later, most of these fossils indicate source rocks of Permian age.

A few pebbles of volcanic material occur in the Shinarump member in the Cameron area and in a possible correlative of the Shinarump member in the Cedar Ranch area, about 25 miles north of Flagstaff, Arizona. The largest volcanic pebble noted is in the Cedar Ranch area and is 205 mm (8.1 inches) in maximum diameter. Most of the volcanic pebbles, however, are 1 or 2 inches in maximum diameter. The volcanic pebbles consist of phenocrysts of quartz, orthoclase, sanidine, and rarely biotite set in an aphanitic groundmass (Schultz, in press). The groundmass contains many relics of glass shards and tuff particles, indicating that many of the pebbles were probably originally vitric and crystal tuffs. Other pebbles are probably vitrophyres. The abundance of orthoclase or sanidine and quartz suggests that these pebbles are of rhyolitic composition (Schultz, in press).

Results of studies of the direction of dip of cross-strata (Poole and Williams, 1956, fig. 50) indicate generally north to northwest current or stream direction during deposition of the

Shinarump member. The Shinarump member in the White Canyon-Elk Ridge area, however, is significantly different from the rest of the member in that the stream directions are west to southwest (F. G. Poole, written communication; Johnson and Thordarson, 1959).

The Shinarump member averages about 30 feet in thickness, although in some areas it is about 50 feet thick along many miles of outcrop. The member is thickest where it fills channels; it is commonly 100 or more feet thick in channels.

The lower contact of the Shinarump member is a surface of erosion. In some areas, the contact is marked by conspicuous channels cut into the Moenkopi formation and filled with the sediments of the Shinarump member. These channels are mostly a few hundred feet wide and 50 to 100 feet deep. Some are as wide as 2,300 feet, and others are as deep as 150 feet (Witkind, 1956). Conspicuous channels are confined to two elongate belts, one extending from the Defiance Uplift in northeastern Arizona, through the Monument Valley area in Arizona and Utah, and ending in the Circle Cliffs and Capitol Reef area in south-central Utah, and the other one extending from the Cameron area to the Lees Ferry area in north-central Arizona. The lower contact is also marked by broad swales which range in width from 1 to 3 miles and have a relief of about 40 feet in the Monument Valley area, Arizona (Witkind, 1956). A swale at Lees Ferry in north-central Arizona is at least 11 miles wide and is about 175 feet deep. In areas where channels and swales are not present, the lower contact of the member is essentially flat and marked only by a few scours a foot or two deep.

The Shinarump member, in most areas, grades upward into and intertongues with claystone, siltstone, or clayey sandstone of overlying units. The upper contact is generally placed at the top of the highest cross-stratified sandstone.

The Aqua Zarca sandstone member and an unnamed sandstone member occur at the base of the Chinle formation in the Nacimiento Mountains, San Pedro Mountain, and Chama River areas in north-central New Mexico (fig. 32). The members are lithologically similar to one another and to the Shinarump member. The unnamed sandstone member occurs in the Nacimiento Mountains area; it may grade laterally into the Aqua Zarca sandstone member or possibly partly into the Poleo sandstone lentil, or both. The Aqua Zarca sandstone member occurs in the northern part of the Nacimiento Mountains area and in the San Pedro Mountain and Chama River areas. The stream directions in the Aqua Zarca sandstone member are dominantly south to southwest (F. G. Poole, written communication), whereas those of the unnamed sandstone member are dominantly north to northwest. Maximum gravel sizes in the conglomerate of the Aqua Zarca sandstone member decrease toward the southwest from a maximum of 330 mm near the Chama River to about 50 mm farther south.

The Gartra member, although not assigned to the lower part of the Chinle formation, is lithologically similar to the Shinarump member. This member occurs in northeastern Utah and northwestern Colorado (fig. 32). Stream directions in the member are dominantly toward the northwest and west.

Monitor Butte and related members

Above the Shinarump member, or related members, occur units, which are generally 50 to 200 feet thick, of slope-forming claystone and clayey sandstone interstratified with thin lenses of ledge-forming sandstone. These units are designated by different member names in different areas, and the names reflect to some extent lithologic differences in the members. The Monitor Butte member is recognized in southeastern Utah and the Monument Valley area, Arizona; the lower red member in the Defiance Uplift area in northeastern Arizona and in the Zuni uplift in west-central New Mexico; the Mesa Redondo member in the St. Johns-Hunt area in east-central Arizona; the sandstone and siltstone member in the Cameron, Echo Cliffs, and Lees Ferry areas in north-central Arizona; and the Salitral shale tongue in north-central New Mexico. Some of these members are the approximate lateral continuation of some of the other members. All of the members have approximately the same stratigraphic position (fig. 29).

The Monitor Butte member occurs throughout most of southeastern Utah and in the Monument Valley area, Arizona. When viewed from a distance, the member appears as a slope-forming, greenish-gray unit at the base of the Chinle formation, or above the Shinarump member, if it is present, and below the variegated beds of the Petrified Forest member or the cliff-forming sandstone of the Moss Back member (fig. 34), whichever is present. The member in many places can be considered a transitional sequence between the sandstone of the underlying Shinarump member and the claystone and clayey sandstone of the overlying Petrified

Figure 34.--Monitor Butte and Moss Back members of Chinle formation at Buchacre Point along Dirty Devil River, Utah. (Trm, Moenkopi formation; Trcb, Monitor Butte member of Chinle formation; Trcm, Moss Back member of Chinle formation; Trc, remainder of Chinle formation; Trw, Wingate sandstone; Jk, Kayenta formation)

Forest member, as it contains beds similar to the Shinarump member and others similar to the Petrified Forest member.

The Monitor Butte member consists dominantly of greenish-gray claystone or clayey sandstone that weathers to form a "frothy" appearing slope. The clay in these rocks is mainly mixed layer illite-montmorillonite in which either (1) the illite layers are slightly more abundant than, or equally abundant to, the montmorillonite layers or (2) the montmorillonite layers are dominant (Schultz, in press). Except for the tendency for the clays to contain more illite in the mixed layer illite-montmorillonite clay, these claystone and clayey sandstone units are very similar to the dominant type of rock in the Petrified Forest member, which are illitic-montmorillonitic claystone and montmorillonitic volcanic or subvolcanic sandstone. The composition of this type of rocks will be discussed in more detail under the section on the Petrified Forest member. The claystone in the Monitor Butte member is structureless; the clayey sandstone is generally cross-stratified in shallow trough sets. Carbonized wood is one of the distinctive characteristics of the member, and identifiable plant remains, including cycadophytes, conifers, and ferns, occur at several localities (Roland Brown, written communication).

Interstratified with the claystone and clayey sandstone are sandstone lenses. Most of these lenses are 1 to 10 feet thick and extend a few hundred to several thousand feet along the outcrop. The lenses commonly form about 5 to 20 percent of the member, but locally they are absent. The sandstone is very fine-grained and composed of quartz and minor amounts of potassium feldspar and volcanic material set in sparse matrix of illitic, kaolinitic, or montmorillonitic clay (Cadigan, 1959a, p. 545). The rocks are well cemented with a carbonate and iron oxide cement (Cadigan, 1959a, p. 545). They are mostly feldspathic sandstone, quartz sandstone, or calcareous feldspathic sandstone. The stratification is distinctive; most of the sandstone lenses are composed of rib-and-furrow structures. A few layers are cross-stratified. Contorted strata is also characteristic of these layers (fig. 27 and 28). These contortions may be intricate folds or irregular blocks of sandstone lying at almost any conceivable strike and dip (see section on contorted strata for detailed description). A few of the sandstone lenses are conglomeratic, with pebbles of limestone, siltstone, and minor amounts of weathered chert pebbles. The sandstone forms ledges, which gives the member a distinctive ledgy appearance (fig. 33 and 34).

The Monitor Butte member ranges in thickness of 0 to 250 feet.

The contact of the Monitor Butte member and the underlying Shinarump member is transitional in most places. Locally the contact is erosional and greenish-gray silty claystone, siltstone, and clayey sandstone of the Monitor Butte member fill channels cut into the underlying Shinarump member. These channels are commonly several hundred feet across and 30 to 40 feet deep. They have been noted by the writer in the northern Monument Valley area and in the Circle Cliffs area. These channels filled with silty and clayey strata, as will be discussed later, are interpreted to be abandoned channel or slough deposits ("clay plugs").

The contact of the Monitor Butte member and the overlying Petrified Forest member is gradational, and in places the Monitor Butte member is difficult or impossible to separate from the Petrified Forest member. In places where the Moss Back member overlies the Monitor Butte member, the upper contact of the Monitor Butte member is a surface of erosion.

The lower red member extends throughout the Defiance Uplift in northeastern Arizona and into the Zuni Uplift in west-central New Mexico. It is lithologically similar to the Monitor Butte member although it is redder, and is, at least in part, a lateral continuation of the Monitor Butte member. It is essentially a transitional sequence between the Shinarump and Petrified Forest members. The lower red member ranges in thickness from 0 to 300 feet.

The Mesa Redondo member occurs in the St. Johns-Hunt area in east-central Arizona. It is in part a lateral continuation of the lower red member, although the two units appear lithologically distinct. The Mesa Redondo member is composed of grayish-red and grayish red-purple, structureless, hackly weathering siltstone and silty claystone and interstratified lenses of grayish-red, medium- to coarse-grained, cross-stratified ledge-forming silty and clayey sandstone. The member does not contain the greenish-gray and red, "frothy" weathering, claystone and clayey sandstone that typify the other units overlying the Shinarump member. The member ranges in thickness from 0 to 160 feet.

The sandstone and siltstone member occurs in the Cameron, Echo

Cliffs, and Lees Ferry areas in north-central Arizona. This member has not been recognized by most other geologists and was previously included as a part of the Shinarump conglomerate by Wanek and Stephens (1953) and of the Shinarump member by Akers and others (1958). As recognized in this report, the sandstone and siltstone member overlies the Shinarump member and grades laterally on outcrops both to the northwest and southeast into the Petrified Forest member. The member consists of complexly interfingering units of sandstone and siltstone. The sandstone is yellowish gray, grayish red, pale red purple, light greenish-gray, fine to coarse grained, cross-stratified, and locally conglomeratic. The granules, pebbles, and cobbles are dominantly quartzite; others are quartz and chert. A few pebbles of volcanic rock are present on most outcrops. The siltstone is grayish red, grayish purple, and light greenish gray and is structureless. The siltstone units weather to form slopes and the sandstone units to form ledges. In the Lees Ferry area, silty and clayey strata of the member fill channels cut into the underlying Shinarump member and are thought to be "clay plugs." The member ranges in thickness from 0 to 280 feet.

The Salitral shale tongue occurs as a slope-forming unit between the ledge-forming Aqua Zarca sandstone member below and the ledge-forming Poleo sandstone lentil above. It is lithologically similar to the Petrified Forest member.

Moss Back member and related units

The Moss Back member is a thin widespread ledge-forming sandstone and conglomerate unit occurring in southeastern Utah and possibly in a part of westernmost Colorado. The lower member of the Dolores formation in the San Juan Mountains region and the Poleo sandstone lentil in north-central New Mexico are lithologically similar to the Moss Back member and may be lateral continuations of that member, although exact correlations cannot be made.

The Moss Back member overlies the Monitor Butte member along its southern margin. It overlaps the Monitor Butte member toward the northeast and is at the base of the formation along its northeast margin (fig. 29). The lower member is at the base of the Dolores formation. The Poleo sandstone lentil overlies the Salitral shale tongue in most places, although locally it rests directly on the Aqua Zarca sandstone member.

The Moss Back member forms a northwest-trending lens about 50 miles wide and 155 miles long (fig. 35), extending from the Elk Ridge and Abajo Mountains areas, Utah, on the southeast to central Utah on the northwest. It covers about 10,000 square miles. A sandstone unit at the base of the Chinle formation in the Lisbon Valley area, Utah, and in the Slick Rock area, Colorado, are probably equivalent, at least in part, to the Moss Back member.

The Moss Back member is composed typically of yellowish-gray and very pale-orange, very fine- to coarse-grained sandstone. Conglomerate and conglomeratic sandstone lenses are common. The pebbles in the conglomerate lenses generally occur in two suites--either (1) light-brown and gray siltstone and limestone or (2) quartz, quartzite, and chert. The member is almost entirely cross-stratified.

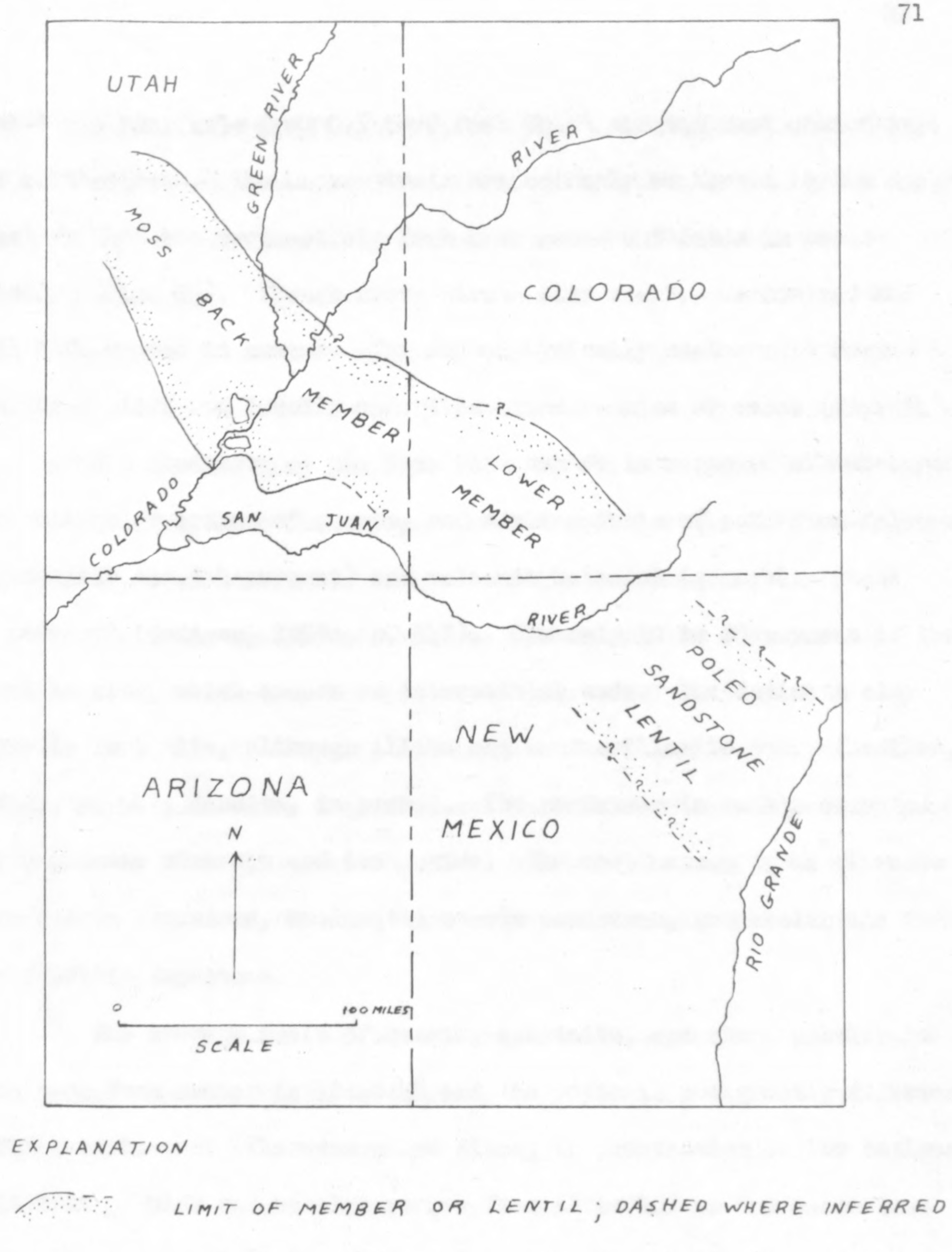


FIGURE 35.--DISTRIBUTION OF THE MOSS BACK MEMBER AND POLEO SANDSTONE LENTIL OF CHINLE FORMATION AND OF THE LOWER MEMBER OF THE DOLORES FORMATION

Tabular planar sets from 0.5 to 2 feet thick are the most common type of cross-strata. The cross-strata are commonly contorted in the upper part so that the cross-strata look like recumbent folds in cross section (fig. 25). Trough cross-strata also occur. Carbonized and silicified wood is common. The member typically weathers to form a vertical cliff and locally underlies broad benches or mesas (fig. 34).

The sandstone of the Moss Back member is composed of subrounded to subangular grains of quartz, and minor amounts of potassium feldspar (generally about 4 percent) and volcanic material (generally about 5 percent) (Cadigan, 1959a, p. 547). Commonly 10 to 20 percent of the rock is clay, which occurs as interstitial wads. The dominant clay type is kaolinite, although illite and montmorillonite occur (Cadigan, 1959a, p. 547; Schultz, in press). The sandstone is weakly cemented by carbonate minerals and iron oxide. The most common types of rocks are quartz sandstone, kaolinitic quartz sandstone, and kaolinitic feldspathic sandstone.

The average ratio of quartz, quartzite, and chert pebbles in the Moss Back member is 12:40:48, and the ratio is not greatly different from area to area (Thordarson and Albee, in preparation). The maximum sizes of pebbles and cobbles ranges from 58 to 102 mm and decreases from the northeast to the southwest, an anomalous situation if the stream directions were to the northwest as is suggested by the direction of dip of cross-strata. Fossiliferous chert pebbles from the Moss Back member contain fusulinids, brachiopods, bryozoa, algae, and, to a lesser extent, gastropods, sponge spicules, ostracods, echinoid spines, corals, and crinoidal material (Thordarson and Albee, in preparation). Most of the fossils, as will be discussed in more detail later, indicate rocks of Permian age; some may indicate rocks of Pennsylvanian or Mississippian age.

Results of studies of the direction of dip of cross-strata (Poole and Williams, 1956, fig. 50) indicate generally northwesterly stream directions during the deposition of the Moss Back member.

The Moss Back member averages about 60 feet in thickness, but is as much as 150 feet thick where it fills channels.

The lower contact of the Moss Back member is a surface of erosion, in most places, and scours as deep as 10 to 20 feet are common. The widest, deepest, and longest channel observed is in the White Canyon area where a channel cut into the Monitor Butte member is a mile wide, 50 to 100 feet deep, and has been traced for 14 miles. The upper contact of the Moss Back member is commonly gradational with the overlying unit.

The lower member of the Dolores formation, a possible lateral continuation of the Moss Back member, is present in the southern part of the San Juan Mountains region (fig. 35). It is composed of light greenish-gray or greenish-gray, very fine- to fine-grained sandstone and subordinate amounts of limestone pebble conglomerate. Some of the conglomerate contains a few granules and pebbles of chert, feldspar, quartz, and possibly granite. Flakes of carbonaceous material are common.

The Poleo sandstone lentil occurs in the northern part of the

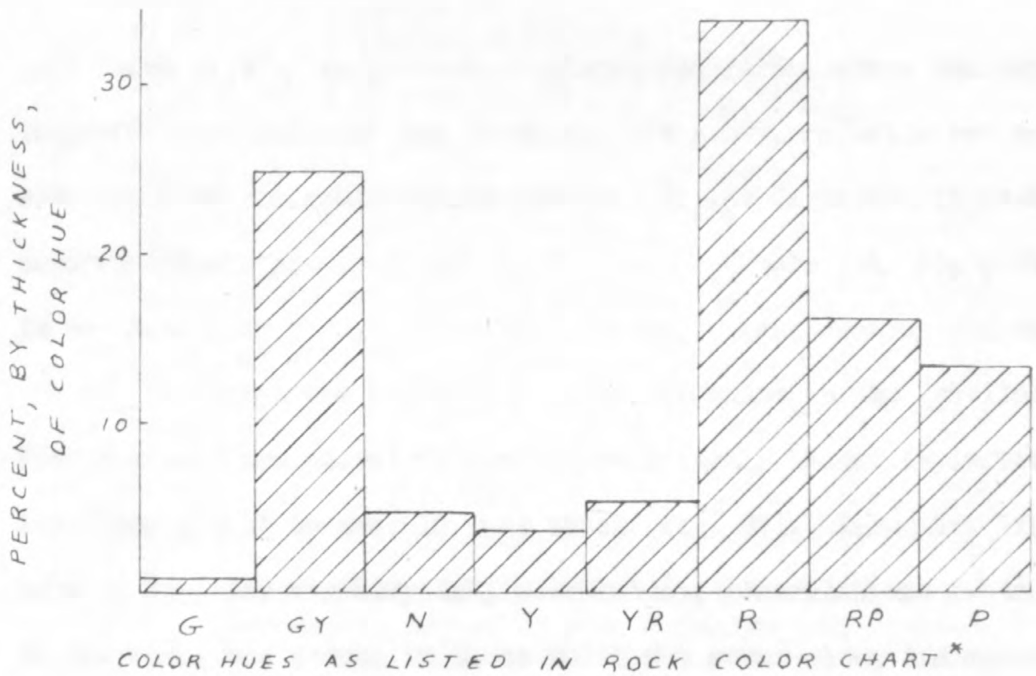
Nacimiento Mountains area, and in the San Pedro Mountain and Chama River area, in north-central New Mexico (fig. 35). It is lithologically similar to the Moss Back member. The pebbles in the Poleo sandstone lentil are dominantly chert and quartz; quartzite is rare (Thordarson and Albee, in preparation). The maximum pebble and cobble sizes at various localities range from 21 to 168 mm, but do not show a systematic increase or decrease from locality to locality (Thordarson and Albee, in preparation). Stream directions in the lentil are mostly north to northwest (F. G. Poole, written communication). The Poleo ranges in thickness from 0 to about 160 feet.

Petrified Forest member

The Petrified Forest member is the thickest and most widespread member in the lower part of the Chinle formation (fig. 29). It is present throughout the southern part of the Colorado Plateau. Toward the west, it extends at least as far as the Spring Mountains in Nevada. On the south, it is present in the most southerly outcrops of the Chinle formation. Much of the Dockum group in eastern New Mexico and adjoining parts of Texas and Oklahoma is lithologically similar to the Petrified Forest member and is probably a lateral continuation, in part, of the Petrified Forest member. The member thins and grades out into other members of the Chinle formation toward the northeast, and reaches a poorly defined northeastern limit in southeastern Utah, southwesternmost Colorado, and northernmost New Mexico.

The Petrified Forest member is composed of three interfingering litholotic types: (1) structureless nonresistant claystone or clayey siltstone, (2) cross-stratified nonresistant clayey sandstone, and (3) cross-stratified ledge-forming sandstone, or locally conglomeratic sandstone. These rocks contain as great a variety of colors as can be expected in any sedimentary formation. Red and green rocks predominate, but the member contains, in addition, rocks with shades of purple, blue, orange, yellow, and gray. The variety of color is shown on figure 36 which is a histogram of the percentages of various hues (based on the Munsel color scheme) occurring in a measured section of the Petrified Forest member at Rockville, near Zion National Park, in southwestern Utah.

Of the three lithologic types present in the formation, the first two, nonresistant claystone or clayey siltstone and cross-stratified nonresistant clayey sandstone, constitute the largest part of the formation. These two lithologic types occur in about equal proportions. Units of claystone or clayey siltstone, which range in thickness from less than a foot to over several hundred feet, occur interstratified with units of clayey sandstone which have a comparable range in thickness. These nonresistant units typically form brightly colored badlands called "painted deserts." The ledge-forming sandstone or conglomeratic sandstone units, the third lithologic type, constitute a small part of the Petrified Forest member. Where most abundant, they probably amount to only 20 percent of the member, and in most areas are less than 5 percent of the member. Over large parts of the Colorado Plateau, they are entirely absent. The most conspicuous and widespread of these ledge-forming units is the Sonsela sandstone bed which occurs



ACTUAL COLORS OCCURRING IN MEASURED SECTION (COLORS IN PARENTHESES OCCUR IN MINOR AMOUNTS)

G LIGHT GREENISH GRAY*
GY LIGHT GREENISH GRAY, GREENISH GRAY (GRAYISH YELLOW GREEN, DARK GREENISH GRAY)
N MEDIUM GRAY TO WHITE
Y YELLOWISH GRAY, DUSKY YELLOW, MODERATE YELLOW (MEDIUM BLUISH GRAY)
YR PALE BROWN, PINKISH GRAY, (DARK YELLOWISH ORANGE, LIGHT BROWNISH GRAY)
R GRAYISH RED, PALE RED, (GRAYISH PINK)
RP GRAYISH RED PURPLE, PALE RED PURPLE, (VERY DUSKY RED PURPLE)
P GRAYISH PURPLE, PALE PURPLE

* COLOR NAMES AND HUES AS LISTED IN ROCK COLOR CHART, PREPARED BY THE ROCK COLOR CHART COMMITTEE, E. N. GODDARD AND OTHERS, NATIONAL RESEARCH COUNCIL, WASHINGTON 25, D.C. (1948)

FIGURE 36.--HISTOGRAM SHOWING PERCENT OF MAIN COLOR GROUPS IN PETRIFIED FOREST MEMBER OF CHINLE FORMATION AT ROCKVILLE SECTION, WASHINGTON COUNTY, UTAH

in a large part of northeastern Arizona and northwestern New Mexico. Ledge-forming sandstone and conglomeratic sandstone units are most numerous near the southernmost outcrops of the formation in east-central Arizona and west-central New Mexico (Cooley, 1959), particularly in the Petrified Forest National Monument in east-central Arizona.

The claystone and clayey siltstone units in the Petrified Forest member are structureless or indistinctly bedded in layers from less than a foot to over 10 feet thick (fig. 37). Irregular limestone nodules (chiefly microcrystalline calcite), generally one or two inches in diameter, are common in these units and occur along individual horizons or scattered irregularly throughout horizontal layers which range in thickness from less than a foot to several feet. The dominant clay in the claystone and clayey siltstone is mixed layer illite-montmorillonite in which montmorillonite layers are dominant. These clays expand readily on contact with water, and as a consequence the rocks containing them weather with a "frothy" or "popcorn" surface (fig. 38). Schultz (in press) reports that "the fine-grained claystones commonly range from a jackstraw mass of clay flakes with tiny, scattered quartz, feldspar, and carbonate grains to fairly well-bedded varieties with parallel arrangement of clay flakes and interstratification of more or less silty bands." Some of the claystone contains fragments with relics of elongate vesicular cavities (Allen, 1930). These fragments are probably relics of pumice. The largest of these fragments observed by Allen has the dimension in thin section of 0.6 by 0.9 mm. Allen also recognized soda sanidine, quartz, biotite,

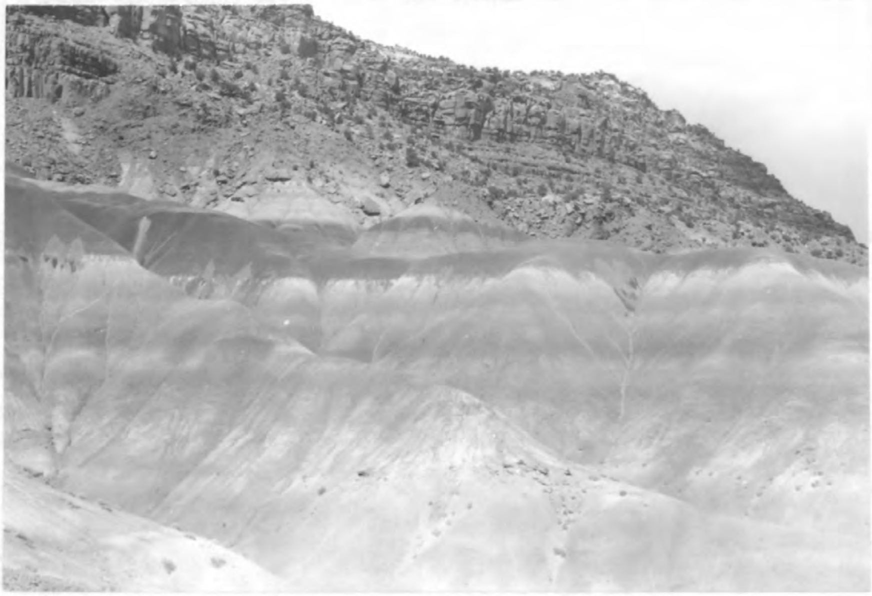


Figure 37.--Horizontally stratified bentonitic claystone in Petrified Forest member of Chinle formation near abandoned town of Paria, Utah. Sandstone in background is part of Glen Canyon group.

Figure 38.--"Frothy" or "popcorn" weathering surface developed on bentonitic claystone in Petrified Forest member of Chinle formation near Joseph City, Arizona. Rock in pit is unweathered.

magnetite, apatite, and zircon in these rocks. The apatite and zircon grains are commonly euhedral (Allen, 1930). Schultz (in press) reported the presence of both low sodic and high sodic varieties of sanidine in the lower part of the Chinle formation. Waters and Granger (1953, p. 6) noted fragments of altered volcanic glass and bits of microlite-filled lava in every section of these rocks cut in oil and more sparingly in the sections cut in water. They observed relics of altered glass shards, pumice, porphyritic lava, spherulitic obsidian, and welded tuff. In order to further test for the presence of volcanic debris in the claystone units, the author examined several thin sections of limestone nodules occurring in these claystone units. The calcite in these nodules apparently has replaced the original rock, and has preserved relics of the original volcanic texture. The probable volcanic material, replaced by calcite, consists of subangular to round, mostly subround, grains ranging in size from a fraction of a millimeter to commonly as large as 3 mm and rarely as large as 5 mm. Many of the grains contain no recognizable internal structures. Other grains, however, contain spherulitic structures which are probably similar to the spherulitic structures in the devitrifying glassy groundmass of a welded tuff illustrated by Enlows (1955, pl. 5, figs. 3 and 4). Still other grains contain minute elongate vesicles, and perhaps elliptical vesicles, and could have been originally pumice fragments. In some of the nodules, these possible volcanic grains constitute a large part of the rock and are set in a structureless matrix.

The claystone in the Petrified Forest member is, with little doubt, derived from the devitrification of volcanic glass, and thus can properly be called bentonite. The volcanic origin of the detrital material in the rock is supported by the abundant remnants of volcanic debris, by the presence of sanidine and other minerals that commonly occur in volcanic rocks, and by the presence of montmorillonite clay which characteristically forms from the devitrification of volcanic debris (Ross and Hendricks, 1945; Grim, 1953). The presently available evidence supports the view that the original rock was mainly water-laid and composed dominantly of subround grains of pumice and other glassy volcanic rock set in a matrix of finer grained glassy material, probably mostly glass shards. The rocks in which pumice fragments cannot be recognized may have originally consisted almost entirely of silt-sized particles of volcanic glass, and contained little, if any, of the larger pumice fragments.

The second lithologic type in the Petrified Forest member is nonresistant clayey sandstone. The clayey sandstone is characteristically cross-stratified; the cross-strata occur in shallow trough sets and dip generally about 5 to 10 degrees, much less of a dip than in the cross-strata in most of the Shinarump and Moss Back members and the Sonsela sandstone bed. The clayey sandstone is generally fine to medium grained and composed of grains of quartz, volcanic rock, and to a lesser extent of potassium feldspar and plagioclase, set in a montmorillonitic clay matrix (Cadigan, 1957b, 1959a, p. 548-549, 1959b; Schultz, in press). The clay matrix commonly constitutes more than

20 percent of the rock. The composition of these rocks is commonly rather varied, but most of them are montmorillonitic volcanic or subvolcanic sandstone. The volcanic material is the most distinctive feature of these rocks; most commonly the volcanic material is a "felsitic igneous rock which contains lath-shaped perfectly euhedral sodic feldspar [plagioclase] phenocrysts in a nearly isotropic groundmass" (Cadigan, 1959b, p. 58). Other grains of less definite volcanic origin include altered rocks with spherulitic structures, which may be relics of spherulitic obsidian or of devitrifying welded tuffs; altered rocks with elongate vesicles, which may be relics of pumice; almost isotropic microcrystalline aggregates of clay minerals; and opaque grains that probably contain a high content of iron oxide. The montmorillonite clay in the rock probably was originally volcanic glass fragments that later devitrified to clay. Red and green microcrystalline chert grains commonly constitute 1 or 2 percent of the sandstone; they are a conspicuous feature of the rock in hand specimen. These chert grains occur in strata in the Chinle formation that contain abundant volcanic material; they are rare or absent in rocks containing only a small amount of volcanic debris. These grains probably were derived originally from irregular silicified masses, chert masses (perhaps siliceous sinter), or chert veins in a siliceous volcanic rock terrain. In addition to the material of certain or probable volcanic origin, the sandstone of the Petrified Forest member commonly contains 1 or 2 percent of microcline grains indicating a probable granitic source rock, a few percent of quartzite grains indicating a

probable metamorphic source rock, and a few quartz grains with secondary overgrowths of quartz indicating a sedimentary source rock.

The third lithologic type in the Petrified Forest member is cross-stratified ledge-forming sandstone and conglomerate. Of this lithologic type, the Sonsela sandstone bed is the most conspicuous unit.

The Sonsela sandstone bed extends throughout 24,000 square miles of northeastern Arizona and northwestern New Mexico. It lies in most areas about 300 feet above the base of the Petrified Forest member. The Sonsela sandstone bed is composed of white, very pale-orange, or yellowish-gray, fine- to coarse-grained, cross-stratified sandstone and conglomerate. Both tabular planar and trough sets of cross-strata occur. Conglomerate layers occur at any position in the Sonsela, but are most abundant near the base. The conglomerate layers are composed of granules, pebbles, and cobbles of chert and small amounts of quartz, quartzite, limestone, and siltstone. The maximum sizes of the pebbles and cobbles range from 33 mm to 152 mm, and decrease gradually toward the north (Thordarson and Albee, in preparation). Pebbles of volcanic rock, probably mostly vitric and crystal tuffs and vitrophyres (Schultz, in press), are locally present, but generally constitute less than 4 percent of the pebbles (Thordarson and Albee, in preparation). The presence of plagioclase (oligoclase) and quartz in these volcanic pebbles suggests that the pebbles are of an intermediate composition, perhaps quartz latite or dacite (Schultz, in press). This composition of volcanic rock contrasts with the rhyolitic composition of those from the Shinarump member. The Sonsela sandstone bed commonly contains bentonitic siltstone and claystone layers, ranging in thickness from less than a foot to over 20 feet, interstratified with the sandstone and conglomerate. The Sonsela is generally 30 to 40 feet thick, although locally it is over 100 feet thick. Stream directions, based on the direction of dip of cross-strata, are north to northeast (F. G. Poole, written communication).

Elsewhere other ledge-forming sandstone units, similar to the Sonsela sandstone bed, occur in the Petrified Forest member (Cooley, 1959). These sandstone units are most numerous in the Petrified Forest National Monument in east-central Arizona. In one of these sandstone units near the boundary of the Petrified Forest National Monument, 66 percent of the gravel fragments are volcanic (Thordarson and Albee, in preparation). The largest one was 132 mm in maximum diameter.

The Petrified Forest member ranges in thickness from 0 to 1400 feet. It is over 1000 feet thick in most of east-central Arizona and west-central New Mexico and thins to the north and northeast.

The Petrified Forest member is the most fossiliferous of the members of the Chinle formation. Extensive amphibian (Colbert and Imbrie, 1956), reptile (Colbert, 1952; Camp, 1930; Colbert, 1947; Colbert and Gregory, 1957; Colbert, 1950; Camp and Welles, 1957), and plant (Daugherty, 1941) remains occur in the member.

Upper part of the Chinle formation

The upper part of the Chinle formation consists of reddish-brown coarse siltstone and minor amounts of limestone, sandstone, and limestone pebble conglomerate. It extends throughout northeastern Arizona, southeastern Utah, western Colorado, and parts of northwestern New Mexico (fig. 39). It is over 1000 feet thick in a part of west-central Colorado and in southwesternmost Colorado, and thins away from these areas where it is thick. The upper part of the Chinle formation is divided into two parts which are, in ascending order, (1) the Owl Rock member and (2) the Church Rock member and related units.

Owl Rock member

The Owl Rock member, distinguished by alternating siltstone and limestone units, occurs in an elliptical area embracing most of northeastern Arizona and southeastern Utah, and small adjacent parts of New Mexico and Colorado. It intertongues and intergrades extensively with overlying and underlying members of the Chinle formation, and in many areas its margin is marked by lateral gradation of the member into other units of the Chinle formation (fig. 29).

The Owl Rock member typically is composed of pale-red and pale reddish-brown coarse siltstone interstratified with pale-red and light greenish-gray limestone beds that form about 5 to 10 percent of the member (fig. 40). The siltstone is indistinctly bedded in layers ranging in thickness from less than a foot to over 10 feet. It appears lithologically similar to the siltstone in the Church Rock member which is illitic arkosic or feldspathic siltstone or in some regions

Figure 40.--Owl Rock member of Chinle formation in southern part of Red Rock Valley in northeastern Arizona. Resistant beds are limestone or limy siltstone. Slope-forming units are reddish-brown horizontally stratified siltstone and sandy siltstone.

illitic-montmorillonitic arkosic siltstone (data from Cadigan, 1959a, p. 551; Schultz, in press).

The limestone in the Owl Rock member occurs as horizontal beds which average a foot in thickness. In some places, the limestone beds appear to have formed by the growth and coalescence of limestone nodules; all gradations can be seen from layers containing a few scattered limestone nodules to layers containing a tight coalescing mass of nodules. Some of the limestone beds, particularly those in the lower part of the member, contain highly irregular masses of reddish orange or gray chert. These masses are generally less than 2 inches across. Some of the chert occurs as irregular stringers in the rock. The composition and texture of the limestone beds is highly varied. Limestone beds at the base of the member are in part probably calcite and dolomite replacement of water laid glassy volcanic sandstone. One such rock studied by the author from the Echo Cliffs area in north-central Arizona consists of subrounded to rounded grains, mostly ranging in diameter from 0.5 to 1.0 mm, set in a finer grained matrix. Many of the grains contain relics of euhedral lath-shaped crystals that very probably are replaced plagioclase. These grains containing the lath-shaped crystal relics appear to be replaced porphyritic volcanic rocks. Other grains have indistinct outlines of elongate or elliptical vesicles, and these grains may be relics of pumice. A few relics of possible, but rather doubtful, glass shards were noted. This same rock contains irregular masses of reddish orange chert ranging in size from microscopic masses to masses about 2 inches in diameter. The chert is microscopic or cryptocrystalline, and much of it is composed of spherulitic aggregates 0.1 mm in diameter. The chert in the rock, as well as the calcite and dolomite, has replaced the original clastic rock composed dominantly or entirely of volcanic material. Cadigan (1959b, p. 55) has noted similar carbonate-silica replacements of volcanic sandstone beds in the Owl Rock member near Fort Wingate in west-central New Mexico. The limestone beds higher in the member, on the other hand, are mostly calcareous arkosic or feldspathic siltstone in which the carbonate cement mineral, probably both calcite and dolomite, constitutes either a large part of the rock or occurs only in irregular patches. Many of these rocks, which appear to be limestone in the field, are actually limy siltstone.

The Owl Rock member also locally contains beds of horizontally laminated and ripple-laminated siltstone and sandstone, cross-stratified sandstone, and limestone and siltstone pebble conglomerate. The member is a moderately resistant unit that weathers to form escarpments. The limestone units weather to form ledges.

The Owl Rock member ranges in thickness from 0 to 450 feet, but in most areas is 250 to 350 feet thick.

The contacts of the Owl Rock member are arbitrary and poorly defined in most areas. The lower contact marks the change from bentonitic claystone and clayey sandstone of the Petrified Forest member below to the largely nonbentonitic reddish-brown siltstone of the Owl Rock member above. Limestone beds in the Owl Rock member also mark this same change. Bentonitic layers and montmorillonitic clays, however, do occur in the Owl Rock member, particularly in the southern part of the Colorado Plateau. In areas where bentonitic beds occur in the Owl Rock member, the contact between the Petrified Forest and Owl Rock members is gradational and may locally be gradational over 100 to 200 feet. The upper contact, in most places, is placed at the top of the highest limestone bed.

The Owl Rock member commonly contains pelecypod and gastropod remains. The pelecypods are most commonly Unio, a fresh water form, and the gastropods belong to several different genera, including the genus Triasamnicola named for forms occurring in the Chinle formation (Yen and Reeside, 1946).

Church Rock member and related units

The Church Rock member and related units are widely distributed on the Colorado Plateau (fig. 29). The Church Rock member itself extends throughout the eastern part of northeastern Arizona, into adjoining parts of northwestern New Mexico, throughout the eastern part of southeastern Utah, most of the west-central and northwestern Colorado, and the easternmost part of northeastern Utah. In addition, the middle and upper members of the Dolores formation in the San Juan Mountains region in southwestern Colorado are lateral continuations of the Church Rock member. A siltstone member of the Chinle formation in the Chama River area in north-central New Mexico is lithologically similar, although not identical, to the Church Rock member.

About 60 to 70 percent of the Church Rock member is composed of pale-reddish-brown and light-brown horizontally stratified coarse siltstone and very fine-grained sandy siltstone (fig. 41). These rocks are distinctly bedded in layers from less than a foot to about 4 feet thick. When viewed closely these rocks mostly appear structureless, but when viewed from a distance the stratification can be seen easily. These rocks (based on data from Cadigan, 1959a, p. 551; Cadigan, 1957a) are classified, in the northern part of the Colorado Plateau, as illitic feldspathic or arkosic siltstone or calcareous feldspathic or arkosic siltstone. In the northern part of the Colorado Plateau, the dominant clay in the Church Rock member is illite; a minor amount of the clay is mixed-layer illite-montmorillonite, in which montmorillonite layers constitute less than a third of the total clay (Schultz, in press). In the northeast part of the Plateau, the member generally contains a few percent of chlorite and in the northwest part a few percent of kaolinite. In the southern part of the Colorado Plateau, the member contains montmorillonite, mixed-layer illite-montmorillonite in which montmorillonite layers are dominant, and illite; chlorite is a common minor constituent.

In addition to the coarse siltstone and sandy siltstone, the Church Rock member contains a variety of other lithologic types including horizontally laminated, ripple-laminated, and wavy-stratified siltstone and sandstone, trough cross-stratified sandstone, limestone and siltstone pebble conglomerate, and planar cross-stratified sandstone.

Horizontally laminated, ripple-laminated, and wavy-stratified siltstone and sandstone occur as very thin to very thick beds

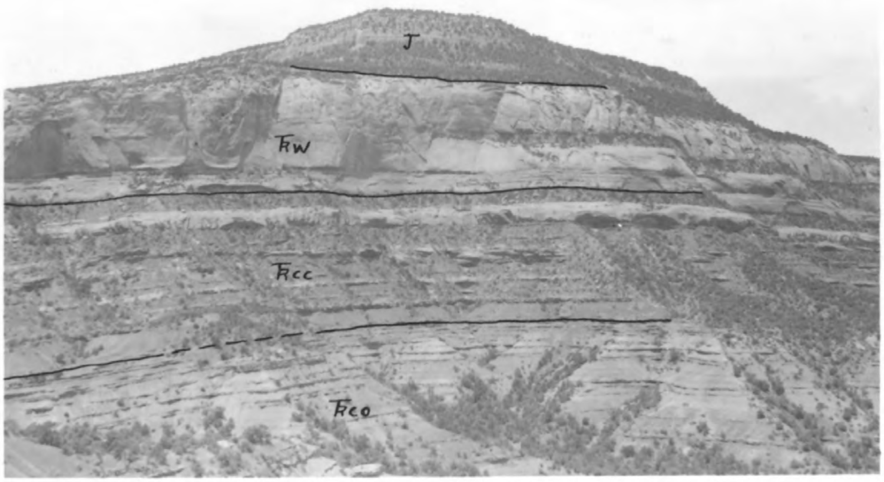


Figure 41.--Owl Rock and Church Rock members of Chinle formation, and Wingate sandstone, in the southern part of Red Rock Valley in northeastern Arizona. Resistant beds in Owl Rock member are limestone or limy siltstone beds. Resistant beds in Church Rock member are sandy siltstone and very fine-grained sandstone, some of which contain planar cross-strata and are interpreted to be eolian in origin. Slope-forming units in both the Owl Rock and Church Rock members are reddish-brown horizontally stratified siltstone and sandy siltstone. (Trco, Owl Rock member of Chinle formation; Trcc, Church Rock member of Chinle formation; Trw, Wingate sandstone; J, formations of Jurassic age)

interstratified with the horizontally stratified siltstone and sandy siltstone and to a lesser extent with the other lithologic types. These rocks exhibit a variety of sedimentary structures including horizontal laminae, parallel and cusp ripple laminae, and thin beds, very thin beds, and laminae that exhibit a vague waviness of low amplitude on their stratification planes. Mudcracked surfaces, worm borings, and raindrop impressions occur locally. Strata included under this heading probably have a variety of origins. Cusp ripple-marked strata occurring in the trough cross-stratified sandstone are probably of fluvial origin; horizontally or parallel ripple-laminated strata are probably mainly deposited by currents in a body of water; wavy-stratified strata may have been deposited by weak currents; some strata exhibit mudcracked surfaces and rain drop impressions indicating some subaerial deposition.

Trough cross-stratified sandstone occurs in the Church Rock member in many areas. It is the dominant lithologic type in a few areas, particularly in a narrow elongate northwest-trending belt extending from southwestern Colorado to central Utah (fig. 42). This belt will be further described in the section on sedimentary facies of the upper part of the Chinle formation. The most conspicuous of the trough cross-stratified sandstone units are given informal names; the Black ledge (Stewart and others, 1959) occurs near the middle of the member in east-central Utah, and the Hite bed (Stewart and others, 1959) occurs at the top of the member in much of southeastern Utah.

The trough cross-stratified sandstone is generally pale red

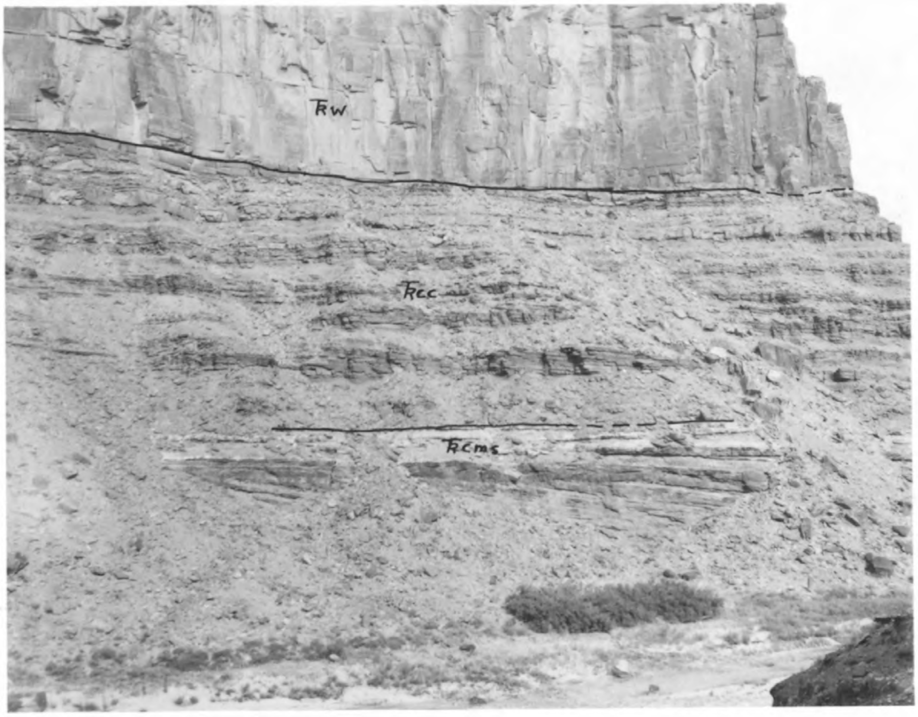


Figure 42.--Chinle formation and Wingate sandstone on Colorado River about 7 miles northeast of Moab, Utah. Mottled strata (Trcms) in this area are anomalous. They are unusually thick and contain a conspicuous angular unconformity. The Church Rock member (Trcc) is typical of the upper part of the Chinle formation in the narrow belt containing abundant fluvial strata that extends from southwestern Colorado to central Utah. The ledges in the Church Rock member are composed of horizontally stratified and cross-stratified sandstone which is probably mostly of fluvial origin. The vertical cliff is the Wingate sandstone (Trw).

or light greenish gray and very fine grained. Although for convenience the strata are designated as trough cross-stratified, actually both trough and tabular planar cross-strata occur. The cross-stratified sandstone is commonly interstratified with horizontally or cusp ripple-laminated siltstone. Carbonaceous material is fairly common in the sandstone. The sandstone occurs either in widespread units, such as the Black ledge or the Hite bed, or in irregular lenses and intertonguing masses. In the narrow elongate belt containing the sandstone, channels filled with sandstone are fairly common.

Stream directions in sandstone units in the area of the elongate belt are dominantly northwest (F. G. Poole, written communication). Stream directions in the Hite bed, based on four studies, are mainly to the northeast and for that reason are anomalous as compared to the rest of the Chinle formation.

Limestone and siltstone pebble conglomerate and their finer grained equivalents, calcarenites, consist of irregular lenses composed of pebbles, granules, or coarse to very coarse grains of limestone, silty limestone, and siltstone set in a limy and silty matrix. Generally the lenses are from 0.5 to 3 feet thick and structureless, although locally some are cross-stratified. They occur interstratified with horizontally stratified siltstone or sandy siltstone, and with ripple-marked or cross-stratified layers. Poorly preserved reptile remains are particularly common in these lenses.

Planar cross-stratified sandstone units occur in the Church Rock member in a region along the Arizona-New Mexico state line in northeastern Arizona and northwestern New Mexico (fig. 41). These units generally are 10 to 50 feet thick and occur, in most places, directly above a horizontally laminated or wavy-laminated sandy siltstone and sandstone unit 5 to 40 feet thick and directly below a unit of typical very thick-bedded siltstone of the Church Rock member. The three types of units, (1) horizontally laminated or wavy-laminated sandy siltstone and sandstone, (2) planar cross-stratified sandstone, and (3) very thick-bedded siltstone, form a cyclic deposit that is repeated as much as four times in one locality. R. F. Wilson first noted the significance of these cyclic deposits and drew the writer's attention to them.

The planar cross-stratified sandstone units are light brown and very fine grained. They are better sorted than sandstone in the rest of the Chinle formation. The sandstone is very similar to that of the Wingate sandstone which overlies the Chinle formation and of which some of these units are tongues. The Wingate sandstone is feldspathic or arkosic sandstone (based on data from Cadigan, 1959a, p. 552 and 1959b, p. 57) and the planar cross-stratified sandstone units in the Chinle formation are probably of similar composition. As the name implies, planar cross-strata are the dominant type of cross-strata; the boundaries of the sets of cross-strata are flat planes (tabular planar cross-strata). Cross-strata with inclined and slightly curved set boundaries (trough cross-strata) also occur. The cross-strata is generally on a medium to large scale. A few studies of the direction of dip of cross-strata (F. G. Poole, oral communication) indicate generally southeasterly inclined cross-strata.

Some of the planar cross-stratified units can be shown to pinch out, or perhaps in part to grade laterally, into the typically very thick-bedded siltstone of the Church Rock member to the northwest in northeastern Arizona (fig. 29). Other units can be shown to be tongues of the Wingate sandstone, tonguing into the Church Rock member toward the southeast. Toward the southeast, the planar cross-stratified sandstone units thicken and coalesce with one another. Along the north side of the Zuni Uplift, the unit that is called the Wingate sandstone is probably composed entirely of a coalesced mass of these planar cross-stratified sandstone units, and the Wingate sandstone as designated here is probably the lateral equivalent of the Church Rock member farther to the northwest (fig. 29).

The Church Rock member ranges in thickness from 0 to over 1000 feet. It is nearly 1000 feet thick in southwesternmost Colorado and over 1000 feet at East Brush Creek in northwestern Colorado. The member is unusually thick in these two areas. In most areas, it is 200 to 400 feet thick.

The lower contact of the Church Rock member is placed, in most places, at the top of the highest limestone unit of the underlying Owl Rock member. The contact of the Church Rock member and the overlying Wingate sandstone is a disconformity except in northwestern Arizona and northwestern New Mexico where the Church Rock member and Wingate sandstone intertongue and in western Colorado where the contact appears conformable. Along the eastern margin of the Colorado Plateau, the

Entrada sandstone (Upper Jurassic) truncates the top part of the Church Rock member or related units.

Fossils in the Church Rock member and related units are rare. The pelecypod Unio occurs in a few areas (Cross, 1907, p. 654; Baker, 1933, p. 40-44). Vertebrate remains include fish and reptiles. Fish occur in the narrow belt of fluvial sandstone extending from southwestern Colorado to central Utah (Hill, 1880, p. 490; Camp, 1930, p. 12-13; Baker, 1933, p. 40-41). Reptile remains occur at scattered localities (Camp, 1930; Colbert, 1950, p. 62; Harshbarger and others, 1957, p. 10). Most of the reptile remains occur in fluvial sandstone and conglomerate units, commonly in limestone pebble conglomerate. The Church Rock member and related units also contain a few remains of cycads (Hills, 1880, p. 490) and of conifers and a possible palm tree (Brown, 1956).

Sedimentary facies of the upper part of the Chinle formation

The upper part of the Chinle formation exhibits marked changes in facies on the Colorado Plateau. These facies relations have been studied by R. F. Wilson (see Stewart and Wilson, 1960) and his illustration is reproduced in figure 43.

In the facies analysis, three lithologic categories are recognized: (1) cross-stratified sandstone and siltstone, probably representing stream channel deposits, (2) horizontally laminated and ripple-laminated siltstone, probably representing deposition from, or reworking by, unrestricted water currents, or "sheet-flow," primarily

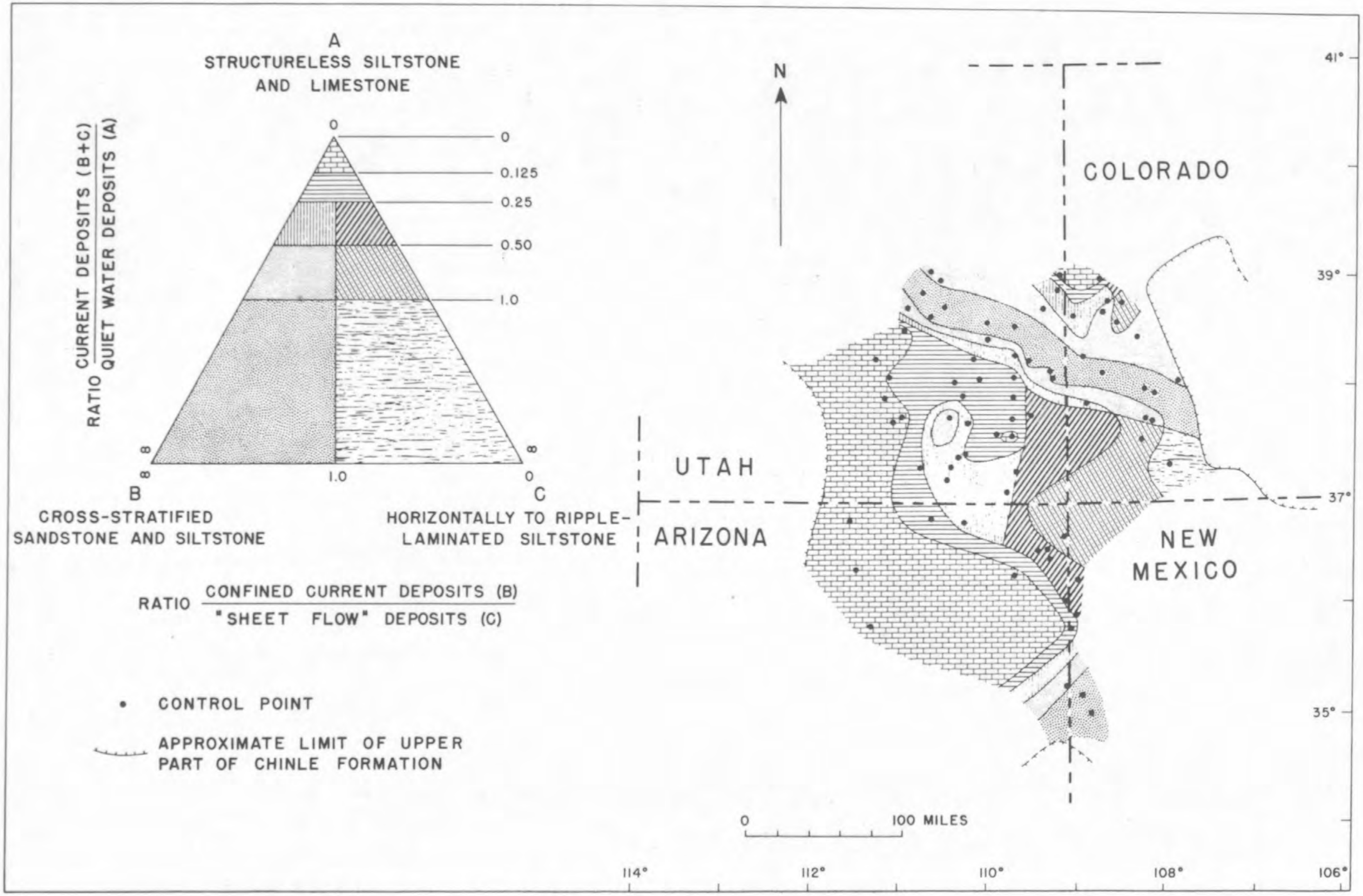


FIGURE 43.- SEDIMENTARY FACIES MAP OF UPPER PART OF CHINLE FORMATION (AFTER STEWART AND WILSON, 1960)

around the margins of lakes and on floodplains, and (3) limestone and horizontally very thick-bedded or structureless siltstone, probably representing quiet water deposition in lakes. The facies map shows a general westward increase in the amount of limestone and structureless siltstone. A large amount of cross-stratified sandstone and siltstone is present in two areas: (1) a relatively narrow belt extending from southwestern Colorado to central Utah, and (2) an area in east-central Arizona and west-central New Mexico.

INTERPRETATIONS

The Chinle formation is a continental deposit containing stream, floodplain, swamp, lake, and eolian sediments. These environmental interpretations are based on the fossil remains and the lithologic types in the formation. Where possible the supposed environments of deposition in the Chinle formation are compared with modern environments. In addition to the environmental interpretations, the paleogeography during the deposition of the formation is discussed, including interpretations as to the location and terrain of the source areas.

Mottled strata

Several ideas have been proposed to explain the origin of the mottled strata. Most geologists (W. L. Stokes, quoted in Johnson, 1957; Johnson, 1957; Schultz, in press; G. M. Richmond, quoted in Finch, 1959, p. 151) consider these strata to be a fossil soil zone. Finch (1953), who first recognized the wide distribution of the mottled zone, which he called the "purple-white" band, had no definite ideas on its origin. Robeck (1956) believed the mottled colors formed by alteration caused by circulating solutions shortly after deposition of the strata. Kerr (1958) believed the mottled color to be an alteration accompanying the formation of ore deposits. He believed the ore deposits to be of hydrothermal origin, and thus apparently also believed the alteration to be hydrothermal.

Available evidence supports the idea that the mottled strata are fossil soil zones. The mottled color is clearly an alteration phenomenon, as it commonly crosses lithologic contacts and locally even crosses the Moenkopi-Chinle contact. The strata occur along the pre-Chinle unconformity where soil would be likely to form. The wide distribution of the mottled strata and their development on different types and ages of rocks, but always in the basal few feet of the Chinle formation or the top few feet of directly underlying rocks, suggests a widespread soil zone developed across rocks of different character. The mottled colors are best developed at the top of a particular group of strata, and the amount of mottling decreases downward and gives way gradually, in most places, into normal colors of the underlying unaltered rock. Such a decrease of intensity of alteration downward is characteristic of the lower part of a soil profile.

The mineralogic differences between the mottled strata and unaltered rock is also indicative of a soil zone. Mottled strata in the Moenkopi formation, for example, commonly contain more kaolinite, more mixed layer illite-montmorillonite in which the illite layers are lightly more abundant than, or about equally abundant to, montmorillonite, and less illite than strata in the unaltered Moenkopi formation. In addition, the mottled strata in the Moenkopi formation rarely contain chlorite and never contain feldspar; both these minerals occur in unaltered rocks of the Moenkopi formation. Such alteration of clay minerals is common in the soil forming processes. The development of kaolinite and destruction of feldspar and chlorite commonly occurs in soils, particularly soils in tropical or subtropical climates.

The fossil soils of the Chinle formation are not unique. Schultz (in press) reports similar mottled soil developed on crystalline rocks just below major unconformities. One such soil reported by Schultz is near Ione, California and occurs on altered Jurassic metamorphic rocks; another is in the bauxite region of Arkansas and occurs on altered syenite. These occurrences are similar to the development of the mottled rocks on the Precambrian crystalline rocks of the Uncompahgre Plateau directly below the Chinle formation. Nikiforoff (1955) reports very similar red mottled strata in southern Maryland which he ascribes to laterization. P. E. Playford (1954) reports similar mottled rocks containing purple and red blotches in an exhumed Tertiary lateritic soil in western Australia.

The soils at or near the base of the Chinle formation developed on a vast smooth surface of erosion. This smooth surface of erosion is indicated by the remarkably featureless contact of the Chinle formation with underlying rocks. Only locally is this smooth surface marked by channels and irregularities. This plain could be called either a peneplain or a pediment. Stokes (1950) has already applied the term pediment to this surface in describing the origin of the Shinarump and related conglomerate units. To most geologists, however, the term pediment is applicable to a much smaller feature developed at the foot of mountains in arid and semi-arid regions. The term peneplain, therefore, seems more desirable and will be used in this report.

The common occurrence of mottled strata in the basal few feet of the Chinle formation indicates that deposition of some of the Chinle formation preceded soil formation. The mottled strata in the basal few feet of the Chinle formation consist of siltstone, which was probably formed by reworking of the underlying Moenkopi formation, and of coarse cross-stratified sandstone and conglomerate, containing debris derived from outside the basin of deposition. These thin alluvial deposits probably formed during the time that the peneplain was developed on top of the Moenkopi formation and other rocks. The streams on this peneplain were probably at grade, but locally they left behind thin alluvial deposits. Mackin (1948) has shown that graded rivers in Wyoming during migration leave behind lateral accretion deposits 15 to 25 feet thick. Although the streams on a peneplain probably would have a lower gradient than those described by Mackin, they undoubtedly would also leave thin alluvial deposits as they migrated laterally. Probably deposition of these thin alluvial deposits and formation of soils went on essentially contemporaneously during the long interval between deposition of the Moenkopi formation and the main part of the Chinle formation. The mottled strata, therefore, are probably much older than most of the strata of the Chinle formation and probably accumulated, at least in part, during Middle Triassic time.

Stokes (1950) has described the formation of thin alluvial deposits during a time of pedimentation. He considered that the Shinarump member formed in such a way and that it was deposited in part during Middle Triassic time. As will be discussed later, however, available evidence suggests that the mottled strata accumulated during the time that the peneplain (or pediment of Stokes) was formed and that the Shinarump member is a later deposit formed by aggrading streams during a depositional cycle following peneplanation.

Lower part of the Chinle formation

The lower part of the Chinle formation consists of variegated claystone, clayey siltstone, and clayey sandstone and thin widespread ledge-forming sandstone and conglomerate units such as the Shinarump and Moss Back members and the Sonsela sandstone bed. As will be discussed below, the lower part of the Chinle formation consists of continental deposits laid down in streams, floodplains, and lakes. The primary source of the detrital material was the Mogollon highland in southern Arizona and adjacent regions. This highland supplied mainly volcanic debris along with some material derived from limestone, sandstone, metasedimentary rocks, and probably granitic rocks. The Uncompahgre highland of western Colorado and adjacent regions supplied a small amount of material, derived from granitic and metamorphic rocks, to this part of the Chinle formation.

Environment of deposition

The types of rocks and fossils that occur in the lower part of the Chinle formation are used in reconstructing the environment of deposition. The fossils quite clearly indicate a continental environment of streams, lakes, and swamps and intervening dry land areas. Based on the types of deposits formed, the Shinarump and Moss Back members and the Sonsela sandstone bed were deposited by shallow braided streams, whereas most of the Monitor Butte, Petrified Forest, and related members were deposited by large, fairly deep, meandering rivers. The structureless and horizontally stratified claystone and clayey siltstone in the Monitor Butte, Petrified Forest, and related members are interpreted to be quiet-water deposits formed in flood basins or lakes.

Fossil evidence

The lower part of the Chinle formation contains an abundant fauna and flora including pelecypods, gastropods, arthropods, fish, amphibians, reptiles, and plants. All these fossils indicate a continental environment of deposition.

Invertebrate remains occurring in the lower part of the Chinle formation are mainly fresh water forms that lived in lakes and streams. *Unio*, a common pelecypod in the formation, is today a fresh water form and probably has been throughout its geologic history. The gastropod *Triasamnicola*, a genus named for specimens found in the Chinle formation, has been referred to as a fresh water form (Yen and Reeside, 1946), because of its association with fresh water invertebrates. Arthropods, including ostracods, branchiopods, and insects are known from the lower part of the Chinle formation. Ostracods occur in either fresh water or marine environments (Moore and others, 1952), but do not give definite evidence for either in this case. Branchiopods are represented by bivalve crustaceans referred to the genus "*Estheria*" (J. B. Reeside, Jr., written communication), a brackish or fresh water form that has been reported from Pleistocene fresh water clays in

Canada (Moore and others, 1952). Insects are represented largely as trails and burrows in petrified wood (Walker, 1938). An "object that appears to be a beetle" has been identified by Roland Brown (written communication) in southeastern Utah.

Vertebrate remains occurring in the lower part of the Chinle formation are aquatic and dry land "upland" forms. The fishes Seminotus and Lepidotus are considered by Colbert (1952) to be fresh water forms that lived in shallow streams and lakes. The lung fish Ceratodus is likewise a fresh water form. The only living form of Ceratodus, or more accurately Neoceratodus, is confined to rivers in Australia where it lives in stagnant pools and water holes (J. W. Bridge, quoted in Lull, 1945). The amphibian Eupelor was generally about 4 to 6 feet long and characterized by an enormous flat skull and small, feeble limbs (Colbert and Imbrie, 1956). It is an aquatic form and may have never left the water (Colbert and Imbrie, 1956; Branson and Mehl, 1929). Reptiles include Hesperosuchus, Typothorax, Machaeroprosopus, Coelophysis, and Placerias, all of which are common and widespread fossils in the Chinle formation. Hesperosuchus (Colbert, 1952) was a lightly constructed, bipedal, carnivorous animal about 4 or 5 feet in length, the smallest of known reptiles from the Chinle formation. Its hind limbs were large, and the fore limbs were small and used for grasping. The animal was adapted to move rapidly and probably was an "upland" form living on firm dry ground. Typothorax was an armored quadrupedal low-lying reptile about 10 feet long with large well-developed, but short, limbs. It probably lived mostly on land (Colbert, 1950, p. 63) where its armor made it practically impregnable to attack. Machaeroprosopus, a phytosaur, was a quadrupedal carnivorous animal closely resembling the present day crocodile (Camp, 1930; Colbert, 1947). Larger individuals probably attained lengths up to 20 feet. This animal probably lived along the banks of streams much like the present day crocodile. Coelophysis was a carnivorous, bipedal animal and one of the first dinosaurs. It was 6 to 8 feet long, lightly built, and probably weighed only 40 or 50 pounds (Colbert, 1955). Its hind legs were very strong and adapted for walking; its front legs were short and bore mobile hands adapted for grasping. Coelophysis must have inhabited fairly dry land over which it could move with agility (Colbert, 1950). Placerias (Camp and Welles, 1956) was a herbivorous, quadrupedal mammal-like reptile about 7 feet long and 3 feet high. It had "tusks" extending out from the upper jaw, and jaws developed for food-mashing and food-grinding. The "tusks" were probably used for raking plant material out of the ground as well as for fighting. Placerias was the chief herbivorous reptile of its time and probably was a dryland upland form.

Plants are the most abundant fossils in the lower part of the Chinle formation and include both land and swamp forms. These plants grew on the depositional plain of the Chinle formation (Daugherty, 1941, p. 29 and 35). Upright stumps with roots traceable more than 10 feet and pith casts of Neocalamites with rhizomes traceable for several feet (Daugherty, 1941, p. 29) indicate that some of the plants are preserved in their original position. Petrified logs of the conifer Araucarioxylon arizonicum, commonly are 3 or 4 feet in diameter and from 60 to 100 feet in length. Judging from the habitat of living araucarians, this conifer lived along the borders of streams or on moist slopes (Daugherty, 1941, p. 30). Remains of Macrotaeniopteris and Neocalamities are abundant, and these plants probably required a swamp environment (Daugherty, 1941, p. 31 and 33). The swollen and fluted bases of Schilderia adamania are similar to the trunks of the bald cypress that grows in today's swamps (Daugherty, 1941, p. 31). The remaining plant fossils including fungi, ferns, lycopods, cordaites, cycadophytes, a ginko, other conifers and shenopsida, most of which require moist land areas.

The fauna and flora of the lower part of the Chinle formation clearly indicate a continental region containing fresh water features such as streams, lakes, and swamps, with intervening dry "upland" areas.

Origin of cross-stratified conglomerate, sandstone, and clayey sandstone

Cross-stratified conglomerate, sandstone, and clayey sandstone occur in two contrasting types in the lower part of the Chinle formation: (1) as thin widespread ledge-forming layers such as the Shinarump and Moss Back members and the Sonsela sandstone bed and (2) as slope-forming units within the Monitor Butte and related members and the Petrified Forest member. The ledge-forming layers generally contain about 10 to 20 percent interstitial clay and silt whereas the slope-forming layers commonly contain more than 20 percent clay and silt. The cross-strata in the ledge-forming layers are both trough and tabular planar types, whereas the cross-strata in the slope-forming units are generally very shallow troughs containing low angle cross-strata.

The combination of cross-strata, channel and scour surfaces, conglomerate layers, abundant plant remains, and locally remains of continental vertebrates clearly indicates that these strata formed by stream action. The origin of the stream deposits in the Chinle formation can be best understood by comparisons with modern stream deposits. The types of deposits formed by recent streams are described below, mostly from descriptions in the literature; the deposits in the Chinle formation are then compared with these recent deposits.

Deposits formed by meandering streams are different from those formed by braided streams, and these two types are described separately below. Leopold and Wolman (1957, p. 59), however, have shown that there is an uninterrupted range of channel patterns from meandering to braided, thus there may be also a complete range in the type of deposits formed by these different types of streams. In the discussion below, ideal cases of meandering and braided streams are considered; the deposits of intermediate types of streams doubtless would combine features of both meandering and braided stream deposits.

<u>Deposits of meandering streams</u>.—Five main types of deposits are associated with meandering streams (fig. 44). These deposits are (1) stream channel deposits, (2) point bar deposits, (3) natural levee and flood plain splay deposits, (4) abandoned channel and slough deposits ("clay plugs"), and (5) river basin deposits (also called

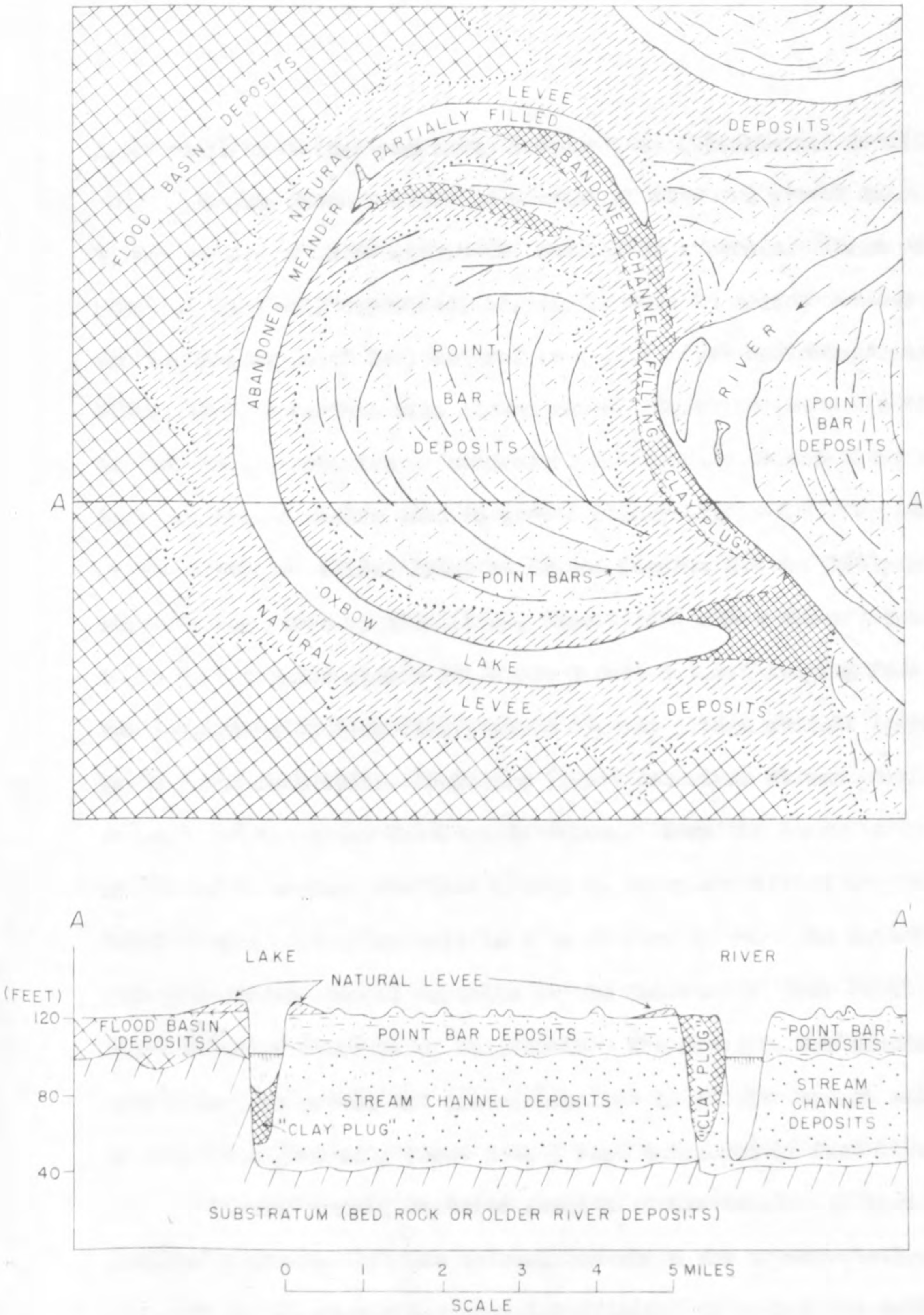


FIGURE 44 — Deposits of a large meandering river. (Modified from Fisk, 1947, Plate 6)

flood plain deposits, backswamp deposits, or interchannel deposits).

Stream channel deposits consist of sand and gravel laid down by the action of water within the channel of a stream. These deposits have not been well described, largely because in recent streams they are covered by point bar, natural levee, or river basin deposits. Fisk (1944; 1947) describes some stream channel deposits (he calls them point bar substratum deposits or sand bars) in the Mississippi River where they consist of medium sand to gravel (Fisk, 1947, fig. 8). They are cross-stratified (Fisk, 1944, p. 18 and figures 14B and 15A); one illustration given by Fisk (1944, figure 15A) shows the cross-strata to be in a tabular planar set about 3 feet thick. Judging from cross-sections given by Fisk (1947, plate 6), the stream channel deposit may be 50 to 60 feet thick. Sundborg (1956) describes stream channel deposits in the river Klarälven in Sweden. Here the deposits consist of coarse to medium sand that occurs in cross-stratified layers. The total stream channel deposit is 7 to 10 feet thick. The author has examined stream channel deposits in the Sacramento River about 75 miles north of Sacramento, California. The deposits are composed of medium sand to gravel and contain shallow lenticular trough sets of cross-strata generally about 3 to 5 feet thick and 15 feet wide.

Stream channel deposits consist of the tractional load (bed load) of a stream that has accumulated where the transportation capacity of the stream has been insufficient to remove the sand as rapidly as it has been deposited (Happ, Rittenhouse, and Dobson, 1940). Stream channel deposits accumulate during the waning stages of a flood, filling in areas that have been excavated during the flood. These deposits also accumulate as bars on the convex sides of meander loops. As a stream migrates laterally, erosion takes place on the convex side of a meander loop and deposition, of a corresponding amount of sediment, takes place on the concave side of the loop. As a result of progressive shifting, or migration, of the stream across an area, a thin tabular layer of stream channel deposits (lateral accretion deposits) is formed on the deposition side of a meander loop (fig. 44). Mackin (1948) has described such deposits left behind by graded streams in Wyoming. These deposits in Wyoming are a thin tabular layer of sand and gravel about 15 to 25 feet thick and include a thin upper silty layer. This silty layer is perhaps a point bar deposit, although it is called a floodplain deposit by Mackin.

Point bar deposits consist of material deposited on top of the stream channel deposits on headlands adjacent to a stream. The sediment consists primarily of material originally suspended in the stream and deposited when the stream spreads across the headlands during times of flood. The top surface of point bar deposits are marked by arcuate ridges, called point bars (fig. 44), from which the name of the deposits is derived. These point bars mark the former positions of the streams (Sundborg, 1956). The deposits consist of coarse silt or very fine-grained sand.

Natural levee and flood-plain splay deposits form along the banks of streams where suspended sediment is deposited during times of flood. As the streams overflow their banks, suspended sediment is dropped where there is a decrease in velocity of the water. Locally the rivers break through the natural levee and form large fan-shaped deposits, similar in shape to alluvial fans, built out into the river basin. These deposits are called flood-plain splays. Natural levee and flood-plain splay deposits consist mostly of silt deposited in irregular laminae or thin beds.

Clay-rich deposits form in abandoned channel or sloughs of a river. Such abandoned channels are common in meandering rivers and form the well known ox-bow lakes. Deposits in abandoned channels are generally clay and fine silt, and are referred to on the Mississippi River as "clay plugs." Such deposits are common on the Mississippi and Sacramento Rivers in the United States (Fisk, 1944, 1947; Lorens and Thronson, 1955).

River basin deposits (flood-plain, backswamp, or interchannel deposits) consist of fine silt and clay deposited in river basins adjacent to a stream. During floods these basins may be largely filled with river water, and sediment held in suspension settles out. River basin deposits are commonly regularly laminated (Jahns, 1947; Dunbar and Rogers, 1957, fig. 17). Wolman and Leopold (1957) have tried to show that flood-plain deposits form only a small part of the deposits in a stream valley. In some streams this is doubtless true, but in others fairly thick river basin deposits have accumulated. Fisk (1944; 1947) illustrates widespread river basin deposits, commonly 50 feet thick, in the valley of the Mississippi River. Extensive river basin deposits also occur in the lower part of the Sacramento River in

California (Lorens and Thronson, 1955). Also, as pointed out by Wolman and Leopold (1957), Jahns (1947) found that one-third of the deposits in the Connecticut River valley are flood-plain deposits.

Deposits of meandering streams consist of several different lithologic types commonly occurring in close association. The deposits show considerable variation laterally; stream channel deposits commonly abut against "clay plugs" which in turn may abut against river basin deposits. "Clay plugs" are particularly common and characteristic of deposits of meandering streams.

Deposits of braided streams.--The alluvial plain of a braided stream consists of a relatively flat surface composed of a network of bars separated by shallow channels (fig. 45). Melton (1936) has described such plains as "bar plains." The bar plains are strikingly different from the alluvial plains of meandering streams. A plain traversed by a meandering stream contains a variety of different types of sediments, such as point bars, natural levees, "clay plugs," and river basin deposits, whereas the bar plains are a monotonous expanse of bars and anastomosing channels. One can imagine that the deposits built up by a braided stream consists of layer upon layer of thin lenticular units of stream channel and bar deposits (fig. 45).

Cross-strata are probably a characteristic feature of most braided stream deposits. Folk and Ward (1957) describe "torrential" cross-strata (probably tabular planar sets) in a bar in the Brazos River of Texas. Doubtless cross-strata also form within the channels of the braided streams, and these cross-strata may closely resemble

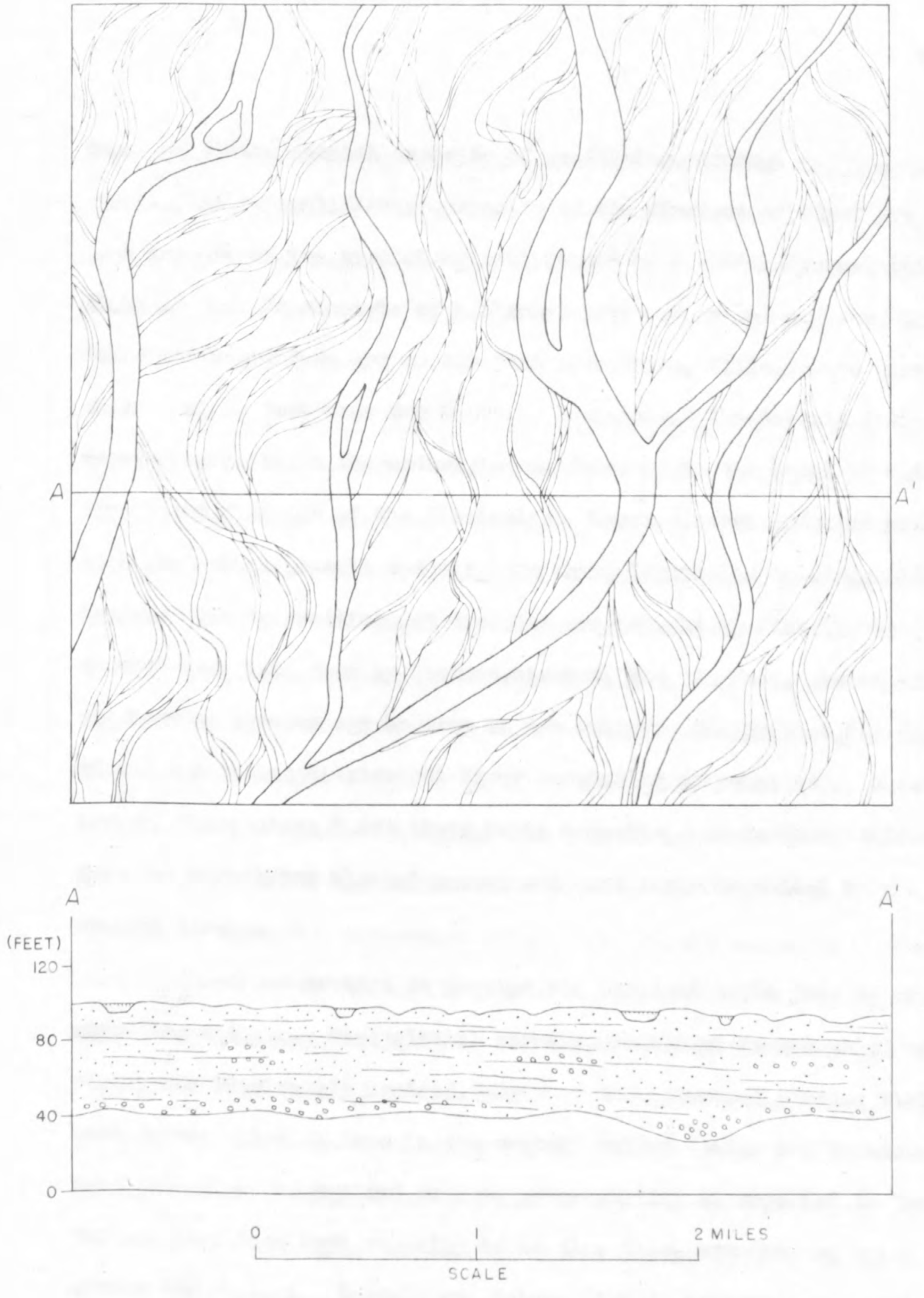


FIGURE 45. — Deposit of braided streams.

those of stream channel deposits of meandering streams.

The "graveliferous" deposits of the Mississippi River are a good example of the type of deposit formed by a braided stream (Fisk, 1944; 1947). It consists of a blanket layer of gravel and sand that thins southward from 100 to 200 feet near Cairo, Illinois, to generally about 50 feet near New Orleans, Louisiana. The deposit fills old erosion channels in the underlying surface and is the basal deposit of the Recent alluvium of the Mississippi River. Gravel and sand constitute the entire deposit and occur in thin, lenticular or tongue-like layers. The "graveliferous" deposits are thought by Fisk (1944; 1947) to have been laid down by braided streams, and, in fact, traces of these old braided streams can be seen in the valley. The present day deposits of the meandering Mississippi River consisting of point bars, natural levees, "clay plugs," and river basin deposits, are markedly different from the underlying blanket gravel and sand layer deposited by the braided streams.

A braided pattern is perhaps the characteristic form of rapidly aggrading streams. Most glacial streams, that have formed thick valley fills, are intricately braided. Many of the ephemeral streams that have formed alluvial fans in the western United States are braided. The streams may be braided because more detritus is supplied to the streams than they have capacity to handle; thus, bars may be built up within the channel. Leopold and Wolman (1957), however, have indicated that braided streams are not necessarily overloaded and that braided streams are one of many channel patterns that streams may adopt in response to the discharge, slope, and other factors of the streams. They show that an increase in the discharge of a stream, for example, may cause the stream to change from a meandering to a braided course. Nonetheless, a large load may also produce a braided stream (Leopold and Wolman, 1957, p. 63).

Comparison of recent stream deposits with the deposits of the Chinle formation.--The Shinarump and Moss Back members and the Sonsela sandstone bed exhibit the typical features of braided stream deposits. They consist of a blanket of sandstone and conglomerate composed of layer upon layer of cross-stratified units. They are laterally homogeneous and are remarkably similar to the blanket "graveliferous" deposits of the Mississippi River that were formed by braided streams.

In a few areas, features that are interpreted to be "clay plugs" occur at the top of the Shinarump member. Such "clay plugs" consist of channels, commonly several hundred feet across and 30 or 40 feet deep, cut into the top of the member and filled with greenish-gray silty claystone, siltstone, and clayey sandstone of the Monitor Butte member or related strata (fig. 33). These features indicate, therefore, that at least locally the streams of the member were of the meandering type. Nonetheless, features than can be interpreted as indicative of meandering streams in the Shinarump member, as well as in the Moss Back member and Sonsela sandstone bed, are extremely rare, indicating further that most of the streams were of the braided type.

The cross-stratified deposits in the Monitor Butte, Petrified

Forest, and related members, on the other hand, could very likely be deposits of meandering streams. The shallow trough sets of low angle cross-strata occurring in these units are different from the cross-strata in the Shinarump and Moss Back members or Sonsela sandstone bed and generally resemble the cross-strata observed by the author in stream channel deposits of the Sacramento River, a meandering river. In addition, these cross-stratified units are lenticular and discontinuous, and are associated with evenly and thick-bedded claystone and siltstone units that could be river basin or associated lake deposits.

Origin of cross-strata.—Cross-strata, as outlined previously, develop in stream channel and bar deposits. Their occurrence in stream deposits has long been recognized by geologists, and considered as one criterion for recognizing stream action, but how cross-strata form is not entirely known and even less is known of why one type develops instead of another. For example, the stream deposits in the Chinle formation contain both tabular planar and trough cross-strata. The different processes by which these contrasting types of cross-strata form can only be understood in a general way.

Tabular planar cross-strata can form in transverse bars in meandering or braided streams or in irregular bars in braided streams. They most likely develop on a flat stream bottom and in fairly shallow water.

The formation of tabular planar cross-strata in transverse bars is fairly well understood because of the work of Sundborg (1956, p. 207, 270-272) on cross-strata in the river Klarälven in Sweden. Sundborg observed that large transverse bars occur almost throughout the meandering course of the river. These bars are 0.05 to 0.5 m high and 2 to 20 m apart. The upstream side of the bars is smooth and dips upstream at an angle of about 1 degree. The downstream side is steep and roughly at the angle of repose. The bars are transverse to the direction of the flow, so that the crests of the bars form long transverse ridges across the stream and the steep downstream sides of the bars appear as linear features at right angle to the stream flow and facing downstream. Sediment is carried up the backside of the bar and deposited on the frontside, in the manner of the foreset beds of a delta building out into a body of water. As the front of the bar is built forward by continued deposition, a tabular layer of cross-strata is left that is very similar, if not identical, to the tabular planar sets of cross-strata of the Chinle formation.

Tabular planar cross-strata probably also form in irregular bars in braided streams. Folk and Ward (1957) have shown that water flowing over a river bar in the Brazos River in Texas alters its flow pattern so that the water runs off the lee side of the bar at right angles to the bar's edge, even though the edge of the bar is not at right angles to the main flow of the stream. Folk and Ward observed "torrential" cross-strata (probably tabular planar cross-strata) developed on the lee side of this bar. These cross-strata apparently were built out on the lee side of the bar in much the same manner as the foreset beds of a delta building out into a body of water.

Tabular planar cross-strata probably tend to develop in streams with flat bottoms and shallow water. Where the water is deeper, swifter, and more turbulent, the regular pattern of transverse bars may be destroyed. Transverse bars are, for example, common in estuaries and on tidal flats where water is shallow and the bottom is flat. Transverse bars would be expected, therefore, in shallow braided streams or in shallow, flat-bottomed portions of meandering streams. The river Klarälven is fairly shallow and has a flat bottom, thus possibly accounting for the numerous transverse bars. The expectation that braided streams commonly contain transverse bars is born out by the common occurrence of tabular planar cross-strata in Pleistocene glacial deposits which, judging from recent glacial streams, are generally braided. Illes (1949) and Longwell, Knopf, and Flint (1932, fig. 122) illustrate beautifully developed tabular planar cross-strata in Pleistocene stratified glacial deposits.

The origin of trough cross-strata is obscure, and, as far as the writer knows, no one has described their formation in a recent stream. In addition, trough cross-strata may form in different ways. The typical cross-strata of the Shinarump member, which occur in shallow elongate troughs, may form by the downstream migration of arcuate "sand waves" or bars. Other trough cross-strata may form by filling in of "holes" on the stream bottom produced by large vertical vortices. Most likely all of the trough cross-strata are formed in areas of deeper, swifter, and more turbulent parts of a river.

The typical cross-strata of the Shinarump member occur in shallow elongate troughs some of which extend for 20 or more feet (fig. 16). These cross-strata are believed to form by the downstream migration of arcuate "sand waves" or bars, similar in shape to barchan dunes. The formation of these trough cross-strata is probably similar to the formation of tabular planar cross-strata in transverse bars except that the bars are crescentic features which are concave downstream and are of limited lateral extent. Also in contrast to transverse bars, erosion takes place on the downstream side of the bar in the area between the two arms of the crescent. In such a manner, the trough is extended downstream as the arcuate bar migrates downstream. The entire bottom of the stream is probably covered by these arcuate bars, and many different sets of cross-strata are thus being created synchronously on the bottom of the stream.

Other cross-strata occurring in the Chinle formation fill deep troughs that appear to be large scours which are later filled in by sediment. These troughs may have been produced by erosion produced by large vertical water vortices in the stream. Mattles (1947) has described large scale turbulence, macroturbulence, in streams, including these large vertical vortices. A vortex consists of an upward rise of a large volume of water in a sort of miniature subaqueous tornado. The vortex starts at the bottom of the stream. It rises quickly off the bottom and ascends to the surface where it forms a sediment laiden boil rising slightly above the surface of the surrounding water. The vortices, according to Mattles, are the most powerful agents of stream scour and produce troughs on the bottom of the streams. Filling of the troughs from the upstream side by bed load material could then produce the cross-strata.

Most likely trough sets of cross-strata develop in deeper, swifter, and more turbulent parts of a river or stream than do the tabular planar cross-strata. Perhaps the river has cross currents or turbulence that breaks up the transverse bars. Kindle (1917, p. 21-22, and plate 13A and B), for example, describes basin-like depressions in estuaries. These depressions cover an area of 3 acres or more in a place where currents were presumably too strong or too irregular to form the more regular linear sand waves (similar to transverse bars of Sundborg, 1956) which are the more common type of wave form in the estuaries. Trough cross-strata may be more characteristic of deep meandering streams than of shallow braided streams, although trough cross-strata probably will also develop in the deeper channels of braided streams.

Origin of channels and swales.--Channels have been interpreted to form in two different ways. According to one idea (McKee and others, 1953), the channels are formed during a two-stage process of channel cutting and much later deposition; the channels are considered as valleys of an old land surface. According to the other idea, the channels are formed and filled with sediment during a one-stage process of cutting and deposition as rejuvenated streams gradually lost their power; the channels are considered as merely the cross-sections of rivers.

Choosing between these two ideas is difficult. Clearly some of the small scours must have formed during flood times and filled during the following slack water time. Such small scour surfaces are common within fluvial units such as the Shinarump and Moss Back members and are common at the bottoms of present day streams. The features considered to be "clay plugs," if interpreted correctly, must also be cross-sections of rivers. The cross-sections of many of the channels filled with sandstone and conglomerate are similar to those filled with finer grained rocks ("clay plugs"). As stated previously, channels at the base of the Shinarump member may be most abundant in the same areas in which the "clay plugs" are most numerous. This association of sandstone- and conglomerate-filled channels with "clay plugs," suggests that both represent the same feature--the cross-section of a river.

The larger channels, on the other hand, are too deep in relation to their width to be merely the cross-sections of rivers. The present day channel of the Mississippi River, for example, is on the average 1,800 feet wide and 56 feet deep and has a width to depth ratio of 37 (calculated from data on plates 23-65 of Fisk, 1947). Channels at the base of the Shinarump member in the Monument Valley area are, on the average, 260 feet wide and 30 feet deep and have a width to depth ratio of 8 (calculated from data on p. 114 of Witkind, 1956). Some of the channels at the base of the Shinarump member are, therefore, much deeper in relation to their width than is the channel of the Mississippi River. In addition, some of the channels at the base of the Shinarump member are 150 or more feet deep, much too deep to be the cross-section of a river. The deeper channels thus are probably valleys on an old land surface in which the stream flowed perhaps as much as 50 feet below the surrounding land. Later aggradation filled these channels.

In summary, some of the shallow channels are considered to be cross-sections of rivers, whereas the deeper channels are considered to be ancient valleys on an old land surface. Telling the difference between these two types in any one channel may be difficult or impossible.

Swales are considered to be broad river valleys which are later partly or entirely filled with sediment. They probably mark the sites of major river systems.

Origin of thin widespread sandstone and conglomerate units.—The fluvial Shinarump and Moss Back members and Sonsela sandstone bed of the Chinle formation have been considered remarkable by many geologists because of their widespread distribution and thinness. The Shinarump member occurs in an area of about 140,000 square miles, although it is absent in several large areas within this region. It may average about 30 feet in thickness, although in some areas it is about 50 feet thick along several miles of outcrop. The Moss Back member is more continuous and covers at least 10,000 square miles. It averages about 60 feet in thickness. The Sonsela sandstone bed covers 24,000 square miles and averages about 30 to 40 feet in thickness.

Such thin widespread fluvial units are fairly common in other parts of the geologic section. Stokes (1950) discusses several units of this nature, including the "Shinarump conglomerate," in the Mesozoic strata of the Colorado Plateau and Rocky Mountain regions. In addition, the Flaxville gravel (Miocene and Pliocene) of the northern Great Plains, the Bishop conglomerate (Miocene ?) of the north flank of the Uinta Mountains in Utah and Wyoming, the upland gravels (Pliocene ?) of Maryland, and the "graveliferous" deposits (Recent) of the Mississippi River are all thin widespread fluvial units. These last four examples are of particular importance as they have been studied extensively and are recent enough that pertinent information concerning their origin can be obtained from present day physiography of the area in which they occur. These four examples are briefly described below, in order to indicate their settling and origin.

The Flaxville gravel occurs on outliers covering 1,800 square miles in northern Montana, and spotty outcrops of the Flaxville gravel and probable equivalents occur in an area of approximately 60,000 square miles in the northern Great Plains (Alden, 1932; Collier and Thom, 1918). It is composed of cross-stratified or irregularly bedded gravel, clay, and sand, and ranges in thickness from a few feet to 100 feet. The Flaxville gravel is a deposit spread out to the east from the Rocky Mountains and lies on one of several vast alluvial terraces formed by stream planation during long periods of tectonic stability. By the end of Miocene time or early Pliocene time the planation proceeded so far that the whole of the northern Great Plains was a vast peneplain (or pediment). The Flaxville gravel is a deposit formed during this planation or deposited on the peneplain.

The Bishop conglomerate (Miocene) occurs on spotty outcrops throughout an area of 2,300 square miles, and consists of sandstone and conglomerate (Bradley, 1936). It rests on the Gilbert Peak surface, a pediment surface, that can be traced southward high up onto the Uinta Mountains. The Bishop conglomerate is thickest in the most concave part of the Gilbert Peak surface where it is 100 or more feet thick. The formation thins towards and pinches out on the north flank of the mountains which were the source of the sediment. Bradley (1936) considers that the Gilbert Peak surface is a pediment cut under semi-arid or arid conditions and that the Bishop conglomerate was deposited on this surface because of stream aggradation due to a change in the regimen of the streams. Most likely the change in regimen was caused by an increase in aridity. Bradley considers the Bishop conglomerate to be too thick to be a pediment gravel.

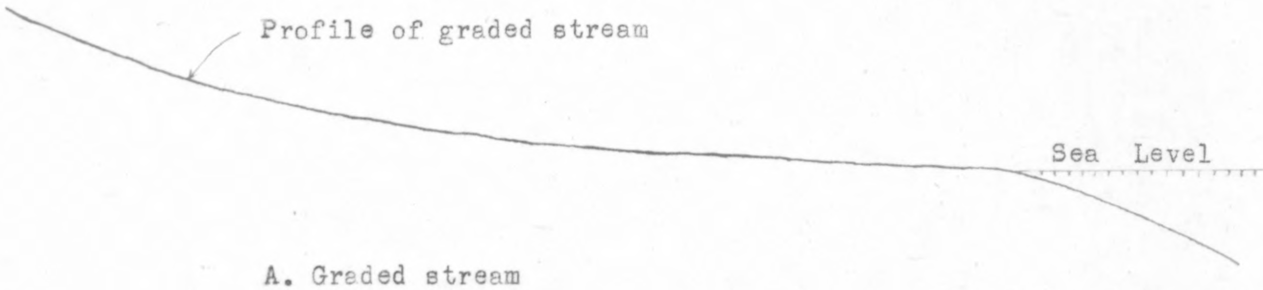
The upland gravels (Pliocene ?) of Maryland cover approximately 600 square miles and consist of cross-stratified or fairly well-stratified gravel and sandy gravel and a thin upper member of loam (Schlee, 1957; Hack, 1955). It has a fairly constant thickness of 25 to 30 feet. The upland gravels were probably deposited by meandering streams that derived their sediment from the Piedmont and flowed eastward onto the easily eroded sediments of the Coastal Plain. The streams were degrading and as they migrated cut the gently sloping surface below the deposits leaving behind the thin lateral accretion deposit of gravel, sandy gravel, and loam.

The Recent "graveliferous" deposits of the Mississippi River cover an area of approximately 40,000 square miles in an elongate belt about 550 miles long and 70 miles wide. These deposits consist of sand and gravel ranging in thickness from 100 to 200 feet near Cairo, Illinois to generally about 50 feet near New Orleans, Louisiana. The deposits fill and cover over an entrenched valley system. Fisk (1944; 1947) has outlined the events leading to deposition of this unit as follows. During accumulation of continental ice sheets of the Late Wisconsin glacial stage, sea level was lowered and an entrenched valley system was cut to a maximum of 400 to 450 feet below the previous level of the river. At the close of the ice age when water level was rising because of the melting of the continental glaciers, the rivers of the Mississippi River began to aggrade due to a rise of base level. The deposits of this aggradation are the "graveliferous" strata.

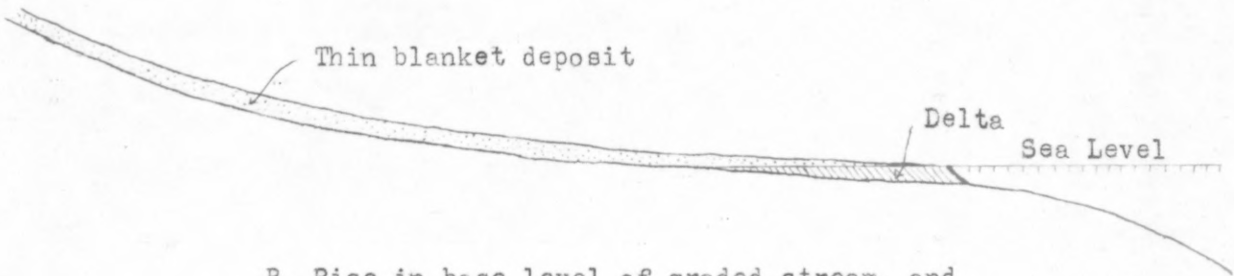
As outlined above, essentially two different hypotheses have been proposed to explain thin widespread sandstone and conglomerate layers such as the Flaxville gravel, Bishop conglomerate, upland gravels of Maryland, and the "graveliferous" deposits of the Mississippi River. These hypotheses are (1) these units represent thin deposits of lateral accretion left behind as a stream migrates across a plain and (2) the units represent deposits formed by aggradation due to a change in the regimen of a graded stream flowing across a pediment or peneplain.

The lateral accretion hypothesis is based on the often observed relation that a stream at grade, or a slowly degrading stream, leaves behind a thin alluvial deposit as it migrates laterally. The flat underlying erosion surface is produced by the stream as it migrates. Mackin (1948) has described this relation very adequately, and the upland gravels of Maryland and, in part, the Flaxville gravel of Montana may have this origin.

The other hypothesis attributes the deposition of thin wide-spread sandstone and conglomerate layers to a change in regimen of a graded, or nearly graded, stream on a peneplaned surface. Suppose that a stream, or streams, have cut a smooth erosion surface or pene-plain and that the streams are at grade on this surface. As defined by Mackin (1948), "a graded stream is one in which, over a period of years, slope is delicately adjusted to provide, with available discharge and with prevailing channel characteristics, just the velocity required for the transportation of the load supplied from the drainage basin. The graded stream is a system in equilibrium; its diagnostic characteristic is that any change in any of the controlling factors will cause a displacement of the equilibrium in a direction that will tend to absorb the effect of the change." Now, if the regimen of the stream is changed, erosion or deposition will occur on the plain in order to restore equilibrium. Suppose, for example, the base level is raised by a rise in sea level (fig. 46), and a delta is built out into the body of water. With these conditions, the overall gradient of the stream is reduced in comparison with its former course. The stream can no longer carry as much load due to the lower gradient and concomitant loss of velocity, and the stream aggrades. The aggrading is



A. Graded stream



B. Rise in base level of graded stream, and deposition of delta and thin blanket deposit

Figure 46.--Development of thin blanket deposit due to rise in base level of graded stream

an attempt of the stream to restore equilibrium and a graded condition. Deposition in the aggrading stream will start in the area of the delta, the area of lowest gradient, and gradually work upstream. The end result of the aggradation will be a thin blanket of sediment extending throughout much of the length of the stream. Such a process has been suggested for the formation of the "graveliferous" deposits of the Mississippi River. The "graveliferous" deposits were, according to Fisk (1944; 1947), built up as a result of rising sea level during melting of the continental glaciers.

The work of Happ (1948) on the middle Rio Grande Valley in New Mexico, affords a good example of the deposition of a thin layer of sand over a long distance in a modern river as a result of a change in regimen of the river. A careful study by Happ showed that the middle Rio Grande River is aggrading its channel at the rate of about 1 foot in 12 years, not at only a few places, but throughout 133 miles of its course. Clearly here a thin sheet of sediment is being deposited over a long distance and if this process were to continue, a thin uniformly thick blanket layer of sediment would be built up. Happ attributes the deposition to an increase in debris supplied to the stream as a result of arroyo cutting in many tributary streams and washes.

Of the two hypotheses, the formation of thin widespread sandstone and conglomerate layers in the Chinle formation is attributed to aggradation due to a change in regimen of a graded stream flowing across a peneplaned surface. The occurrence of soils, as represented by the mottled strata, below the Shinarump member indicates that a smooth flat surface--a peneplain--was developed before deposition of the Shinarump member. If the lateral accretion hypothesis is considered, the flat surface below the Shinarump member would be cut by the same streams that deposited the member, and soils would not be expected. In addition, the mottled strata themselves contain sandstone and conglomerate strata that probably represent lateral accretion deposits laid down during the time that the peneplain was cut. The Shinarump member is seemingly a deposit formed by a later period of aggradation.

In addition, the Shinarump and Moss Back members and the Sonsela sandstone bed are generally thicker than deposits usually ascribed to lateral accretion deposits. The upland gravels of Maryland, which are considered to be lateral accretion deposits, are 25 to 30 feet thick. The Shinarump member, although it averages about 30 feet in thickness, is commonly 50 feet thick along many miles of outcrop, and is 100 or more feet thick in some areas. The Moss Back member averages about 60 feet in thickness and is commonly 100 feet thick. The Sonsela sandstone bed averages 30 to 40 feet in thickness, although it is commonly over 100 feet thick. The thicknesses of the sandstone and conglomerate layers in the Chinle formation are, therefore, commonly thicker than deposits usually considered to have formed by lateral accretion.

The arguments for the origin of the sandstone and conglomerate layers in the Chinle formation by aggradation on a peneplaned surface are not strong. Possibly both aggradation and lateral accretion played a part in the formation of these units; one process might have been active in one part of the basin of deposition while the other was active in some other part. In areas where the Shinarump member is thin, the lateral accretion hypothesis is particularly inviting. Nonetheless, the available evidence seems to support the idea that the sandstone and conglomerate layers in the Chinle formation formed on top of an earlier formed peneplain during distinct periods of aggradation and are not typical deposits of lateral accretion.

The deposition of the thin widespread sandstone and conglomerate units in the Chinle formation may have started as the result of both an increase in load of the streams and a rise in base level. Uplift of the source area probably supplied a greater load to the streams. Possibly almost synchronously with uplift of the source area, the basin of deposition was slightly downwarped (see section on Tectonic Control of Deposition) and the base level thus effectively raised. Deposition was thus started as a response to a change in the regimen of the stream and an attempt by the stream to restore equilibrium conditions.

Slope of depositional surface.—An estimate of the depositional slope of the lower part of the Chinle formation, particularly the Shinarump and Moss Back members, can be made by comparisons with depositional slopes of Recent or late Tertiary deposits and with slopes of modern rivers and pediment or peneplain surfaces (Table 2). Some of the depositional plains of the Late Tertiary probably have been uplifted and the slopes are higher than the original slopes. The figure for the Gilbert Peak surface is measured near the Uinta Mountains and is much too high for a depositional surface away from a source area.

Table 2.--Depositional slopes of Late Tertiary or Recent deposits and gradients of modern rivers

Locality	Slope (feet per mile)	Source of formation
Gilbert Peak surface (Miocene), Wyo., 35 miles from crest	55	Bradley, 1936
High Plains of Kansas (Late Tertiary)	10	Frye and Leonard, 1952
Flaxville gravel (Miocene or Pliocene), eastern Montana	8.9	Alden, 1932
Upland gravels of Maryland (Pliocene ?)	5	Schlee, 1957
Aggrading course of middle Rio Grande River, New Mexico	4.4	Happ, 1948
Streams on Great Plains, Kansas and Nebraska	1.5 to 6*	Leopold and Wolman, 1957
Colorado River delta, California	2.2	Sykes, 1937
Hwang Ho delta, China	1.3	Dunbar and Rogers, 1957
Klarälven River, Sweden	0.9	Sundborg, 1956
Top surface of "graveliferous" deposits (Recent) of Mississippi River	0.75	Fisk, 1947
Valley slope of Mississippi River	0.55	Fisk, 1944 (Plate 18)
Yukon River, Alaska	0.47*	Leopold and Wolman, 1957
Ganges River, India	0.35*	Leopold and Wolman, 1957

*Measured on meandering course of river; valley slope is slightly greater.

On the other hand, the recent deposits of rivers like the Mississippi are finer grained than those of the Chinle formation, and the gradients are probably lower than that of the Chinle formation. The slope of "graveliferous" deposits of the Mississippi River (0.75 feet per mile) and of the aggrading course of the middle Rio Grande River (4.4 feet per mile) are in the middle range of those listed and probably the most significant. They are in the general range of the depositional slopes of deltas and of streams flowing on the Great Plains. As a guess, the depositional slope of the Chinle formation was probably greater than 10 feet per mile close to the high source areas, but on the order of 4 feet per mile 50 miles from the source. Farther out from the source area the gradient probably slowly decreased to about 1 foot per mile, at a distance of 300 miles or more from the source area.

At first glance, it is difficult to imagine streams on a plain of such low gradient having the competence to carry gravel. Pebbles one or two inches in diameter are common throughout most of the depositional area of the Shinarump and Moss Back members. These pebbles are not locally derived but were transported long distances across the plain from the source areas. Streams on low slopes, however, apparently have the competence to move quite coarse debris. The Kansas River at Lecompton, Kansas, has a gradient of 2 feet per mile (Leopold and Wolman, 1957, Appendix H) and had a flood velocity in 1903 of 8.05 feet per second or 245 cm per second (Murphy, 1904, p. 73). At least two other floods during recorded history have been greater than that of 1903 (Water Resource Division, 1952). A velocity of 8.05 feet per second is sufficient to transport fragments about 2 inches in diameter according to Hjulstrom (1935, fig. 18; also see Bucher, 1919, Table 1). Further, the "graveliferous" deposits of the Mississippi River, which apparently had a depositional slope of 0.75 feet per mile, contains abundant gravel as large as 3 inches in diameter (Fisk, 1944; 1947). The writer has observed many one inch pebbles on the Sacramento River where it has a gradient of 1.6 feet per mile. Apparently rather coarse gravel can be transported in a stream on a low gradient.

Origin of ripple-laminated sandstone and of rib-and-furrow structures

Very fine-grained ripple-laminated sandstone units are common at the top of the Shinarump member, in the Monitor Butte and lower red members, and locally elsewhere in the lower part of the Chinle formation. These units locally contain parallel ripples, both of the symmetrical and asymmetrical types, but the dominant structure is rib-and-furrow markings that are thought to consist of layer upon layer of cusp ripple marks.

These ripple-laminated strata are interpreted to be a special type of stream channel deposit, formed in shallow, flat-bottomed, probably low velocity streams with a silt or fine sand bottom. In the middle portion of the Colorado River delta, the dominant depositional type is ripple-laminated fine sand and mud (McKee, 1939). Cusp ripple marks are common in this portion of the delta (McKee, 1939, Plate 1A and B). Many of the streams in the Colorado Plateau region contain broad surfaces of cusp ripple marks. These streams are generally flat-bottomed, and the ripple-marked sediments are composed of silt and very fine sand.

Cusp ripple marks are the characteristic type of ripple mark developed in stream deposits. Apparently they develop in response to fluctuating and variable currents in comparison to the parallel ripple marks which develop in smooth and steady currents. McKee (1957, p.142) noted that cusp ripple marks on the tidal flats at Cholla Bay occur largely in the channels of concentrated water movement, whereas parallel ripples occur elsewhere.

Rib-and-furrow structures, as mentioned previously, are interpreted to be the product of deposition of layer upon layer of cusp ripple marks. Their origin is believed to be analogous to that of trough cross-strata formed by downstream migration of arcuate "sand waves" or bars. The cusp ripples advance downstream by deposition on their lee sides. Layer upon layer of cusp-shaped laminae are left forming the rib-and-furrow structure.

Origin of structureless or horizontally stratified claystone and clayey siltstone

Structureless or horizontally stratified claystone and clayey siltstone constitute a large part of the Petrified Forest member and a lesser part of the Monitor Butte and related members. As described previously, they are composed of montmorillonitic clay and contain relics of probable rounded pumice fragments, many as large as 3 mm in diameter, and some silty material composed of quartz and feldspar.

The clay in the rock is derived from the devitrification of volcanic ash and pumice. Limestone nodules are common in the claystone and clayey siltstone.

The claystone and clayey siltstone could have been deposited in one of two ways, (1) from ash falls or (2) from ash and pumice carried in streams and deposited in lakes or river basins. The ash fall hypothesis is apparently ruled out for a number of reasons. First, the rock itself contains large fragments of probable pumice fragments that could not have been transported in the air for the necessary distances. Second, wind directions in rocks of Permian, Triassic, and Jurassic age (Poole and Williams, 1956) on the Colorado Plateau are consistently southerly, a direction directly opposite from that necessary to transport debris from the presumed source area in southern Arizona and adjacent states (see section on Paleogeography). Third and last, the amount of claystone and clayey siltstone is so great and the distances so far that ash falls seem to be an inadequate method of transportation. In the ash and pumice fall associated with the explosion that produced Crater Lake in Oregon, 85 percent of the fall, which consisted of 3.5 cubic miles of material, was within 75 miles of the volcano (Williams, 1942). If the deposits in the Chinle formation are ash falls, most of the material would have been transported more than 75 miles, and large ash falls at distances greater than 75 miles seem unlikely.

As ash falls seem unlikely, transport of the glassy volcanic material in streams and deposition of it in river basins or lakes is indicated. Quiet water deposition is suggested by the even horizontal bedding of the claystone and clayey siltstone. The presence of limestone nodules is also consistent with the idea of subaqueous deposition. Most lake deposits consist dominantly of clay and fine silt (Emery, 1954; Hunt and others, 1953; Hamilton, 1951; Pofte, 1957; Hough, 1958), so that the occurrence of large fragments of volcanic debris in these possible lake deposits in the Chinle formation seems anomalous. The large fragments, however, can be explained as pieces of buoyant pumice that floated out into the lake, and then were waterlogged and sank. Such large pumice fragments in Tertiary lake deposits in California have been so explained (Bateman, 1953; Chesterman, 1956). The grains of quartz and feldspar in the claystone and clayey siltstone are mostly of fine silt size; this grain size is in accord with that expected in a lake deposit. Probably much of the original glassy material in these rocks was also of silt size.

Origin of contorted strata

A characteristic feature of many of the cross-strata in the Shinarump and Moss Back members is a peculiar deformation in the upper part of the cross-stratified set, in which the cross-strata are drawn downstream, so that individual laminae in cross-section have the shape of a U laid on its side (fig. 25). This deformation probably is caused by a flowage of sand shortly after deposition of the cross-strata. Observations in modern streams show that the top few feet of the sand bed often has the consistency of "quicksand" and that it commonly moves downstream as a viscous "fluid" mass. This motion of the top few feet of the sand bed has been observed by "a civil engineer who descended in a diving bell to the bottom of the Mississippi at a point where the depth was 65 feet and the bottom of sand. Stepping to the bed, he sank into it about 3 feet, and then thrusting his arm into the yielding mass, could feel its flowing motion to a depth of 2 feet, the velocity diminishing downward" (Gilbert, 1914, p. 156). A similar motion down to 3 m was observed on the gravel beds of the upper part of the Rhine and one of its small tributary streams (Bucher, 1919, p. 169-170). This motion would very likely cause the laminae to deform. The diminishing downward velocity could account for the observed diminishing "bending" of the strata downward.

A characteristic feature of the Monitor Butte and lower red members are contorted strata consisting of intricately folded layers (fig. 27 and 28) or of irregular blocks of rippled-marked sandstone lying at almost any conceivable strike and dip. Some of the contortions are similar to those in recent landslide blocks which also occur in strata of the lower part of the Chinle formation, but many of the contortions are primary features caused by slumping during deposition of the Chinle formation. The best evidence of primary slumping is that the deformed strata are truncated at the top and overlain, with a sedimentary contact, by undeformed horizontal beds. Further, many of the contorted layers occur in flat country where recent landslides are unlikely to occur.

The most likely explanation of the contorted strata is that they are large landslips into stream channels from adjacent flat lands.

Stream channels, which may have been locally 40 to 50 feet deep, extended below the surface of the plain and caused instability in the mass of fine-grained sediment on the adjoining flat lands. At some critical time, perhaps when the adjoining land area was saturated with water, sediment slumped and "flowed" from the flat lands and filled the channel. A "clay plug" in the northern Monument Valley area (fig. 33) which contains contorted strata may be an example of such a slump.

Dawson (1899, also in Sharpe, 1938) has described a landslip in Quebec, Canada that may be very similar to those of the Chinle formation. The landslip occurred on a small stream occupying a shallow valley about 1000 feet wide with sloping banks 25 to 35 feet high. The surrounding country is flat and practically level, and composed of Pleistocene clay deposits. The landslip covered an area of 86 acres with greatest width of 1,700 feet and greatest length of 3,000 feet. The maximum difference in elevation between the original land surface at the head of the landslip and the level of the stream is 20 to 25 feet. The mass of clay filled the valley adjacent to the slide and "flowed" down the valley for nearly two miles. Although the slumping in the Chinle formation may not have been exactly like that in Quebec, the Quebec landslide indicates that such slumping can occur in almost flat country similar to the supposed depositional plain of the Chinle formation.

Paleogeography

The depositional plain of the lower part of the Chinle formation sloped northward and northwestward away from a broad source area in southern Arizona and adjacent states--the Mogollon highland, a name proposed by Harshbarger and others (1957). In addition, a highland in western Colorado and northernmost New Mexico--the Uncompahgre highland, a name long in use--supplied some detritus to the formation. The Mogollon highland was dominantly a volcanic terrain, but also contained cherty limestone, sandstone, metasedimentary rocks and probably granitic rocks. The Uncompahgre highland was an igneous and metamorphic terrain.

Location of source areas

The location of the source areas of the lower part of the Chinle formation is indicated by the direction of stream flow, by the distribution of coarse gravel and of unusual lithologic pebble types, and by stratigraphic hiatuses in the supposed source areas.

A southern source area, the Mogollon highland (fig. 31), is clearly indicated by the direction of stream flow in the lower part of the Chinle formation. Stream directions, as determined from orientation of cross-strata, are mostly northwest to north-northeast throughout the lower part of the Chinle formation (fig. 31). The north to north-northeast flowing streams indicate a highland to the south.

The mean and maximum sizes of pebbles, cobbles, and boulders in most of the Shinarump member decrease gradually northward, in the direction of stream flow, from the presumed southern source area (Thordarson and Albee, in preparation). The mean sizes of gravel in the St. John, Holbrook, and southern Defiance Uplift areas in east-central Arizona is 21 mm and the maximum sizes range up to 284 mm. In the Circle Cliffs area 200 miles to the north, the mean size is 15 mm and the maximum sizes range up to only 63 mm. The mean and maximum sizes in the Sonsela sandstone bed show a similar, although less pronounced, decrease northward. This decrease in size northward is interpreted to be due to decreasing competence of the streams northward away from the southern source area.

Volcanic pebbles occur almost exclusively along the southern margin of the basin of deposition, indicating further the likelihood of a source area to the south.

Additional evidence of a southern source is that erosion is known to have taken place during early Mesozoic time in the area of the supposed source in southern Arizona and adjacent regions. In southern Arizona, southwestern New Mexico, and in Sonora, Mexico, rocks of Cretaceous age rest unconformably on Paleozoic or Precambrian rocks (Darton, 1925, p. 135; Butler and Wilson, 1938, p. 13; Ross, 1925; Gilluly, 1956, p. 123). The hiatus in southern Arizona represents all of Triassic and Jurassic time. Clearly then, southern Arizona and adjacent regions were land areas during early Mesozoic time.

A source area to the northeast of the main depositional area, the Uncompahgre highland, contributed a small amount of sediment to the lower part of the Chinle formation. The Shinarump member in the

White Canyon-Elk Ridge area, the Aqua Zarca sandstone member in north-central New Mexico, and possibly some other material in the lower part of the Chinle formation along its northeastern limit probably had a source in the Uncompahgre highland. This source area is indicated by stream directions as well as by other evidence. The stream directions in the Shinarump member in the White Canyon-Elk Ridge area are to the west indicating a highland somewhere to the east. Stream directions in the Aqua Zarca sandstone member in north-central New Mexico are to the south and southwest indicating a highland to the north or northeast. Pebble sizes decrease to the west or southwest in both these units (Thordarson and Albee, in preparation). In addition, the pebbles in the Shinarump member in the White Canyon-Elk Ridge area are, on the average, over 80 percent quartz whereas in other areas quartz is generally nearly or less than 50 percent (Thordarson and Albee, in preparation). Apparently the different source area provided a different type of pebble assemblage than in other areas. The Aqua Zarca sandstone member contains large quartzite and quartzite conglomerate pebbles and cobbles which may have been derived from known metamorphic rocks to the north (Gabelman and Brown, 1955). The Uncompahgre highland is easily delimited in Utah, Colorado, and New Mexico as an area where Triassic or younger strata rest on Precambrian rocks.

Terrain of source areas

The types of rocks exposed in the source areas is indicated by the type of detrital material in the formation. The terrain also can be determined directly from study of the type of rocks present in the supposed source areas themselves. Where possible, an attempt has been made to locate rocks within the presumed source area that could have supplied the rocks that occur as pebbles and other detrital material in the formation.

The Mogollon highland was predominantly a volcanic terrain but also contained cherty limestone or dolomite, sandstone, metasedimentary rocks, and probably granitic rocks. Volcanic debris in the lower part of the Chinle formation, as described before, consists of volcanic pebbles, pumice fragments, and montmorillonite probably derived from the devitrification of volcanic debris. The volcanic pebbles are vitric and crystal tuffs and vitrophyres and are of rhyolitic composition in the Shinarump member and of an intermediate composition, perhaps quartz latite or dacite, in the Petrified Forest member. The dominant clay in the lower part of the Chinle formation is montmorillonitic, and judging from the amount of this clay and other volcanic debris, at least half, and possibly much more, of the lower part of the Chinle formation was derived from volcanic source rocks.

The nearest possible Triassic igneous activity to the Colorado Plateau is in southern Arizona. The likelihood of this source is supported by the occurrence of volcanic pebbles only near the southern margin of the Chinle formation. The best documented occurrence of Mesozoic igneous activity in southern Arizona is in central Cochise County in southeastern Arizona. Gilluly (1956) recognized granite, granite porphyry, quartz monzonite, monzonite porphyry, and associated alaskite that intrude Paleozoic formations and are unconformably overlain by sedimentary rocks of Early Cretaceous age. These intrusive rocks, therefore, could be of Triassic or Jurassic age. Similar igneous rocks that intrude formations of Paleozoic age but not those of Cretaceous age occur in mountains in Santa Cruz and Pima Counties, that adjoin Cochise County on the west (Schrader and Hill, 1915, p. 57-70). Other intrusive rocks that have been tentatively assigned to the Mesozoic occur widely in southern Arizona (Schrader, 1909; Bancroft, 1911; Jones, 1915, 1916; Ransome, 1919; Ross, 1922; Bryan, 1923; Darton, 1925). These intrusive rocks, however, cannot be dated with any assurance and could be most any age. Anderson and others (1955) report intrusive and extrusive rocks of Late Cretaceous (?) or early Tertiary (?) age in the Bagdad area in Yavapai County in central Arizona. Here again, these rocks cannot be dated accurately but could possibly be of Triassic age. The rocks described by Anderson and others (1955) include the Grayback Mountain rhyolite tuff formed by pelean eruptions. This rock is similar in texture and composition to the rhyolitic tuff pebbles occurring in the Chinle formation from 125 to 150 miles to the northwest and east.

The presence of fossiliferous chert pebbles is definite evidence that marine cherty limestone or dolomite occurred in the southern source area. The fossils include fusulinids, brachiopods, bryozoa and, to a lesser extent, gastropods, pelecypods, corals, algae, crinoidal material, sponges, ostracods, and echinoid spines (Thordarson and Albee, in preparation; McKee, 1936). Fusulinids include Parafusulina (Leonard and

Word age, Permian), which is one of the most abundant fossils, and Schwagerina (middle and upper Wolfcamp and Leonard age, Permian). The most common brachiopod is the productid Dictyoclostus (Mississippian-Permian), which is particularly common in the Kaibab limestone and equivalent formations of Leonard age on and adjacent to the Colorado Plateau. The most common bryozoa are Fenestella (Ordovician-Permian) and Rhabdomeson (Carboniferous and Permian). Most of the bryozoa are types that occur in the Kaibab limestone; a few bryozoa may be from an upper Mississippian rock. The algae include Mizzia sp. which is known only from the Permian.

The large majority of the fossils in the pebbles are from rocks of Permian age; a few may be from rocks of Pennsylvanian or Mississippian age. Most of the brachiopods and bryozoa are types that occur in the Kaibab limestone (Thordarson and Albee, in preparation; McKee, 1936). Fusulinids, on the other hand, are not known from the Kaibab limestone (L. C. Henbest, written communication), and, therefore, the source of the fusulinid-bearing pebbles must have been either a fusulinid-bearing facies of the Kaibab limestone in areas where it has since been eroded, or strata approximately equivalent to the Kaibab limestone in other areas. Limestone and dolomite of approximately the same age as the Kaibab limestone probably originally were present throughout southern Arizona, southwestern New Mexico, and adjacent regions (McKee, 1947, 1951). These approximate equivalents of the Kaibab limestone of Permian age are (1) limestone of Permian age in southeastern California (Thompson and Hazzard, 1946) and southwestern

Arizona (McKee, 1947), (2) the Colina limestone, Epitah dolomite, and Concho limestone of Wolfcamp, Leonard, and possibly Word age in southeastern Arizona (Gilluly, Cooper, and Williams, 1954), (3) the San Andres formation of Guadalupe and perhaps Leonard age (McKee and others, in preparation) of southwestern New Mexico, and (4) the Wolfcamp formation, Leonard formation, Bone Springs limestone, Hueco formation, and Word formation of Wolfcamp, Leonard, and Word age in the Trans-Pecos region of Texas (King, 1934). Fusulinids are reported in these rocks in southeastern California, in southeastern Arizona, and in the Trans-Pecos region of Texas; the fusulinids include <u>Schwagerina</u> and <u>Parafusulina</u>. Roth (1943) has identified species of <u>Schwagerina</u> in pebbles of the Dockum group (a probable equivalent of the lower part of the Chinle formation) in Texas. These species are know today only in the Hueco and Wolfcamp formations of the Trans-Pecos region. Other more westerly fusulinid-bearing limestone units, however, probably contributed sediment to the Chinle formation.

Quartz and quartzite pebbles may have been derived directly from quartz veins and quartzite formations in the source areas or from conglomerate layers in sedimentary rocks in the source areas. The Precambrian rocks of southern Arizona, composed of gneiss, schist, granite, and various types of metasedimentary rocks including quartzite (Darton, 1925) are a likely source. Some of the quartz and quartzite pebbles could have been derived from conglomerate layers which occur in the Precambrian and Cambrian rocks (Darton, 1925), and to a lesser extent in higher Paleozoic formations in Arizona. Stoyanow (1942) mentions conglomerate layers of Pennsylvanian age containing boulders of Precambrian quartzite in Gila County, Arizona.

The occurrence of microcline in the lower part of the Chinle formation indicates that some material came from a granitic source rock in the southern source area, or less likely from reworking of microcline-bearing sedimentary rocks in the source area.

The occurrence of quartz overgrowth on some quartz grains in the lower part of the Chinle formation indicates that some quartz-bearing sandstone units were exposed in the source area.

The Uncompahgre highland was a Precambrian terrain containing granitic and metasedimentary rocks including granite, gneiss, schist, and quartzite. These rocks are exposed in the rejuvenated parts of this highland (Shoemaker, 1956; Larsen and Cross, 1956; and Montgomery, 1956) and in many areas directly underlie the Chinle formation. Locally around the flanks of the highland Paleozoic rocks were exposed during Late Triassic time.

Tectonic control of deposition

Fluvial and lacustrine sediments, similar to those of the lower part of the Chinle formation, commonly occur in basins at the foot of a large mountain range (fig. 47). The Tertiary basin of the Great Plains is filled with fluvial and lacustrine sandstone, siltstone, and claystone and lies directly east of the Rocky Mountains. The alluvial deposits of the Indo-Gangetic plain of India are in a basin directly south of the high Himalaya Mountains (Wadia, 1953; Krishnan, 1956). The lower part of the Chinle formation, as outlined previously, was

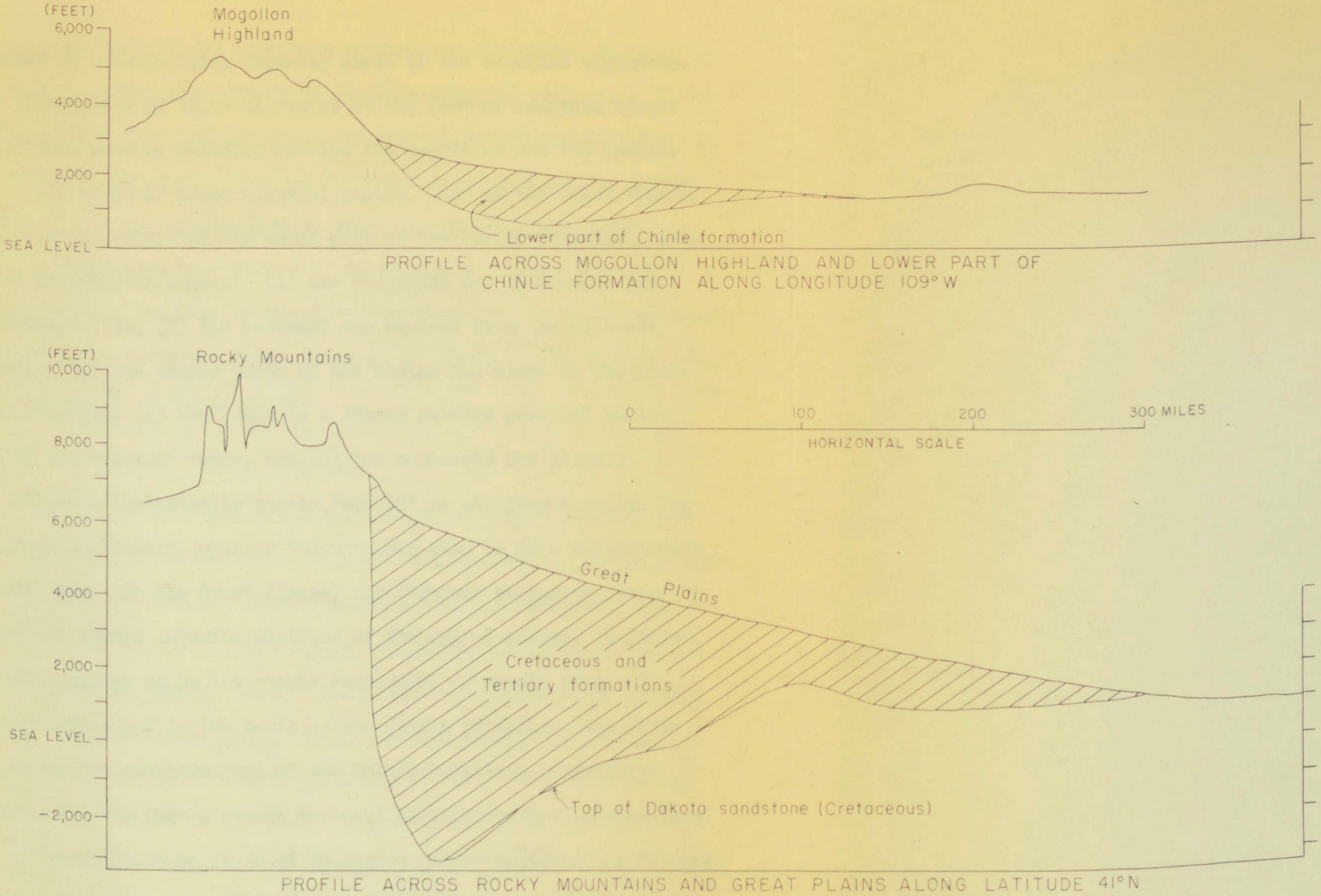


FIGURE 47.— Comparison of basin of Lower Part of Chinle formation and that of the Great Plains.

deposited in a basin lying directly north of the Mogollon highlands. These occurrences of alluvial basins at the foot of mountain ranges suggest some genetic relation between the mountains and the basins.

All three of these alluvial basins, that of the Great Plains, Indo-Gangetic plain, and the lower part of the Chinle formation, have several features in common: (1) the sediments are of fluvial and lacustrine origin, (2) the sediment was derived from the adjacent mountains, (3) the deeper parts of the basins lie close to the mountains (fig. 47), (4) the basin is a linear feature parallel to the trend of the mountain range, and (5) the sediments are largely deposited on a tectonically stable "shield" or platform area, or lap up on such a platform, whereas the mountain area is in a tectonically unstable area. On the Great Plains, the Tertiary sediments cover part of the stable interior platform of the United States; in India, the sediments lap up on the stable Peninsula, or shield area, of southern India; and in the basin of the Chinle formation, the sediments cover the southern part of the Colorado Plateau, a geologic division that has been a stable tectonic block since Precambrian time.

These basins in front of mountains can be explained as forming by deep seated thrusting of the mountains toward the stable platform area and concomitant down-buckling of the edge of the platform. In this respect, the mountain range and basin corresponds respectively to the island arcs and fore-deeps of the continental margins which have been interpreted by some geologists as being produced by deep seated thrusting of the continent seaward, and overriding and downwarping of the crust in front of the thrust to produce the foredeep. Suess (in Krishnan, 1956) has considered the Indo-Gangetic basin as a "foredeep" formed in front of the resistant Precambrian shield of India as the rocks comprising the Himalaya Mountains were uplifted and thrust southward. If this explanation is true, the basin in the Chinle formation could have been formed by down-buckling associated with thrusting of the Mogollon highland northward toward the resistant block of the Colorado Plateau. Downwarping of the basin would, in this case, accompany uplift of the highland.

Climate

The fauna and flora provide, perhaps, the only reliable clue of the climate during deposition of the lower part of the Chinle formation. Water must have been ample, judging from the abundance of such aquatic animals as the pelecypod Unio, the amphibian Eupelor, and the reptile Machaeroprosopus. Plants such as the cycadophyte Macrotaeniopteris, the sphenopsida Neocalamites, and the gymnosperm Schilderia probably grew in swamps (Daugherty, 1941, p. 33). Other animals and plants probably lived on dry land. Daugherty (1941, p. 33) believed the climate to have been of the savanna type, a tropical or subtropical climate with ample rainfall and a distinct dry season. Daugherty based his interpretation on the types of plants that include large ferns and conifers, as well as swamp plants, all indicating ample rainfall. In addition, Daugherty considered that the climate had a distinct dry season because of the presence of growth layers (tree rings) in the large conifers, because of the modification of some of the plants for wind pollination, and because of the presence of bark-boring beetles, most of which today require dry fallen logs for hosts. Daugherty's interpretation of a savanna climate with ample rainfall and a distinct dry season agrees with the evidence from vertebrate remains which indicates large water courses and adjoining dry land areas.

Upper part of the Chinle formation

The upper part of the Chinle formation consists dominantly of reddish-brown horizontally bedded siltstone, and minor amounts of limestone, horizontally laminated, ripple-laminated, or wavy-stratified siltstone or sandstone, cross-stratified (fluvial) sandstone, irregularly bedded limestone pebble conglomerate, and cross-stratified (eolian) sandstone. This part of the formation was probably deposited mostly in a large shallow lake. Locally fluvial and eolian deposits formed. The main source areas were the Uncompahgre and Front Range highlands of Colorado, although some debris was supplied from the Mogollon highland of southern Arizona and adjacent regions.

Environment of deposition

The types of rocks and fossils that compose the upper part of the Chinle formation are used in reconstructing the environment of deposition. The fossils indicate mainly a fresh water environment of lakes and streams. Based on the rock types present, the upper part of the formation contains lake, stream, and eolian deposits. Much of the sediment may have been deposited in a subaqueous "delta" formed in a shallow lake.

<u>Fossil evidence</u>

The fauna and flora of the upper part of the Chinle formation is much impoverished as compared to that in the lower part of the formation. Some types of fossils that occur in the lower part of the formation do not occur in the upper part; the number of fossil individuals in the upper part is much less than in the lower part. The fossils that do occur in the upper part are, with a few exceptions, the same types as those that occur in the lower part. The decreased fauna and flora of the upper part has been interpreted as an indication of increased aridity (Daugherty, 1941, p. 28). Although the climate may have been, at times, more arid, the most likely explanation of the impoverished fauna and flora is poor conditions for preservation in the upper part of the formation.

Pelecypods and gastropods occurring in the upper part of the Chinle formation are fresh water forms. These fossils are the most abundant types occurring in the upper part of the formation. The pelecypod <u>Unio</u>, which occurs in limestone of the Owl Rock member and in siltstone, sandstone, and limestone pebble conglomerate elsewhere in the upper part of the formation, lives today in lakes and rivers, and probably has been a fresh water form throughout its geologic history. The gastropod <u>Triasamnicola</u>, which occurs in the limestone of the Owl Rock member, has been called a fresh water form by Yen and Reeside (1946).

Vertebrate remains, including fishes and reptiles, indicate aquatic and to a lesser extent dry land forms. Fish remains, which occur in the narrow belt of fluvial sandstone of the upper part of the formation extending from southwestern Colorado to central Utah, are apparently fresh water forms, based on their association with other fresh water fossils. Reptile remains include the phytosaur Machaeroprosopus, which is a fairly common fossil, and Typothorax and Coelophysis which are known from only one locality each. Machaeroprosopus was a crocodile-like form that lived in streams and along their banks (Camp, 1930; Colbert, 1947). Typothorax was a low-lying reptile that probably lived mostly on land (Colbert, 1950, p. 63). Coelophysis, which occurs in beds transitional between the lower and upper parts of the formation in north-central New Mexico, was probably an agile dry land form (Colbert, 1950). The reptile remains mostly occur in fluvial sandstone and conglomerate units, commonly in limestone pebble conglomerate.

Fossil plants are rare in the upper part of the Chinle formation indicating either that most of the basin was under water at this time or that conditions for preservation were poor, or both. Plant fossils in the upper part of the Chinle formation include only a few remains of cycads (Hills, 1880, p. 490), and of conifers and a possible palm tree (Brown, 1956). These fossils indicate that a few dry land areas existed during deposition of this part of the formation.

Most of the fossils in the upper part of the Chinle formation are aquatic fresh water forms and only a few are dry land or "upland" forms. Perhaps most of the sedimentary basin was under water at this time and land areas were minor. The lower part of the Chinle formation, in comparison, probably contained many dry land areas, judging from the common occurrence of "upland" vertebrate remains.

Origin of horizontally stratified siltstone and sandy siltstone

The dominant lithologic type in the upper part of the Chinle formation is pale reddish-brown horizontally stratified coarse siltstone and very fine-grained sandy siltstone. The siltstone and sandy siltstone occurs in poorly defined horizontal bed from less than a foot to 4 feet thick. No fossils are known from these rocks.

The horizontally stratified siltstone and sandy siltstone occur interstratified, in the Owl Rock member, with limestone that commonly contains fresh water fossils such as the pelecypod Unio. The limestone beds are considered to be lake deposits, and the horizontally stratified siltstone and sandy siltstone, because of their association with the limestone beds, may also be lake deposits.

The type of bedding in the siltstone and sandy siltstone is indicative of quiet water deposition as would occur in a lake. Some lake sediments contain delicately layered sediments (Bradley, 1929; Hough, 1958) whereas others contain indistinctly bedded or unlaminated sediments (Hunt and others, 1953; Hough, 1958) similar to that in the siltstone and sandy siltstone in the upper part of the Chinle formation.

The siltstone and sandy siltstone are, on the other hand, not typical lake deposits. Most lake deposits are extremely fine-grained. The grain size is generally from 1 to 5 microns in present day or Quaternary lake deposits (Emery, 1954; Hunt and others, 1953;

Hamilton, 1951; Hofte, 1957; Hough, 1958) whereas the grain size in the siltstone and sandy siltstone of the upper part of the Chinle formation ranges from about 40 to 80 microns (Harshbarger and others, 1957; Cadigan, 1957a). The grain size of the material in the siltstone and sandy siltstone is approximately that which is carried in suspension in streams (Sykes, 1937; Colby and others, 1953; Fisk and McFarlan, 1954; Johnson, 1921). In addition, most lake bottoms contain abundant decaying vegetation that produces reducing bottom conditions (Hutchinson, G. E., 1957). Bottom sediments, therefore, contain ferrous iron instead of ferric iron (Twenhofel, 1926), and the colors of the sediments in this reduced state are gray, greenish gray, or black (Hough, 1958; Hunt and others, 1953; Bradley, 1929) and not red. The siltstone and sandstone of the upper part of the Chinle formation, in comparison, contains ferric iron in the mineral hematite, and the color of these sediments is red because of the presence of this hematite pigment. No organic material is present. The hematite in the sediments and lack of organic material indicate oxidizing conditions during deposition, in contrast to reducing conditions in most lakes. Red sediments, however, do occur rarely in lake sediments. Hough (1958, p. 69-75) describes red clay containing hematite in a core of the sediments of Lake Michigan from 350 to 1050 cm below the surface; the overlying layer is gray.

If the sandstone and sandy siltstone are not typical lake deposits, how then did they form? This question cannot be answered with any assurance, but the following hypothesis is presented as a possibility. The lake or lakes in which the sediments were deposited may have been extremely shallow, and the amount of water entering it large. The deposits of fluvial sandstone that extend in a narrow elongate belt from southwestern Colorado to central Utah are probably the deposits of the largest river entering this lake. The deposits of this river extended well out into the lake, in a manner similar to the present subdelta of the Mississippi River which extends about 65 miles out into the Gulf of Mexico and is only about 15 miles wide. The narrow elongate belt is also very similar to the "Red Bedford Delta" described by Pepper and others (1954) in the Mississippian rocks of Ohio. The siltstone and sandy siltstone may have been deposited as a sort of subaqueous delta deposit extending away from the delta or subdelta of this major river. The subaqueous delta, instead of being composed of bottomset, foreset, and topset beds, consisted almost entirely of bottomset beds; the lake was too shallow for the development of foreset and topset beds. The idea of a deltaic origin of the siltstone and sandy siltstone is supported by the grain size of these rocks which is in the range of that carried in suspension by streams and that commonly occurring on deltas (Johnson, 1921; Fisk and McFarlan, 1954).

Oxidizing conditions would probably exist in a lake like that hypothesized because of its shallow depth and because of frequent inflow of aerated river water. Thus bottom conditions would be oxidizing, and vegetation would decay rapidly. Bottom faunas would be scarce because of frequent additions of turbid waters, and their remains even scarcer because of easy decay in the oxidizing waters.

The siltstone and sandy siltstone of the Chinle formation may have been deposited in a manner similar to the subaqueous part of the Hwang Ho delta of China which is built out into the Yellow sea, except that in the Chinle formation the water body was fresh and not marine. The subaqueous plain of the Hwang Ho delta is flat and of a very low gradient; water depths of 30 feet occur 80 or 90 miles off shore (Kuelegan and Krumbein, 1949; Dunbar and Rogers, 1957, p. 83-84). The delta is composed of silt and sandy silt that is carried as suspended sediment in the Hwang Ho river. The suspended sediment is carried, presumably, for long distances out to sea across the flat subaqueous part of the delta before settling out. The shore has no distinguishing features, such as beach sands, because the gradient of the sea is so low that the energy of the waves is dissipated before reaching the shore and breakers do not form. The Hwang Ho delta, therefore, is forming in the manner hypothesized for the deposits in the upper part of the Chinle formation; a large silt-laden river is emptying into an extremely shallow body of water.

Origin of limestone

Limestone occurs in thin to thick horizontal beds interstratified with reddish-brown horizontally stratified siltstone and sandy siltstone in the Owl Rock member. The limestone in places is a calcite and dolomite replacement of water laid glassy volcanic sandstone composed originally of subrounded to rounded grains of porphyritic volcanic rock and pumice set in a finer grained matrix. Other limestone is calcareous arkosic or feldspathic siltstone in which the carbonate cement mineral, which fills interstitial spaces, constitutes a large part of the rock, or in other rocks occurs only as irregular patches. Some of the limestone beds may have formed by the growth and coalescence of limestone nodules; all gradations occur from layers containing a few scattered limestone nodules to layers containing a tight coalesced mass of nodules.

The limestone is considered to have been deposited in lakes, based on the contained fresh water fauna including the pelecypod Unio. The formation of the limestone is not by the usual methods of precipitation of carbonate minerals or of accumulation of calcareous organic debris. Apparently the limestone formed in a lake which was nearly saturated with carbonate minerals; during a diagenetic process the carbonate minerals in some places replaced the preexisting tuff beds or in other places filled the interstitial spaces in siltstone. Perhaps only rarely did the carbonate minerals precipitate directly to form a pure or nearly pure limestone bed.

Origin of horizontally laminated, ripple-laminated, and wavy-stratified siltstone and sandstone

Horizontally laminated, ripple-laminated, and wavy-stratified siltstone and sandstone occur as thin to very thick beds interstratified with the horizontally stratified siltstone and sandy siltstone, and to a lesser extent with other lithologic types in the upper part of the Chinle formation. Both parallel, generally asymmetrical, and cusp ripple marks occur. Some beds are evenly laminated; in other beds the laminae or thin beds exhibit a vague waviness on their stratification planes. Mudcracked surfaces, worm borings, and raindrop impressions occur locally.

These siltstone and sandstone beds probably were deposited mostly in shallow bodies of water, perhaps in the subaqueous near-shore part of a delta, and locally in streams. Current action was marked as is indicated by asymmetrical parallel ripple marks and cusp ripple marks. The wavy-stratified rocks may also have been produced by current action; the wavy-stratification may be disrupted ripple marks. The even horizontally laminated strata probably was formed by reworking of sediments in shallow water by currents. The cusp ripple-laminated strata probably formed in shallow streams with a silt or fine sand bottom much in the same manner that the cusp ripple-marked strata in the lower part of the formation formed. The presence of mudcracked surfaces and raindrop impressions indicate that at times the sediments were exposed to the air.

As outlined before, the upper part of the Chinle formation may have formed as a sort of subaqueous delta in a large shallow lake. The horizontally laminated, ripple-laminated, and wavy-stratified siltstone and sandstone may have formed in places where the surface of the delta was covered by only a thin layer of water or at times when the delta surface was dry land. Current action in the shallow water led to the development of evenly laminated, ripple-laminated, and wavy-stratified layers. Some of the ripple-marked layers, particularly those with cusp ripples, may have formed in streams flowing across the top of the deltaic plain.

Origin of limestone and siltstone pebble conglomerate and calcarenite

Limestone and siltstone pebble conglomerate and their finer grained equivalents, calcarenites, consist of irregular lenses, generally from 0.5 to 3 feet thick, composed of pebbles, granules, or coarse to very coarse grains of limestone, silty limestone, and siltstone set in a limy and silty matrix. Reptile remains are common in these lenses.

The limestone and siltstone pebble conglomerate and calcarenite are composed of material of local derivation. They probably represent reworking of underlying layers by stream action at times when the delta of the upper part of the Chinle formation was a subaerial plain. Johnson (1921, p. 33) has described the breaking up of hard silt layers by stream action on the delta of the Fraser River, Canada, into small masses which are then rolled along the bottom of the stream and become subangular to rounded fragments. The original hard silt layers are older deposits of the same delta. This breaking up and rounding of fragments noted by Johnson may be similar to the formation of limestone and siltstone pebble conglomerate and calcarenite in the Chinle formation.

Origin of trough cross-stratified sandstones

Trough cross-stratified sandstone is fairly common in the upper part of the Chinle formation, but has a spotty distribution. It forms the main part of the Black ledge and Hite bed, both thin widespread units, in southeastern Utah and, in addition, is the dominant lithologic type in the narrow elongate belt extending from southwestern Colorado to central Utah and in some other areas of the Colorado Plateau. The trough cross-stratified sandstone represents stream deposits, based on the type of cross-strata, on the types of contained fossils, and on the presence of channels in some areas. The Black ledge and Hite bed are probably braided stream deposits similar to the Shinarump and Moss Back members, although they are finer grained and contain more siltstone layers than the Shinarump and Moss Back members. The narrow elongate belt of sandstone extending from southwestern Colorado to central Utah contains irregular lenses or tabular-shaped sandstone units, which commonly fill channels and commonly are associated with ripple-laminated (commonly cusp ripples) or thin-bedded siltstone and sandstone. This type of deposit is more suggestive of a meandering stream deposit than of a braided stream deposit. As suggested previously, the river or rivers flowing in this belt may have been the main river around which the supposed delta of the upper part of the Chinle formation formed. The amount of cross-stratified and ripple-laminated siltstone and sandstone gradually decreases outward on both sides of this belt (fig. 43) suggesting gradually deepening water on both sides of the major river.

Origin of planar cross-stratified sandstone

Units of cross-stratified very fine- to fine-grained sandstone occur in the Church Rock member in a region along the Arizona-New Mexico state line in northeastern Arizona and northwestern New Mexico. The sandstone occurs in units from 10 to 50 feet thick containing cross-strata of the tabular planar type, although trough cross-strata also occur. The sandstone in these units is similar, or identical, to that in the Wingate sandstone of which some units are tongues. The Wingate sandstone is, therefore, considered in the discussion below along with the units in the Chinle formation.

These cross-stratified sandstone units are considered to be eolian sand dune deposits. This interpretation is based largely on the type of cross-strata which is similar to those in modern dunes. Modern dunes are composed dominantly of cross-laminae which have been deposited on the downwind side of the dune and dip generally 31 to 33 degrees downwind. The cross-strata in the Wingate sandstone and related units in the Chinle formation are similar to cross-strata in modern dunes illustrated by McKee (1957, fig. 20-22, Plate 3), Bagnold (1943, Plate 14b), Jones (1953, fig. 3), Thompson (1937), Huntington (1907, Plate 36, fig. 2; Plate 38, fig. 2), and Beadnell (1910, fig. 11). In addition, ripple marks (fig. 12) which are interpreted to be eolian in origin occur in the Wingate sandstone in west-central New Mexico. This sandstone is apparently composed of units laterally equivalent to the Church Rock member (see discussion under Stratigraphy section). The ripple marks have a ripple index of 1:45 which is characteristic of eolian ripple marks; aqueous ripple marks have indexes generally between 1:4 and 1:10 (Kindle and Bucher, 1926). Further evidence of eolian origin is the occurrence of small vertebrate trackways (Harshbarger and others, 1957, p. 12) in the Wingate sandstone in the Defiance uplift. These trackways occur on cross-strata and indicate that the cross-strata formed under subaerial conditions. Some of the fossils in the upper part of the Chinle formation indicate dry land conditions showing that a suitable environment existed, at least at times, for the development of sand dunes.

Tabular planar cross-strata are the dominant type of cross-strata in the eolian sandstone units in the Chinle formation, Wingate sandstone, and many other ancient and modern eolian units (Shrock, 1948, fig. 219; Bagnold, 1943, Plate 15; Gregory, 1950, fig. 41, 42, 48c and 119; Huntington, 1907, Plate 37). These cross-strata probably formed in transverse dunes much in the same way that tabular planar cross-strata develop in transverse bars in streams. Transverse dunes, as used in this report, are long linear ridges extending in a direction at right angles to the wind direction. They have a long gently dipping upwind slope and a short steeply dipping downwind slope, on which the sand rests at the angle of repose. Transverse dunes occur in some present day sand dune areas (see Cooper, 1958, p. 27, for listing of occurrences), but do not appear to be the dominant type of present day dune. The formation of the cross-strata is similar to that in transverse bars in streams; the sand is transported up the gentle back slope of the transverse dune and deposited on the steep downwind face. As the dune advances forward by deposition on its downwind face, a cross-stratified deposit is left behind.

Trough cross-strata are a secondary type of eolian cross-strata in the Chinle formation and Wingate sandstone but are fairly common in other ancient and modern eolian deposits (McKee, 1957, fig. 22; Huntington, 1907, Plate 36, fig. 2; Gregory, 1917, Plate 12A; Gregory, 1950a, fig. 48A and B; Almeida, 1953, fig. 5-8). These cross-strata probably form in barchan dunes much in the same way that trough cross-strata presumably develop in arcuate bars in streams. Barchan dunes are crescent-shaped features with the two arms of the crescent extending downwind. Barchan dunes or dune complexes are common in present day dune areas. In plan view, eolian trough cross-strata are arcuate with the curvature downwind, as are also the cross-strata in modern barchan dunes (McKee, 1957, fig. 20). The cross-strata are formed by transportation of the sand up the back side of the barchan dune and deposition on the downwind side. The cross-stratified deposit is left behind as the dune is built forward.

Transverse dunes and tabular planar cross-strata most likely develop in regions of steady winds that blow in approximately the same direction throughout the year. Barchan dunes and trough cross-strata, on the other hand, most likely develop in regions with gusty winds or winds with variable directions during the year. This explanation of the different conditions that form tabular planar and trough cross-strata is similar to the explanation of the conditions that produce the different types of ripple marks and aqueous cross-strata. Parallel ripple marks and both aqueous and eolian tabular planar cross-strata are formed under conditions of smooth and steady currents. Cusp ripple marks and both aqueous and eolian trough cross-strata form under conditions of fluctuating and variable currents.

Horizontal stratification planes are common in the Wingate sandstone. In addition, the tops of the eolian units in the Chinle formation are flat planes. These flat planes cut across the top surface of the dune deposits truncating the cross-strata. These planes commonly extend for a mile or more. In places, they are doubtless produced by subaqueous planation, as for example in the upper part of the Chinle formation where dune areas were covered by lakes. In other places, particularly in the Wingate sandstone and similar units, the flat surfaces may have been produced by wind erosion that leveled off the tops of the dunes and produced a level surface or sand sheet (Bagnold, 1943, p. 243). Such level areas (sand sheets) are particularly common in the sand dune region of the Libyan Desert (Bagnold, 1943).

The source of the sand in most widespread sand dune regions is upwind arenaceous sedimentary rocks. In the Libyan Desert, for example, the sand that forms the vast sand dune region is derived from Tertiary arenaceous deposits to the windward (northward) of the dune deposits (Beadnell, 1910, p. 382). Similarly, the sand that formed the eolian units in the Chinle formation was probably derived from upwind deposits of siltstone and sandstone in the upper part of the Chinle formation. Wind directions in the eolian units are generally to the southeast indicating a source of the sand to the northwest. Apparently at times when the lake or lakes of the upper part of the Chinle formation partly or entirely dried up, winds blowing to the southeast picked up sand and deposited it in a dune region in the southeastern part of the basin of deposition.

The eolian units show a cyclic pattern of deposition (fig. 48). The cycle starts with deposition of a thin to very thick horizontally

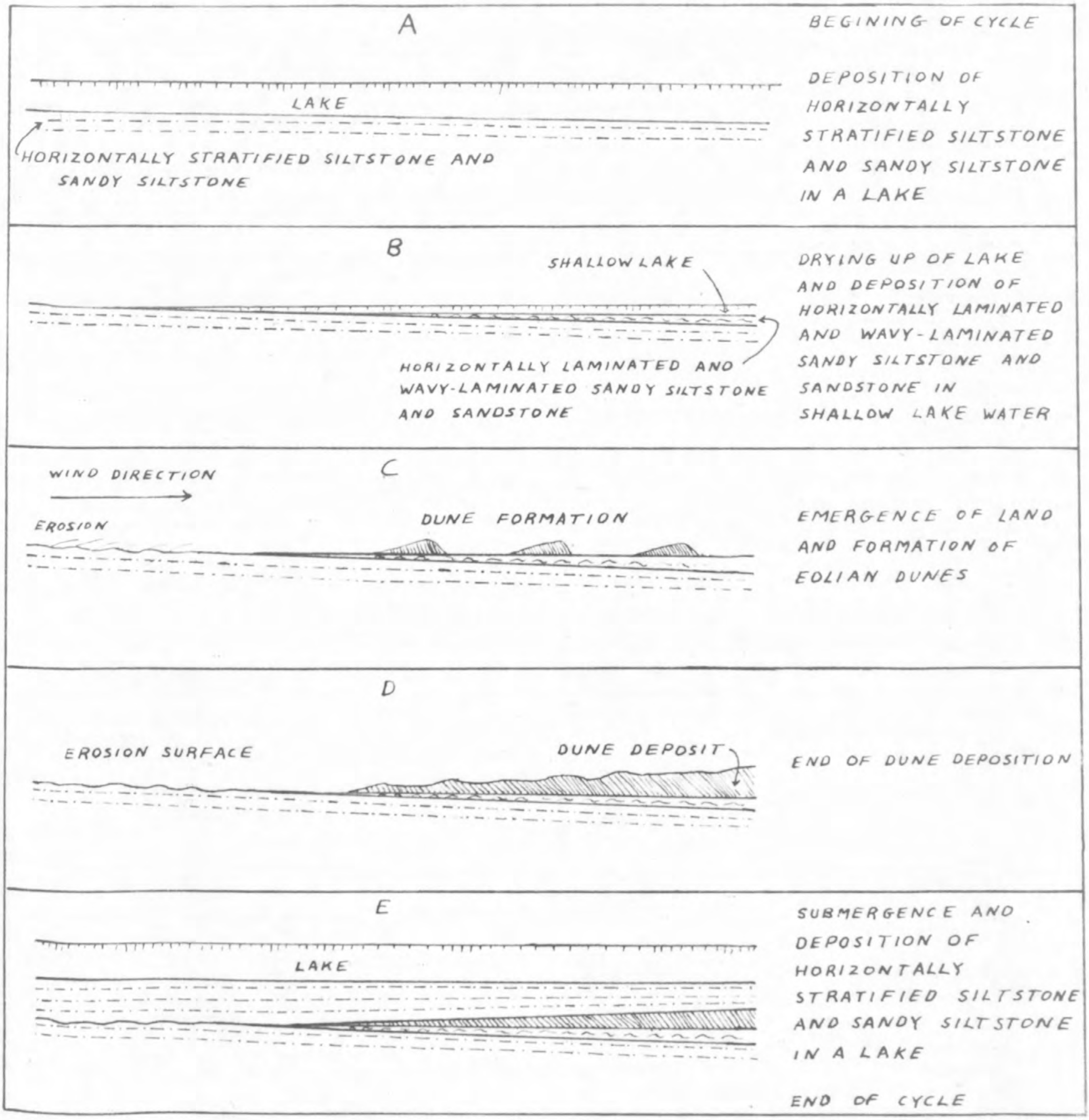


FIGURE 48.--DEVELOPMENT OF CYCLIC DEPOSITS IN CHINLE FORMATION

bedded siltstone and sandy siltstone unit which is commonly 50 to 100 feet or more thick. This is followed by deposition of a unit of horizontally laminated or wavy-laminated sandy siltstone and sandstone generally from 5 to 40 feet thick. The latter layer locally contains a few ripple-marked strata. Finally, an eolian unit is deposited. It ranges in thickness from 10 to 50 feet thick. The next cycle is then started by deposition of a unit of a thin to very thick horizontally bedded siltstone and sandy siltstone. At Chee Dodge in the Defiance Uplift, this cycle is repeated four times in the Church Rock member. This cycle is interpreted to be caused by the gradual drying up of a body of water, followed by dune formation, and finally by the return of the lake waters (fig. 48). The horizontally laminated layers below the eolian units are probably shallow water deposits formed in the dissipating waters of the lake.

Paleogeography

The main source areas of the upper part of the Chinle formation were the Uncompahgre and Front Range highlands of Colorado and adjacent regions and the Mogollon highland of southeastern Arizona and southwestern New Mexico. The Mogollon highland apparently supplied much less detritus than during deposition of the lower part of the formation. The Uncompahgre and Front Range highlands are composed of igneous and metasedimentary rocks; the Mogollon highland probably mostly of volcanic rocks. Perhaps a land area rose in northwestern Arizona and southwestern Utah at the close of Chinle deposition.

Location of source areas

The location of the source areas of the upper part of the Chinle formation is indicated by the direction of stream flow, by the distribution of clay types, by the presence of sandstone layers, and by stratigraphic breaks in the supposed source areas.

A source of sediment in the Uncompahgre highland of western Colorado and north-central New Mexico is indicated by the direction of stream flow (fig. 39). Stream directions are to the northwest in and near the narrow elongate belt containing abundant fluvial sandstone that extends from southwestern Colorado to central Utah. These stream directions indicate a highland to the southeast. In addition, the strata in the upper part of the Chinle formation in its eastern area of deposition contain chlorite as a minor constituent, whereas to the west they do not. The chlorite was, most likely, derived from erosion of Precambrian igneous and metamorphic rocks in the source area. Finally, the Uncompahgre highland is known to have been a land mass during much of late Paleozoic and early Mesozoic time (Heaton, 1933, 1950), and thus its existence during deposition of the upper part of the Chinle formation is likely.

The Front Range highland of central Colorado also contributed sediment to the upper part of the Chinle formation. Stream directions in the Gartra member of the Chinle formation are to the northwest, indicating a source area to the southeast, most likely in central Colorado. The Gartra member directly underlies typical red beds of the upper part of the formation. Chlorite also occurs along the eastern margin of deposition of the upper part of the formation near the Front Range, but not in areas farther to the west. The chlorite probably indicates closeby erosion of igneous and metamorphic rocks. In addition, the Front Range highland was a persistent land mass throughout Paleozoic and Mesozoic time (Lovering and Johnson, 1933, p. 372), and thus the presence of a highland here during deposition of the upper part of the formation is likely.

Some material was also derived from the Mogollon highland. This source is indicated by the presence of an increasing amount of montmorillonitic clay in the upper part of the formation toward the south. As outlined before, the montmorillonitic clay in the lower part of the formation was mostly derived from the devitrification of glassy volcanic detritus, and this volcanic material had a source in the Mogollon highland. In addition, fluvial sandstone units occur in the upper part of the formation near its southern limit along the Arizona-New Mexico state line (fig. 43) indicating a closeness to a source area. The amount of montmorillonitic clay in the upper part of the formation, however, is much less than in the lower part, indicating that the Mogollon highland was less extensive, and had less relief and volcanic activity during deposition of the upper part of the formation than during deposition of the lower part.

A land area may have formed in northwestern Arizona and southwestern Utah in the closing stages of deposition of the Chinle formation. Stream directions in the Hite bed, a fluvial unit at the top of the Chinle formation, are, based on four cross-strata studies, to the northeast. Although the evidence is inconclusive, the Hite bed may have been deposited by streams flowing to the northeast and derived from a land area in northwestern Arizona and southwestern Utah. In support of this view, an unconformity exists between the Chinle formation and the overlying units of the Glen Canyon group in the region of the supposed land area in northwestern Arizona and southwestern Utah.

Terrain of source areas

The Uncompahgre and Front Range highlands consisted of Precambrian igneous and metasedimentary rocks, based on the type of rocks now exposed in these areas (Shoemaker, 1956; Larsen and Cross, 1956; Montgomery, 1956; Lovering and Goddard, 1950). Granite, gneiss, schist, and quartzite are the main types of rocks exposed. Paleozoic sedimentary rocks were locally exposed around the flanks of these highlands during Late Triassic time.

The main type of rock formed from the material supplied from the Uncompahgre and Front Range highlands is illitic arkosic siltstone and sandstone. The arkosic character of these rocks is a reflection of the source terrain which contains feldspar-rich rocks such as granite and gneiss. The abundance of illite, however, cannot be so easily explained. Some of this illite was probably derived from weathering of the granitic and metasedimentary rocks of the source areas. The soils developed on these rocks directly below the Chinle formation, however, contain mixed layer illite-montmorillonite as the dominant clay. Apparently these mixed layer clays which were eroded and carried into the basin of deposition were changed by diagenetic alteration to illite. This diagenetic change may have taken place in the lake hypothesized for the upper part of the Chinle formation. As outlined by Milne and Earley (1958) and Keller (1956, p. 2703) illite is commonly altered to other clays in a marine environment and presumably could also take place in a fresh water environment. In addition, some of the clay may have been derived from older red beds that flank the Uncompahgre and Front Range highlands. The most common clay in red beds is illite (Van Houten, 1948; Schultz, in press; Hooks and Ingram, 1955), and the Moenkopi formation that underlies the Chinle formation close to these highlands contains illite as the dominant clay (Schultz, in press).

The Mogollon highland probably was mainly a volcanic terrain similar to what it was during deposition of the lower part of the Chinle formation. The supposed land area in northwestern Arizona and southwestern Utah probably supplied a small amount of material derived from the uppermost beds of the Chinle formation.

Climate

The climate during deposition of the upper part of the Chinle formation was probably essentially the same as that during the deposition of the lower part of the formation; that is, a savanna climate with ample rainfall and a distinct dry season. The fossil types in the upper part of the Chinle formation are mainly the same, although less in number, as those in the lower part of the formation, indicating a similar climate.

The presence of sand dune deposits in the upper part of the

Chinle formation suggests some increase in aridity as compared to the lower part of the formation. Further, the Chinle formation is overlain by, and in some areas intertongues with, the thick widespread eolian deposits of the Wingate sandstone. Such extensive deposits of eolian sandstone are forming today only in vast desert regions lying generally between latitude 30° N. and 30° S. Perhaps, therefore, the climate during deposition of the upper part of the Chinle formation was initially a tropical savanna type and changed gradually with time to a desert type. The desert climate probably prevailed throughout the deposition of the overlying eolian deposits of the Wingate sandstone.

At first glance, the change from a tropical savanna climate to a contrasting desert climate seems illogical. One might expect a change to a more moderate climate and not to the extreme desert climate. However, the part of the earth that today lies between latitudes 30° N. and 30° S. exhibits this same juxtaposition of contrasting climate types. The region of the earth between latitudes 30° N. and 30° S. consists of 50 percent tropical rainy climates (savanna and rainforest climates), 37 percent dry climates (desert and semiarid or steppe climates), 7 percent humid subtropical climates, and 5 percent of undifferentiated climates in highlands (calculated from Trewartha, 1954, Pl. 1). Some of the most extreme desert regions of the earth today lie adjacent to tropical regions. A climatic shift from tropical to desert in the upper part of the Chinle formation, therefore, need not be considered unusual.

Origin of red beds

The origin of red beds has been a subject of debate for many years (Van Houten, 1948, gives a summary of many of the theories of origin of red beds). In recent years, Van Houten (1948) and Krynine (1935, 1949, 1950) have made comprehensive studies of the origin of red beds. Both of these authors have pointed out that the fossils contained in red beds indicate a humid tropical climate. Krynine (1935) showed that red beds are forming today in humid parts of southern Mexico. Here the red color in the sediments is derived from erosion of red lateritic soils rich in hematite. Krynine (1935, 1949, 1950) concludes that the red color in most red beds is derived from erosion of red hematite-rich soils in the source area, and transportation of the hematite into the area of deposition. The presence of mud cracks and casts of salt crystals indicated to Krynine (1949, 1950) that occasionally the area of deposition was dry. Krynine concludes that the most likely climatic conditions for the formation of the Triassic red beds of Connecticut was a savanna climate.

In general, Krynine's and Van Houten's ideas on the origin of red beds seem to apply to the upper part of the Chinle formation. As outlined previously, the climate was probably of a savanna type, although desert conditions may have prevailed during deposition of the uppermost part of the formation. The source rock on the Uncompahgre and Front Range highlands was dominantly granitic rock and metasedimentary gneiss and schist in which the iron content is probably fairly high and on which hematite-rich soils could easily form. The mottled strata which occur directly below or in the bottom few feet of the Chinle formation, are considered to be a soil zone, and contain abundant hematite. In addition, many of the Pennsylvanian and Permian rocks which were exposed on the flanks of the Uncompahgre and Front Range highlands are red beds. These red beds could have supplied hematite-rich detrital material directly to the Chinle formation, or hematite-rich soils could have been developed on them.

A possible objection to Krynine's and Van Houten's ideas is that the main clay mineral in red beds is illite, whereas kaolinite is the dominant clay of red soils developed in a tropical or subtropical climate (Grim, 1953) and for that reason should be the dominant clay in sediments derived from these soils. In addition hydrated aluminum oxide minerals, which are characteristic of tropical lateritic soils, have not been reported in the Chinle formation. The dominant clay in the upper part of the Chinle formation is illite, although montmorillonite and mixed layer montmorillonite-illite also occur. Illite is also the dominant clay in the red beds of the Moenkopi formation of the Colorado Plateau (Schultz, in press), in other red beds in the western United States (Van Houten, 1948), and in the Triassic rocks in the eastern United States (Hooks and Ingram, 1955). Hooks and Ingram (1955) have suggested that the clays in the Triassic rocks in the eastern United States were derived from immature red soils of a savanna climate, and this explanation appears to have application to the Chinle formation. Schultz (in press) has shown that the soils developed on Precambrian rocks directly below the Chinle formation on the Uncompahgre

Plateau in Colorado, from which some of the sediments of the upper part of the Chinle formation were derived, contain mixed layer montmorillonite-illite as the dominant clay. Supposedly this mixed layer clay was converted by diagenetic processes into illite in the basin of deposition of the upper part of the Chinle formation. Schultz also found that locally pre-Chinle soils developed away from the source area within the basin of deposition of the Chinle formation contain kaolinite as the dominant clay. Perhaps erosion in the source area was too rapid to allow the development of mature kaolinitic soils and partial weathering produced instead immature soils with mixed layer montmorillonite-illite clay. Where weathering was more complete, as locally away from the source area, kaolinite was formed. Thus the climate might be characterized as capable of producing mature kaolinitic soils but only rarely and locally did weathering continue long enough to reach this end product.

Another possible objection to Krynine's and Van Houten's ideas is the presence of possible dry climate features, such as gypsum and eolian deposits, in red beds, whereas these writers suggest that the climate is tropical. Gypsum is characteristically associated with red beds (Krumbein, 1951), and although gypsum is absent in the upper part of the Chinle formation, the presence of sand dune deposits indicates probable local dry conditions during the deposition of this part of the formation. Perhaps, as suggested by Dunham (1953) in his discussion of red beds of Permian and Triassic age in Britain, the source area had a distinctly different climate from that of the adjoining basin of deposition. The source area may have had a tropical climate whereas the basin of deposition had a dry, perhaps locally desert, climate. The source area stood higher than the basin of deposition, and thus could have received more rainfall than the basin of deposition. Such a situation exists today in California where the average annual precipitation in the higher parts of the Sierra Nevada is 50 to 70 in. as compared with 15 to 20 in. in the adjoining Sacramento Valley (Sprague, 1941, p. 795). Also, as discussed previously, in the low latitude regions of the world today, desert regions commonly occur adjacent to tropical regions.

Although the evidence is not conclusive, the red beds of the upper part of the Chinle formation probably derived their color from erosion of immature red soils developed in a tropical or subtropical climate and also probably from preexisting red beds in the source area. The climate in the basin of deposition may at times have been drier than the supposed tropical climate in the source area, thus accounting for local eolian deposits in the uppermost part of the Chinle formation.

Summary of interpretations

The surface below the Chinle formation is a peneplain on which soils and thin alluvial deposits formed during a long time of tectonic stability prior to the deposition of the Chinle formation. The mottled strata on this peneplain are probably remnants of a soil. These strata locally contain conglomerate and sandstone that are believed to be the deposits of the rivers that flowed across the peneplain.

The lower part of the Chinle formation represents an alluvial deposit laid down on a gently sloping surface. The initial deposits, the Shinarump and related members, consist of cross-stratified conglomerate and sandstone laid down in braided streams. Aggradation in these streams was caused by a change in the regimen of the graded streams flowing across the peneplaned surface. Such aggradation in graded streams leads to the deposition of thin widespread units such as the Shinarump and related members. Locally the Shinarump member fills channels cut into the underlying units. Some of these channels probably are remnants of valleys that developed in the land surface below the Shinarump member; others are the scour surfaces of the streams that deposited the Shinarump member.

The Monitor Butte and related members are transitional between the braided stream deposits of the Shinarump and related members and meandering stream and lake deposits that characterize the Petrified Forest member. Locally the Monitor Butte and related members contain lenses of sandstone similar to that in the Shinarump member; these lenses are probably braided stream deposits. The members also contain clayey sandstone composed of shallow trough sets of low angle cross-strata; these deposits may be those of meandering streams. Horizontally stratified claystone and clayey siltstone occur and are considered to be river basin or lake deposits. Finally, very fine-grained sandstone with rib-and-furrow structure is fairly abundant. This sandstone is probably formed in shallow, flat-bottomed, probably low velocity streams with a silt and very fine-grained sand bottom. Contorted strata on a large scale are common in the Monitor Butte and related members. These contorted strata probably formed by large landslips of material into stream channels from adjacent flat lands.

The Moss Back member and related units are widespread braided stream deposits similar to the Shinarump member.

The Petrified Forest member is composed of deposits of large meandering rivers and of deposits formed in river basins or lakes. Locally braided stream deposits, such as the Sonsela sandstone bed, extended out across the basin of deposition. The fossils in the member indicate both fresh water features such as rivers and lakes and intervening dry land "upland" areas.

The depositional plain of the lower part of the Chinle formation sloped northward and northwestward away from a broad source area in southern Arizona and adjacent states--the Mogollon highland. The rocks exposed in this highland were dominantly volcanic; minor quantities of cherty limestone or dolomite, sandstone, metasedimentary rocks, and probably granitic rocks also occurred in the highland. In addition, a small amount of material was supplied to this part of the formation from the Uncompahgre highland in western Colorado and northernmost New Mexico. This highland had an igneous and metamorphic terrain.

The upper part of the Chinle formation represents a large deltaic deposit formed in a shallow lake. The Owl Rock member at the base of the upper part of the formation contains pale-red and light greenish-gray limestone and reddish-brown siltstone. The limestone formed in lakes by calcite replacement of glassy volcanic sandstone beds or by filling the interstitial spaces in siltstone. The reddish-brown siltstone is considered to be a lake deposit.

The Church Rock member and related units at the top of the Chinle formation contain horizontally stratified siltstone and sandy siltstone, horizontally laminated, ripple-laminated, and wavy-stratified siltstone and sandstone, limestone and siltstone pebble conglomerate, and cross-stratified sandstone. The horizontally stratified siltstone and sandy siltstone are considered to be lake deposits. The horizontally laminated, ripple-laminated, and wavy-stratified siltstone and sandstone are probably lake deposits which formed in shallow and current-swept parts of lakes. Some of this siltstone and sandstone probably formed under subaerial conditions. The limestone and siltstone pebble conglomerate units formed by fluvial reworking of locally derived material. Some of the cross-stratified sandstone units are stream deposits; others are sand dune deposits.

Most of the upper part of the Chinle formation was formed as a vast subaqueous deltaic deposit laid down in a shallow lake. Fluvial sandstone in the upper part of the formation is abundant in a narrow belt extending from southwestern Colorado to central Utah. This sandstone probably represents the deposit of a major river entering the lake. This river may have supplied most of the material that formed the deltaic deposit.

The source areas of the upper part of the Chinle formation are the Uncompahgre and Front Range highlands of Colorado and adjacent regions, and the Mogollon highland of southeastern Arizona and southwestern New Mexico. The Uncompahgre and Front Range highlands contained mostly igneous and metamorphic rocks. The Mogollon highland probably was mainly a volcanic terrain. The Mogollon highland was not as extensive during deposition of the upper part of the formation as it was during deposition of the lower part.

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