Eocene tectonics and depositional setting of west-central New Mexico and eastern Arizona

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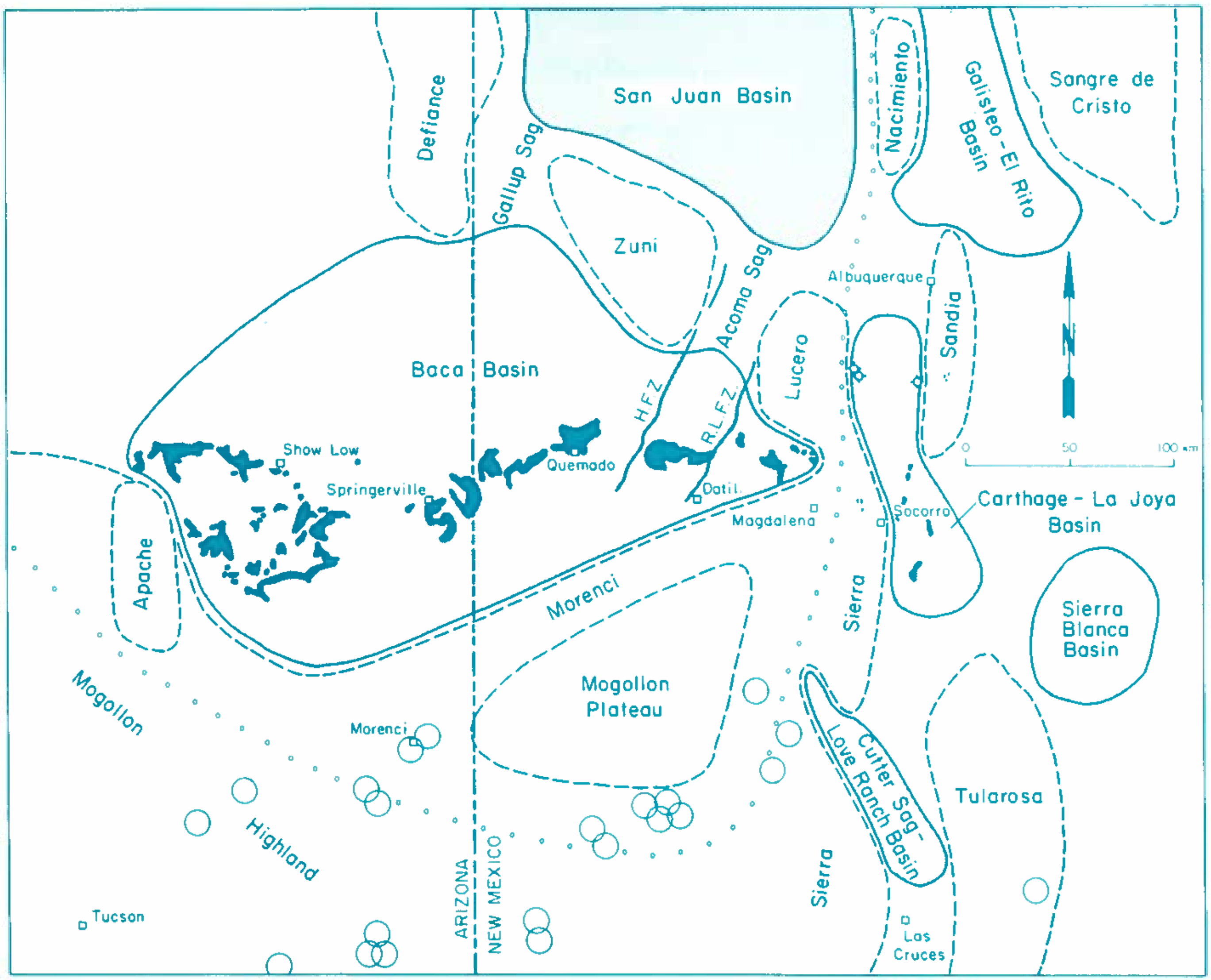

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FRONT COVER—Map showing distribution of Eocene uplifts and basins in western New Mexico and eastern Arizona (see Fig. 2).

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Abstract

The deposits of the Baca and Carthage—La Joya basins record late Laramide (Eocene) sedimentation in west-central New Mexico and eastern Arizona. Sedimentation in these basins was influenced by tectonism (both intrabasinal and in surrounding uplifts) and by the prevailing semiarid climate.

A well-exposed facies tract through the central part of the Baca basin reveals a broad spectrum of depositional paleoenvironments. The braided alluvial-plain system, consisting of the deposits of coalesced humid alluvial fans, is widespread throughout the basin. Sedimentary structures present in these deposits indicate that discharge was increasingly flashy toward the basin center. Meanderbelt sedimentation was dominantly restricted to an actively subsiding, fault-bounded block in the eastern part of the basin; examples of both fine- and coarse-grained point-bar deposits are present in this area. Lacustrine sedimentation occurred in small, floodplain lakes and in a large, shallow, closed lake near the eastern end of the basin. Cyclic progradation, abandonment, and subsidence of fine-grained deltas and fan deltas well out into the lake basins prevented development of extensive lacustrine basin-center deposits.

Only rocks of the braided alluvial-plain system are present in the poorly preserved deposits of the Carthage—La Joya basin. Updip retreat of facies tracts in both the Carthage—La Joya and Baca basins reflects waning Laramide deformation in adjacent uplifts during late Eocene time.

Introduction

The Eocene Baca Formation of New Mexico and correlative Eagar Formation and Mogollon Rim gravels of Arizona comprise a sequence of conglomerate, sandstone, mudstone, and claystone that crops out in a discontinuous west-trending belt from near Socorro, New Mexico to the Mogollon Rim of Arizona (Fig. 1). These rocks were deposited in two separate, yet structurally related, basins that formed during late Laramide time. The largest and westernmost of these basins, herein termed the Baca basin, is represented by the sedimentary rocks of the Mogollon Rim gravels, the Eagar Formation, and those portions of the Baca Formation that crop out west of the Rio Grande. The maximum exposed thickness of the Eocene deposits in the Baca basin is approximately 580 m, although as much as 760 m has been reported in the subsurface (Snyder, 1971). The deposits of the Carthage—La Joya basin, the smaller of the two basins, are represented by a series of small exposures of the Baca Formation that crop out to the east of the Rio Grande and attain a maximum exposed thickness of approximately 315 m. In this paper, we delineate the depositional setting within these two basins and describe the Eocene structural elements, both intrabasinal and extrabasinal, that influenced sedimentation within the basins. To facilitate discussion of various parts of the Eocene outcrop belt, we have divided the exposure areas into geographic segments (Fig. 1).

The deposits of the Baca basin unconformably overlie strata ranging in age from Late Cretaceous to Pennsylvanian. Rocks of Late Cretaceous age underlie the majority of the Baca basin, but basin-fill deposits onlap progressively older rocks along the basin margins (such as in the Mogollon Rim and Lemitar Mountains areas). Subcrop relations within the Carthage—La Joya basin appear to be similar to those of the Baca basin; however, interpretation of many of the geologic features of the Carthage—La Joya basin is greatly hindered by erosional stripping and extensive structural disruption of the basin during subsequent development of the Rio Grande rift. Throughout most of west-central New Mexico and eastern Arizona, the Baca Formation and its equivalent units are conformably and transitionally overlain by andesitic volcaniclastic deposits of the early Oligocene Spears Formation (Cather, in prep.). In eastern Arizona, however, an unknown thickness of the upper portion of the Mogollon Rim gravels was removed by erosion beginning in Oligocene(?) time (Pierce and others, 1979). In this area, the rim gravels are locally disconformably overlain by volcanic rocks of late Oligocene and younger age.

The Baca and Eagar Formations and the Mogollon Rim gravels have been assigned ages ranging from Late Cretaceous to early Oligocene. Most determinations, however, indicate an Eocene age for these units (see summaries in Johnson, 1978; Cather, 1980; Lucas and others, 1981). Volcanic clasts in the lower portion of the overlying Spears Formation have been radiometrically dated at 38.6 ± 1.5 m.y. and 39.6 ± 1.5 m.y. in the western Gallinas Mountains (C. E. Chapin, unpub. dates), 38.6 ± 1.5 m.y. in the Datil Mountains (Bornhorst and others, 1982), and 37.1 ± 1.5 m.y. in the Joyita Hills (Burke and others, 1963). Thus, in these areas, the Baca Formation can be no younger than earliest Oligocene. In their summary of the Eocene biostratigraphy of New Mexico, Lucas and others (1981, p. 962) state ". . . the fossiliferous outcrop of the Baca Formation east of the Rio Grande is of Bridgerian age, whereas the fossiliferous outcrops of the Baca Formation west of the Rio Grande are significantly younger, being of Duchesnean age." Paleontologic data for the Baca Formation, however, are sparse, and temporal relations between the Baca and Carthage—La

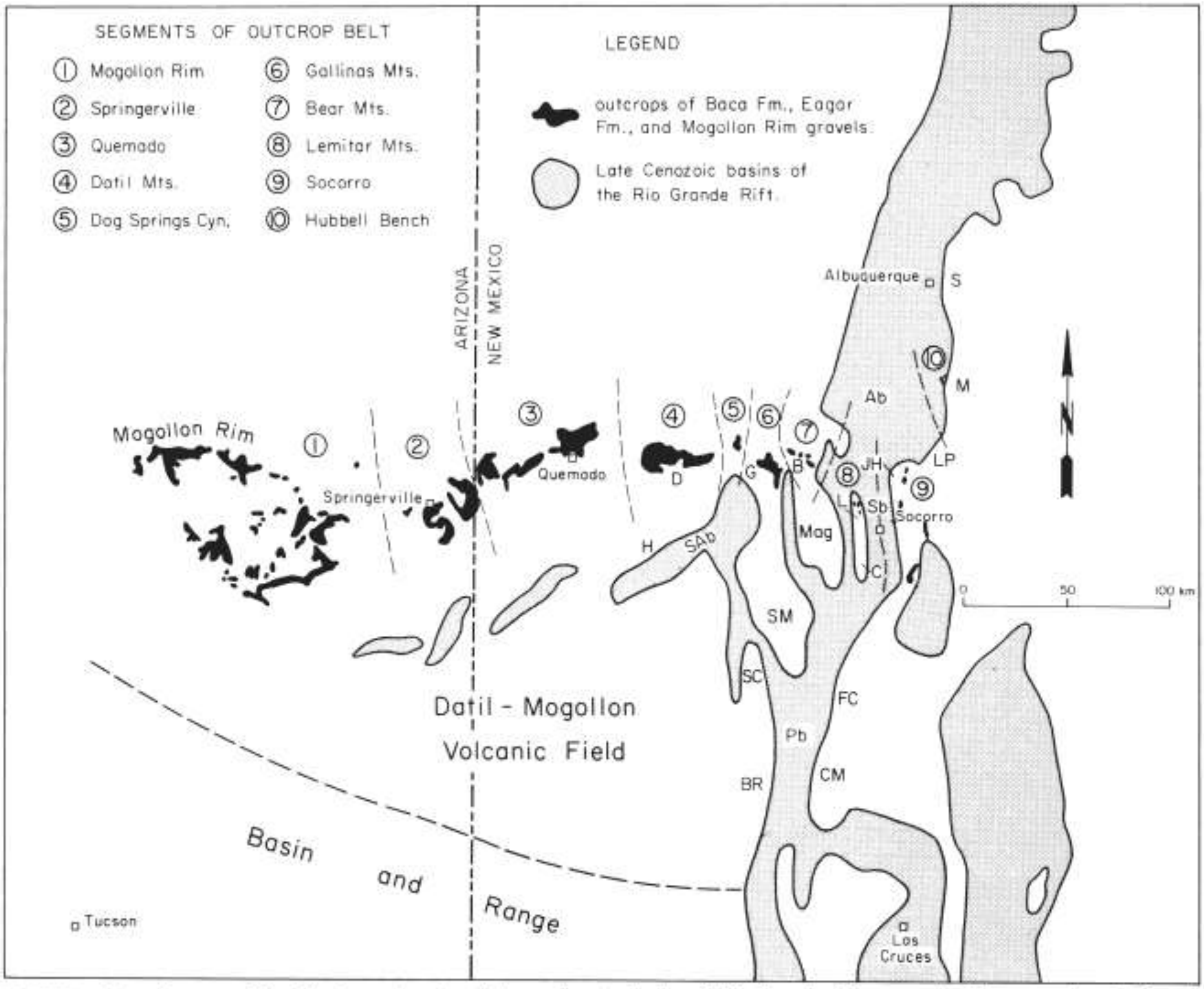

FIGURE 1—Map of western New Mexico and eastern Arizona showing basins of Rio Grande rift, segments of Eocene outcrop belt, and localities mentioned in text. S = Sandia Mountains, M = Manzano Mountains, Ab = Albuquerque Basin, LP = Los Pinos Mountains, JH = Joyita Hills, D = Datil Mountains, G = Gallinas Mountains, B = Bear Mountains, L = Lemitar Mountains, Sb = Socorro Basin, Mag = Magdalena Mountains, C = Chupadera Mountains, H = Horse Mountain, SAb, San Agustin Basin, SM = San Mateo Mountains, SC = Sierra Cuchillo, FC = Fra Cristobal Mountains, Pb = Palomas Basin, BR = Black Range, CM = Caballos Mountains. Base map after Chapin and others (1978).

Joya basins cannot be firmly established at this point. Indeed, the fact that deposits in parts of both the Carthage—La Joya and Baca basins were derived from the same uplift (the Sierra uplift; Fig. 2) and that the deposits of both these basins are transitionally overlain by the early Oligocene Spears Formation implies that sedimentation within these two basins may have been largely synchronous.

The outcrop quality of sedimentary rocks within the Carthage—La Joya basin, and the sedimentologic perspective afforded by these exposures, contrast sharply with those of the Baca basin. The majority of the exposures of Carthage—La Joya basin deposits occur in a north-trending belt of very discontinuous, small outcrops that approximately parallels the depositional strike of the western flank of the basin (Fig. 2). Due to the effects of subsequent rift-related tectonism, large parts of Carthage—La Joya basin (especially distal environments) have been either erosionally removed or buried beneath Neogene basins of the Rio Grande rift.

In contrast, the deposits of the Baca basin are well preserved and exposures representative of the majority of the facies tract are present along the outcrop belt. The high exposure quality of the sedimentary rocks within the Baca basin is the result of several factors:

1. Most of the Baca basin developed on relatively stable, weakly deformed crust of the southeastern Colorado Plateau, and thus was not strongly disrupted by Laramide or subsequent tectonism.
2. The Baca Formation and equivalent units are, in most areas, conformably and gradationally overlain by a thick sequence of early Oligocene volcanic and volcaniclastic rocks that allowed preservation of even the stratigraphically highest Eocene basin-fill sediments.

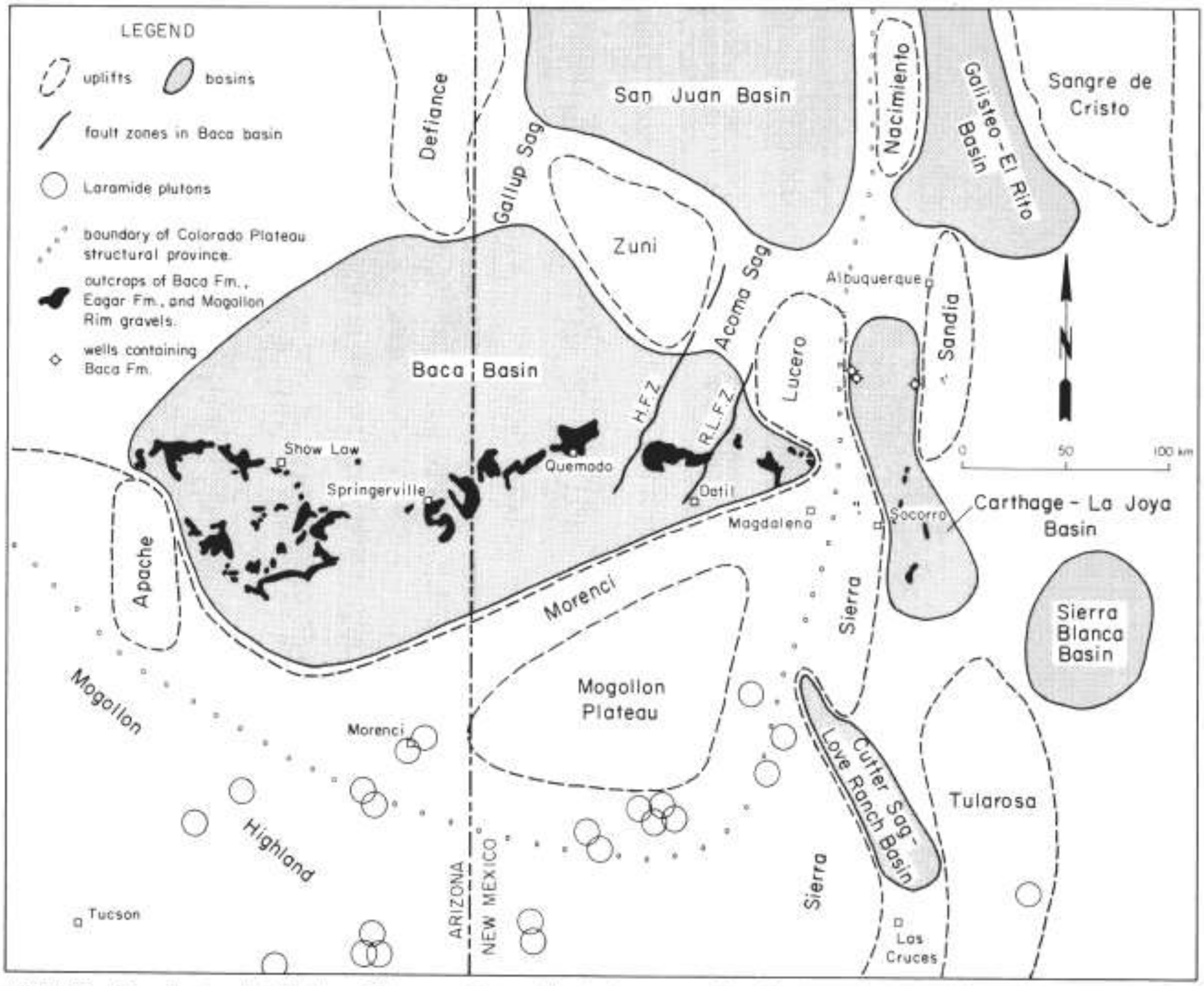

FIGURE 2—Map showing distribution of Eocene uplifts and basins in western New Mexico and eastern Arizona (map coverage same as Fig. 1). Base map and outcrop data modified from Dane and Bachman (1965) and Wilson and others (1969); Laramide pluton data from Chapin and others (1978); Red Lake fault zone (RLFZ) and Hickman fault zone (HFZ) simplified from Wengerd (1959) and Chamberlin (1981).

However, this is not the case in eastern Arizona, where Oligocene(?) and later erosion removed an unknown thickness of Mogollon Rim gravels (Pierce and others, 1979) and locally stripped portions of the Eagar Formation (Sirrine, 1956).

3. With the exception of those in eastern Arizona, the deposits of the Baca basin were tilted gently southward during post-Eocene time (the Mogollon slope of Fitzsimmons, 1959), possibly in response to volcanism-related crustal loading and subsidence in the Datil-Mogollon field to the south.
4. The onset of general uplift of the Colorado Plateau and environs beginning in early Miocene time (Chapin and Seager, 1975) caused the updip, northern portion of the Baca basin-fill deposits and the overlying volcanic rocks to be erosionally stripped.

Therefore, due to a fortuitous series of geologic events, a relatively continuous cross-section view of the facies tract exists within the Baca basin. The Baca outcrop belt transects the axial portion of the basin and trends approximately parallel to the depositional dip of the basin-fill sediments, thus affording us with an informative, perhaps unique, perspective of the downdip succession of paleoenvironments within a Laramide foreland basin.

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Eocene tectonic framework

The Laramide orogeny occurred between about 80 and 40 m.y. ago (Coney, 1971) and resulted in the creation of numerous uplifts and basins in the western North American foreland. Maximum strain rates and the structural culmination of Laramide deformation began in latest Paleocene—earliest Eocene time (Chapin and Cather, 1981); development of the Baca and Carthage—La Joya basins is related to this late phase of Laramide tectonism. Unlike many of the Laramide basins in the western United States, the Baca and Carthage—La Joya basins do not contain any Paleocene deposits.

Sedimentation within the Baca and Carthage—La Joya basins was influenced by intrabasin structures and by adjacent uplifts. The following is a brief synopsis of these structural features.

Lucero uplift

The northern boundary of the Baca basin is defined by the Lucero, Zuni, and Defiance uplifts (Fig. 2), which are monocline-bounded flexures typical of Laramide deformation on the Colorado Plateau (Kelley, 1955). The Lucero uplift lies on the eastern margin of the Colorado Plateau and consists of broadly arched, west-dipping strata except along its eastern flank which is marked by a complexly faulted and folded section of steeply east-tilted strata (Callender and Zilinski, 1976). Sedimentary rocks ranging in age from Pennsylvanian to Triassic compose the majority of the exposures on the uplift. The structurally low area between the Lucero and Zuni uplifts is termed the Acoma sag (Kelley, 1955).

Zuni uplift

The Zuni uplift trends northwest and is markedly asymmetric; its northeastern flank dips gently into the San Juan Basin, while its southwestern flank (the Nutria monocline) dips steeply toward the Baca basin and exhibits as much as 1,400 m of structural relief (Edmunds, 1961). The majority of the exposures on the Zuni uplift consist of sedimentary rocks of Permian to Triassic age. However, Precambrian granite, metarhyolite, gneiss, schist, and metaquartzite (Fitzsimmons, 1967; Goddard, 1966) crop out along the crest of the uplift. The structurally low area between the Zuni and Defiance uplifts is termed the Gallup sag. More than 2,450 m of structural relief exists between the highest part of the uplift and the Gallup sag. Structural relief between the uplift and the deepest part of the San Juan Basin is more than 4,000 m (Kelley, 1955).

Defiance uplift

The north-trending Defiance uplift is asymmetrical, with its steep eastern flank (the Defiance monocline) facing the Gallup sag and the San Juan Basin. The Defiance monocline is highly sinuous due to the presence of a series of en echelon, southeast-plunging anticlines and synclines in the eastern part of the uplift (Kelley and Clinton, 1960). These en echelon folds are suggestive of right-slip along the Defiance monocline, and Kelley (1967) has documented approximately 13 km of dextral offset of Jurassic facies in this area. Although the precise timing of Laramide deformation along the Defiance monocline has not been established, it may be related to dextral wrench faulting in the southern Rocky Mountains that culminated in early Eocene time (Chapin and Cather, 1981).

With the exception of a few small outcrops of Precambrian quartzite, the majority of the exposures on the Defiance uplift are of Permian to Triasic sedimentary rocks. Structural relief between the highest part of the uplift and the Gallup sag is at least 2,150 m (Kelley, 1955).

Mogollon Highland and Apache uplift

The Mogollon Highland of southern Arizona and southwestern New Mexico forms the southwestern boundary of the Baca basin and was the dominant contributor of detritus to the basin. The early Tertiary structural style of the Mogollon Highland is the subject of much ongoing controversy. Interpretations of the nature of Laramide tectonism in this area include large-scale overthrusting (Drewes, 1976, 1978), crustal extension and differential uplift (Rehrig and Heidrick, 1972, 1976; Lowell, 1974), and the vertical uplift of large, basement-cored blocks during strong regional compression (Davis, 1979; Keith and Barrett, 1976). More recently, S. B. Keith (1981, written commun.) has compiled evidence for large-scale Eocene underthrusting of the Colorado Plateau in this area.

The magnitude of Laramide structural relief in the Mogollon Highland is difficult to estimate, due to superposition of numerous episodes of orogeny in this area. However, the dominance of Mogollon Highland-derived sediments in the Baca basin suggests that topographic relief, and hence structural relief, was considerable. Lithologies exposed in southern Arizona during Baca time were highly diverse and ranged in age from early Tertiary to Precambrian.

The Apache uplift is a monocline-bounded uplift with at least 1,800 m of Laramide structural relief (Davis and others, 1982) and appears to be contiguous with the region of general Laramide uplift to the south (the Mogollon Highland). The Apache uplift may have influenced sedimentation in the westernmost part of the Baca basin; however, such effects cannot be documented with the available data.

Morenci uplift

The probable existence of a Laramide drainage divide along the southeastern margin of the Baca basin, which was subsequently obscured by thick accumulations of mid-Tertiary volcanic rocks of the Datil-Mogollon field, was pointed out to us by C. E. Chapin (1980, oral commun.). The presence of large exposures of Laramide volcanic and plutonic rocks in southwestern New Mexico and adjacent Arizona (Fig. 2) and the nearly complete lack of volcaniclastic detritus in the eastern part of the Baca outcrop belt (Cather, 1980) imply the existence of an intervening uplift that prohibited input of volcanic detritus into the eastern part of the Baca basin. The Morenci lineament (Chapin and others, 1978) provides a likely site along which such a drainage divide may have developed.

A series of en echelon, northeast-trending fault zones and grabens, the largest of which is the San Agustin Basin (Fig. 1), presumably represent portions of the Morenci uplift that collapsed during middle and late Tertiary extension. Stratigraphic relations also support the existence of a Laramide positive area along the Morenci lineament. At Horse Mountain, along the northwestern margin of the San Agustin Basin, volcaniclastic rocks of the early Oligocene Spears Formation overlie strata of Triassic and Permian age with apparent unconformity (Stearns, 1962). In the Magdalena Mountains, at the northeastern end of the Morenci uplift, Upper Paleozoic strata are unconformably overlain by the Spears Formation (Krewedl, 1974; Blakestad, 1978). Thus, in these areas, a period of Laramide uplift and erosion preceded the onset of mid-Tertiary volcanism.

The Morenci uplift separated two regions of relative crustal stability: the Mogollon Plateau and the Colorado Plateau. In this paper, the Mogollon Plateau is considered a subregion of the Laramide Colorado Plateau based on the relatively undeformed nature of its central portion (Coney, 1976) and its dissimilarity to adjacent, more structurally complex areas to the south and east. The structural style of the Morenci uplift is poorly known due to extensive cover by mid-Tertiary volcanic rocks and partial collapse during middle and late Tertiary extension. However, due to its association with relatively stable, Colorado Plateau-type crust, the Morenci uplift may have been monoclinal. Davis (1978) stated that Colorado Plateau monoclines were produced by reactivation along basement fracture zones (lineaments) during early Tertiary tangential compression, and he recognized two dominant trend directions (N. 20° W. and N. 55° E.) of lineaments that controlled monocline development. The Morenci lineament of Chapin and others (1978), which appears to control the Morenci uplift, closely parallels Davis' northeasterly trend direction. In detail, however, the Morenci uplift may have consisted of several en echelon segments, as suggested by the en echelon arrangement of the San Agustin Basin and its associated extensional features. If the en echelon pattern of late Tertiary extensional structures in this area were inherited from a Laramide compressional precursor, right shift may have taken place along the Morenci lineament during Laramide time. Magnitude of Laramide structural relief along the Morenci uplift is impossible to estimate at present, although it must have been appreciable, because significant amounts of Precambrian detritus are present in Baca Formation deposits derived from this uplift (Cather, 1980).

The existence of a Morenci uplift would also explain the increased abundance of volcanic clasts in the western part of the Baca basin (Johnson, 1978; Pierce and others, 1979). Laramide volcanic rocks in the vicinity of Morenci, Arizona, would not have been isolated by the postulated drainage divide and may have supplied detritus to the western part of the Baca basin.

Sierra uplift

Eardley (1962) proposed the existence of a Laramide positive area adjacent to the eastern margin of the Colorado Plateau in southern New Mexico and termed it the Sierra uplift. Sedimentologic data from the Baca Formation indicate that the uplift extended northward into the Socorro vicinity, approximately 70 km beyond the northern boundary of the uplift as depicted by Eardley. Large portions of this uplift subsided to form basins of the Rio Grande rift during middle and late Tertiary time; thus, documentation of a Laramide positive area in this region must depend upon evidence derived from synorogenic sediments, regional stratigraphic relations, and the structural style of remaining, unsubsided portions of the uplift.

The only detailed sedimentologic studies and facies analyses of sediments derived from the Sierra uplift are those of the Baca Formation by Snyder (1971), Johnson (1978), and Cather (1980, ongoing research). Upper Cretaceous to Eocene deposits of the Cutter Sag—Love Ranch basin (McRae and Love Ranch Formations; Seager, 1975, 1981) are present adjacent to the central and southern parts of the uplift; to date, however, no systematic study of these rocks has been attempted. Data from the Baca Formation clearly demonstrate that the northern portion of the Sierra uplift encompassed much of what is now the Socorro Basin of the Rio Grande rift. Evidence includes:

1. Paleocurrent and facies data from the Baca outcrops east of the Rio Grande near **Socorro**

indicate generally eastward paleoflow in proximal- and mid-fan environments, thus suggesting a source nearby to the west.

2. Paleoflow was dominantly to the southwest during Baca deposition in the Bear Mountains vicinity, and presumably represents drainage off the western flank of the Sierra uplift and the southwestern flank of the Lucero uplift.
3. Local, lenticular units of limestone-clast conglomerates, which are correlative with the Baca Formation (Chamberlin, 1982), crop out along the western margin of the Socorro Basin in the Lemitar Mountains. Pebble imbrication indicates westward paleoflow, and these deposits are interpreted as local fill in paleocanyons that drained the western slope of the Sierra uplift.
4. The presence of a large, persistent, closed lacustrine system in the Bear—Gallinas Mountains area indicates the presence of a damming element to the east.

Stratigraphic relations suggest that much of the rift south of Socorro was a Laramide positive area (the central and southern portions of the Sierra uplift). In the ranges bounding portions of the west side of the rift (Magdalena Mountains, San Mateo Mountains, and Sierra Cuchillo), early Oligocene volcanic rocks typically overlie strata of late Paleozoic age (Krewedl, 1974; Blakestad, 1978; Farkas, 1969; Jahns and others, 1978). In the Black Range, similar relations exist but are complicated by the local presence of a suite of older volcanic and hypabyssal rocks, some of which are as old as Late Cretaceous (Hedlund, 1974). Along the east shoulder of the Palomas Basin, volcanic rocks of late Eocene to early Oligocene age unconformably overlie Paleozoic strata in the southern Caballos Mountains (Kelley and Silver, 1952; Kottlowski and others, 1969; Seager, 1975), and deposits possibly correlative to the McRae Formation overlie Precambrian gneiss in the northern Fra Cristobal Mountains (Kelley and McCleary, 1960). These stratigraphic relations differ greatly from those in adjacent Laramide basins, where thick deposits of early Tertiary sediments typically overlie strata of Late Cretaceous age.

The Sierra uplift implies a Laramide compressional phase prior to the onset of middle and late Tertiary extension. This early phase of compressive tectonism has been noted by Kelly and Clinton (1960, p. 55). They state:

> . . there is much evidence along the Rio Grande trough of an early tectonic compressional history. The structures of the Caballos Mountains (Kelley and Silver, 1952, p. 136-146) are clearly early Tertiary and compressional. Low-angle (15°) faults in which gneiss rests on Cambrian limestone are clearly displayed within a few hundred years of the Rio Grande trough in the Fra Cristobal Range of Sierra County, New Mexico (Jacobs, 1957, p. 257). Thrusts and overturns have been mapped in many places in the Sandia and Manzano uplifts. . ."

Thrust faults of Laramide age have been also reported along the eastern margin of the rift in the vicinity of Socorro (Wilpolt and Wanek, 1951), and Seager (1975, 1981) documents Laramide compressive structures and uplifts in the Las Cruces area.

Baca sediments derived from the northern part of the Sierra uplift are arkosic and contain locally abundant clasts of Precambrian granite, schist, and metaquartzite. Only a few small exposures of Precambrian rocks are present today along the Rio Grande rift near Socorro. Thus, Precambrian detritus in the Baca Formation in this area must have been derived largely from source terranes that were subsequently down-faulted and buried in the rift. In the Chupadera Mountains, the southern part of an intrarift horst adjacent to the Socorro Basin, Precambrian rocks are locally overlain by early Oligocene volcanic and volcaniclastic deposits (Eggleston, 1982), indicating at least 1.5 km of Laramide uplift and erosional stripping in that area.

In the Laramide Cutter Sag—Love Ranch basin to the south, Precambrian-derived detritus is present in the McRae Formation (Kelley and Silver, 1952; Bushnell, 1955) and in an unnamed early Tertiary(?) sandstone overlying the Love Ranch Formation (Seager, 1981). Clasts of Precambrian lithologies also occur in deposits possibly equivalent to the McRae Formation in the northernmost part of the Fra Cristobal Mountains (Kelley and McCleary, 1960). Although these deposits may have been derived largely from the Sierra uplift, determination of the provenance and precise dating of these units awaits further study.

Detailed analysis of the structural style and geometry of the Sierra uplift is not possible with the data available at present. However, on the basis of trend, geographic location, the presence of thrust faults and associated compressional structures along its east flank, and evidence for significant exposures of Precambrian rocks along its crest, the Sierra uplift probably represents a basement-cored, anticlinal uplift (or series of uplifts) similar in structural style to the classic Laramide Front Range uplifts of Colorado and southern Wyoming. Chapin and Cather (1981) have recently proposed that north—northeastward translation of the Colorado Plateau, culminating during early Eocene time, resulted in the creation of an en echelon series of uplifts and basins along its eastern margin (including the Sierra uplift and Carthage—La Joya and Cutter Sag—Love Ranch basins) and caused the severe crustal shortening in the Wyoming province to the north.

Sandia uplift

The existence of a Laramide positive area, herein termed the Sandia uplift, in the vicinity of the modern

Sandia, Manzano, and Los Pinos Mountains has been proposed by several workers (Eardley, 1962; Kelley and Northrop, 1975; Kelley, 1977; Chapin and Cather, 1981) on the basis of structural data and stratigraphic and regional-tectonic relationships. The Sandia uplift is separated from the Lucero uplift by the northern part of the Carthage-La Joya basin, which contains as much as 1,146 m of Baca Formation red beds in the subsurface beneath the southern Albuquerque Basin (Foster, 1978).

Sedimentologic evidence for the Sandia uplift is restricted to a few isolated outcrops of the Baca Formation on the Hubbell bench (Kelley, 1977, 1982). These exposures occur near the center of T. 5 N., R. 4 E. and consist of conglomerates and sandstone composed of detritus derived solely from upper Paleozoic units (dominantly Abo Formation sandstones and siltstones). The highly conglomeratic nature of the Baca deposits on the Hubbell bench and the presence of well-developed pebble imbrication indicative of westward paleoflow suggest a source nearby to the east. Kelley (1982) has postulated that the axis of Laramide uplift passed somewhat east of the present crest of the Manzano Mountains, in the vicinity of the Montosa and Paloma reverse faults. The lack of Precambrian detritus in Baca exposures on the Hubbell bench implies that structural relief was only moderate. The Sandia uplift may have been monoclinal in nature, with its steep, east-facing limb locally broken by reverse faults (Kelley, 1977). The deposits of the Galisteo-El Rito basin to the north (Gorham and Ingersoll, 1979), as well as the Baca Formation in the Socorro area, show no evidence of derivation from the Sandia uplift.

Intrabasin structures

Structural features within the Baca basin were locally important determinants of synorogenic sedimentation style and thickness of basin-fill deposits. Throughout most of the Baca basin, the Laramide style of deformation is characterized by low-amplitude, gently plunging, broad folds and associated minor faulting. These features, which are best developed in the Cretaceous and older rocks that underlie the basin, also occur to a lesser extent in the Eocene basin-fill deposits. Although these low-amplitude folds and faults appear to have had only minor effect on Eocene sedimentation, Chamberlin (1981) suggests that they may have provided important controls on uranium mineralization in a lateritic weathering profile that underlies the Baca Formation in the Datil Mountains area.

Two fault zones in the eastern part of the Baca basin, the Hickman and Red Lake fault zones (Fig. 2; Wengerd, 1959; Chamberlin, 1981), were major intrabasin structures active during Eocene time. The Hickman and Red Lake fault zones are two north-northeast-trending systems of normal faults and folds that form the western and eastern boundaries, respectively, of a large, late Cenozoic "synclinal horst" (Wengerd, 1959) in the Datil Mountains area. Thickness variations in the Baca Formation, however, demonstrate that the area between these fault zones was downthrown during Eocene time. At Mariano Mesa, about 25 km west of the Hickman fault zone, the Baca Formation is approximately 185 m thick (Guilinger, 1982). East of the Red Lake fault zone, in the Dog Springs Canyon, Gallinas Mountains, and Bear Mountains areas, the Baca attains thicknesses of approximately 335 m, 285 m, and 230 m, respectively. The greatest exposed thickness of the Baca basin-fill deposits (approximately 580 m; Chamberlin, 1981), however, occurs *between* the Hickman and Red Lake fault zones, in the Datil Mountains area. The increased thickness of the Baca between these fault zones, as well as the nearly complete restriction of low-gradient meanderbelt deposits to this same area (see below), strongly suggests that the block bounded by the Hickman and Red Lake fault zones was actively subsiding during Baca sedimentation.

The Hickman and Red Lake fault zones apparently represent two north-northeast-trending systems of Laramide reverse faults that have been reactivated as late Cenozoic normal faults. The south-plunging syncline present in the now-uplifted block between these fault zones may be a relict Laramide feature, and appears to merge northward with the Acoma sag (Wengerd, 1959; Chamberlin, 1981). An en echelon series of northwest-trending folds present between the Hickman and Red Lake fault zones (Wengerd, 1959, fig. 4) suggests that right-slip may have accompanied Laramide reverse faulting, although this cannot be confirmed with available data.

The Jaralosa Creek fault system, a north-trending structural zone between the Bear Mountains and the Gallinas Mountains, also may represent a Laramide intrabasin wrench fault (Johansen, 1983). However, deformation along this zone appears to have had little effect on Baca sedimentation.

Documentation of Laramide intrabasin structures in the Carthage-La Joya basin is difficult due to poor exposures and the discontinuous nature of the Baca Formation outcrops in this area. However, many of the complex faults and folds along the eastern shoulder of the Rio Grande rift in the Socorro area probably are relict or reactivated Laramide structures (e.g., Smith, 1983).

Depositional systems

A depositional system is a genetically defined, three-dimensional, physical stratigraphic unit composed of a contiguous set of process-related sedimentary facies (Fisher and McGowen, 1967; Fisher and Brown, 1972; Galloway, 1977). Depositional systems are the stratigraphic manifestation of major ancient geomorphic features, such as barrier islands, eolian-dune fields, and lakes. Criteria utilized in the discrimination of paleoenvironments include lithofacies geometry, lateral and vertical variation in grain size and sedimentary structures, nature of contacts between lithofacies, petrographic data, and fossils. Because subsurface data in the study area are extremely limited, our method of study consisted of: (1) measurement, description, and interpretation of numerous partial and complete stratigraphic sections throughout the outcrop belt; (2) local detailed mapping, especially in the Gallinas Mountains area (Cather, 1980); (3) petrographic and x-ray study of sandstones and mudstones; (4) measurement and analysis of paleocurrent indicators; (5) conglomerate lithology and clast-size studies; and (6) regional reconnaissance.

Braided alluvial-plain system

The braided alluvial-plain system constitutes the most widespread depositional system within the Baca basin, and is the only system present in those portions of the Carthage—La Joya basin that have been preserved. We have arbitrarily divided this system into proximal-fan facies (dominantly conglomerate), mid-fan facies (subequal sandstone and conglomerate), and distal-fan fades (dominantly sandstone). This classification has been applied to an ancient alluvial-fan sequence by McGowen and Groat (1971), but differs from that utilized in studies of modern proglacial outwash fans (Boothroyd and Ashley, 1975; Boothroyd and Nummedal, 1978) that delineate proximal-, mid-, and distal-fan environments on the basis of clast size. A variant of the mid-fan facies, herein termed the arroyo-fill facies, is also discussed.

Six segments of the braided alluvial-plain system are exposed along the Eocene outcrop belt of west-central New Mexico and eastern Arizona:

1. Widespread exposures of proximal-, mid-, and distal-fan deposits occur in the Mogollon Rim, Springerville, Quemado, and Datil Mountains areas. These deposits represent an extensive, east-facing, braided alluvial plain that drained part of the Mogollon Highland in Arizona.
2. Mid- and distal-fan deposits that occur in the basal part of the Datil Mountains section were derived from the Zuni uplift by a southeast-flowing tributary system.
3. Lenticular, canyon-filling conglomerates in the Lemitar Mountains (Chamberlin, 1982) and fan-delta and braided-stream deposits in the Bear Mountains area represent the extreme proximal and distal ends of a poorly preserved, coalesced fan system that drained the western part of the Sierra uplift and the southwestern flank of the Lucero uplift.
4. Distal-fan deposits of northeast-flowing braided streams are present in the Gallinas Mountains area and consist of detritus derived from the Morenci uplift.
5. Discontinuous exposures of proximal- and

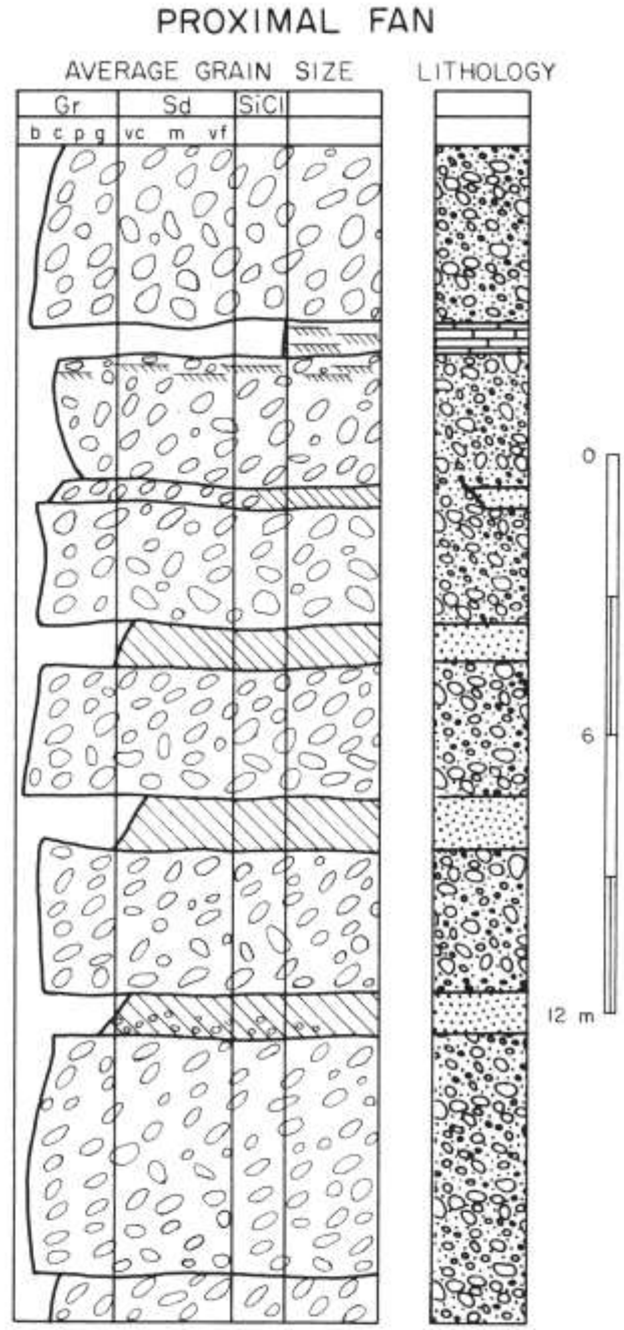

FIGURE 3—Typical vertical section of proximal-fan facies (data from Mogollon Rim area). See Fig. 4 for explanation of symbols.

mid-fan deposits that occur in the Socorro area are indicative of eastward drainage off the Sierra uplift into the Carthage–La Joya basin.

6. Small, isolated deposits of westward-flowing proximal braided streams in the Hubbell bench area represent drainage from the Sandia uplift.

Proximal-fan facies

Characteristics—This facies is exposed in the Mogollon Rim, Socorro, and Hubbell bench areas. The canyon-filling conglomerates exposed in the Lemitar Mountains are also included in this facies, although they apparently represent feeder-channels updip of a fan system. A typical vertical section of the proximal-fan facies (Fig. 3) shows a dominance of massive to horizontally stratified conglomerate, with interbedded lenses of very coarse grained sandstone. Figure 4 is an explanation of symbols used in this and subsequent vertical sections. Conglomerate clasts are generally pebble- to boulder-sized, clast supported, crudely imbricated, and lie within a sand matrix. Conglomerate units are laterally extensive and are as much as 6 m thick in the Mogollon Rim area (Fig. 5). Their thickness in the Socorro, Lemitar, and Hubbell bench areas, however, is generally less than 1.5 m.

Sandstone occurs in lense-shaped units as much as 1 m thick and dominantly exhibits horizontal lamination and medium- to large-scale planar foreset bedding. Contacts between units are generally sharp and do not show high-relief scour surfaces. Local, dense layers of microcrystalline calcium carbonate occur in proximal-fan deposits in the Mogollon Rim area; however, these were not observed in such deposits in other areas.

Depositional processes—The conglomerates of the proximal-fan facies are equivalent in structure, texture, and depositional-unit geometry to perennial

FIGURE 5—Horizontally stratified conglomerate and sandstone lenses of the proximal-fan facies, Mogollon Rim area. Jacob-staff increments are 0.3 m.

braided-stream deposits in the upper reaches of modern proglacial outwash fans (Boothroyd and Ashley, 1975; Nummedal and Boothroyd, 1976; Boothroyd and Nummedal, 1978). Using these modern fans as a depositional analogue, the proximal-fan conglomerates are interpreted as deposits of laterally extensive, longitudinal bars that were active during infrequent high floods. The thickness of individual gravel depositional units suggests that the duration of individual flood events may have been greater in the Mogollon Rim area than in the Socorro, Lemitar, and Hubbell bench regions.

Deposition of sandstone units occurred by: (1) construction of foresets on longitudinal-bar margins (Rust, 1972; Boothroyd and Ashley, 1975); (2) infilling of shallow scours and channels; and (3) downstream migration of linguoid bars (Miall, 1977, 1978; Collinson, 1970). Sand deposition took place during waning flood stages.

Matrix-supported debris-flow deposits, typical of

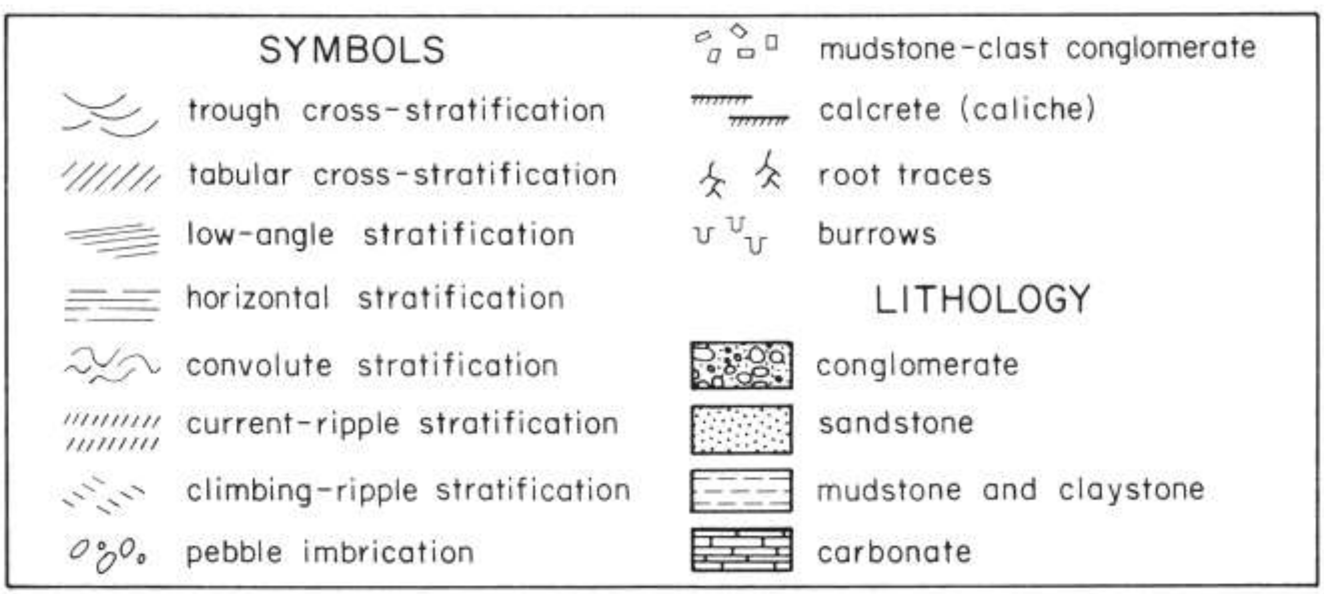

FIGURE 4—Explanation of symbols used in vertical sections.

alluvial fans in arid regions (Bull, 1972; Hooke, 1967), are very rare in the proximal-fan facies. Local, dense calcium-carbonate layers exposed in the Mogollon Rim area are interpreted as paleocaliche horizons. These zones formed pedogenically, presumably resulting from a shift in the axis of sedimentation and long-term exposure of the fan surface.

Local, lenticular units of pebble to boulder conglomerate and minor sandstone, correlated with the Baca Formation by Chamberlin (1981), are present in several isolated exposures in the Lemitar Mountains. Because stratification types, depositional-unit geometries, and textural relations in these conglomerates are similar to those of other exposures of the proximal-fan facies, depositional processes are also inferred to be similar. The exposures in the Lemitar Mountains are channel shaped in transverse cross section, with thicknesses as great as 20 m and widths as much as 400 m. The conglomerates show evidence of westward paleoflow, consist dominantly of upper Paleozoic limestones and sandstones, and grade upward into the volcaniclastic sandstones and conglomerates of the Spears Formation. The base of each of these lenticular, conglomeratic deposits rests disconformably on strata of Pennsylvanian age. We interpret these deposits to represent feeder-channel complexes within paleocanyons that supplied detritus to alluvial fans in the eastern part of the Baca basin.

Precise interpretation of the depositional setting of the proximal braided-stream deposits on the Hubbell bench is greatly hindered by their small outcrop area and poorly known contact relations. However, due to the fact that these deposits apparently overlie strata of Permian age (Kelley, 1982), they probably represent paleocanyon-fill, such as in the Lemitar Mountains, or an erosional remnant of the proximal portion of a fan that had onlapped its source region.

Mid-fan facies

Characteristics—This facies is present in the Springerville, Datil Mountains, Bear Mountains, and Socorro areas. In contrast to the proximal-fan facies, mid-fan conglomerates occur as lense- or channel-shaped units as much as 30 m wide and 5 m thick. Conglomerates typically exhibit crude horizontal stratification and imbrication (Fig. 6), but large-scale crossbedding is locally present (Fig. 7).

Sandstones are very coarse to fine-grained and occur in laterally extensive sheet-like deposits and in channel-shaped units as wide as 50 m. Sedimentary structures present within sandstones include horizontal stratification, trough cross-stratification, low-angle stratification, and rare planar cross-stratification. Graded laminae and graded beds as much as 2 m thick were commonly observed. A typical vertical section of the mid-fan facies is presented in Fig. 8.

Depositional processes—The horizontally stratified conglomerates of the mid-fan facies are longitudinal-bar and channel-lag deposits, and are analogous to such deposits in modern braided streams described by Williams and Rust (1969), Rust (1972), and Church and Gilbert (1973). Large-scale tabular crossbedding in conglomerate units was formed by gravel transverse-bar migration (Hein and Walker, 1977; Miall, 1977), gravel point-bar accretion (Martini, 1977), and, more rarely, by deposition on slip faces of longitudinal bars (Rust, 1978; Boothroyd and Nummedal, 1978). Channel-margin accretionary bedding, such as that shown in Fig. 7, was probably formed by deposition on a gravel point bar. Conglomerate units that exhibit tabular cross-stratification and are not associated with channel margins are considered to be deposits of transverse gravel bars. Gravel accretion on longitudinal-bar margins is represented by foreset-bedded conglomerate that grades laterally into horizontally bedded conglomerate (Rust, 1978).

The mid-fan facies contains more channel-shaped conglomerate units than does the proximal-fan facies. This observation is consistent with the downdip bifurcation and increase in number of channels noted

FIGURE 6—Horizontally stratified gravel-bar deposits with sandstone foresets developed on margin, mid-fan facies, Springerville area. Jacob-staff increments are 0.3 m.

FIGURE 7—Channel-fill conglomerate with large-scale foresets, mid-fan facies, Datil Mountains area. Jacob-staff increments are 0.3 m.

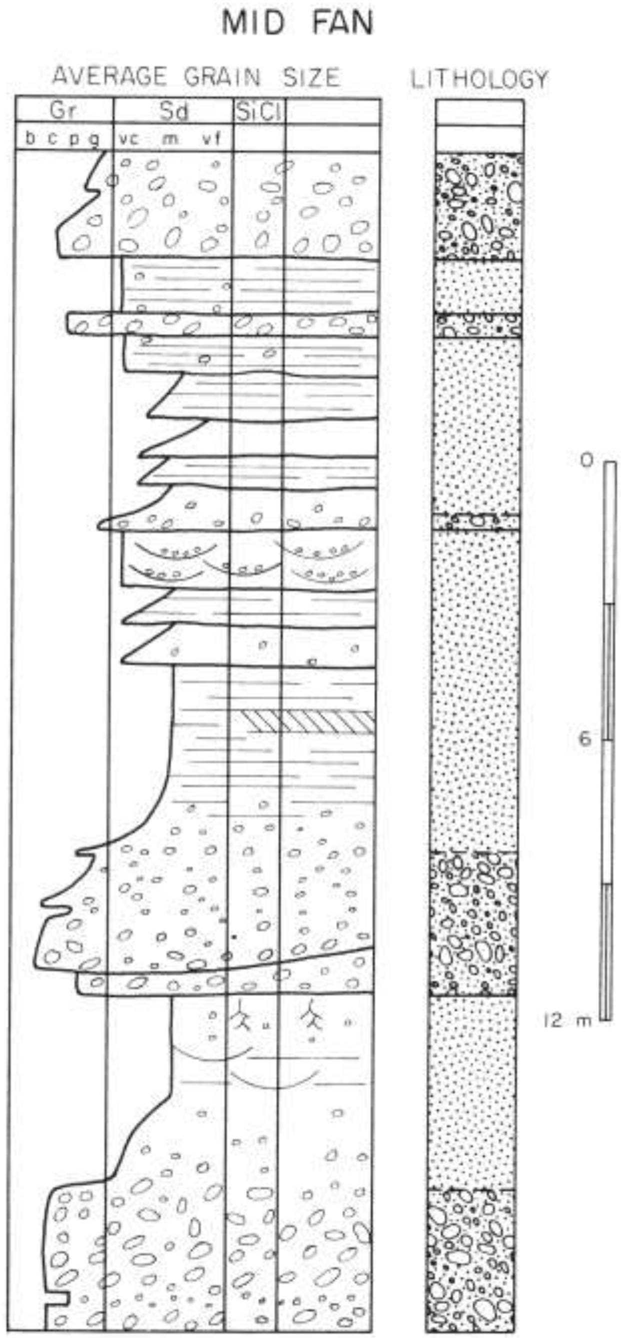

FIGURE 8—Typical vertical section of mid-fan facies (data from Springerville area). See Fig. 4 for explanation of symbols.

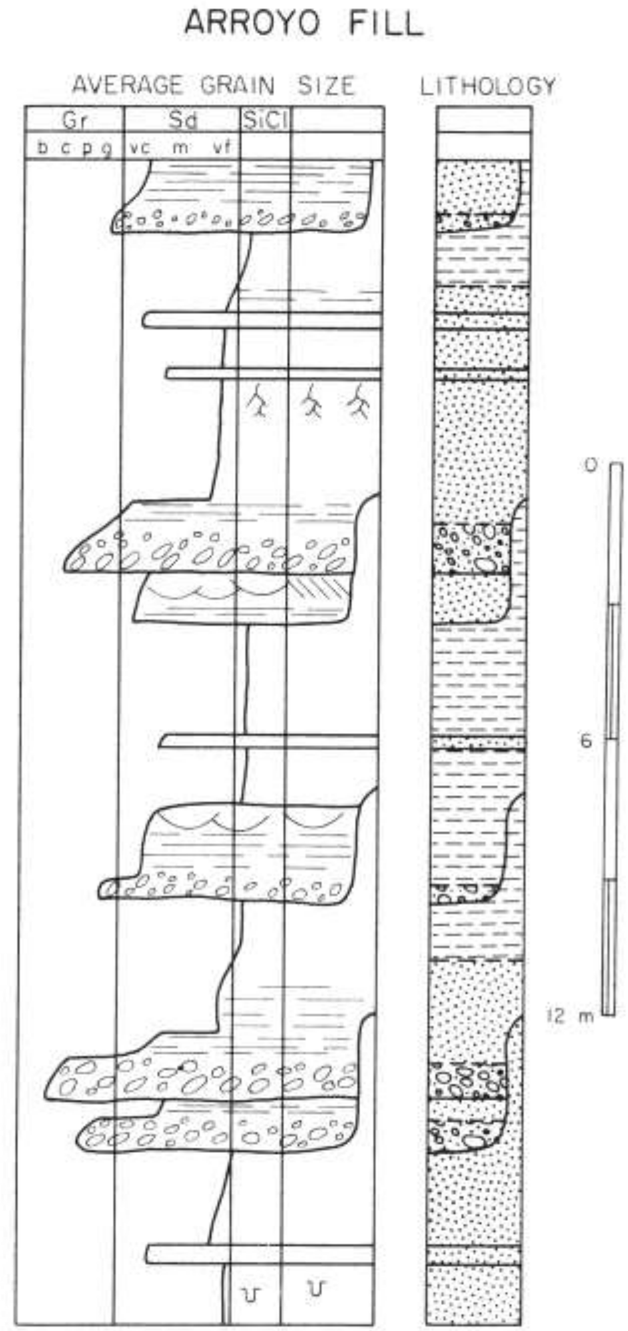

FIGURE 9—Typical vertical section of arroyo-fill facies (data from Socorro area). See Fig. 4 for explanation of symbols.

on proglacial outwash fans by Boothroyd and Nummedal (1978).

Sand was deposited dominantly by migrating bedforms (primarily dunes) in shallow braided channels and by plane-bed aggradation in overbank areas. The sandstone units of the mid-fan facies are very similar to the distal-fan facies, and will be discussed in detail in that section.

Arroyo-fill facies

Characteristics—The arroyo-fill facies is exposed in the southern part of the Socorro area, near the Carthage coal field. Sandstone and conglomerate occur in subequal proportions in this facies; thus, it is considered to be a variant of the mid-fan facies. A typical vertical section (Fig. 9) shows fine sandstone and sandy mudstone interbedded with very coarse to medium sandstone and conglomerate. The common occurrence of steep-walled, rectangular channel-shaped units of conglomerate and coarse sandstone as much as 100 m in width (Fig. 10) distinguish this facies from the mid-fan facies. Channel-fill deposits are poorly sorted and exhibit abrupt lateral changes in texture and sedimentary structures. Sedimentary structures within

FIGURE 10—Coarse-grained, arroyo-fill deposits incised into laterally continuous fine-grained sandstones and mudstones, Socorro area. Paleoflow was toward viewer. Socorro Peak on skyline at center.

channel units include crudely stratified conglomerate, horizontally laminated sandstone, and, more rarely, trough cross-stratified sandstone and medium-scale tabular cross-stratified sandstone. Convolute stratification in sandstones was also noted at several localities.

Channel-shaped units are incised into laterally persistent, sandy mudstones and thin sandstone beds. Burrows and root traces are common in the fine-grained deposits and may be largely responsible for the general lack of stratification in these beds. Faint horizontal stratification, however, is locally present in some of the non-channelized deposits.

Depositional processes—The arroyo-fill facies is interpreted to represent deposition by unconfined vertical aggradation of the fan surface alternating with episodes of arroyo incision and infilling. The vertical-walled aspect of the channels in this facies and the coarse grained nature of the channel-fill is similar to modern arroyos throughout the study area and to the arroyos described by Love (1977), Leopold and others (1964), Leopold and Miller (1956), and Schumm and Hadley (1957). Love (1977) described alternating arroyo cutting and unconfined flooding in Chaco Canyon, New Mexico (Fig. 11), which may be a modern analogy to the depositional processes of the arroyo-fill facies. Love (1977) suggests that arroyo entrenching may be related to increased stream power during periods of increased rainfall. Various other causes of arroyo formation have been proposed, including changes in rainfall intensity (Leopold, 1951) and climatic and tectonic effects (Leopold and others, 1964). Schumm and Hadley (1957), however, suggest that alternating arroyo entrenchment and infilling are part of the natural semiarid erosion cycle.

Based on preserved sedimentary structures, gravel-bar formation and plane-bed aggradation during upper-flow-regime conditions were the dominant sedimentation processes in the arroyo channels. Dune

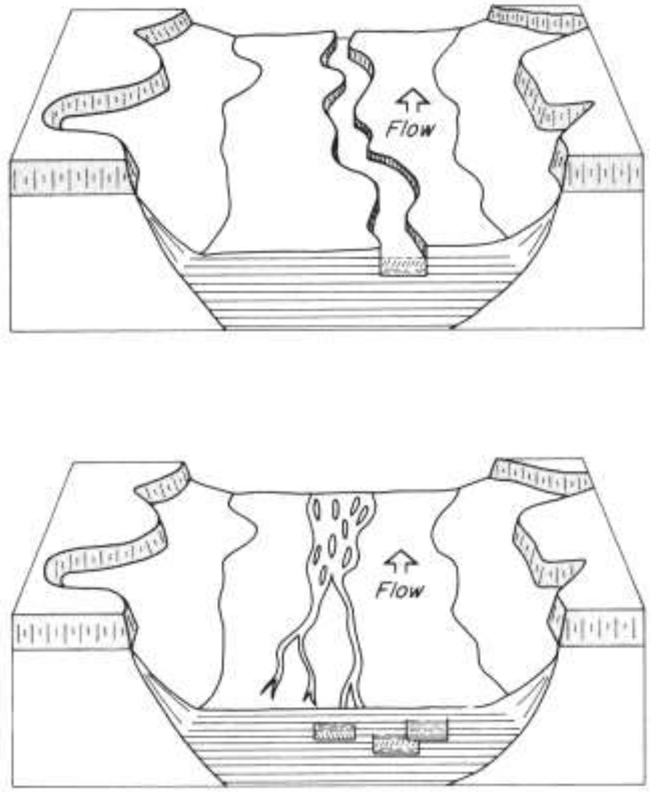

FIGURE 11—Alternating arroyo incision (above) and unconfined flooding (below) in Chaco Canyon, New Mexico. Modified from Love (1977).

migration and progradation of transverse bars occurred locally. The poor sorting of the channel-fill deposits suggests high rates of deposition, possibly due to rapid infiltration of flood waters. Lack of large-scale lateral accretion surfaces, such as those described by Shepard (1978) in the degradational deposits of Tapia Canyon, New Mexico, implies that the incised channels of the arroyo-fill facies were not highly sinuous.

The fine-grained, laterally continuous units of the arroyo-fill facies were deposited as overbank flood deposits and sheet wash during periods of unconfined flow. The abundance of root traces in some of these beds indicates at least seasonally wet conditions.

Distal-fan facies

Characteristics—The distal-fan facies is exposed in the Springerville, Quemado, Datil Mountains, Gallinas Mountains, and Bear Mountains areas. A typical vertical section (Fig. 12) shows a dominance of very coarse to fine sandstone with subordinate conglomerate, mudstone, and claystone. The prevalent mode of occurrence of sandstone in this facies is sheet-like bodies that contain abundant horizontal laminations (Fig. 13). Individual laminae are generally not graded, but graded laminae, as well as larger scale, upward

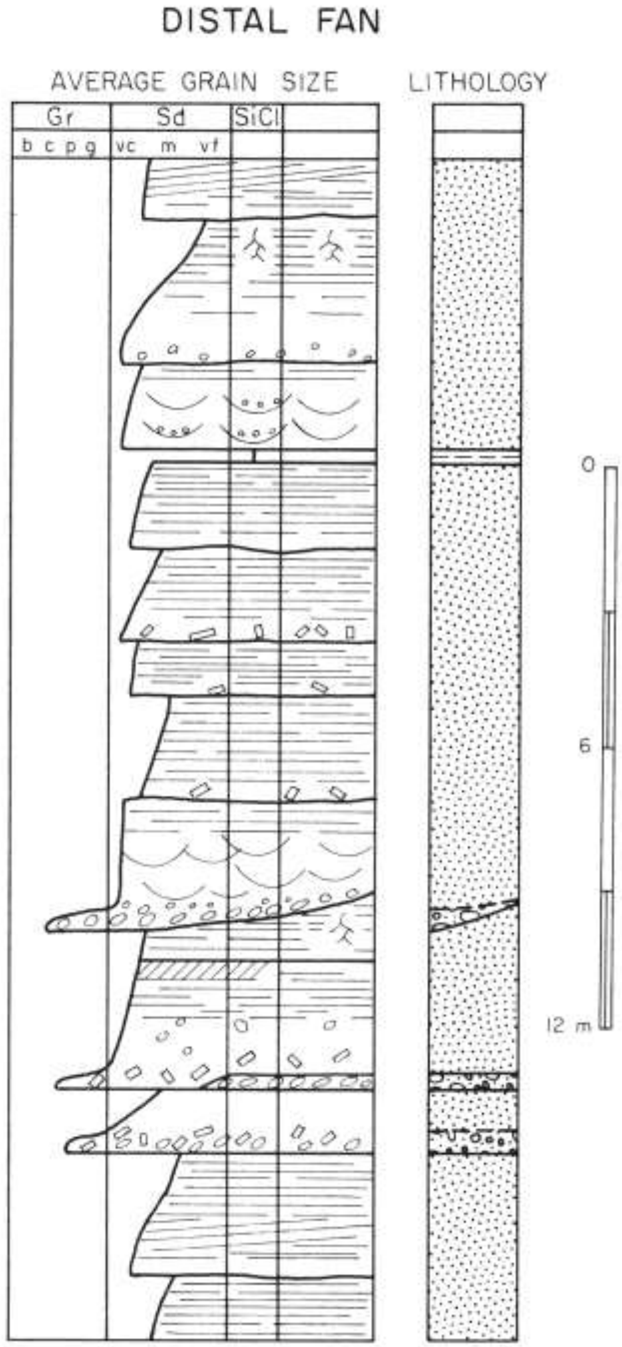

FIGURE 12—Typical vertical section of distal-fan facies (data from Quemado area). See Fig. 4 for explanation of symbols.

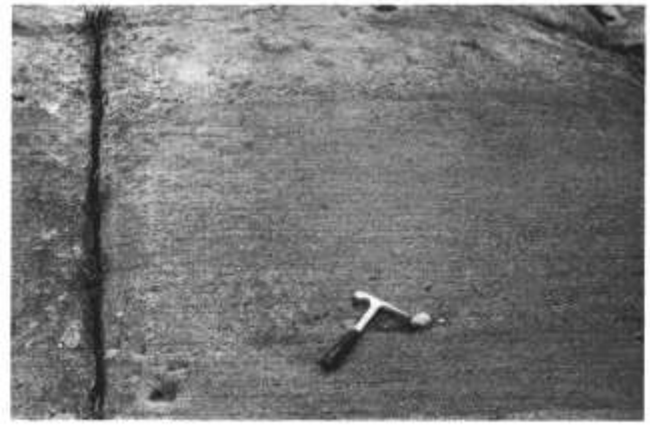

FIGURE 13—Horizontally stratified flood deposits, distal-fan facies, Quemado area. Several graded intervals are present.

fining units as much as 2 m thick, were sometimes observed. Low-angle stratification and parting lineation are commonly associated with the horizontally stratified sandstones.

Trough cross-stratification is common in the distal-fan facies and generally occurs as cosets of mutually cross-cutting, medium- to large-scale festoon cross-bedding in broad, shallow, channel-shaped sandstone units as much as 50 m wide (Fig. 14). More rarely, solitary trough-shaped scour-fill, as well as minor planar cross-stratification, occurs within the horizontally stratified sandstone bodies. Conglomerates form very thin channel- and sheet-like deposits as much as 15 m wide, and also occur as small lenses at the base of trough-shaped scours. Conglomerates locally contain abundant clasts of intraformational mudstone and claystone, commonly exhibit crude horizontal stratification and imbrication, and may be clast supported or occur as scattered pebbles in a stratified sand matrix.

In contrast to the proximal- and mid-fan facies, the distal-fan facies locally contains appreciable amounts of mudstone and claystone that occur as thin, laterally continuous beds. Rare horizons of nodular caliches and locally abundant root traces also occur.

Depositional processes—Vertical textural trends and stratification types in this facies are comparable to those described by McKee and others (1967) in recent flood deposits along ephemeral Bijou Creek, Colorado. Large-scale, upward-fining units in horizontal stratified sandstones record deposition by plane-bed aggradation during waning flood stages. Low-angle stratification represents filling of shallow scours during high-velocity flow (Rust, 1978). Graded laminae,

FIGURE 14—Trough cross-stratified channel-fill, distal-fan facies, Quemado area.

common in the horizontally stratified sandstones of the distal-fan facies, may be the result of minor velocity fluctuations during plane-bed aggradation (perhaps including the "burst and sweep" phenomenon of Jackson, 1976a, and Bridge, 1978) or the migration of very low amplitude sandwaves in shallow water (Smith, 1971). Similar small-scale, graded beds have been reported in the Lower Platte River (Smith, 1971) and in recent ephemeral streams in Kenya (Frostick and Reid, 1977).

The dominance of medium- to large-scale trough crossbedding in channel-fill deposits indicates that dune migration was the primary depositional process in the distal-fan channels. Dune development in channels probably occurred dominantly during waning flood stages (McKee and others, 1967) as did minor dune and transverse-bar activity on the adjacent floodplain. Recent analogues of dune-dominated channel systems include the shallow, ephemeral braided streams of Colorado (McKee and others, 1967), central Australia (Williams, 1971), and Israel (Karcz, 1972). The scarcity of planar crossbedding in channel deposits of the distal-fan facies differs from perennial braided-stream models (Smith, 1970; Rust, 1972; Cant and Walker, 1976; Costello and Walker, 1972).

Thin mudstones and claystones present in the distal-fan fades represent floodplain and ephemeral-pond deposits. Reworking of these fine-grained sediments by subsequent floods produced the clay clasts commonly observed in conglomerates deposited as channel-bottom lags and in scours leeward of advancing dunes.

Braided alluvial-fan model

The braided alluvial-plain system consists of the deposits of a series of coalesced humid alluvial fans. Although discrete fans probably existed directly adjacent to surrounding uplifts, the lack of radial paleocurrent indicators within basin-fill deposits suggests basinward integration of fans into broad, braided alluvial plains or piedmont slopes.

The most commonly cited recent examples of humid alluvial fans are the proglacial braided-outwash fans of Alaska and Iceland (Boothroyd and Ashley, 1975; Gustayson, 1974; Boothroyd and Nummedal, 1978) and the Kosi tropical fan of India (Gole and Chitale, 1966). Humid alluvial fans differ from arid fans in several respects:

1. Flow in braided streams on humid fans is perennial or seasonal, in contrast to the ephemeral-discharge characteristics typical of arid fans.
2. Humid alluvial fans exhibit smooth, gentle downstream gradients, ranging from about 6 m/km (proximal) to 3 m/km (distal) in proglacial outwash fans (Boothroyd and Nummedal, 1978) and from 1 to 0.2 m/km in the Kosi fan (Gole and Chitale, 1966).
3. Humid fans are much more areally extensive than their arid counterparts. The relatively uniform flow characteristics of humid fans result in steady, long-term deposition, which favors development of large, low-gradient surfaces on these fans.
4. Unlike arid alluvial fans, humid fans consist almost exclusively of braided-stream sediments and characteristically lack significant debris-flow and sieve deposits. The importance of debris-flow and sieve deposits in arid fans is due to rapid erosion and runoff resulting from sporadic, intense rainfall events typical of arid climates.
5. Humid fans exhibit systematic downstream decrease in grain size. In arid fans, the sporadic supply of coarse, unsorted sediments to all parts of the fan by debris flows hinders the development of such downstream textural trends (J. C. Boothroyd, 1979, oral commun.).

Certain aspects of the proglacial outwash fans of Alaska and Iceland and the Kosi fan of India provide useful modern analogues for parts of the braided alluvial-plain system discussed in this paper. However, several important differences exist between these modern fans and the ancient braided alluvial-plain systems present in the Baca and Carthage—La Joya basins. First, although the downdip extent of the large alluvial plain present in the western part of the Baca basin is of a size (about 150 km) similar to that of the Kosi fan, it is much larger than typical proglacial braided-outwash fans (10-35 km). The smaller size of the outwash fans can probably be explained, however, by the relatively short duration during which they have been active (late Pleistocene to Recent), in contrast to the Baca fan system which was presumably active for several million years during the Eocene. The size of modern outwash fans is roughly similar to that of the poorly exposed fan systems present in the Carthage—La Joya basin and the eastern part of the Baca basin.

Second, important sedimentologic differences exist between the Baca fan systems and the recent examples described in the literature. Although the proximal portions of the Kosi River fan contain pebble to boulder gravels similar to the proximal-fan fades, the lower reaches of the Kosi fan contain much more fine-grained sediment than analogous parts of the Baca braided alluvial-plain system. Similarly, the textures and stratification types present in the upper reaches of modern proglacial outwash fans are very much like the proximal-fan facies of the Baca and Carthage—La Joya basins. However, the mid- and distal-fan facies show evidence of flashy, perhaps ephemeral, flow and thus differ markedly from proglacial examples.

Sedimentary structures present in braided alluvial-plain deposits of the Baca and Carthage—La Joya basins indicate that discharge was increasingly flashy basinward. Proximal-fan conglomerates are similar to the perennial "Scott-type" of Miall (1977, 1978), whereas the distal-fan facies is much like Miall's ephemeral "Bijou Creek-type," although containing somewhat more trough crossbedding and conglomerates. The mid-fan facies seems to be transitional between these two end-members. The downdip change in paleohydrologic regime implied by this facies succession may be due to: (1) the presence of a wetter microclimate in the highlands surrounding the Baca and Carthage–La Joya basins, giving rise to seasonal or perennial flow near the mountain fronts; and (2) increased infiltration and evapotranspiration within the basins, which would tend to increase the discharge peakedness in the distal-fan environment. Evidence for downfan increase in discharge flashiness is best developed in the large fan system present in the western part of the Baca basin, although it also occurs to a lesser extent in other areas of the Baca and Carthage–La Joya basins.

The climate in west-central New Mexico during Baca time was semiarid and savanna-like (see paleoclimate section). Climatic effects seem to best explain the differences between the braided alluvial-plain system discussed in this paper and the proglacial outwash model, as well as to provide controls on other modes of deposition within the Baca basin.

Meanderbelt-fluvial system

Fine-grained meanderbelt facies

Characteristics—The fine-grained meanderbelt fades is exposed in the Datil Mountains and Dog Springs Canyon segments of the outcrop belt. This facies consists of tabular-shaped sandstone units, as much as 30 m thick and 1 km wide, interbedded with mud-stone and siltstone; fine-grained rocks generally make up greater than 50% of this facies. A typical vertical section (Fig. 15) shows a pronounced stacking of genetic cycles within the sandstone lithofacies (multistory sandstone units). Individual cycles average approximately 4-6 m in thickness and display upward-fining textural trends and relatively consistent vertical arrays of sedimentary structures. Thin interbeds of bioturbated, fine-grained deposits locally occur between superposed sandstone cycles. The base of an ideal genetic cycle consists of an erosional, high-relief scour surface overlain by massive to horizontally stratified conglomerate (commonly intraformational) and very coarse sandstone. This is overlain by upward-fining sandstone displaying, in order of ascending stratigraphic position, large-scale trough cross-stratification, medium-scale trough cross-stratification and horizontal laminations, and current- and climbing-ripple laminations. Where present, the uppermost part of a typical cycle consists of fine to very fine sandstone and siltstone that exhibit abundant root traces and common burrows. However, due to

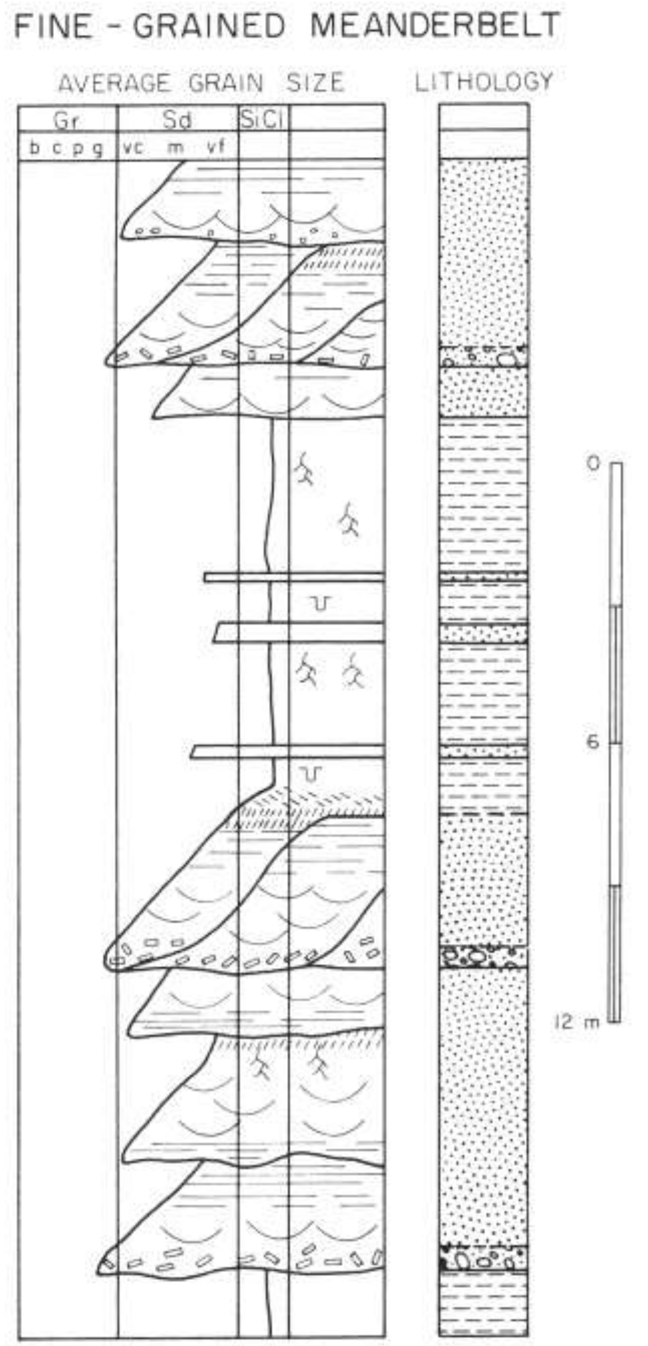

FIGURE 15—Typical vertical section of fine-grained meanderbelt facies (data from Datil Mountains area). Heavy sigmoidal lines indicate epsilon cross beds. See Fig. 4 for explanation of other symbols.

the stacked, multistory nature of genetic units in the fine-grained meanderbelt facies, the upper portions of individual cycles are commonly erosionally truncated by the base of superjacent cycles. Where inter-fingering of the sandstone lithofacies and the mudstone lithofacies occurs, individual sandstone units representing complete genetic cycles are commonly preserved. Large-scale epsilon crossbedding (Allen, 1965) is locally well developed (Fig. 16).

The mudstone lithofacies consists dominantly of thick, tabular-shaped mudstone units with minor siltstone and thin interbeds of very fine to medium sand-

FIGURE 16—Point-bar sandstone showing large-scale epsilon crossbeds. Fine-grained meanderbelt facies, Datil Mountains area. Sandstone unit is approximately 4 m thick.

stone, although local channel-shaped mudstone units are also present. Mudstone and siltstone are typically structureless, but locally may show current- and climbing-ripple lamination, burrows, and root traces. Sandstone beds range from 5 cm to 0.6 m in thickness and exhibit horizontal lamination, current-ripple lamination, and rare root traces and burrows.

Depositional processes—Numerous criteria for the recognition of point-bar accretion in a meandering-river environment have been proposed in the literature. The fine-grained meanderbelt facies exhibits all of the major characteristics, including upward-fining textural sequences (Allen, 1964; Visher, 1972; Masters, 1967), vertical decrease in scale of sedimentary structures (Allen, 1964; Visher, 1972; Belt, 1968; Bernard and others, 1970), and the presence of large-scale epsilon crossbedding (Allen, 1965; Cotter, 1971; Moody-Stuart, 1966). Bernard and others (1970) described the distribution of grain sizes and sedimentary structures in a modern meander of the Brazos River, Texas. The vertical sequence of structures and textures resulting from point-bar accretion in the Brazos River is similar to that exposed in sandstones of the fine-grained meanderbelt facies.

The fine-grained rocks locally interbedded between point-bar sandstones are topstratum deposits (Allen, 1965). The highly bioturbated, horizontally laminated and ripple-laminated deposits locally preserved at the top of point-bar cycles probably represent levees. They commonly show abrupt textural variation between fine sandstone, siltstone, and mudstone that characterizes modern levees (Fisk, 1944).

Truncation of upper parts of genetic sandstone cycles results from multistory stacking of point-bar deposits during downdip migration of individual meander loops within the meanderbelt system. Comparable ancient, stacked point-bar sequences have been described by Belt (1968), Steel (1974) and Allen (1964). Puigdefabregas and van Vliet (1978) suggest that the occurrence of multistory sandstone sequences, as opposed to isolated point-bar deposits encased within floodplain mudstones, indicate that the meanderbelt occupied a somewhat stable position within the floodbasin and that vertical aggradation rates were not high relative to the rate of meander migration.

Laterally persistent mudstone and siltstone beds represent vertically accreted floodbasin deposits. The thin, fine-grained sandstone interbeds are interpreted as crevasse-splay deposits that entered the floodbasin through breaches in levees during floods. The thickness of these sandstone beds, as well as the sedimentary structures and root traces within them, is similar to recent crevasse splays (Happ and others, 1940). Channel-shaped mudstone units, commonly associated with point-bar sandstone sequences, represent abandoned-channel fill. Interbedding of lacustrine delta deposits and point-bar sequences in the Datil Mountains area suggests that the floodbasins were periodically occupied by lakes.

Size of epsilon crossbedding can be used to estimate paleochannel width and depth. The width of the crossbed set is approximately two-thirds of the bankfull channel width (Allen, 1965; Moody-Stuart, 1966; Cotter, 1971). Bankfull depth in straight reaches of the paleochannel can be estimated by the following relationship (Ethridge and Schumm, 1978):

Bankfull depth = epsilon cross-bed
height x 0.585/0.9

The width of multistory sandstone units provides a rough estimate of average meanderbelt width.

Using these criteria in conjunction with paleocurrent data, we delineate two distinct fine-grained meanderbelt paleoenvironments in the Datil Mountains area. Their paleohydrologic characteristics are summarized in Table 1. We interpret the smaller, stratigraphically lower point-bar sequences as deposits of southeast-flowing tributary streams that drained the Zuni uplift. The larger paleomeanderbelt system represents an east-flowing, integrated river system that occupied the axial part of the Baca basin. In addition to these fluvial trends, Chamberlin (1981) noted the presence of a northeast-flowing tributary system in the southwestern part of the Datil Mountains area.

TABLE 1—Paleohydrologic characteristics of the fine-grained meanderbelt facies. *Based on 3 examples; **based on 4 examples; ***based on 16 examples.

Type of paleostream	Bankfull paleochannel size Width	Depth	Paleomeanderbelt width
Axial (east-flowing)	65 m*	4.2 m*	525 m***
Tributary (southeast-flowing)	35 m**	2.2 m**	185 m***

Coarse-grained meanderbelt facies

Characteristics—This facies is restricted to the upper-middle part of the Baca Formation in the Datil Mountains area and is overlain and underlain by the

fine-grained meanderbelt facies. A typical vertical section of the coarse-grained meanderbelt facies is presented in Fig. 17. The sandstone lithofacies shows poorly developed stacking of genetic cycles (multilateral sandstone units), scarcity of ripple-laminated fine-grained sandstone and siltstone, poorly developed upward-fining textural trends, abundance of horizontally laminated sandstones, and local large-scale foreset bedding as much as 2 m thick (Fig. 18). These features distinguish the coarse-grained meanderbelt facies from its fine-grained counterpart. The presence of significant amounts of mudstone and the local occurrence of epsilon crossbedding in the coarse-grained meanderbelt facies differs from the braided-stream deposits found elsewhere in the Baca basin. The mudstone lithofacies of the coarse-grained meanderbelt facies is similar to, but less abundant than, that of the fine-grained meanderbelt facies.

FIGURE 18—Cross-stratified chute-bar deposit overlain by horizontal stratification. Coarse-grained meanderbelt facies, Datil Mountains area.

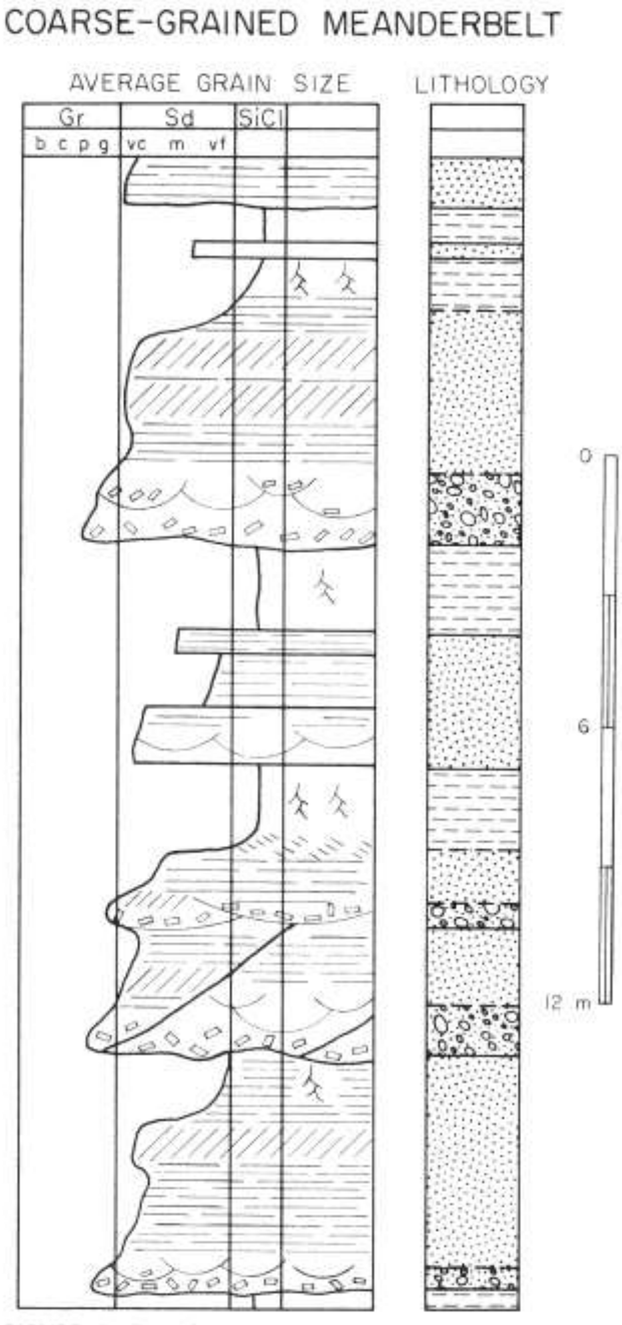

FIGURE 17—Typical vertical section of coarse-grained meanderbelt facies (data from Datil Mountains area). Heavy sigmoidal lines indicate epsilon cross beds. See Fig. 4 for explanation of other symbols.

Depositional processes—The sedimentologic characteristics of sandstones in this facies indicate deposition by chute-modified point bars (McGowen and Garner, 1970); these are commonly termed coarse-grained point bars because the fine-grained, upper part of the point bars is replaced by coarse-grained chute-fill and chute-bar deposits. The large-scale foreset crossbedding in this facies records avalanche-face accretion on chute and transverse bars that formed in response to sediment transport across the point bar during floods. Similar foresets can form in large bedforms in a fine-grained meanderbelt environment, such as scroll bars (Jackson, 1976b) and spillover lobes (Bernard and others, 1970). However, the large foresets in the coarse-grained meanderbelt facies probably cannot be attributed to these types of bedforms. The overall change in point-bar characteristics and the decrease in sandstone/mudstone ratio in this facies, as well as the restriction of this facies to a specific stratigraphic interval, support a change to a coarse-grained meanderbelt regime.

The abundant horizontally laminated sandstones of this facies were deposited during upper-flow-regime, plane-bed aggradation on point bars and floodplains. McGowen and Garner (1970) described similar horizontally laminated flood deposits on modern coarse-grained point bars of the Amite River, Louisiana. Isolated scour troughs associated with the horizontally laminated sandstones may have been generated by turbulence around vegetation on point-bar surfaces and on the floodplain. This interpretation is supported by the abundance of root traces in the horizontally laminated sandstones.

Coarse-grained point bars occur in rivers characterized by unstable, rapidly shifting channels (Mc-

Gowen and Garner, 1970). Lack of channel stabilization may explain the poorly developed stacking of chute-bar deposits over lower-point-bar sandstones in the coarse-grained meanderbelt facies. Instead, the sandstone lithofacies commonly consists of multilateral sequences of (1) trough cross-stratified, lower-point-bar sandstones, (2) large-scale foreset deposits of chute bars, and (3) horizontally laminated upper-point-bar and floodplain sandstones.

Conditions favoring chute and chute-bar development in the modern Amite and Colorado Rivers are high bedload-to-suspended-load ratios, moderate gradients, short peak-flood duration, and sandy banks stabilized by vegetation (McGowen and Garner, 1970). Sediment load and discharge characteristics would support a braided-stream regime, but a meandering pattern is preserved by vegetative stabilization of point bars. Fine-grained fluvial systems, on the other hand, tend to be favored by relatively uniform flow, low gradients, low bedload-to-suspended-load ratio, fine-grained, cohesive bank materials, and vegetatively well-stabilized channels (Leopold and Wolman, 1957; Schumm and Kahn, 1972).

The coarse-grained meanderbelt facies in the Baca basin is both overlain and underlain by fine-grained meanderbelt deposits. Possible reasons for the temporary shift to a coarse-grained meanderbelt regime are: (1) a pulse of tectonism in source regions adjacent to the Baca basin or in intrabasin areas, giving rise to steeper stream gradients and greater supply of bed-load sediments; and (2) a change to a drier climate causing less uniform discharge characteristics and decreased stabilization of channels by vegetation. Although neither of these possible mechanisms can be ruled out, climatic change appears to be the probable cause of a major regressive phase of Eocene lacustrine sedimentation in the eastern part of the Baca basin. These lacustrine low-stand deposits occupy a stratigraphic position similar to that of the coarse-grained meanderbelt facies in the Datil Mountains area; however, evaluation of their time-stratigraphic relations is not yet possible due to sparse paleontologic control.

Lacustrine system

The lacustrine system is widespread throughout the Baca basin. Lacustrine sedimentation took place in two general settings: (1) small, impermanent lakes situated on fluvial-system floodplains in the Springerville, Quemado, and Datil Mountains areas; and (2) a large, shallow, persistent lake located in the Dog Springs Canyon, Gallinas Mountains, and Bear Mountains areas.

In the Bear Mountains area, Potter (1970) divided the Baca Formation into three informal members that he termed, in ascending stratigraphic order, the lower red, middle sandstone, and upper red units. Cather (1980, 1982) extended this terminology to the Gallinas Mountains area and emphasized that these informal units are genetically related to lake-level fluctuations in a large Eocene lacustrine system in the eastern part of the Baca basin. In the Bear Mountains and Gallinas Mountains areas, the lower and upper red units record intervals of dominantly lacustrine sedimentation, while the middle-sandstone unit represents an interval of braided-stream sedimentation during a low stand of the lake (Fig. 19). Although complicated by later faulting and cover, a similar tripartate division of the Baca Formation in the Dog Springs Canyon area is possible. However, in this area the middle-sandstone unit comprises the stacked-sandstone units and intervening mudstones of the fine-grained meanderbelt facies (Robinson, 1981).

The lacustrine deposits of the Baca basin consist of both delta and basin deposits. We have divided the delta deposits into two types (fine-grained delta and fan delta), although a continuous spectrum of delta characteristics exists between these two end-members.

Fine-grained delta facies

Characteristics—Exposures of the fine-grained delta facies are present in the Datil Mountains, Dog Springs

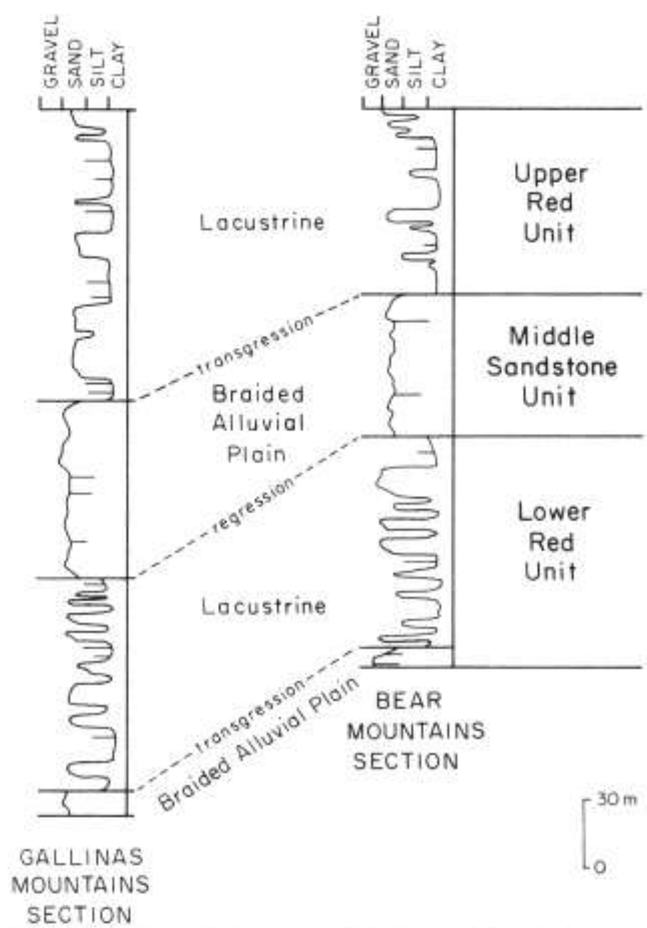

FIGURE 19—Generalized stratigraphic columns of Baca Formation in Gallinas Mountains and Bear Mountains areas, showing relationships between informal units, shoreline migrations, and depositional systems.

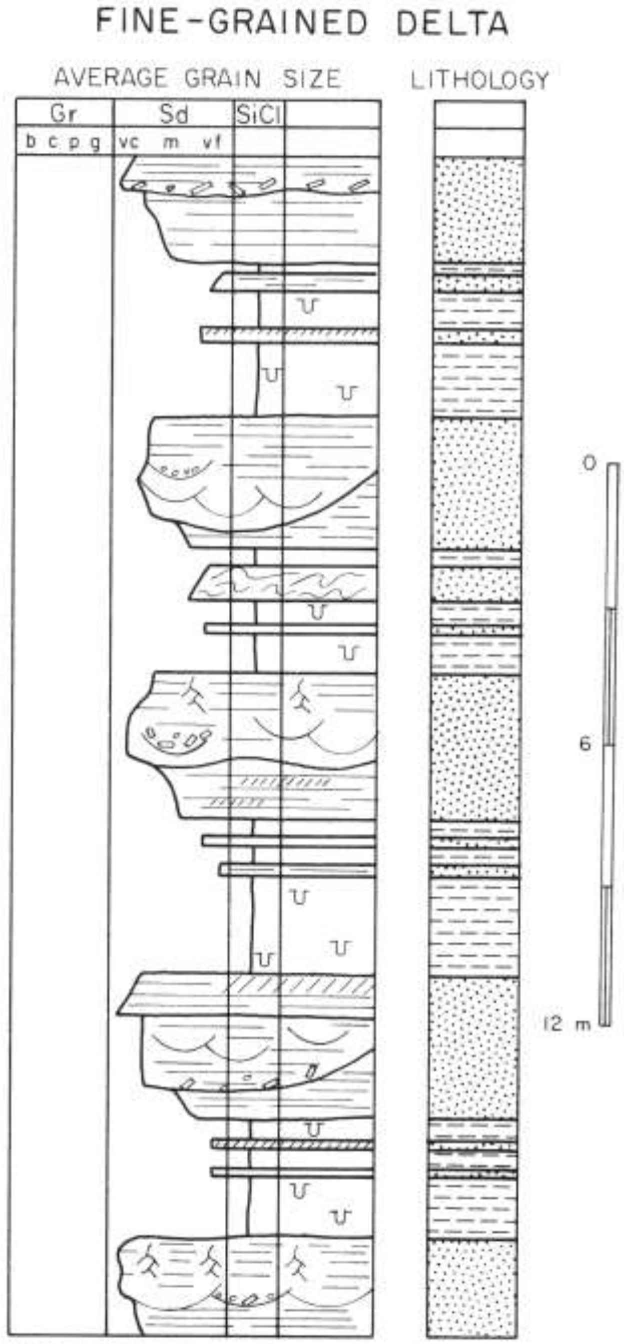

FIGURE 20—Typical vertical section of the fine-grained delta facies (data from Gallinas Mountains area). See Fig. 4 for explanation of symbols.

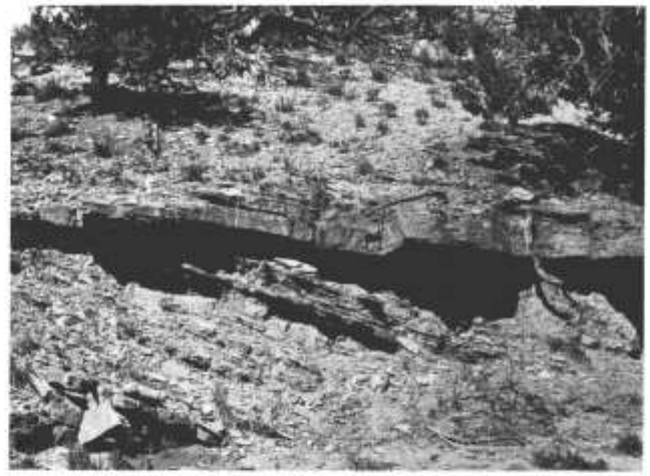

FIGURE 21—Fine-grained delta deposit showing unusually steeply inclined foreset beds, Gallinas Mountains area. Hammer handle points in direction of delta progradation.

Canyon, Gallinas Mountains, and Bear Mountains areas. Cyclical upward-coarsening sequences characterize this facies (Fig. 20). Individual cycles range from approximately 6 to 35 m and average approximately 9 m in thickness.

The basal portion of an idealized cycle consists of a laterally persistent, calcareous mudstone or clay-stone intercalated with thinly bedded (generally less than 25 cm thick), very fine- to medium-grained sandstones. Carbonate content of mudstones, determined by weighing samples before and after acidization with cold, dilute hydrochloric acid, generally ranges between 10 and 15% by weight. Mudstones rarely exhibit laminations and are usually structureless and homogeneous, with the exception of burrows. Burrowing in the mudstones is pervasive. In contrast to sandstones in the lower portions of deltaic cycles, mudstones do not exhibit well-defined burrows, but rather show a churned, curdled texture both megascopically and microscopically, which gives rise to the homogeneous nature of the mudstones. Rare horizons of mudcracks and pedogenic calcite nodules were also observed.

Structures present within the thinly bedded sandstones intercalated with the above-described mudstones include horizontal laminations, current-ripple laminations, parting-step lineation, occasional normal-graded beds, and burrows. Burrows are vertical, horizontal, and oblique, range in diameter from 1 to 5 cm, and sometimes exhibit knobby surface ornamentation and scoop-shaped backfill laminae. These burrows are similar to those created by *Scoyenia sp.*, which is common in non-marine red beds (Hantzschel, 1975), and are believed to have been formed by polychaete worms. The thinly bedded sandstones in the lower portions of cycles are often inclined (Fig. 21), forming large-scale foresets with dip angles ranging up to 15° in rare instances. Dip angles are more commonly only a few degrees (Fig. 22) and are often so gently inclined that the angularity is not readily apparent in a solitary exposure. Soft-sediment deformation is locally well developed in the lower part of deltaic cycles.

The above-described units are transitionally overlain by a horizontal- and current-ripple-laminated, laterally continuous, fine- to coarse-grained sandstone that averages approximately 1.5 m in thickness. Orientation of current-ripple cross laminations usually indicates a direction of flow at high angles to that shown by other paleocurrent indicators within the same deltaic cycle. The well-sorted, nearly homogeneous nature of these sandstones gives rise to a quasi-spheroidal weathering habit (Fig. 22). Coloration of

FIGURE 22—Gently inclined prodelta foreset beds overlain by quasi-spheroidally weathering, delta-front deposits. Fine-grained delta facies, Gallinas Mountains area. Hammer handle points in direction of delta progradation.

this and superjacent sandstones within a given cycle may be red or yellowish gray, whereas the previously described intercalated mudstones and sandstones are almost always red.

Upsection, the next part of an ideal cycle will commonly exhibit large, symmetrical, channel-shaped sandstone units with erosional bases. Channels may be as much as 15 m wide and 5 m deep, but are generally much smaller. Intrachannel sedimentary structures include medium-scale trough crossbedding and plane beds. Where symmetrical channel-shaped units are not present, the base of the upper part of each cycle is represented by an irregular, low-relief, erosional surface. The remainder of the upper portion of individual deltaic cycles comprises the deposits of either the distal-fan facies or the fine-grained meanderbelt facies.

Depositional processes—Cyclical deposits of the fine-grained delta facies are interpreted as recording alternate deltaic progradation and abandonment in a shallow lake. Geometry of lithofacies and vertical and lateral sequences of textures and sedimentary structures are similar to those found in lobate, high-constructive marine deltas (Fisher and others, 1969). Depositional processes are inferred to be the same as in these marine deltas.

The lower intercalated mudstones and thinly bedded sandstones were deposited in a prodelta and distal delta-front environment. Silt and clay were deposited by settling from suspension. The thin sandstone beds are frontal-splay deposits, probably representing prodelta turbidites. Normal-graded beds, a typical feature of turbidites, are occasionally seen in these sandstones. Mudcracks and caliches in the lower portion of deltaic cycles are interpreted as representing lake-level low stands.

The laterally continuous, quasi-spheroidally weathering, horizontally laminated and rippled sandstones are delta-front deposits. These sandstones were deposited in channel-mouth bars or in longshore-current-redistributed bars. Primary processes on the delta front were plane-bed aggradation and ripple migration. The large divergence between paleocurrent directions shown by delta-front ripples and that of other indicators within a given deltaic cycle suggests that ripple-migration direction was predominantly controlled by longshore currents.

The large, symmetrical, channel-shaped sandstones commonly present in the upper part of deltaic cycles are distributary-channel deposits. Sedimentary structures indicate that plane beds and subaqueous dunes were the dominant bedforms within distributary channels. Vegetative stabilization of delta-platform braided-stream channels near the lake may have allowed the development of large, symmetrical, relatively low width-to-depth-ratio distributary channels. Deltaic deposits lacking distributary channels indicate non-stabilization of delta-platform stream channels, as is common in classic "Gilbert-type" delta or fan-delta deposits (Gilbert, 1885; Theakstone, 1976). In the Gallinas Mountains and Bear Mountains areas, the remainder of the upper portion of fine-grained delta deposits is composed of delta-platform sandstones and minor conglomerates deposited by braided-stream processes identical to those of the distal-fan facies. In the Datil Mountains and Dog Springs Canyon areas, however, delta-platform deposits comprise rocks of the meanderbelt system.

Fan-delta facies

Characteristics—The lacustrine fan-delta facies is exposed in the Springerville, Quemado, and Bear Mountains sections of the outcrop belt. Cyclical upward-coarsening deposits characterize this facies (Fig. 23). Sediment caliber and sandstone/mudstone ratios are generally greater than those of the fine-grained delta facies.

The interbedded red sandstones and mudstones in the lower part of upward-coarsening.fan-delta cycles range from subhorizontal beds to steeply inclined foresets (20^0-301 as much as 4 m thick. These deposits exhibit bedding geometries and sedimentary structures similar to those of analogous parts of fine-grained deltas. In the Quemado area, steeply inclined, lower-delta deposits commonly appear to be filling depressions scoured into older strata.

Upsection, the above-described sandstones and mudstones are transitionally overlain by horizontal to current-ripple-laminated sandstones and thin-bedded conglomerate that are as much as 2 m thick and form massive, rounded outcrops. These are, in turn, overlain by sandstones and conglomerates of the mid- or distal-fan facies (Fig. 24).

Depositional processes—The upward-coarsening genetic units of this facies record deltaic progradation of bedload (braided) streams into lakes, forming small fan deltas. Fan-delta sequences with large-scale, steeply inclined bedding are classic Gilbert-type deltas (Gil-

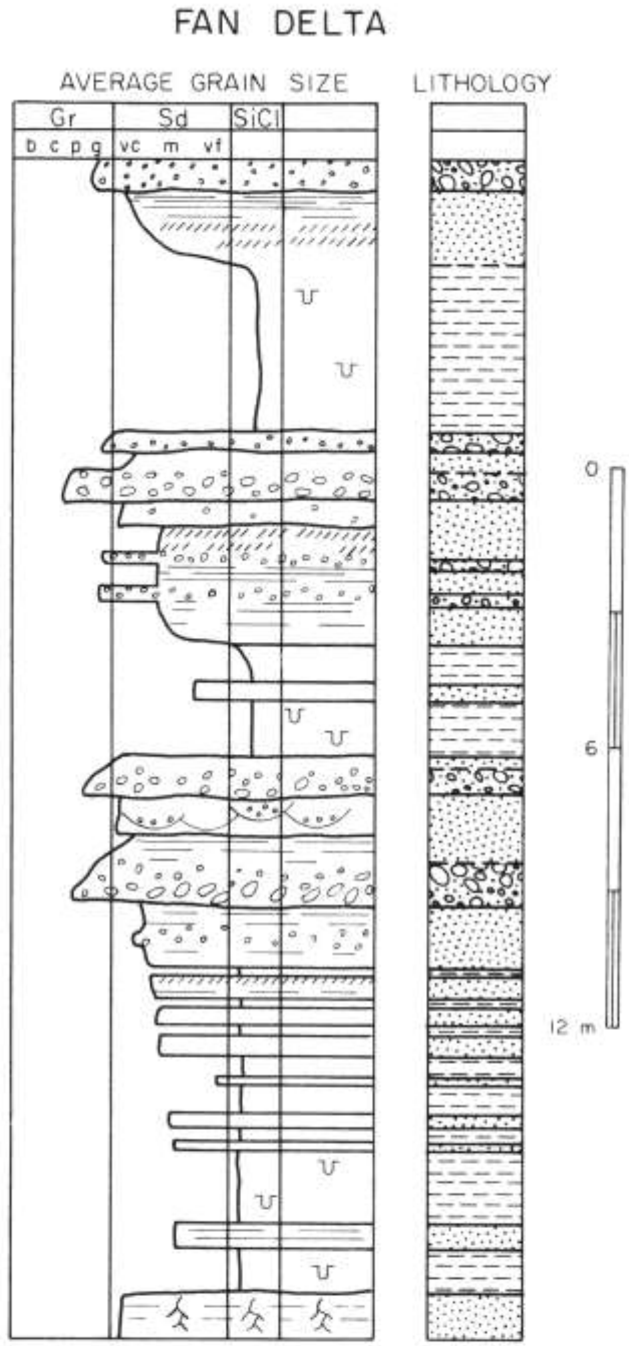

FIGURE 23—Typical vertical section of fan-delta facies (data from Bear Mountains area). See Fig. 4 for explanation of symbols.

Bert, 1885). These deltas typically display well-defined bottomset, foreset, and topset beds. Steep foreset bedding, however, is present in only a few localities in the Baca lacustrine rocks; possible controls on deltaic foreset development are discussed below.

Depositional-unit geometry and vertical sequence of sedimentary structures and textures in the lower part of fan-delta deposits are similar to, although somewhat coarser-grained than, those of the prodelta and delta-front portion of the fine-grained delta facies. Therefore, depositional processes are also inferred to be similar. The style of fluvial sedimentation on the fan deltas, however, differed significantly from

FIGURE 24—Upward-coarsening prodelta deposits (foreground) overlain by delta-front and braided-stream deposits (across arroyo). Fan-delta facies, Bear Mountains area.

that of the fine-grained delta end-member. Well-developed, distributary-channel deposits, which are in common in the fine-grained deltas, were rarely observed in the fan-delta deposits. This may result from decreased vegetative stabilization of braided-stream channels of the fan deltas or from the coarser-grained, less-cohesive nature of the fan-delta deposits.

Basin facies

Characteristics—Basin-facies deposits occur in the Bear Mountains and Gallinas Mountains areas, and are particularly well developed in the lower part of the Dog Springs Canyon section. When present, the lacustrine-basin facies occurs directly beneath the above-described delta facies and consists of calcareous, generally structureless mudstone and claystone with sparse, thin interbeds of very fine-grained sandstone and siltstone. Evidence of burrowing is abundant. The basin facies is identical to the lower prodeltaic portion of the fine-grained delta deposits, with the exception that the sand and silt interbeds of the basin facies are never inclined and are generally thinner and less abundant than those of the delta facies. The boundary between the two facies is arbitrary.

Depositional processes—Basin depositional processes are exactly the same as those of the lower pro-deltaic portion of the fine-grained delta facies, and include settling of silt and clay from suspension, deposition of silt and sand by turbidity currents, and homogenization of mudstones and claystones by burrowing. The thin, sparse nature of the sand and silt interbeds indicates deposition far from nearshore sources of coarse sediments.

Lacustrine-system characteristics

Delta and basin sediments in the large lake that existed in the Dog Springs Canyon, Gallinas Mountains, and Bear Mountains areas show evidence of deposition in a closed lacustrine basin. High concentrations of early authigenic (precompaction) calcite and dolomite in lower-delta and basin sandstones and mudstones suggest evaporative concentration of solutes in a closed-lake environment. The restricted megafaunal assemblage in the Baca lacustrine rocks also favors a closed-lake system. Only a few scattered ostracods were observed. Langbein (1961) notes that closed lakes usually exhibit widely fluctuating water levels and are found exclusively in arid and semiarid regions. The climate during Baca time in west-central New Mexico was probably semiarid and savanna-like. Rare caliches and mudcracked horizons in Baca prodelta and basin deposits are indicative of fluctuating lake levels. The presence of the fluvially dominated middle sandstone unit between the predominately lacustrine lower and upper red units indicates a large-scale regression that was probably caused by a drastic drop in lake-level due to climatic change (see below).

Steeply inclined Gilbert-type foresets are rare in the fine-grained delta facies and only locally well developed in the fan-delta facies. Factors contributing to the development of foresets include: (1) homopycnal flow (inflow density approximately equal to lake-water density) that causes three-dimensional mixing and an abrupt decrease in current velocity, resulting in rapid deposition of sediment (Bates, 1953); (2) deltaic pro-gradation into deep water (McGowen, 1970; Axelsson, 1967); and (3) transport of coarse bedload sediments, of which Gilbert-type foresets are composed, to the distributary mouth (Smith, 1975; Axelsson, 1967). The conditions listed above were uncommon during deposition of the fine-grained deltas in the large lake present in the eastern part of the Baca basin. Sediment caliber was rarely coarser than coarse sand and water depth, as indicated by the thickness of prodelta deposits, was generally less than 6 m. The probable closed nature of the lake implies that lake waters were more dense (due to salinity) than inflowing river water, producing hypopycnal flow and plane-jet formation (Bates, 1953). Lack of three-dimensional mixing causes the plane jet to maintain its velocity over a relatively long distance basinward, resulting in deposition of suspended sediments over a considerable distance from the distributary mouth. This leads to the development of a gently sloping prodelta surface, which produces the typical shallowly inclined prodelta foresets seen in the fine-grained delta facies.

The Gilbert-type deltas in the Baca Canyon area probably formed in response to possible deeper water conditions and the input of coarse, conglomeratic sediments derived from the nearby Sierra uplift. Many of the fan-delta deposits in the Quemado area that display well-developed bottomset, foreset, and topset beds appear to have filled deep scours along the margins of impermanent, interfluvial lakes. These deposits probably record local drainage incision due to a drop in lake level, followed by infilling of the scour by prograding fan-delta sediments during subsequent high stands. Foreset development in these deposits thus appears to have been related to the presence of local deep-water conditions and perhaps also to the probable occurrence of homopycnal flow and three-dimensional mixing in these small, fresh-water lakes.

The laterally persistent nature of the delta-front and delta-platform sandstones and the general lack of destructional-phase features indicate that Baca fine-grained deltas were mainly high-constructive lobate (Fisher and others, 1969), which suggests progradation into relatively shallow water. Thin, destructional-phase shoreface sequences produced by reworking of upper-delta sediments by waves and longshore currents following deltaic abandonment and subsidence are rarely seen. The paucity of destructional-phase deposits suggests relatively low-energy conditions within the lake basin. Attenuation of wave energy resulting from shallow water depths may explain the rarity of destructional-phase features.

The large, shallow lake system present during Eocene time in the eastern part of the Baca basin area probably was polymictic (frequent overturn), because surface mixing due to eddy diffusion would be expected to penetrate at least several tens of feet. Evidence of widespread burrowing in basin and prodelta deposits indicates that lake-bottom sediments were oxygenated, supporting a polymictic regime. Preservation of laminations in lake-bottom sediments is usually restricted to oligomictic and meromictic (permanently stratified) lakes, in which lack of oxygen inhibits the activities of burrowing organisms.

The majority of the red coloration of the Baca Formation (in both lacustrine and fluvial deposits) appears to be due to intrastratal solution of iron-bearing minerals, precipitation of hydrated iron oxides, and subsequent dehydration of these oxides resulting in the development of hematite pigment (e.g., Walker, 1967). The diagenetic reddening of Baca lacustrine sediments was initiated by dissolution of unstable, iron-bearing minerals in a positive-Eh setting. This may have taken place in the oxygenated, lake-bottom environment or via subaerial exposure during periods of lake-level low stands.

In contrast to many lacustrine systems, Baca deltas were not restricted to the periphery of the lake but appear to have prograded well out into the shallow lake basin. The occurrence of intermittant deltaic deposition throughout the lake basin precluded the development of widespread, chemically precipitated and/or strongly reduced, fine-grained clastic deposits typical of lacustrine basin-center environments.

Three possible explanations exist for the large-scale transgressions and regressions represented by the lower red, middle sandstone, and upper red units in the eastern part of the Baca basin: (1) shifting of the locus of lacustrine sedimentation due to tectonism within the basin; (2) increased erosion caused by tectonic activity in the source area, with resultant large-scale regression due to progradation of alluvial aprons basinward; (3) climatic fluctuations with resultant large-scale transgressive and regressive phases. The relative importance of each of the above-listed hypotheses is difficult to evaluate. However, one line of evidence suggests that climatic changes were a major cause of the lake-level fluctuations. The lacustrine mudstones in the Gallinas Mountains area, with the exception of the prodelta mudstone of the basal deltaic cycle of the upper unit, are highly calcareous and indicative of probable saline, closed-lake conditions. The essentially noncalcareous nature of the basal mudstone of the upper red unit (2% calcite by weight as compared with 10-15% in other mudstones) implies a temporary change to more fresh-water conditions during the beginning of the second transgressive phase. If the transgression were due to a change to a wetter climate, the low-saline characteristics could be easily explained by the introduction of large amounts of fresh water to the lake. Neither of the tectonic alternatives can explain the noncalcareous nature of the basal mudstone of the upper red unit. Large-scale lake-level fluctuations in contemporaneous Eocene Lake Gosiute, Wyoming, have been attributed to climatic changes by Surdam and Wolfbauer (1975).

Several other lines of evidence support a lacustrine interpretation for large parts of the Baca Formation in the Bear—Gallinas Mountains area. These include (1) the presence of rare oolites and ostracods and common intraclasts observed in thin sections from the Gallinas Mountains area (Cather, 1980); (2) the occurrence of limestone beds in the lower red unit of the Baca in the Bear Mountains area, which are interpreted by Massingill (1979) to be of lacustrine origin; and (3) palynologic data. In his work on the palynology of the Baca Formation in the Bear Mountains vicinity, Chaiffetz (1979) states that the predominantly pink-gray Baca deposits (i.e., the lower and upper red units, which we interpret to be dominantly lacustrine) yielded only a few poorly preserved pollen grains and the fresh-water alga, *Pediastrum*. However, a greenish-gray shale sample from a thin mudstone in the fluvially dominated middle sandstone unit produced spores and pollen from a wide variety of upland flora, including conifera. Chaiffetz (1979) further states that the presence of *Pediastrum* " . . . perhaps requires some standing fresh-water bodies in the region at that time."

The existence of deltaic deposits in the basal Spears Formation indicates that the lake persisted for a short time during the beginning of Oligocene volcanism. The cause of the final demise of the lake is not known. Likely possibilities include climatic change and rapid infilling of the lake with volcaniclastic sediments.

Lacustrine model

The limited exposures of the Baca lacustrine system make comparison to modern and ancient analogues difficult. However, any model of the large lake present in the eastern part of the Baca basin must account for the following characteristics: (1) shallow water depth; (2) fluctuating lake levels; (3) rarity of lacustrine megafauna; and (4) high concentrations of early authigenic, carbonate cements in lacustrine deposits.

Eugster and Surdam (1973), Eugster and Hardie (1975), and Surdam and Wolfbauer (1975) have proposed a closed-basin, playa-lake model for Eocene Lake Gosiute in Wyoming. Many aspects of the marginal, clastic-dominated areas of this lake adequately fit the characteristics of the Baca lacustrine system. Lake Gosiute was a large, shallow, closed lake that exhibited widely fluctuating lake levels. With the exception of periods of lake-level low stands during which large volumes of trona were deposited in basin-center areas, chemical sedimentation within Lake Gosiute was dominated by precipitation of calcite and dolomite. Megafaunal diversity in Lake Gosiute was greater than that of the Baca Formation lacustrine system, and included ostracods, molluscs, algal reefs, and fish. The more restricted assemblage of the Baca lacustrine system may be due to lack of potentially more fossiliferous basin-center deposits and/or higher salinity resulting from higher evaporation rates in the more southerly Baca lacustrine system. Surdam and Wolfbauer (1975) suggest that modern Deep Springs Lake in Inyo County, California, may be a modern (although much smaller) analogue of Lake Gosiute.

Certain aspects of Lake Chad, Africa (Mothersill, 1975), are similar to those inferred for the Baca lacustrine system, including a polymictic regime, the closed nature of the lake, and shallow water depth (less than 5 m).

Paleoclimate

Evidence in the Baca Formation indicates that the climate in west-central New Mexico during the Eocene was semiarid and savanna-like. Snyder (1971) cites the presence of caliche and the scarcity of organic materials in the Baca as evidence for arid or semiarid conditions. Goudie (1973) states that caliche is favored in semiarid climates where rainfall is seasonal and is below a threshold value of about 50 cm/yr. Caliches are rare yet widespread throughout the deposits of both the Baca and Carthage—La Joya basins.

Organic material is rarely preserved outside the lacustrine deposits of the Baca basin; even there it is restricted to a few occurrences of carbonaceous debris in non-red deltaic sandstones. Locally, however, abundant root traces are common in the deposits of the-Baca and Carthage—La Joya basins, indicating that vegetation was not climatically restricted. The scarcity of preserved carbonaceous debris in these deposits suggests that organic materials may have been decomposed in a dry, oxidizing environment of deposition.

Sedimentologic data from the Baca Formation also favor a non-humid climate. Closed-lake systems, such as the one that existed in the eastern part of the Baca basin, are restricted to arid and semiarid climatic regimes (Langbein, 1961). Evidence for flashy, perhaps ephemeral discharge in streams of the distal-fan facies probably also indicates arid to semiarid conditions, because modern ephemeral streams typically prevail in such climates. Schumm (1968), however, speculates that before extensive colonization of the land surface by grasses (pre-Miocene), river discharges were flashier, and runoff and sediment yield were greater. Although the effects of decreased continental vegetation during the Eocene may have had some influence on discharge peakedness, the rarity of ephemeral-flow indicators (debris-flow and sieve deposits) in the proximal-fan facies suggests that the ephemeral nature of flow in the distal facies was due to a basinward increase in aridity and, hence, evapotranspiration and infiltration. In this view, a wetter microclimate probably prevailed in the highlands adjacent to the Baca and Carthage—La Joya basins. The common occurrence of root traces and the nearly complete lack of eolian deposits in the Baca Formation and equivalent units suggest that highly arid conditions did not exist during Eocene time in west-central New Mexico.

The widespread red coloration of sediments in both the Baca and Carthage—La Joya basins is predominantly diagenetic in origin. Walker (1967) states that formation of diagenetic red beds is favored by warm temperatures, alkaline groundwaters, and an oxidizing environment; such conditions typically prevail in arid and semiarid climates.

Although paleontologic data for the Baca Formation are sparse, remains of fossil vertebrates (titanothere, cameloid, creodont, entelodont, and oreodont) have been recovered from sites in the Quemado and Datil Mountains areas (Schiebout and Schrodt, 1981; Lucas and others, 1981; Lucas, 1983). This fauna suggests that a savanna-type climate existed during Baca deposition in west-central New Mexico (J. A. Wilson, 1977, oral commun.; Schiebout and Schrodt, 1981).

Summary

Figure 25 is a model for the regional distribution of paleoflow and depositional environments for both the Baca and Carthage—La Joya basins that is based on paleocurrent and fades-distribution data obtained from outcrops of Eocene rocks in west-central New Mexico and eastern Arizona. An east—west cross section (Fig. 26) depicts facies geometries and subcrop relations in these basins.

The interplay of two factors, climate and tectonism, can account for most of the sedimentologic attributes of the Baca and Carthage—La Joya basins. Tectonism (both intrabasinal and extrabasinal) was the fundamental determinant of basin geometries, regional topography (and therefore paleoflow), subsidence rates and thickness variations of basin-fill deposits, and many aspects of the gross distribution of paleoenvironments within the basins. Climate provided strong controls on styles of fluvial and lacustrine sedimentation through effects on rates of erosion and sediment supply, volume and peakedness of discharge, evaporation rates, and amount of vegetative cover.

Sedimentation in the Baca and Carthage—La Joya basins began in Eocene time upon a low-relief erosion surface developed on Upper Cretaceous rocks. Initiation of sedimentation in the Carthage—La Joya basin may have occurred somewhat earlier than in the Baca basin, but establishment of precise temporal relations between the two basins is not possible with available data. Although the uplifts adjacent to these basins probably began to rise prior to Baca time, petrographic data (Cather, 1980) indicate that maximum structural development and widespread exposure of Precambrian rocks in at least two of the uplifts (the Morenci and Sierra uplifts) did not occur until early Baca time. Many aspects of the timing and style of Eocene deformation and sedimentation in west-central New Mexico are consistent with the late Laramide wrench-fault model of Chapin and Cather (1981).

Braided alluvial-plain, meanderbelt, and lacustrine environments of deposition are represented in a well-exposed facies tract through the axial portion of the Baca basin. These deposits record lower stream gradients, decreased sediment caliber, and decreased sand/mud ratios toward the basin center. Only the rocks of the braided alluvial-plain system are present in the poorly preserved remnants of the Carthage—La Joya basin. Figure 27 depicts clast-size trends in the lower part of the basin-fill deposits of the Baca and Carthage—La Joya basins, based on the average size of the ten largest clasts from a given conglomeratic outcrop. Maximum clast size in the Baca basin (especially in the western part) decreases systematically toward the basin center. Clast-size trends in the Carthage—La Joya basin are not readily discernible, probably

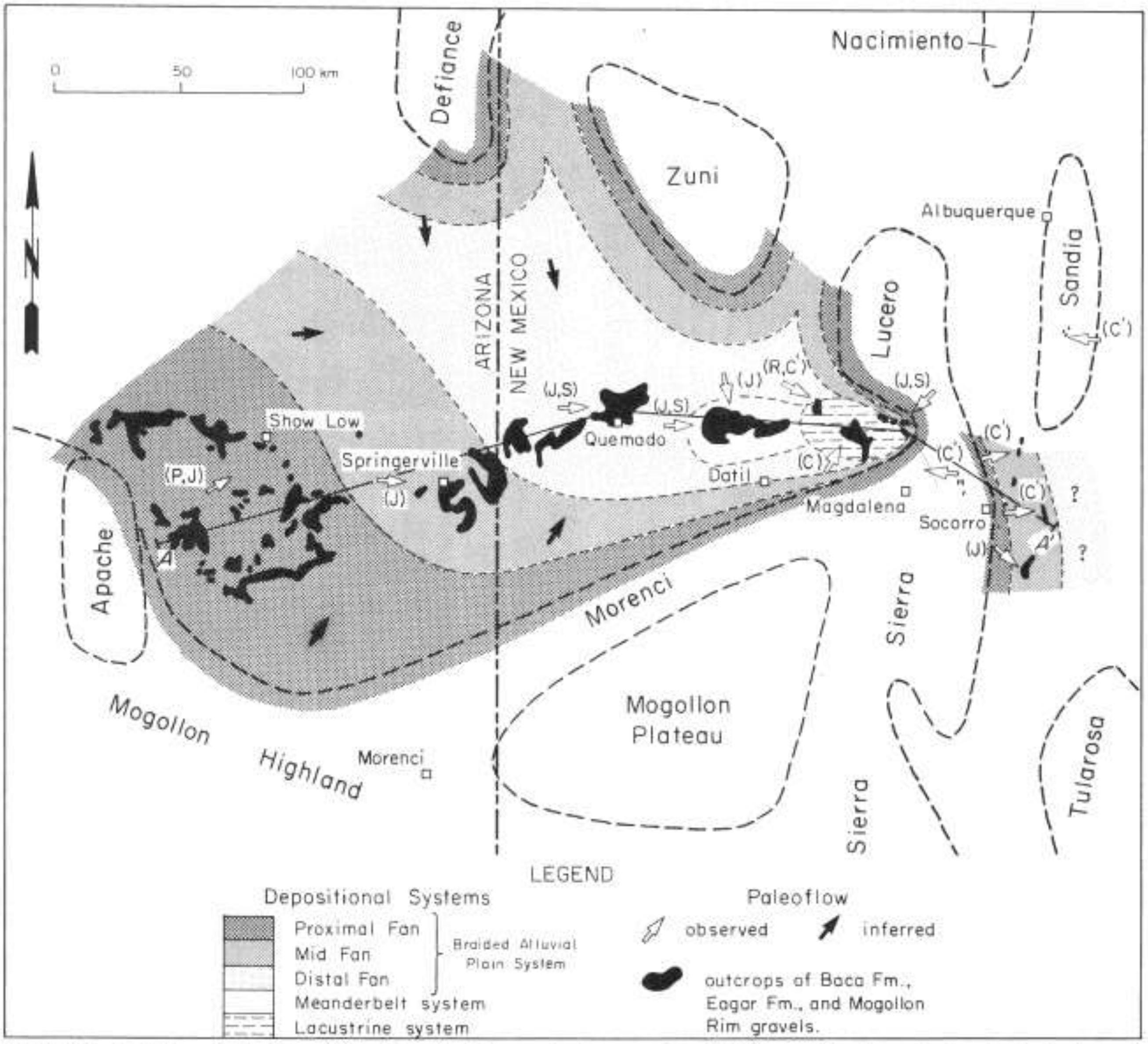

FIGURE 25—Model for distribution of depositional systems and paleocurrents in the Baca and Carthage–La Joya basins during early Baca time (high lake stand). Paleocurrent data sources (in parentheses): J = Johnson (1978), P = Pierce and others (1979), S = Snyder (1971), R = Robinson (1981), C = Cather (1980), C' = Cather (ongoing research). A–A' is section line for Fig. 26.

because most of the basin-fill exposures occur in a north-trending belt that approximately parallels the western margin of the basin.

Deposits of the braided alluvial-plain system are widespread in both the Baca and Carthage—La Joya basins. These deposits show evidence that perennial or seasonal flow in the proximal-fan facies gave way to more flashy, perhaps ephemeral flow in the distal environs. An extensive braided alluvial plain that dominated the western two-thirds of the Baca basin represents detritus shed from the Mogollon Highland.

Examples of both coarse- and fine-grained point-bar deposits are present in the Baca basin. Meander-belt deposits are dominantly restricted to a structurally low area between the Red Lake fault zone and the Hickman fault zone that was actively subsiding during Baca sedimentation (Fig. 26). Paleohydrologic reconstruction in this area indicates the presence of one or more large, east-flowing, integrated meandering-river systems in the axial part of the basin as well as smaller, tributary meandering streams that drained the Zuni uplift.

Lacustrine sedimentation took place in small, temporary lakes on fluvial-system floodplains and in a large, shallow, closed lake in the eastern part of the Baca basin. Sedimentation in these lakes was characterized by cyclical progradation, abandonment, and subsidence of both fine-grained deltas and fan deltas. Deltas prograded well out into lake basins and precluded the development of widespread lacustrine basin-center deposits. Major fluctuations in lake level in the large, closed lake present in the eastern part of the Baca basin may have resulted from climatic change.

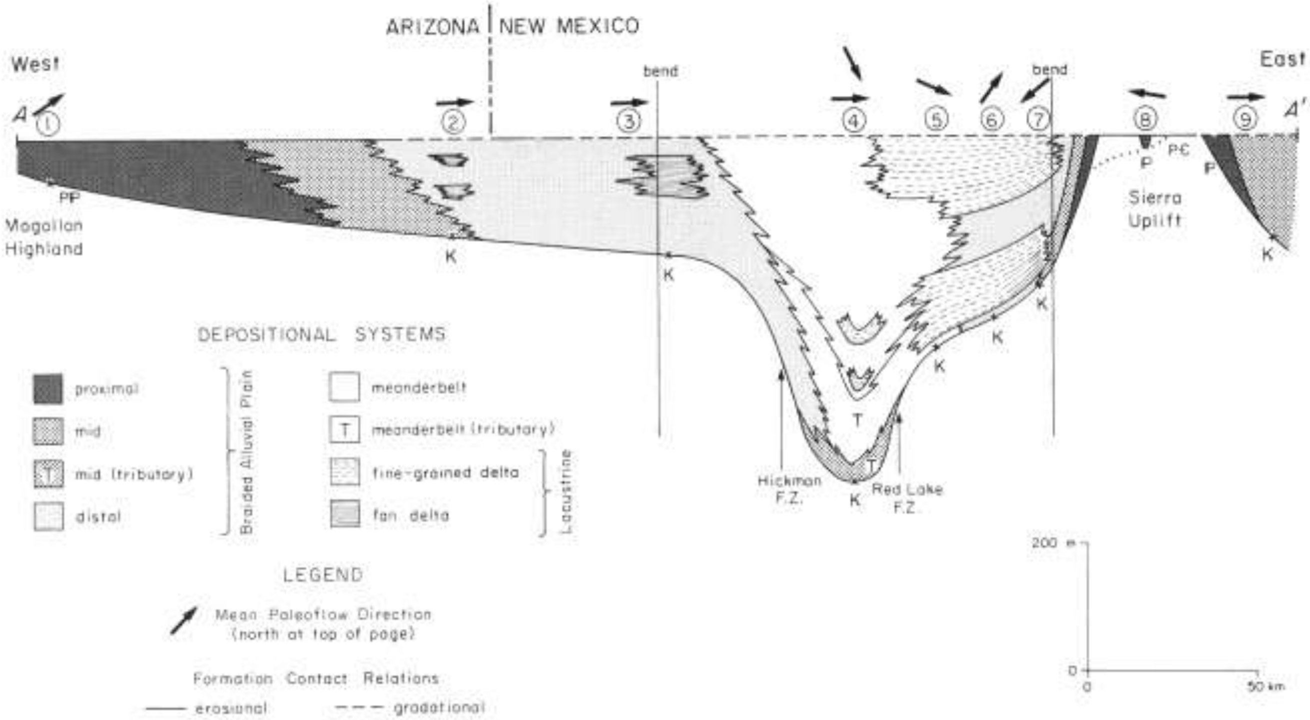

FIGURE 26—Cross section through Baca and Carthage–La Joya basins showing facies relationships. Circled numerals are keyed to outcrop-belt segments in Fig. 1. Line of section A–A' is shown in Fig. 25. Crosses at lower contacts of basins show thickness-control points; symbols beneath basins indicate age of subcrop.

Interestingly, the deposits of this lake did not occupy the structurally lowest part of the basin. This apparently is a reflection of the general easterly paleoslope in the central and western parts of the Baca basin that was largely due to the dominance of Mogollon Highland-derived sediments in those areas.

The deposits of the Baca and Carthage—La Joya basins record the waning of the Laramide deformation in adjacent uplifts. A cross section through both basins (Fig. 26) shows an onlapping relationship of component facies of the basin-fill. This probably resulted from the updip retreat of the facies tract in response to gradual erosion of source areas. Decreased source-area relief during late Baca time is further supported by the general upsection decrease in grain size and sandstone/mudstone ratios in the deposits of both basins. Laramide positive areas persisted during the onset of mid-Tertiary volcanism (Gather, in prep.), and the distribution of sedimentary facies in the basal volcaniclastic rocks of the Spears Formation is very much like that of the underlying Laramide deposits in most areas.

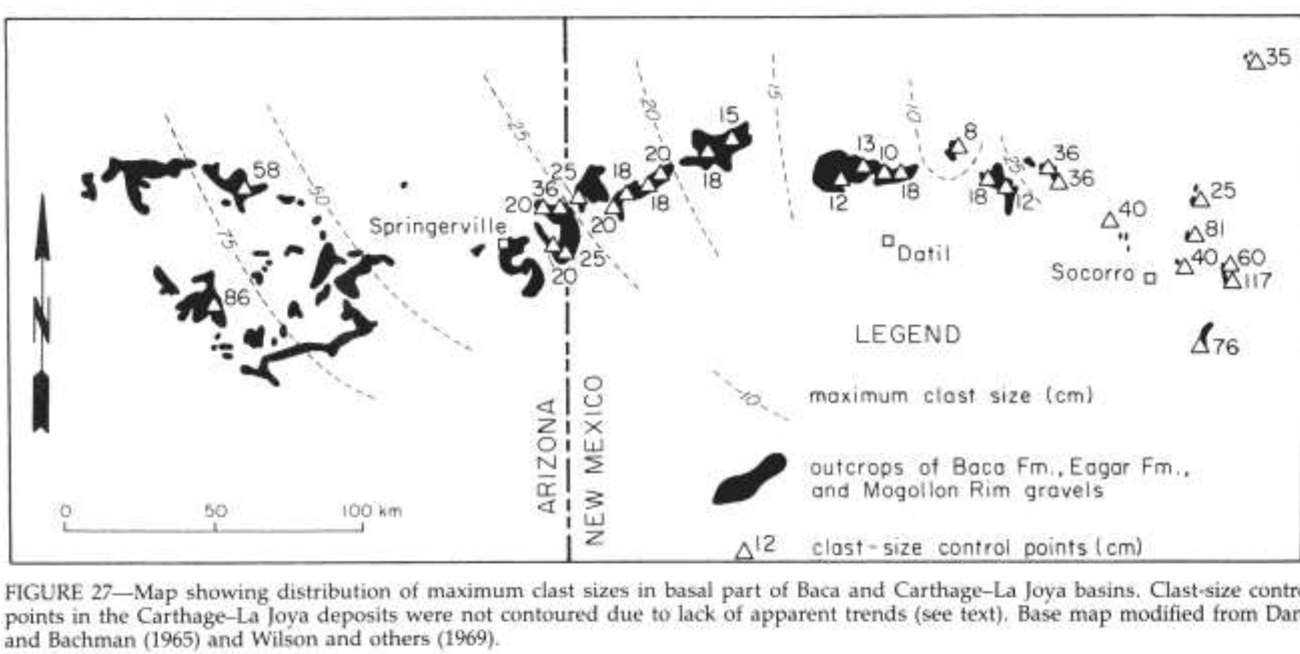

FIGURE 27—Map showing distribution of maximum clast sizes in basal part of Baca and Carthage–La Joya basins. Clast-size control points in the Carthage–La Joya deposits were not contoured due to lack of apparent trends (see text). Base map modified from Dane and Bachman (1965) and Wilson and others (1969).

References

Allen, J. R. L., 1964, Studies in fluviatile sedimentation: Six cyclothems from the Lower Old Red Sandstone, Anglo-Welsh Basin: Sedimentology, v. 3, pp. 163-198.

, 1965, The sedimentation and palaeogeography of the Old Red Sandstone of Anglesey, North Wales: Yorkshire Geological Society Proceedings, v. 35, pp. 139-185.

Axelsson, V., 1967, The Laiture delta-a study in deltaic morphology and processes: Geografiska Annaler, v. 49, pp. 1-127.

Bates, C. C., 1953, Rational theory of delta formation: American Association of Petroleum Geologists, Bulletin, v. 37, pp. 2119-2162.

Belt, E. S., 1968, Carboniferous continental sedimentation, Atlantic Provinces, Canada; *in* Klein, G. deV. (ed.), Late Paleozoic and Mesozoic continental sedimentation, northeastern North America: Geological Society of America, Special Paper 106.

Bernard, H. A., Major, C. F., Jr., Parrott, B. S., and Leblanc, R. J., 1970, Recent sediments of southeast Texas: University of Texas (Austin), Bureau of Economic Geology, Guidebook 11, 120 pp.

Blakestad, R. B., 1978, Geology of the Kelly mining district, Socorro County, New Mexico: M.S. thesis, University of Colorado, 127 pp.

Boothroyd, J. C., and Ashley, G. M., 1975, Processes, bar morphology, and sedimentary structures on braided-outwash fans, northeastern Gulf of Alaska; *in* McDonald, B. C., and Jopling, A. V. (eds.), Glaciofluvial and glaciolacustrine sedimentation: Society of Economic Paleontologists and Mineralogists, Special Publication 23, pp. 193-222.

, and Nummedal, D., 1978, Proglacial braided outwash: a model for humid alluvial-fan deposits; *in* Miall, A. D. (ed.), Fluvial sedimentology: Canadian Society of Petroleum Geologists, Memoir 5, pp. 641-668.

Bornhorst, T. J., Jones, D. P., Elston, W. E., Damon, P. E., and Shafiqullah, M., 1982, New radiometric ages on volcanic rocks from the Mogollon-Datil volcanic field southwestern New Mexico: Isochron/West, no. 35, pp. 13-15.

Bridge, J. S., 1978, Origin of horizontal lamination under turbulent boundary layers: Sedimentary Geology, v. 20, pp. 1-16.

Bull, W. B., 1972, Recognition of alluvial-fan deposits in the stratigraphic record, *in* Rigby, J. K., and Hamblin, W. K. (eds.), Recognition of ancient sedimentary environments: Society of Economic Paleontologists and Mineralogists, Special Publication 16, pp. 63-83.

Burke, W. H., Kenny, G. S., Otto, J. B., and Walker, R. D., 1963, Potassium-argon dates, **Socorro** and Sierra Counties, New Mexico: New Mexico Geological Society, Guidebook to 14th field conference, p. 224.

Bushnell, H. P., 1955, Mesozoic stratigraphy of south-central New Mexico: New Mexico Geological Society, Guidebook to 6th field conference, pp. 81-87.

Callender, J. F., and Zilinski, R. E., Jr., 1976, Kinematics of Tertiary and Quaternary deformation along the eastern edge of the Lucero uplift, central New Mexico, *in* Woodward, L. A., and Northrop, S. A. (eds.), Tectonics and mineral resources of southwestern North America: New Mexico Geological Society, Special Publication 6, pp. 53-61.

Cant, D. J., and Walker, R. G., 1976, Development of a braided-fluvial facies model for the Devonian Battery Point Sandstone, Quebec: Canadian Journal of Earth Sciences, v. 13, pp. 102-119.

Cather, S. M., 1980, Petrology, diagenesis, and genetic stratigraphy of the Eocene Baca Formation, Alamo Navajo Reservation, Socorro County, New Mexico: M.A. thesis, University of Texas (Austin), 243 pp.

, 1982, Lacustrine sediments of the Baca Formation (Eocene), western Socorro County, New Mexico: New Mexico Geology, v. 4, pp. 1-6.

, in preparation, Volcano-sedimentary evolution and tectonic implications of the lower Oligocene Datil Group, west-central New Mexico: Ph.D. dissertation, University of Texas (Austin).

Chaiffetz, M. S., 1979, Palynological age and paleoecology of the Baca Formation, northwestern Socorro County, central-western New Mexico (abs.): Geological Society of America, Abstracts with Programs, v. 11, p. 268.

Chamberlin, R. M., 1981, Uranium potential of the Datil Mountains-Pie Town area, Catron County, New Mexico: New Mexico Bureau of Mines and Mineral Resources, Open-file Report 138, 51 pp.

, 1982, Geologic map, cross sections, and map units of the Lemitar Mountains: New Mexico Bureau of Mines and Mineral Resources, Open-file Report 169, 3 sheets.

Chapin, C. E., and Cather, S. M., 1981, Eocene tectonics and sedimentation in the Colorado Plateau-Rocky Mountain area, *in* Dickinson, W. R., and Payne, W. D. (eds.), Relations of tectonics to ore deposits in the southern Cordillera: Arizona Geological Society, Digest, v. 14, pp. 173-198.

, and Seager, W. R., 1975, Evolution of the Rio Grande rift in Socorro and Las Cruces areas: New Mexico Geological Society, Guidebook to 26th field conference, pp. 297-321.

, Chamberlin, R. M., Osburn, G. R., White, D. W., and Sanford, A. R., 1978, Exploration framework of the Socorro geothermal area, New Mexico, *in* Chapin, C. E., and Elston, W. E. (eds.), Field guide to selected cauldrons and mining districts of the Datil-Mogollon volcanic field, New Mexico: New Mexico Geological Society, Special Publication 7, pp. 115-130.

Church, M., and Gilbert, R., 1973, Proglacial fluvial and lacustrine environments; *in* Jopling, A. V., and McDonald, B. C. (eds.), Glaciofluvial and glaciolacustrine sedimentation: Society of Economic Paleontologists and Mineralogists, Special Publication 23, pp. 22-100.

Collinson, J. D., 1970, Bedforms in the Tana River, Norway: Geografiska Annaler, v. 52A, pp. 31-55.

Coney, P. J., 1971, Cordilleran tectonic transitions and motion of the North American plate: Nature, v. 233, pp. 462-465.

, 1976, Structure, volcanic stratigraphy, and gravity across the Mogollon Plateau, New Mexico; *in* Elston, W. E., and Northrop, S. A. (eds.), Cenozoic volcanism in southwestern New Mexico: New Mexico Geological Society, Special Publication 5, pp. 29-41.

Costello, W. R., and Walker, R. G., 1972, Pleistocene sedimentology, Credit River, southern Ontario: a new component of the braided river model: Journal of Sedimentary Petrology, v. 42, pp. 389-400.

Cotter, E., 1971, Paleoflow characteristics of a Late Cretaceous river in Utah from analysis of sedimentary structures in the Ferron Sandstone: Journal of Sedimentary Petrology, v. 41, pp. 129-138.

Dane, C. H., and Bachman, G. 0., 1965, Geologic map of New Mexico: U.S. Geological Survey, 2 sheets, scale 1:500,000.

Davis, G. H., 1978, Monocline fold pattern of the Colorado Plateau; *in* Matthews, V., III (ed.), Laramide folding associated with basement block faulting in the western United States: Geological Society of America, Memoir 151, pp. 215-233.

, 1979, Laramide folding and faulting in southeastern Arizona: American Journal of Science, v. 279, pp. 543-569.

, Schowalter, S. R., Benson, G. S., McCalmont, L. S., and Faulds, J. E., 1982, The Apache uplift, heretofore unrecognized Colorado Plateau uplift (abs.): Geological Society of America, Abstracts with Programs, v. 14, p. 472.

Drewes, H., 1976, Laramide tectonics from Paradise to Hells Gate, southeastern Arizona: Arizona Geological Society, Digest, v. 10, pp. 151-167.

, 1978, The Cordilleran orogenic belt between Nevada and Chihuahua: Geological Society of America, Bulletin, v. 89, pp. 641-657.

Eardley, A. J., 1962, Structural geology of North America: 2nd edition, Harper & Row, New York, 743 pp.

Edmunds, R. J., 1961, Geology of the Nutria Monocline, McKinley County, New Mexico: M.S. thesis, University of New Mexico, 100 pp.

Eggleston, T. L., 1982, Geology of the central Chupadera Mountains, Socorro County, New Mexico: M.S. thesis, New Mexico Institute of Mining and Technology, 161 pp.

Ethridge, F. G., and Schumm, S. A., Reconstructing paleochannel morphologic and flow characteristics: methodology, limitations, and assessment; *in* Miall, A. D. (ed.), Fluvial sedimentology: Canadian Society of Petroleum Geologists, Memoir 5, pp. 703-722.

Eugster, H. P., and Hardie, L. A., 1975, Sedimentation in an ancient playa-lake complex: the Wilkins Peak Member of the Green River Formation of Wyoming: Geological Society of America, Bulletin, v. 86, pp. 319-334.

______ , and Surdam, R. C., 1973, Depositional environment of the Green River Formation of Wyoming: a preliminary report: Geological Society of America, Bulletin, v. 84, pp. 1115-1120.

Farkas, S. E., 1969, Geology of the southern San Mateo Mountains, Socorro and Sierra Counties, New Mexico: Ph.D. dissertation, University of New Mexico, 137 pp.

Fisher, W. L., and Brown, L. F., Jr., 1972, Clastic depositional systems-a genetic approach to facies analysis: University of Texas (Austin), Bureau of Economic Geology, Annotated Outline and Bibliography, 211 pp.

, and McGowen, J. H., 1967, Depositional systems in the Wilcox Group of Texas and their relationship to the occurrence of oil and gas: American Association of Petroleum Geologists, Bulletin, v. 53, pp. 30-54.

, Brown, L. F., Jr., Scott, A. J., and McGowen, J. H., 1969, Delta systems in the exploration for oil and gas: University of Texas (Austin), Bureau of Economic Geology, Research Colloquium, 212 pp.

Fisk, H. N., 1944, Geological investigation of the alluvial valley of the lower Mississippi River: Vicksburg, Mississippi River Commission, 78 pp.

Fitzsimmons, J. P., 1959, The structure and geomorphology of west-central New Mexico: New Mexico Geological Society, Guidebook to 10th field conference, pp. 112-116.

, 1967, Precambrian rocks of the Zuni Mountains: New Mexico Geological Society, Guidebook to 18th field conference, pp. 119-121.

Foster, R. W., 1978, Selected data for deep drill holes along the Rio Grande rift in New Mexico: New Mexico Bureau of Mines and Mineral Resources, Circular 163, pp. 236-237.

Frostick, L. E., and Reid, I., 1977, The origin of horizontal laminae in ephemeral stream channel-fill: Sedimentology, v. 24, pp. 1-9.

Galloway, W. E., 1977, Catahoula Formation of the Texas coastal plain: depositional systems, composition, structural development, ground-water flow history, and uranium distribution: University of Texas (Austin), Bureau of Economic Geology, Report of Investigations 87, 59 pp.

Gilbert, G. K., 1885, The topographic features of lake shores: U.S. Geological Survey, Annual Report 5, pp. 69-123.

Goddard, E. N., 1966, Geologic map and sections of the Zuni Mountains fluorospar district, Valencia County, New Mexico: U.S. Geological Survey, Miscellaneous Investigations Map 1-454.

Gole, C. V., and Chitale, S. V., 1966, Inland delta building activity of the Kosi River: Proceedings of the American Society of Civil Engineers, Journal of the Hydraulics Division, HY-2, pp. 111-126.

Gorham, T. W,. and Ingersoll, R. V., 1979, Evolution of the Eocene Galisteo basin, north-central New Mexico: New Mexico Geological Society, Guidebook to 30th field conference, pp. 219-224.

Goudie, A., 1973, Duricrusts in tropical and subtropical landscapes: Oxford, Clarendon Press, 174 pp.

Guilinger, D. R., 1982, Geology and uranium potential of the Tejana Mesa-Hubbell Draw area, Catron County, New Mexico: M.S. thesis, New Mexico Institute of Mining and Technology, 129 pp.

Gustayson, T. C., 1974, Sedimentation on gravel outwash fans, Malaspina Glacier Foreland, Alaska: Journal of Sedimentary Petrology, v. 44, pp. 374-389.

Hantzchel, W., 1975, Treatise on invertebrate paleontology, Part W, Trace fossils and problematica: University of Kansas (Lawrence), 2nd edition, 269 pp.

Happ, S. C., MacKenzie, D. B., and McCubbin, D. G., 1940, Some aspects of accelerated stream and valley sedimentation: U.S. Department of Agriculture, Technical Bulletin 695, 134 pp.

Hedlund, D. L., 1974, Age and structural setting of base-metal mineralization in the Hillsboro-San Lorenzo area, southwestern New Mexico (abs.): New Mexico Geological Society, Guidebook to 25th field conference, pp. 378-379.

Hein, F. J., and Walker, R. G., 1977, Bar evolution and development of stratification in the gravelly, braided, Kicking Horse River: Canadian Journal Earth Sciences, v. 14, pp. 562-570.

Hooke, R. L., 1967, Processes on arid region alluvial fans: Journal of Geology, v. 75, pp. 453-456.

Jackson, R. G., II, 1976a, Sedimentological and fluid-dynamic implications of the turbulent flow bursting phenomenon in geophysical flows: Journal of Fluid Mechanics, v. 77, pp. 531-550.

_______, 1976b, Depositional model of point bars in the lower Wabash River: Journal of Sedimentary Petrology, v. 46, pp. 579-594.

Jahns, R. H., McMillan, D. K., O'Brient, J. D., and Fisher, D. L., 1978, Geologic section in the Sierra Cuchillo and flanking areas, Sierra and Socorro Counties, New Mexico; *in* Chapin, C. E., and Elston, W. E. (eds.), Field guide to selected cauldrons and mining districts of the Datil-Mogollon volcanic field, New Mexico: New Mexico Geological Society, Special Publication 7, pp. 130-138.

Johansen, S., 1983, Abrupt changes in the Gallup Sandstone (Cretaceous) across the Puertecito fault system in northwest Socorro County: New Mexico Geological Society, Guidebook to 34th field conference, pp. 45-46.

Johnson, B. D., 1978, Genetic stratigraphy and provenance of the Baca Formation, New Mexico, and the Eagar Formation, Arizona: M.S. thesis, University of Texas (Austin), 150 pp.

Karcz, I., 1972, Sedimentary structures formed in flash floods in southern Israel: Sedimentary Geology, v. 7, pp. 161-182.

Keith, S. B., and Barrett, L. F., 1976, Tectonics of the central Dragoon Mountains: a new look: Arizona Geological Society, Digest, v. 10, pp. 169-204.

Kelley, V. C., 1955, Regional, tectonics of the Colorado Plateau and relationships to the origin and distribution of uranium: University of New Mexico, Publications in Geology, no. 5, 120 pp.

, 1967, Tectonics of the Zuni-Defiance region, New Mexico and Arizona: New Mexico Geological Society, Guidebook to 18th field conference, pp. 28-31.

,1977, Geology of the Albuquerque Basin, New Mexico: New Mexico Bureau of Mines and Mineral Resources, Memoir 33, 59 pp.

, 1982, Diverse geology of the Hubbell bench, Albuquerque basin: New Mexico Geological Society, Guidebook to 33rd field conference, pp. 159-160.

, and Clinton, N. J., 1960, Fracture systems and tectonic elements of the Colorado Plateau: University of New Mexico, Publications in Geology, no. 4, 104 pp.

, and McCleary, J. T., 1960, Laramide orogeny in south-central New Mexico: American Association of Petroleum Geologist, Bulletin, v. 44, pp. 1419-1420.

, and Northrop, S. A., 1975, Geology of the Sandia Mountains and vicinity, New Mexico: New Mexico Bureau of Mines and Mineral Resources, Memoir 29, 136 pp.

, and Silver, C., 1952, Geology of the Caballos Mountains: University of New Mexico, Publications in Geology, no. 5, 286 pp.

Kottlowski, F. E., Weber, R. H., and Willard, M. E., 1969, Tertiary intrusive-volcanic episodes in the New Mexico region (abs.): Geological Society of America, Abstracts with Programs, part 7, pp. 278-280.

Krewedl, D. A., 1974, Geology of the central Magdalena Mountains, Socorro County, New Mexico: Ph.D. dissertation, University of Arizona, 128 pp.

Langbein, W. B., 1961, Salinity and hydrology of closed lakes: U.S. Geological Survey, Professional Paper 412, 50 pp.

Leopold, L. B., 1951, Rainfall frequency-an aspect of climatic variation: American Geophysical Union, Transactions, v. 32, pp. 347-357.

, and Miller, J. P., 1956, Ephemeral streams-hydraulic factors and their relation to the drainage net: U.S. Geological Survey, Professional Paper 282-A, pp. 1-37.

and Wolman, M. G., 1957, River channel patterns: braided, meandering, and straight: U.S. Geological Survey, Professional Paper 282-B, 85 pp.

__________, and Miller, J. P., 1964, Fluvial processes in geomorphology: W. H. Freeman and Co., San Francisco, 522 pp.

Love, D. W., 1977, Dynamics of sedimentation and geomorphic history of Chaco Canyon National Monument, New Mexico: New Mexico Geological Society, Guidebook to 28th field conference, pp. 291-298.

Lowell, J. D., 1974, Plate tectonics and foreland basin deformation: Geology, v. 2, pp. 275-278.

Lucas, S. G., 1983, The Baca Formation and the Eocene-Oligocene boundary in New Mexico: New. Mexico Geological Society, Guidebook to 34th field conference, pp. 187-192.

, Schoch, R. M., Manning, E., and Tsentas, C., 1981, The Eocene biostratigraphy of New Mexico: Geological Society of America, Bulletin, v. 92, pp. 867-951.

Martini, I. P., 1977, Gravelly flood deposits of Irvine Creek, Ontario, Canada: Sedimentology, v. 24, pp. 603-622.

Massingill, G. L., 1979, Geology of the Riley-Puertecito area, southeastern margin of the Colorado Plateau, Socorro County, New Mexico: Ph.D. thesis, University of Texas (El Paso), 301 pp.

Masters, C. D., 1967, Use of sedimentary structures in determination of depositional environments, Mesaverde Formation, Williams Fork Mountains, Colorado: American Association of Petroleum Geologists, Bulletin, v. 51, pp. 2033-2043.

McGowen, J. H., 1970, Gum Hollow fan delta, Nueces Bay, Texas: University of Texas (Austin), Bureau of Economic Geology, Report of Investigations 69, 91 pp.

, and Garner, L. E., 1970, Physiographic features and stratigraphic types of coarse-grained point bars: Sedimentology, v. 14, pp. 77-111.

, and Groat, C. G., 1971, Van Horn Sandstone, west Texas: an alluvial-fan model for mineral exploration: University of Texas (Austin), Bureau of Economic Geology, Report of Investigations 72, 57 pp.

McKee, E. D., Crosby, E. J., and Berryhill, H. L., Jr., 1967, Flood deposits, Bijou Creek, Colorado, June 1965: Journal of Sedimentary Petrology, v. 37, pp. 829-851.

Miall, A. D., 1977, A review of the braided-river depositional environment: Earth Science Reviews, v. 13, pp. 1-62.

1978, Lithofacies types and vertical profile models in braided-river deposits: a summary; *in* Miall, A. D. (ed.), Fluvial sedimentology: Canadian Society of Petroleum Geologists, Memoir 5, pp. 597-604.

Moody-Stuart, M., 1966, High- and low-sinuosity stream deposits, with examples from the Devonian of Spitsbergen: Journal of Sedimentary Petrology, v. 36, pp. 1102-1117.

Mothersill, J. S., 1975, Lake Chad: geochemistry and sedimentary aspects of a shallow polymictic lake: Journal of Petrology, v. 45, pp. 295-309.

Nummedal, D., and Boothroyd, J. C., 1976, Morphology and hydrodynamic characteristics of terrestrial fan environments: Coastal Research Division, Department of Geology, University of South Carolina, Technical Report No. 10-CRD, 61 pp.

Pierce, H. A., Damon, P. E., and Shafiqullah, M., 1979, An Oligocene(?) Colorado Plateau edge in Arizona: Tectonophysics, v. 61, pp. 1-24.

Potter, S. C., 1970, Geology of Baca Canyon, Socorro County, New Mexico: M.S. thesis, University of Arizona, 54 pp.

Puigdefabregas, C., and van Vliet, A., 1978, Meandering stream deposits from the Tertiary of the southern Pyrenees; *in* Miall, A. D. (ed.), Fluvial sedimentology: Canadian Society of Petroleum Geologists, Memoir 5, pp. 469-486.

Rehrig, W. A., and Heidrick, T. L., 1972, Regional fracturing of Laramide stocks of Arizona and its relationship to porphyry copper mineralization: Economic Geology, v. 67, p. 198-213.

, and, 1976, Regional tectonic stress during the Laramide and late Tertiary intrusive periods, Basin and Range Province, Arizona: Arizona Geological Society, Digest, v. 10, pp. 205228.

Robinson, B. R., 1981, Geology of the D-Cross Mountain Quadrangle, Socorro and Catron Counties, New Mexico: Ph.D. dissertation, University of Texas (El Paso), 213 pp.

Rust, B. R., 1972, Structure and process in a braided river: Sedimentology, v. 18, pp. 221-245.

1978, Depositional models for braided alluvium; *in* Miall, A. D. (ed.), Fluvial sedimentology: Canadian Society of Petroleum Geologists, Memoir 5, pp. 605-625.

Schiebout, J. A., and Schrodt, A. K., 1981, Vertebrate paleontology of the lower Tertiary Baca Formation of western New Mexico: Geological Society of America, Bulletin, v. 92, pp. 976-979.

Schumm, S. A., 1968, Speculations concerning paleohydrologic controls of terrestrial sedimentation: Geological Society of America, Bulletin, v. 79, pp. 1573-1588.

, and Hadley, R. F., 1957, Arroyos and the semiarid cycle of erosion: American Journal of Science, v. 255, pp. 161-174.

_______, and Kahn, H. R., 1972, Experimental study of channel patterns: Geological Society of America, Bulletin, v. 83, pp. 17551770.

Seager, W. R., 1975, Cenozoic tectonic evolution of the Las Cruces area: New Mexico Geological Society, Guidebook to 26th field conference, pp. 241-250.

, 1981, Geology of the Organ Mountains and southern San Andres Mountains, New Mexico: New Mexico Bureau of Mines and Mineral Resources, Memoir 36, 97 pp.

Shepard, R. G., 1978, Distinction of aggradational and degradational fluvial regimes in valley-fill alluvium, Tapia Canyon, New Mexico; *in* Miall, A. D. (ed.), Fluvial sedimentology: Canadian Society of Petroleum Geologists, Memoir 5, pp. 277-286.

Shrine, G. K., 1956, Geology of the Springerville-St. Johns area, Apache County, Arizona: Ph.D. dissertation, University of Texas (Austin), 248 pp.

Smith, C. T., 1983, Structural problems along the east side of the Socorro constriction, Rio Grande rift: New Mexico Geological Society, Guidebook to 34th field conference, pp. 103-109.

Smith, N. D., 1970, The braided-stream depositional environment: comparison of the Platte River and some Silurian clastic rocks, north-central Appalachians: Geological Society of America, Bulletin, v. 81, pp. 2993-3014.

, 1971, Pseudo-planar stratification produced by very low amplitude sand waves: Journal of Sedimentary Petrology, v. 41, pp. 69-73.

, 1975, Sedimentary environments and late Quaternary history of a low-energy mountain delta: Canadian Journal of Earth Sciences, v. 12, pp. 2004-2013.

Snyder, D. 0., 1971, Stratigraphic analysis of the Baca Formation, west-central New Mexico: Ph.D. dissertation, University of New Mexico, 158 pp.

Stearns, C. E., 1962, Geology of the north half of the Pelona quadrangle, Catron County, New Mexico: New Mexico Bureau of Mines and Mineral Resources, Bulletin 78, 46 pp.

Steel, R. J., 1974, New Red Sandstone floodplain and piedmont sedimentation in the Hebridean Province, Scotland: Journal of Sedimentary Petrology, v. 44, pp. 336-357.

Surdam, R. C., and Wolfbauer, C. A., 1975, The Green River Formation: a playa-lake complex: Geological Society of America, Bulletin, v. 86, pp. 335-345.

Theakstone, W. H., 1976, Glacial lake sedimentation, Austerdalsisen, Norway: Sedimentology, v. 23, pp. 671-688.

Visher, G. S., 1972, Physical characteristics of fluvial deposits; *in* Rigby, J. K., and Hamblin, W. K. (eds.), Recognition of ancient sedimentary environments: Society of Economic Paleontologists and Mineralogists, Special Publication 16, pp. 84-97.

Walker, T. R., 1967, Formation of red beds in modem and ancient deserts: Geological Society of America, Bulletin, v. 78, pp. 353368.

Wengerd, S. A., 1959, Regional geology of the Lucero region, west-central New Mexico: New Mexico Geological Society, Guidebook to 10th field conference, pp. 121-134.

Williams, G. E., 1971, Flood deposits of the sand-bed ephemeral streams of central Australia: Sedimentology, v. 17, pp. 1-40.

Williams, P. F., and Rust, B. R., 1969, The sedimentology of a braided river: Journal of Sedimentary Petrology, v. 39, pp. 649-679.

Wilpolt, R. H., and Wanek, A. A., 1951, Geology of the region from Socorro to San Antonio east to Chupadera Mesa, Socorro County, New Mexico: U.S. Geological Survey, Oil and Gas Investigation Map OM-121.

Wilson, E. D., Moore, R. T., and Cooper, J. R., 1969, Geologic map of Arizona: U.S. Geological Survey, scale 1:500,000.

Selected conversion factors*

TO CONVERT	MULTIPLY BY	TO OBTAIN
Length		
inches, in	2.540	centimeters, cm
feet, ft	3.048×10^{-1}	meters, m
yards, yds	9.144×10^{-1}	m
statute miles, mi	1.609	kilometers, km
fathoms	1.829	m
angstroms, Å	1.0×10^{-8}	cm
Å	1.0×10^{-4}	micrometers, μm
Area		
in^2	6.452	cm^2
ft^2	9.29×10^{-2}	m^2
yds^2	8.361×10^{-1}	m^2
mi^2	2.590	km^2
acres	4.047×10^{3}	m^2
acres	4.047×10^{-1}	hectares, ha
Volume (wet and dry)		
in^3	1.639×10^{1}	cm^3
ft^3	2.832×10^{-2}	m^3
yds^3	7.646×10^{-1}	m^3
fluid ounces	2.957×10^{-2}	liters, l *or* L
quarts	9.463×10^{-1}	l
U.S. gallons, gal	3.785	l
U.S. gal	3.785×10^{-3}	m^3
acre-ft	1.234×10^{3}	m^3
barrels (oil), bbl	1.589×10^{-1}	m^3
Weight, mass		
ounces avoirdupois, avdp	2.8349×10^{1}	grams, gr
troy ounces, oz	3.1103×10^{1}	gr
pounds, lb	4.536×10^{-1}	kilograms, kg
long tons	1.016	metric tons, mt
short tons	9.078×10^{-1}	mt
oz mt^{-1}	3.43×10^{1}	parts per million, ppm
Velocity		
ft sec^{-1} (= ft/sec)	3.048×10^{-1}	m sec^{-1} (= m/sec)
mi hr^{-1}	1.6093	km hr^{-1}
mi hr^{-1}	4.470×10^{-1}	m sec^{-1}

TO CONVERT	MULTIPLY BY	TO OBTAIN
Pressure, stress		
lb in^{-2} (= lb/in^2), psi	7.03×10^{-2}	kg cm^{-2} (= kg/cm^2)
lb in^{-2}	6.804×10^{-2}	atmospheres, atm
lb in^{-2}	6.895×10^{3}	newtons (N)/m^2, N m^{-2}
atm	1.0333	kg cm^{-2}
atm	7.6×10^{2}	mm of Hg (at 0° C)
inches of Hg (at 0° C)	3.453×10^{-2}	kg cm^{-2}
bars, b	1.020	kg cm^{-2}
b	1.0×10^{6}	dynes cm^{-2}
b	9.869×10^{-1}	atm
b	1.0×10^{-1}	megapascals, MPa
Density		
lb in^{-3} (= lb/in^3)	2.768×10^{1}	gr cm^{-3} (= gr/cm^3)
Viscosity		
poises	1.0	gr cm^{-1} sec^{-1} *or* dynes cm^{-2}
Discharge		
U.S. gal min^{-1}, gpm	6.308×10^{-2}	l sec^{-1}
gpm	6.308×10^{-5}	m^3 sec^{-1}
ft^3 sec^{-1}	2.832×10^{-2}	m^3 sec^{-1}
Hydraulic conductivity		
U.S. gal day^{-1} ft^{-2}	4.720×10^{-7}	m sec^{-1}
Permeability		
darcies	9.870×10^{-13}	m^2
Transmissivity		
U.S. gal day^{-1} ft^{-1}	1.438×10^{-7}	m^2 sec^{-1}
U.S. gal min^{-1} ft^{-1}	2.072×10^{-1}	l sec^{-1} m^{-1}
Magnetic field intensity		
gausses	1.0×10^{5}	gammas
Energy, heat		
British thermal units, BTU	2.52×10^{-1}	calories, cal
BTU	1.0758×10^{2}	kilogram-meters, kgm
BTU lb^{-1}	5.56×10^{-1}	cal kg^{-1}
Temperature		
°C + 273	1.0	°K (Kelvin)
°C + 17.78	1.8	°F (Fahrenheit)
°F − 32	5/9	°C (Celsius)

**Divide by* the factor number to reverse conversions.
Exponents: for example 4.047×10^{3} (see acres) = 4,047; 9.29×10^{-2} (see ft^2) = 0.0929.

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