Climate and ephemeral-stream processes: Twentieth-century geomorphology and alluvial stratigraphy of the Little Colorado River, Arizona

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ABSTRACT

During the first 40 years of the twentieth century, erosion was the dominant geomorphic process affecting the morphology of the Little Colorado River channel. The discharge regimen was one of frequent large floods and high annual discharge that created a wide sandy channel free of vegetation. In the 1940s and early 1950s, average annual precipitation declined, reducing annual discharge to about 57% of that of the preceding period as well as reducing the frequency of large floods. The channel adjusted to the new hydrologic regimen by reducing its width. Parts of the channel were frequently dry, and riparian vegetation, primarily nonnative salt cedar, became established on the higher channel surfaces. Precipitation and discharge thereafter increased and aggradation by overbank deposition was the primary geomorphic process, as indicated by accretion of 2 to 5 m of flood-plain alluvium between 1952 and 1978. Events of 1980, however, suggest that the flood plain has ceased to accrete, although climate has not fluctuated. The flood plain has probably reached a critical height above the channel, beyond which further accretion is unlikely under the existing discharge regimen. The recent history of the Little Colorado broadly suggests that flood-plain development was initiated by climatically induced hydrologic fluctuations. Flood-plain deposits in the stratigraphic column of such ephemeral streams may record repeated adjustments to altered hydrologic conditions.

INTRODUCTION

A problem facing sedimentologists, stratigraphers, geomorphologists, and other students of modern and ancient fluvial processes is that the influence of climate on such fundamental stream processes as erosion and aggradation is not well understood. Several facies models are widely used to interpret alluvial sequences, but they do not account for climatic control over stream processes (Miall, 1980). The absence of climatic-facies models, the limited understanding of how climate modifies alluvial channels, and the complex temporal and spatial patterns of sediment transport result in little consensus among recent workers regarding the role of climate in the origin of alluvial deposits (Baker and Penteado-Orellano, 1977; Schumm, 1977; Lewin, 1978; Graf, 1982; McDowell, 1983). Yet, in many geologic studies, such deposits provide the only potential record of past climate.

The problem of how climate influences stream processes is unusually difficult for the late Holocene of the southwestern United States. In this region of numerous ephemeral streams, a complex debate (reviewed by Cooke and Reeves, 1976, and Graf, 1983) has developed over the cause of the several widespread cycles of late Holocene erosion and aggradation. Three general hypotheses (Cooke and Reeves, 1976, p. 6–15) have been formulated to explain recent ephemeral-stream processes: (1) poor land use, such as overgrazing; (2) changes in climate; and (3) random variations within the fluvial system that are not related to climate. More recent investigations have put forth similar interpretations. Cooke and Reeves (1976) concluded that the widespread occurrence of accelerated erosion after 1880 in many cases resulted from poor land use. They also found that arroyos could have been produced in several other ways and that, given only the final form, it was rarely possible to determine the specific cause. Other workers (Emmett, 1974; Leopold, 1976; Euler and others, 1979) acknowledged some human impact but concluded that a change in climate was largely responsible for the post-1880 cycle of erosion and subsequent aggradation later in the twentieth century. On the other hand, Patton and Schumm (1981) reported from their observations of several streams, one of which was studied during the same period by Emmett (1974) and Leopold (1976), that erosion and sedimentation in this century have occurred independently of climatic fluctuations. There seems to be a wide divergence of thought among recent workers regarding the causes of fluvial erosion and sedimentation in the Southwest.

The conclusions drawn from these studies, to further complicate matters, are used as models for interpreting Holocene (Emmett, 1974; Leopold, 1976; Euler and others, 1979) or Quaternary (Patton and Schumm, 1981; Womack and Schumm, 1977) geomorphology throughout the arid to semiarid western United States. The landscape in the alluvial valleys of this region abounds with terraces or abandoned flood plains. According to Patton and Schumm (1981), as well as Womack and Schumm (1977), this landscape developed by the episodic transport of sediment from the drainage system; other investigators (Emmett, 1974; Leopold, 1976; Euler and others, 1979) believed that the terraced landscape reflects several climatic oscillations from wet (aggradation) to dry (erosion and flood-plain incision) simultaneously influencing the entire region. Perhaps the major difficulty confronting previous investigators has been the lack of runoff hydrographs and sediment-transport data; consequently, the effect of runoff variations on erosional and depositional processes could not be determined. It would seem that studies of modern stream processes are limited by the absence of long-term discharge information. The present study differs from earlier ones in that discharge records are available and were utilized in analyzing the historic changes in the alluvial channel of the Little Colorado River.

This study examines the influence of climate on twentieth-century geomorphology and alluvial stratigraphy of the Little Colorado River. An ephemeral stream, the Little Colorado flows in a sandy channel that drains 44,000 km^2 of the arid to semiarid southern Colorado Plateaus province (Fig. 1). From about 1900 to the early 1940s, erosion was the main geomorphic process and the river was incising its older deposits. The erosive phase ended in the early 1940s, and aggradation, resulting in the development of a

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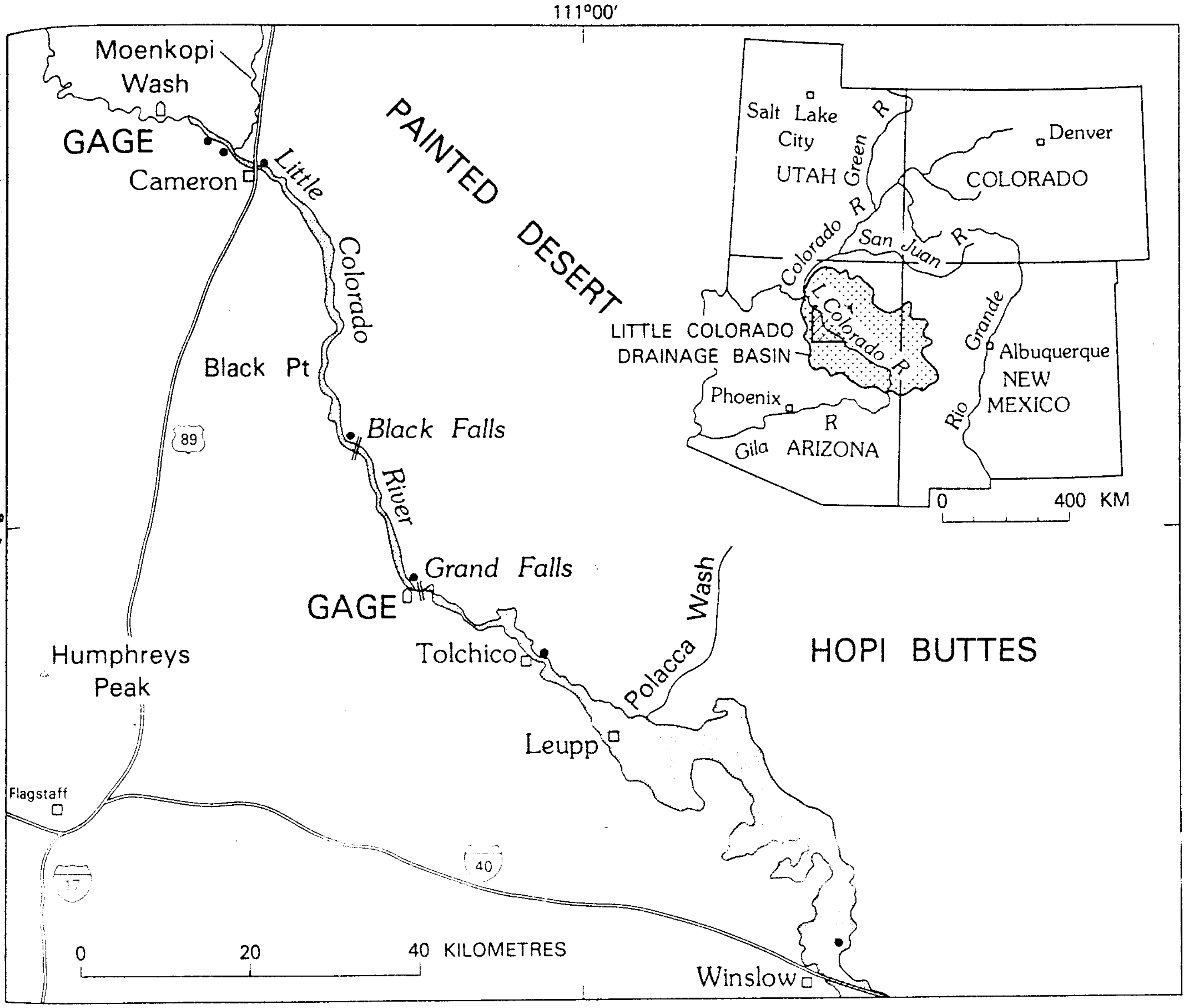

Figure 1. Little Colorado River drainage basin in the southwest United States and the study area in the Painted Desert region of northern Arizona. Solid circles are the localities listed in Table A. The distribution of recent alluvial deposits as shown were modified from the channel and youngest fluvial deposits mapped by Hereford (1979).

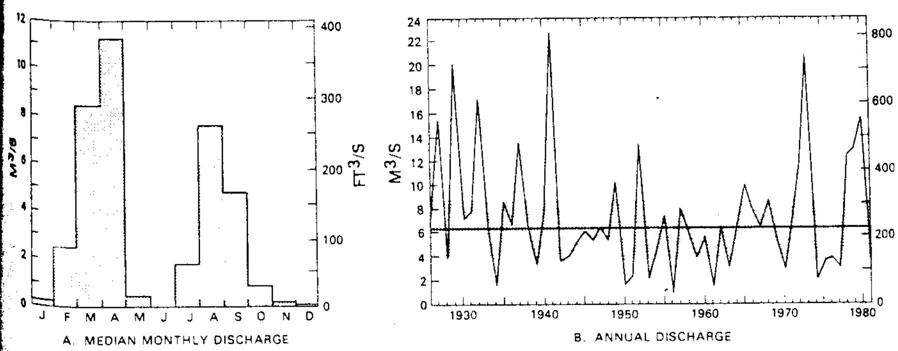

Figure 2. A. Median monthly discharge (median daily discharge by month) of the Little Colorado River, 1926 to 1981. B. Annual discharge (average daily discharge by year), 1926 to 1981. Horizontal line is the median; pattern is the 25th to 75th percentiles. Discharge data in this paper are from various annual reports and summaries published by the U.S. Geological Survey and daily discharge figures supplied by F. C. Brewsaugh (U.S. Geological Survey, Flagstaff, Arizona).

flood plain by vertical accretion, was until very recently the dominant stream process. Numerous streams in the southwestern United States have a similar history of erosion and subsequent aggradation since about 1880. Despite a voluminous literature, the causes of this widely observed stream behavior are not clear. The results of this study indicate that the historic erosional and aggradational events along the Little Colorado River occurred simultaneously with subtle changes in climate.

GEOMORPHOLOGY

Physical and Climatic Characteristics of the Drainage Basin

The physical characteristics of a drainage basin such as size, shape, and relief are important factors that moderate runoff processes. The Little Colorado River basin is relatively large and has the shape of a northwest-oriented ellipse (Fig. 1). The ellipse is 400 km in length and lies athwart the prevailing southwesterly storm track. The Mogollon Rim, a major southwest-facing escarpment with as much as 1 km of relief, lies below the southwest edge of the drainage basin. This prominent orographic barrier concentrates precipitation on the southwest side of the basin, where vegetation (mainly ponderosa pine forest) increases infiltration and reduces runoff. The elevation of the drainage basin ranges from 1,200 m in the lowest part of the valley to 3,850 m on Humphreys Peak, the highest point in Arizona; median elevation is 1,830 m. Despite the considerable range in absolute elevation, average relief is quite low. Runoff from intense, local precipitation is reduced by the low relief, which promotes infiltration, and by the large size of the basin, which requires widespread precipitation to produce significant flooding.

Bedrock is widely exposed in the Little Colorado drainage basin and consists of late Paleozoic limestone and Mesozoic sandstone, siltstone, and minor shale. Cenozoic strata are mainly weakly consolidated to unconsolidated formations of clay, silt, and sand, as well as lava flows and unconsolidated pyroclastic material. These deposits, the limestone, sandstone, and unconsolidated alluvial and eolian formations, and the abundantly distributed pyroclastic material are relatively permeable and absorb rainfall. This factor complements the low relief of the basin in reducing runoff.

The climate of the drainage basin is cool and dry, mainly arid to semiarid, and 80% of the basin receives less than 300 mm of precipitation annually. The basin lies within two very similar climatic regions charaterized by winter and summer precipitation maxima (Sellers and Hill, 1974). Precipitation and vegetation communities are closely related to elevation. The two major vegetation communities, occupying approximately 70% of the basin between 1,700 and 2,300 m elevation, are pinyon-juniper woodlands and plains and desert grasslands (U.S. Department of Agriculture, 1981). These vegetation communities are utilized for livestock production, the main economic activity in the basin.

The annual hydrograph and annual discharge of the Little Colorado are shown in Figure 2. An ephemeral stream, the river is dry in June, the driest month of the year (Fig. 2A). The number of days annually without measurable discharge has ranged from 33 to 247. The annual hydrograph reflects the biseasonal precipitation pattern. Winter runoff results from frontal precipitation as snow or rain in the higher elevations. Summer runoff is produced during the July to September rainy season after several days to several weeks of widespread convective precipitation in which rainfall occurs daily at almost all stations in the basin. Annual discharge varies widely from year to year; nevertheless, the period 1942–1961 is distinctive because 15 yr were below the long-term median (Fig. 2B). This period of reduced discharge, although of marginal statistical significance, is of practical

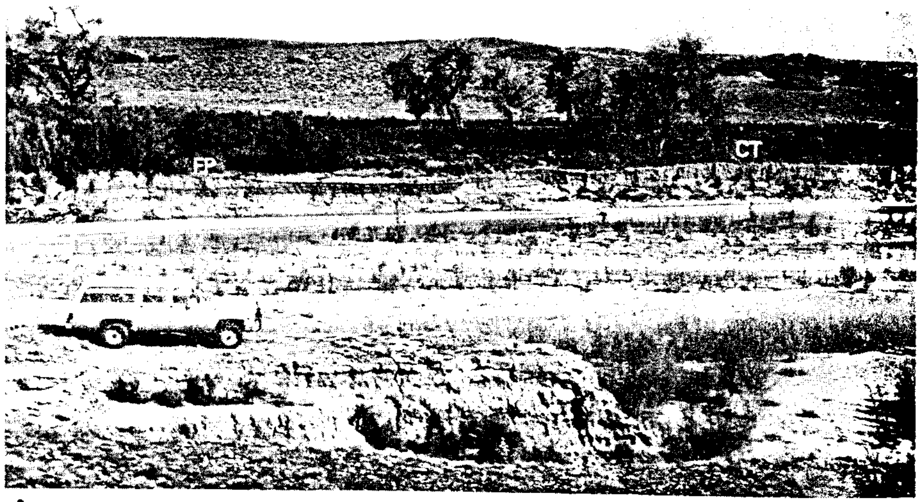

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Figure 3. A. East bank of the river at the Tolchico study area (Table A). FP = flood plain; CT = cottonwood terrace. Large cottonwoods on terrace germinated between 1880 and 1905. Large salt cedars on flood plain germinated in the late 1940s. Salt cedar in channel to right of automobile germinated in 1980. B. Stratigraphic contacts (excavated), from right to left, between cottonwood terrace alluvium (CTA) and older channel alluvium (OCA); note that the flood-plain alluvium (FPA) overlaps the older channel alluvium. Scale is 1.4 m in length.

importance because geomorphic and stratigraphic evidence indicates that the morphology of the channel changed during this period.

Geomorphology of the Historic Alluvial Deposits

Two surfaces and related deposits record the historic geomorphic activity of the Little Colorado River. The surfaces are the cottonwood terrace and the flood plain; the deposits, which postdate the cottonwood terrace and predate the flood plain, are, from oldest to youngest, the older channel alluvium and the flood-plain alluvium. The appearance of the surfaces and the stratigraphic relations between the deposits as typically exposed in the study area are shown in Figure 3.

The cottonwood terrace and the flood plain are present throughout the study area and are recognized on the basis of their height above the channel and by their vegetative cover, which indicates their age (Fig. 3A). The cottonwood terrace typically is 2 to 5 m above the channel and supports a growth of large, old cottonwood trees (*Populus fremontii* Wats.). Cottonwood is thought to produce annual growth rings (Everitt, 1968, p. 420–421; Fritts, 1976, p. 14), and ring counts of cores obtained with an increment borer indicate that most of the cottonwoods germinated between 1880 and 1905, although trees germinating as early as 1820 are also present. Photographs taken in the early 1900s (available at Special Collections Library, Northern Arizona University, Flagstaff) at the abandoned Tolchico settlement (Fig. 1) indicate that the terrace formed the banks of the river and may have been flooded during high water. The terrace probably was the active flood plain until about 1905, when cottonwoods ceased to germinate on its surface. The present flood plain is 2 to 4 m above the channel and 1 to 3 m below the cottonwood terrace. Vegetation on the flood plain is dominated by salt cedar, a nonnative riparian tree or shrub, although native cottonwood is also present. According to Turner and Karpiscak (1980), the taxonomic status of salt cedar (*Tamarisk chinensis* Lour.) in Arizona is uncertain.

The stratigraphic relationship between the deposits is commonly exposed in the cutbanks. An erosional unconformity, indicated by truncation of stratification, separates the deposits underlying the cottonwood terrace and the older channel alluvium (Fig. 3B). The flood-plain alluvium was deposited on the older channel alluvium, and the stratigraphic relationship between them is one of onlap, as illustrated in Figure 3B, caused by progressive aggradation of the flood plain. Deposition of flood-plain alluvium has covered the older channel deposits; consequently, they are not present at the surface.

The origin of the older channel alluvium was inferred from aerial photographs taken in November 1936 that show broad expanses of unvegetated sand throughout the study area in the then-active channel of the river. Where presently exposed, the deposits occur as a narrow elevated strip of sand that slopes upward, terminating against the cottonwood terrace (Fig. 3B). The alluvium has been covered with flood debris and a thin veneer of sediment indicating that it has been incorporated into the aggrading flood plain.

The geomorphology of the flood plain from the channel to its outer edge typically consists of three elements: the banks, ranging from 2 to 4 m in height; levees, rising 20 to 75 cm above the adjoining flood plain; and overbank channels. The width of the flood plain varies from less than 100 m to as much as 12 km, depending on the width of the valley. Most of the areas studied in detail (Table A; Fig. 1)[1] are in relatively narrow rock-walled canyons, where the flood plain and related deposits could be mapped at large scale. The geomorphic elements of the flood plain as typically present in a rock-walled canyon are shown in Figure 4. At this locality, the reduction in the width of the channel due to flood-plain aggradation is striking, because in the November 6, 1937, aerial photographs, the unvegatated channel extended to the outer edges of the present wooded flood plain.

The relationship between alluvium and vegetation indicates that most sediment near the surface of the flood plain was deposited by recent floods, associated with the large runoff years of the 1970s (Fig. 2B). Large trees and bushes on the flood plain have been partly buried by alluvium since they germinated, but the trunks were not bent by the weight of the sediment or by the force of the flood water, and the trunks can be traced directly to the germination layer (Fig. 5). This is somewhat different from the "natural layering" of salt cedar illustrated by Everitt (1980, Fig. 4). The partially buried trees look like poles or posts stuck in the ground at a high angle and resemble the buried trees described by Sigafoos (1964). Immature as well as mature salt cedar, cottonwood, and willow trees are partly buried, suggesting that deposition occurred recently. In addition, flood debris and alluvium near the surface of the flood plain

[1]Tables A and B may be secured free of charge by requesting Supplementary Data 84-15 from the GSA Documents Secretary.

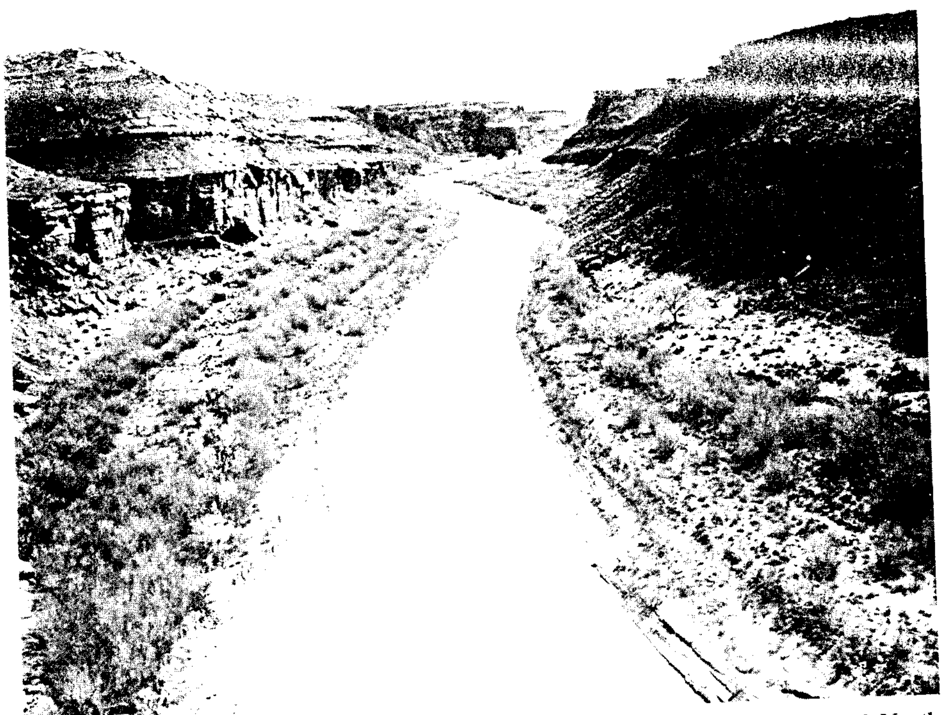

Figure 4. Flood plain and channel below Moenkopi Wash (Table A, Moenkopi Wash Northwest) on February 26, 1982 (view is upstream). Channel is 51 to 56 m wide; banks are 3 to 4 m high; discharge is rather low at about 35 m^3/sec. L = levee; OBC = overbank channel. Vegetation is mainly salt cedar, but native cottonwood trees are also present.

Figure 5. Typical occurrence of partially buried and exhumed salt-cedar tree in the right cutbank near Black Falls. The tree germinated in 1966 and was partially covered several times between 1969 or 1970 and 1978. Note the numerous adventitious roots developed from the trunk and stems after burial.

contain plastic artifacts such as a dish-soap bottle with a push-pull cap and a chemical container with an expiration or manufacture date of August 9, 1969. In several places (Winslow and Black Falls; Fig. 1; Table A), recently constructed barbed-wire fences, one having steel posts with unfaded paint, are covered by alluvium.

In short, the geomorphology of the historic alluvial deposits suggests that from about 1905 to 1937, erosion was the dominant geomorphic process. This is indicated by the absence of aggradational land forms in the channel and the widespread erosional unconformity between the cottonwood terrace deposits and the older channel alluvium (Fig. 3B). Sometime after 1937, however, the flood plain began to aggrade (the actual year is probably 1952, the date of the oldest flood-plain deposit) and aggradation has until very recently been the dominant process.

Other streams in the region have a similar history of erosion and aggradation within the past 100 yr. On Black Mesa in the northeast portion of the Little Colorado basin, Euler and others (1979) dated a sequence of alluvial deposits that are approximate temporal equivalents of the cottonwood terrace and flood plain ("z" and "a" deposits of Euler and others, 1979). East of the Little Colorado near Santa Fe, New Mexico, the recently aggraded channel deposits are inset beneath a terrace that was incised not long after 1880 (Leopold, 1976). The Chaco River in Chaco Canyon, New Mexico, developed a flood plain after 1939, following a period of entrenchment (Love, 1979). In addition, Emmett (1974) reported that several streams in the western United States aggraded their channels between 1962 and 1973. The latter part of this period was a time of significant flood-plain accretion on the Little Colorado.

Historic Changes in Channel Morphology and the Spread of Salt Cedar

The morphology of the channel when erosion was the dominant process was considerably different from its geometry during the succeeding aggradational period. A variety of evidence indicates that the channel was broad, sandy, and unvegetated for at least the first 40 yr of this century. At some time shortly after 1937, the width of the channel decreased and vegetation began to occupy and stabilize higher parts of the older channel. These events marked the end of the erosional period and the beginning of flood-plain development and probably were caused by a decrease in the frequency of large floods.

From the late 1890s until at least the late 1930s, the channel was approximately 50% wider than at present. Photographs (available at Special Collections Library, Northern Arizona University, Flagstaff) taken at several river crossings between 1890 and 1926 show a broad unvegetated channel. Describing the river between Cameron and Holbrook, Colton (1937) reported that it had a broad sandy channel with steep cutbanks. Topographic maps of the river as surveyed in 1926 and 1934 (Lewis, 1948; Fischer, 1947) confirmed Colton's observations and showed a channel about 50% wider than at present. Many of the channel bars in the 1926 and 1934 maps have been incorporated into the flood plain. Aerial photographs taken in November 1936 by the U.S. Soil Conservation Service (approximate scale 1:27,000) showed that the channel was broad and free of vegetation throughout the study area. More recent aerial photographs, taken in 1954 by the U.S. Army Map Service (approximate scale 1:54,000) and in 1968 by the U.S. Geological Survey (approximate scale 1:24,000), showed that the width of the channel had been considerably reduced and that a vegetated flood plain occupied parts of the older channel. This evidence indicates that the flood plain began to form between 1937 and 1954. Figure 6 illustrates the changes in the channel between 1936 and 1968 at Grand Falls and Black Point. These changes are similar in type and timing to those reported on the Chaco River in northwest New Mexico (Love, 1979) and to those along the Green, Colorado, and Fremont Rivers in the central Colorado Plateau (Graf, 1978, 1982).

Associated with the decrease in channel width of the Little Colorado was the appearance of riparian vegetation, chiefly non-native salt cedar, in areas that formerly were part of the channel. A tree or shrub native to the Mediterranean area, salt cedar was introduced into the western United States as early as 1854 (Robinson, 1965, p. A4). Since about 1925, the plant has become naturalized, often displacing native species in many stream valleys of the arid to semiarid western United States (Christensen, 1962; Robinson, 1965). Owing to complex environmental interactions (Everitt, 1980), the direct effect of salt cedar on the morphology of the alluvial channels in these stream valleys is unclear. Hadley (1961), Robinson (1965), and Graf (1978, 1979) concluded that the changes in channel morphology resulted largely from the spread of salt cedar into the channel system. Everitt (1979, 1980), however, pointed out that the spread of salt cedar into many channels occurred simultaneously with other environmental changes, and that salt cedar is no more effective at stabilizing sediment than are native riparian species.

The history of salt cedar in the Little Colorado River valley is summarized in Table B.[2] The plant was growing under cultivation very close to the river at Winslow in 1909 but was unknown in the river system until 1937. Considering the propensity of salt cedar for the

[2]See footnote 1.

riparian environment, particularly its ability to spread rapidly through the channel system (Graf, 1978), this delay of more than 28 yr is unexpected. The absence of salt cedar in the channel system probably was caused by the large floods that were typical during the early part of this century. Colton (1937, p. 19) reported that since the late 1890s there had occurred a number of disastrous floods that uprooted many of the cottonwood trees along the banks (presumably trees on the cottonwood terrace). The five largest recorded floods, moreover, occurred between 1923 and 1938. This period of large floods probably delayed the spread of salt cedar and prevented native plants from occupying the channel system.

Many workers have noted that the vertical zonation of riparian vegetation is related to flood levels: many species will not grow below a level in the channel that marks the height of recent floods (Sigafoos, 1964; Everitt, 1968; Wolman, 1971). The relationship between flood levels and the spread of salt cedar was investigated near the Grand Falls gaging station (Fig. 1; Table A), where discharge records overlap in time with the growth of a small grove of salt cedars. Figure 7, a transverse cross section through the channel about 200 m downstream from the station, shows the relationship between discharge levels and the germination position of salt cedar in the channel. The trees germinated on and stabilized a surface corresponding to a rather low flood discharge. Judging from the flood marks (Fig. 7), this level of the channel was deeply inundated during the 1923 and 1929 floods. Further analysis indicates that the plants became established during a subsequent period of generally low floods. Ring counts from eight living trees yielded germination dates ranging from 1947 to 1961. This time interval and the

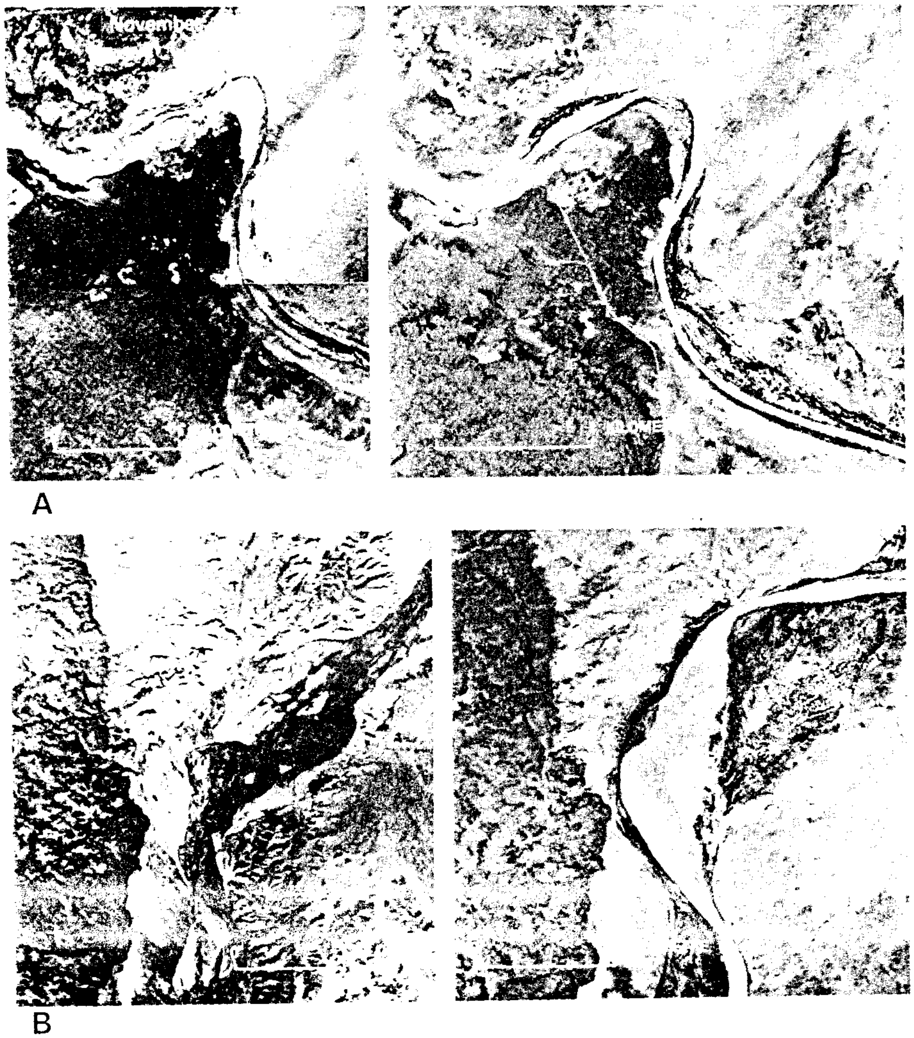

Figure 6. Aerial photographs illustrating changes in the channel of the Little Colorado River at Grand Falls (A) and Black Point (B). The left side of each pair of photographs shows the condition of the channel in November 1936, and the right side illustrates the channel and flood plain in 1968.

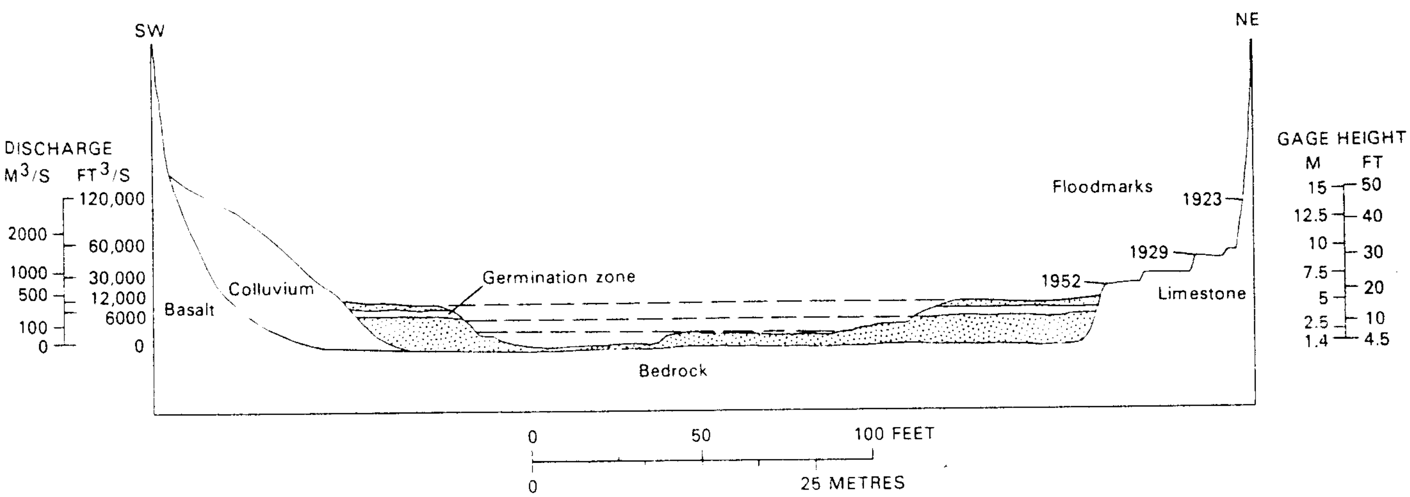

Figure 7. Cross section of the Little Colorado River 200 m below the Grand Falls gaging station (Table 1). Horizontal and vertical scales are equal. Velocity measurements for the gage were made at this locality.

range in flood stage of the germination zone are plotted with the stage of the annual flood (largest flood during the water year from October to September) in Figure 8, which indicates that these trees and others of similar age along the Little Colorado germinated during a period of relatively low annual floods and discharge (Fig. 2B). It is reasonable to conclude, therefore, that the naturalization of salt cedar on the Little Colorado was delayed until the frequency of large floods decreased in the early 1940s.

Additional evidence indicating that the spread of salt cedar is dependent on flood conditions comes from the Grand Canyon, where Turner and Karpiscak (1980) found that salt cedar spread rapidly through the canyon after Glen Canyon Dam was closed in 1963. The regulated release of water apparently has provided a dependable, nondestructive source of moisture that promoted salt-cedar growth on sand bars.

The decrease in channel width and the spread of vegetation into the channel of the Little Colorado were related events that quite likely occurred simultaneously in response to reduced discharge. The smaller floods and discharge occurring after the early 1940s required a narrower channel in which sand bars and other elevated areas were not deeply flooded or mobilized. Native vegetation would have colonized this newly formed riparian environment, but salt cedar, because it displaces native species (Turner, 1974), became the dominant plant. Once established, the plants trapped sediment during times of overbank flow and in this manner the flood plain aggraded. In conclusion, the recent changes in channel geometry along the Little Colorado River probably were initiated by a reduction in the number of large floods and annual discharge that took place in the 1940s and 1950s.

These conclusions are consistent with the findings of other studies. Historic changes in channel shape on a number of rivers and in a variety of climates were reviewed by Stevens and others (1975), who found that channel morphology generally was a function of the sequence of flood events. A specific example of how sand-bed streams in a semiarid environment responded to reduced discharge is provided by the Platte and North Platte Rivers in Nebraska (Williams, 1978). Since the early 1900s, both stream beds narrowed considerably and vegetation spread into the channels. Decreased channel width was caused by reduced annual peak floods and average yearly discharge. The annual discharge and peak floods were diminished by the regulating effects of major upstream dams and greater water use by man. The situation on the Platte River that eventuated under controlled discharge is analogous to the naturally occurring events on the Little Colorado.

Effect of Human Activity on the Historic Changes

Could the changes in flood intensity and discharge that altered the channel morphology of the Little Colorado River have resulted from human activity? The question merits consideration because the effects of human activity on stream behavior can be similar to those caused by climatic change (Leopold, 1976). Extensive overgrazing could modify the rainfall-runoff relationship by soil compaction and destruction of vegetation, thereby reducing infiltration and increasing runoff. On the other hand, increased water storage and consumption could reduce discharge and floods, causing channel stability and flood-plain development.

The erosional episode was associated with introduction into the drainage basin of large herds of domestic animals. Large-scale cattle and sheep ranching began in the southwestern one-half of the basin shortly after settlement of the area in the 1870s and 1880s and probably reached its peak after completion of the transcontinental railroad in the mid-1880s (Colton, 1937). In the northwestern half of the basin, sheep and cattle have been raised in large numbers by the native inhabitants for several hundred years, and the size of these herds grew steadily, probably reaching a maximum by 1930 (Reno, 1981). The widespread stream erosion observed by early workers in the study area was attributed mainly to climate by Gregory (1917) and to overgrazing by Colton (1937), Lockett and Snow (1939), and Thornthwaite and others (1942).

On the basis of his observations of stream processes over most of the southern Colorado Plateau during the early 1900s, Gregory (1917) believed that small changes in the amount of rainfall, or its distribution, or the character of storms were sufficient to cause the widespread stream erosion. He noted that the meager vegetation had little effect on runoff and that erosion occurred along streams that had not been utilized for grazing. In their extensive study of Polacca Wash (Fig. 1), however, Thornthwaite and others (1942) could not detect a change in climate and concluded that stream erosion was associated with and largely caused by overgrazing.

The severe overgrazing described by the early workers (Colton, 1937; Lockett and Snow, 1939; Thornthwaite and others, 1942) may not have been typical of the entire basin. Today, just as during the early 1900s, the effects of overgrazing are conspicuous along fence lines, around water holes, and near settlements (see Figs. 41 and 42 in Thornthwaite and others, 1942), but other areas do not show such extensive overuse. The extent of overgrazing may have been overestimated by these workers, when averaged over the entire drainage basin.

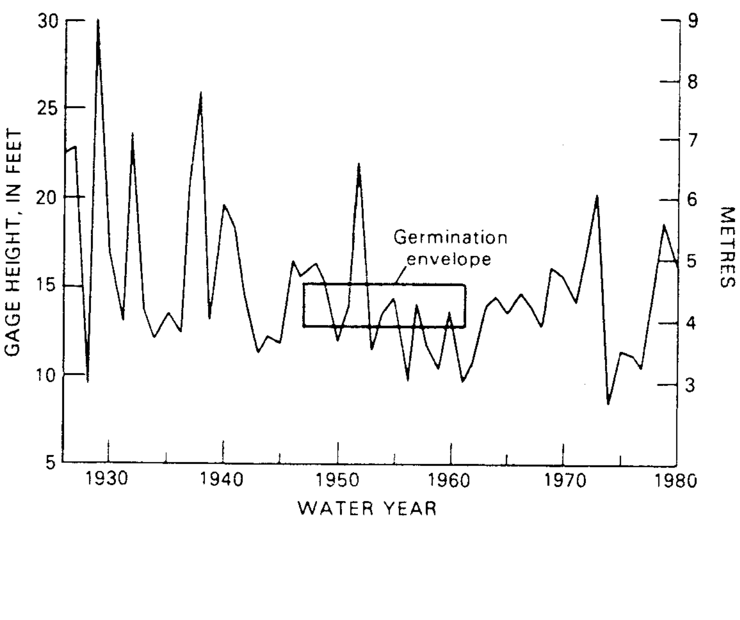

Figure 8. Stage of the annual flood at the Grand Falls gaging station, 1926 to 1980. The germination envelope is the range of elevations (gage height) of the oldest and youngest partially covered salt cedars. Measurements are not available for this station for 1961–1969, 1971, and 1973–1980. The stage was estimated in the missing years by adjusting the recorded value at the Cameron station to the equivalent stage at Grand Falls. Analysis of the partial duration flood series of the two records during the overlap period (1947–1959) indicates that statistically they are indistinguishable.

There is insufficient information to adequately evaluate the hydrologic impact of grazing in the Little Colorado River basin, even though grazing and stream erosion began at about the same time. Most workers in the Little Colorado basin during the early 1900s noted the poor condition of vegetation in the desert grasslands, which they associated with overgrazing and stream erosion (Colton, 1937; Lockett and Snow, 1939; Thornthwaite and others, 1942). Recent studies of range hydrology reviewed by Gifford and Hawkins (1978), however, showed that estimates of range condition (an expression of the status of the present plant community relative to that of the natural community) are generally inadequate to evaluate the hydrologic impacts of grazing (Gifford and Hawkins, 1978). Range condition, moreover, is not independent of climate. In a study of grazed and ungrazed desert grasslands like those in the Little Colorado River basin, Davey (1980) found that moisture (annual and seasonal precipitation) was the primary factor limiting grass productivity and that a soil moisture threshold must be surpassed for approximately three days for a summer growth pulse to occur. The poor range conditions and stream erosion, therefore, may have resulted as much from climatic conditions as from poor land use.

The subtle changes in climate that Gregory (1917) suggested as the cause of the increased erosion during the early 1900s probably did occur, as indicated by the rise in mean surface air temperature over the Northern Hemisphere from about 1900 to the 1940s (Bryson, 1974; Kalnicky, 1974). This period was characterized by atmospheric circulation patterns different from those existing since about 1950 (Kalnicky, 1974), and it is likely that precipitation patterns also were different, as suggested by Leopold (1951) for the early 1900s and by Douglas and others (1982) for recent decades.

An artificial reduction in discharge by increased water consumption and storage, as exemplified by the Platte Rivers (Williams, 1978), will cause a decrease in channel width resulting in flood-plain development. The Little Colorado, unlike the Platte Rivers, is largely free-flowing: the several reservoirs that have been constructed are relatively small and close to the head waters. Furthermore, the largest reservoirs were built early in the century. Water use by man has not significantly reduced the discharge characteristics of the river because annual discharge and flood stage have not steadily declined over the measurement period (see Figs. 3B, 8, and 15A below). It seems reasonable to conclude that flood-plain aggradation probably occurred without significant human interference; however, the erosional episode was associated with and perhaps aggravated by changes in land use.

FLOOD-PLAIN STRATIGRAPHY

Sedimentology of the Flood Plain

In the preceding section of the discussion, the chronology, effects, and general causes of the erosional and aggradational processes were deduced from geomorphic evidence. By itself, however, this type of information provides little insight into the stratigraphic record or the functional details of the aggradational process. In this section of the discussion, the mechanics of flood-plain aggradation are inferred from the sedimentology and stratigraphy of the flood-plain alluvium.

The sedimentologic distinction between channel and flood-plain deposits is produced largely by the loss of stream power when flood water overtops the banks and spreads across the flood plain (Allen, 1970, p. 138–139). Compared to channel sediment, flood-plain sediment is generally finer grained and has sedimentary structures indicative of less stream power. The histograms in Figure 9 show the average grain-size distribution of channel and flood-plain sediment from the Little Colorado. Sand in the channel is on the average 0.8 to 1 phi-size coarser than sand in the levees and overbank channels. Specifically, channel sand is mainly fine to medium sand, whereas the flood-plain alluvium is characterized by fine to very fine sand.

The channel and flood-plain alluvium of the Little Colorado are further distinguishable by the occurrence in each of distinctive sedimentary structures. Channel bars are characterized by trough and low-angle cross-stratification as well as by horizontal stratification. These structures are typical of sandy alluvial channels of perennial and ephemeral streams (Allen, 1965; Picard and High, 1973), and they form in the upper-lower to upper flow regime (Harms and Fahnestock, 1965). In contrast, flood-plain alluvium of the Little Colorado has features such as climbing ripple structure and ripple lamination that typically form in the lower flow regime. These sedimentary structures generally occur on river flood plains, particularly levees, where the suspended load of sand and finer material is quickly deposited during overbank floods (McKee, 1966). Another distinctive sedimentary structure, centriclinal cross-stratification, which forms when flood waters overtop and then recede from a wooded flood plain (Underwood and Lambert, 1974), is present encircling salt-cedar trunks buried by the flood-plain alluvium.

The flood-plain alluvium is made up of a number of beds that accumulated by vertical accretion (Fig. 10), and each bed has a distinctive stratification sequence that developed during and after a single overbank flood. A typical stratification sequence in the levees consists of a basal scour surface, clayey to silty sand with ripple lamination, climbing ripple structure, or centriclinal cross-stratification, and commonly a clay drape or duff zone composed of salt-cedar leaves. The stratification sequence in the overbank channels is similar to that of the levee except that the clay drape is consistently well developed. Fining-upward cycles are present only within sequences that include the clay drape; otherwise, there is no upward decrease in grain size. Scour surfaces probably form when flood waters spread rapidly across the flood plain or by wind deflation. Clay drapes at the top of the sequence probably form from waning floods. Duff results from leaf drop between overbank floods, mainly during the fall months, when salt cedar sheds its leaves. Individual flood deposits have been recognized by other workers using similar criteria. Ray (1976) found that the deposits of individual floods in Mississippi River overbank alluvium typically were bounded by clay drapes and erosional surfaces. On the Cimarron River, Schumm and Lichty (1963) noted that the stratification sequence produced by individual floods consists of a basal scour surface overlain by sand that is capped by a clay layer.

These erosional surfaces and organic accumu-

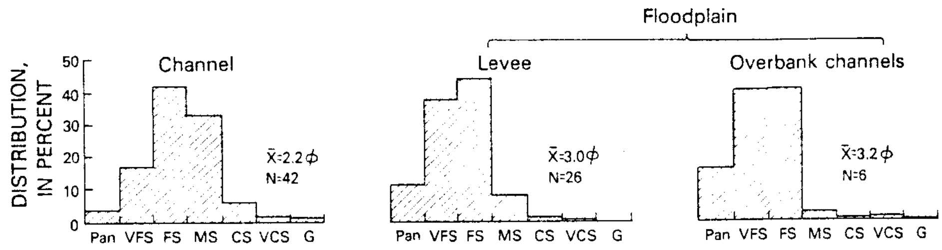

Figure 9. Grain-size distributions of channel and flood plain (levees and overbank channels) sediments from samples collected at the localities listed in Table A.

Figure 10. Flood-plain stratification in a levee on the south bank at the Moenkopi Wash Southeast section (Table A). Scale is 1.4 m in length with 10-cm divisions. Six beds deposited between 1969 or 1970 and 1978 are exposed in the face. Bedding planes were excavated for emphasis.

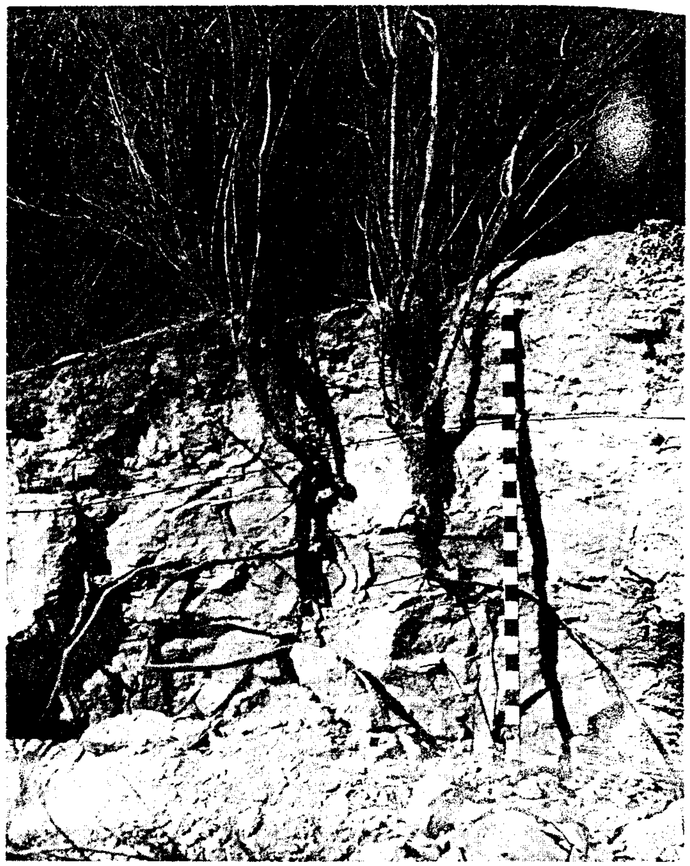

Figure 11. Cutbank exposure of buried salt cedar. The trees germinated on a scoured surface locally associated with duff. The root collar is 35 cm below the top of the scale (5-cm divisions).

lations in the stratification sequence of the Little Colorado River flood-plain alluvium are evidence of periods of nondeposition and exposure of the flood plain. Root collars, marking the germination position of buried salt cedars (Fig. 11), occur with scour surfaces, clay drapes, and duff zones, indicating that the flood-plain surface was stable long enough for trees to establish themselves. The root collars are an important constituent of the stratification sequence because, in the case of living trees, the duration of the nondepositional interval can be determined by counting rings to the germination date.

Dendrochronology of Salt Cedar

Are salt-cedar growth rings produced annually? In the following section of the discussion, considerable accuracy is claimed in dating and correlating the flood-plain deposits on the assumption that ring counts accurately determine the year of germination. Although many trees are known to produce annual rings, salt cedar is not listed among them by Fritts (1976, p. 14). Within a single growing season, a tree potentially can add one ring, add multiple rings, skip a ring, or produce a discontinuous ring. Because of these false and missing rings, ring counts do not necessarily determine the actual germination date. Nevertheless, there is abundant evidence that the growth rings of salt cedar are produced annually in most cases.

Salt cedars in the Little Colorado River valley are deciduous, shedding their leaves during the latter half of October. At Winslow, freezing temperature normally occurs first on October 21 and the last date of 0 °C is frequently around April 28 (Sellers and Hill, 1974). During this period of recurrent freezing temperatures, salt cedar is dormant and probably cannot produce rings until growth resumes with the return of warmer weather in late spring. Temperature, therefore, modulates growth in an annual cycle, and the resulting growth rings are also annual (Fritts, 1976).

Additional evidence that salt cedar typically produces annual rings comes from the following observations. At Tolchico, young trees, 1 to 1.5 m in height, germinated on the surface of bars and other deposits that were mobilized during the high spring runoff of 1980 (Fig. 2B). These trees had one ring in late 1980, after they had lost their leaves. By late 1981, the same group of trees had two rings and by late 1982, they had three rings. They thus produced rings annually for at least the first three seasons of their growth. In addition, the germination dates of mature trees agree well with independent evidence that salt cedar was not present before 1937. Ring counts of mature salt cedars growing at several localities between Winslow and Cameron where none appear in the November 1936 aerial photographs indicate that they germinated after that time. Where mature salt cedar and cottonwood trees occur together on the flood plain, their ring-count ages differ by at most a few years, and cottonwood is known to produce annual rings (Fritts, 1976). Finally, the oldest salt cedar found during this study germinated in 1937, although most mature trees are somewhat younger, having germinated in the mid-1940s. If salt cedar regularly produced intra-annual rings (which is not the case), there would be many pre-1937 germination dates.

False or missing rings, however, can be recognized reliably only through the process of cross dating, whereby ring-width patterns are matched within the tree and among other trees in nearby and neighboring stands (Stokes and Smiley, 1968; Fritts, 1976). The specimens used in this study were not cross-dated. However, J. S. Dean (Laboratory of Tree-Ring Research, University of Arizona) analyzed 16 samples and found that they are cross-datable, although the trees are too young to allow unequivocal identification of the ring-width patterns (J. S. Dean, 1981, personal commun.). These results suggest that salt-cedar trees in the study area (and perhaps salt cedar in general) are a reliable dendrochronologic tool.

Physical Stratigraphy and Age of the Flood Deposits

Three major breaks indicating periods of flood-plain stability and salt-cedar germination are recorded in the physical stratigraphy of the flood-plain alluvium. These and associated deposits are called, respectively, the older, intermediate, and younger germination horizon or alluvium. The ages of the horizons, determined by time-stratigraphic analysis, are 1937–1951 (older), 1953–1969 or 1970 (intermediate), and 1974–1977 (younger). The intervening years (namely, 1952, 1969 or 1970 to 1973, and 1978) were periods of flood-plain deposition.

Figure 12 illustrates the stratigraphy of the germination horizons as well as the relation between the flood plain and other deposits at four of the localities in Table A. In most places, the germination horizons and associated flood-plain deposits are stacked on top of each other in standard stratigraphic succession (Fig. 10). In places, however, such as Tolchico (left bank), Cameron (right bank), and Moenkopi Wash Northwest (right bank), the deposits appear to have a cut-and-fill relationship, suggesting separate periods of erosion and aggradation rather than progressive vertical accretion. This positioning, however, probably resulted from different flood heights and from relief on the flood-plain surface. Levees, for instance, build up above the general level of the flood plain; eolian processes redistribute sediment and locally increase relief. Vertical accretion, therefore, rather than successive episodes of cutting and infilling, has been the dominant sedimentary process indicated by the physical stratigraphy of the flood-plain deposits.

The youngest deposits, those in the channel (Fig. 12) such as side bars and channel bars that were active during the February 1980 runoff, have a cut-and-fill relationship with the flood-plain alluvium. Relationships of this type between fluvial deposits suggest erosion, perhaps caused by lateral movement of the channel, followed by, or associated with, migration of channel bed forms. Activity of this sort is a process of lateral accretion whereby sedimentation is associated with channel migration and bank erosion. Such lateral accretion, evidenced by the physical stratigraphy of the youngest sediments, is in marked contrast to the vertical accretion recorded by the flood-plain deposits.

Time-stratigraphic analysis of the flood deposits suggests they are of the same age throughout the study area. Table 1 lists the stratigraphic position and germination dates of salt cedars collected at stratigraphic sections measured at the localities in Table A. The range of dates in Table 1 (except for 1971 Moenkopi Wash Southeast) is plotted as a correlation chart in Figure 13. The three periods of flood-plain stability occurred at approximately the same times throughout the study area, with the possible exception of the Grand Falls locality, where the lack of cutbank exposures prevents identification of the older and intermediate horizons (Figs. 7, 13). The uncertainty in the chart regard-

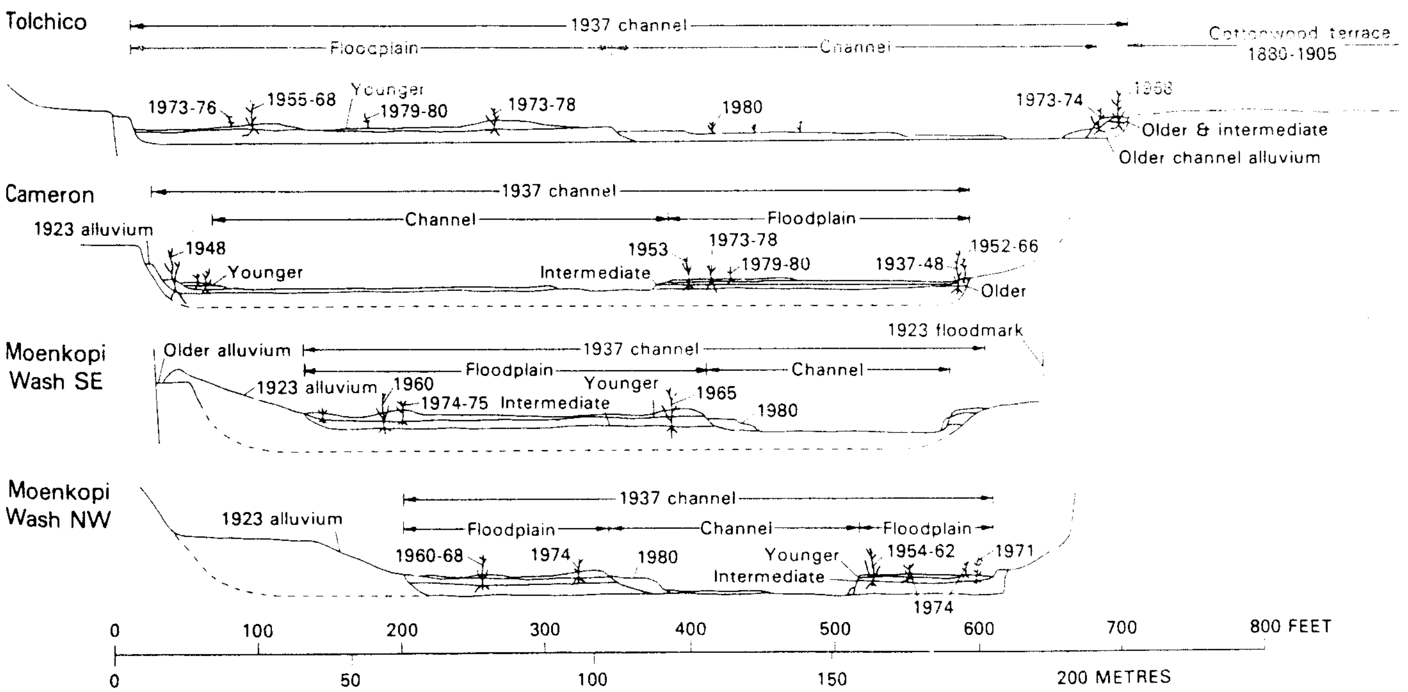

Figure 12. Cross sections, constructed from maps made by plane table and alidade at the scales listed in Table A, illustrating the surface morphology and stratigraphy of the flood plain and older deposits. Extent of the 1937 channel was determined by its morphologic expression and by reference to the November 1936 aerial photographs. "Older," "intermediate," and "younger" refer to salt-cedar germination horizons. Dated trees on flood plain indicate approximate age of salt-cedar groves. The stratigraphic position of buried root collars was inferred from position of trees of similar age exposed in cutbanks (Table 1). Surveyed in January 1981; last flow, September 1980.

TABLE 1. STRATIGRAPHIC POSITION OF GERMINATION HORIZONS AND GERMINATION DATES

Locality	Height above river bed (cm)	No. samples	Germination date(s)
Winslow	100-120	4	1973, 1974(2), 1977
Tolchico	183-194	4	1945, 1946, 1947, 1951
	140*	1	1968
	208	2	1973, 1974
Black Falls	244	4	1958, 1964, 1968, 1969
	307-310	15	1974(2), 1975(8), 1977(3), 1978(2)
Grand Falls	290-366	8	1947, 1950, 1951, 1952, 1953, 1956, 1959, 1961
Cameron	..†	3	1937, 1948, 1952
	107-110	1	1969
	136-139	4	1973(3), 1977
Moenkopi Wash Southeast	127	4	1956, 1963, 1964, 1966
	170	1	1971
	340-350	6	1975(4), 1977(2)
		Total 57	

*Apparent inset stratigraphic relation.
†Germination horizon not exposed for two older samples; youngest sample germinated on surface.

ing the age and correlation of the deposits is probably caused by the limited number of samples; it is likely that the samples analyzed do not bracket the entire nondepositional interval at each locality. Nonetheless, the duration of the nondepositional intervals can be estimated, within the constraints imposed by the physical stratigraphy, by comparing these intervals with the flood record, because only the larger floods in the period significantly overtop the banks, causing aggradation or flood-plain instability.

The stratigraphic column on the right side of Figure 13 shows the age of the deposits as deduced from the flood record. Germination dates (Table 1; Fig. 13) and the stratigraphy at Cameron and Tolchico (Fig. 12) suggest that the older alluvium was deposited in 1952, when the largest flood since 1938 occurred (Fig. 8). After 1952, there were no significant overbank floods until 1969; several large floods between 1969 and 1973 (Fig. 8) deposited the intermediate-age alluvium. From 1974 until 1977, the river did not overtop its banks; 1974 had the lowest annual flood ever recorded at the gage near Cameron, and the following three years were only slightly higher (Fig. 8). Finally, trees that germinated in the summer of 1978 are covered by the younger alluvium (Figs. 12, 13). The river was last out of its banks on December 21, 1978, when the younger alluvium most likely was deposited. Despite the uncertainties in correlation indicated by Figure 13, the dated horizons calibrate reasonably well with the flood history of the river, and the deposits are reasonably correlative throughout the study area.

As previously discussed, the intermediate alluvium was deposited by several floods. Figure 14 is a section of intermediate and younger alluvium at the Moenkopi Wash Southeast section (Tables 1 and A; Figs. 10 and 12). The germination dates shown in the figure were calibrated to the flood record as follows. At the gage near Cameron, the largest flood since 1952 occurred on September 7, 1970; but a year earlier, on September 11, the river crested only 15 cm below the 1970 mark. Both floods probably overtopped the banks; consequently, deposition began in either 1969 or 1970. The bed covering the tree germinating in 1971 (Fig. 14) probably was deposited by the flood of October 3, 1971, the largest since 1970 at the Grand Falls and Cameron gages. The remaining three beds of the intermediate alluvium were deposited by two overbank floods in October 1972 and by the sustained high runoff of spring 1973 that crested on April 18.

In summary, seven floods occurring between 1952 and 1978 were the primary cause of flood-plain aggradation. Table 2 lists the floods and the climatic conditions associated with them. For the most part, floods occurred during wet years in the Little Colorado basin and the western United States. The germination horizons also developed during periods of distinctive climatic patterns of regional extent. The older and intermediate horizons partially correspond with the drought conditions that existed in the southwest United States between 1942 and 1956 (Thomas, 1963). The younger germination horizon and the corresponding period of low discharge (Fig. 2B) in the mid-1970s likely reflect a period of widespread drought that elsewhere was nearly as severe as any since the dust-bowl era of the mid-1930s (American Meteorological Society Bulletin, 1977; Felch, 1978; Matthai, 1979).

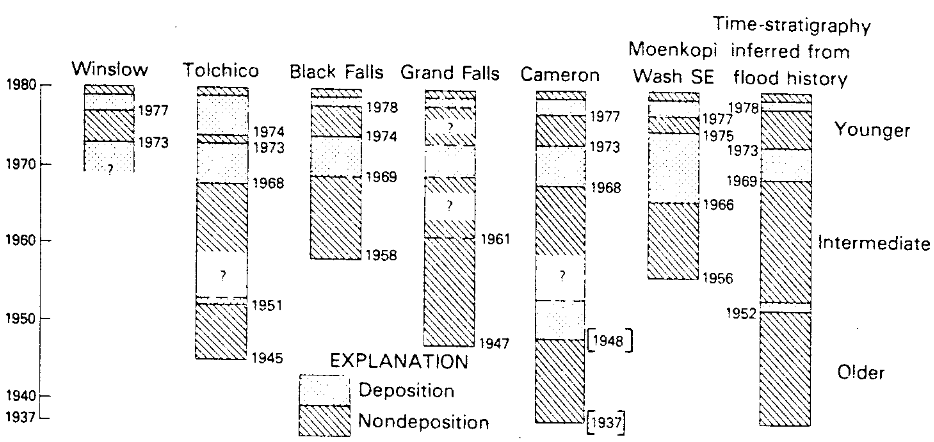

Figure 13. Time-stratigraphic correlation of germination horizons. Diagonal lines indicate the range of dates from each of the horizons in Table 2 and represent periods of nondeposition. Queried where the beginning of nondeposition is unknown because of limited number of samples; boundary dashed where inferred. Dates from trees with unexposed root zones shown in brackets. Column on right shows age of deposits inferred from the flood record.

Interpretation

The recent alluvial history of the Little Colorado suggests that sedimentation was largely controlled by the discharge characteristics of the river that are linked to the climate of the drainage basin. Variations in climate altered discharge, causing a variety of responses in the channel. For the most part, sedimentation was allocyclic—that is, under the control of processes originating outside the drainage basin (Miall, 1980). Erosion predominated during the early history of the river, when numerous floods eroded the cottonwood terrace and swept the channel free of vegetation. After a decrease in discharge, beginning in 1942 (Fig. 2B), the flood plain began to develop and subsequently accreted by overbank deposition during large

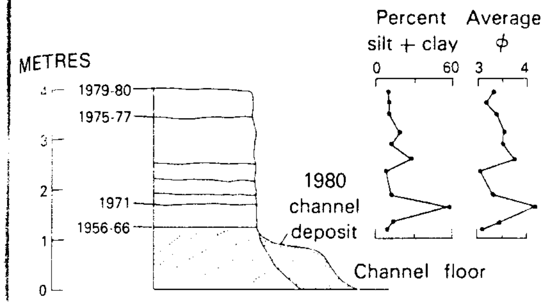

Figure 14. Stratigraphic section of a levee in the flood plain near the confluence of Moenkopi Wash with the Little Colorado River (Table A). Youngest dates (1979–1980) are from trees rooted on surface of the levee; older dates are from Table 1.

floods. In 1980, however, sedimentation changed from vertical to lateral accretion, but this shift in sedimentary style was not associated with a change in discharge. The flood and runoff of 1980 were as large as others that in previous years overtopped the banks and deposited sediment on the flood plain (see Figs. 2B and 8), yet in 1980 the banks and flood plain were not overtopped and instead were eroded, suggesting that the flood plain had become a terrace.

The change from vertical to lateral accretion probably was caused by increased energy in the channel and is an example of an autocyclic sedimentary process (Miall, 1980). Vertical accretion is limited by the elevation of the flood plain above the channel; beyond a certain height, unless the channel aggrades, successively larger floods will be necessary to overtop the banks. Once a flood plain aggrades to this critical height, floods of a given frequency will be confined within the channel rather than overtopping the banks and dissipating a portion of their energy across the flood plain. As a result, the average velocity in the channel is higher, because reduced hydraulic roughness and wetted perimeter provide greater channel efficiency, and the capacity of the stream to do work is greater, resulting in bank erosion and lateral migration of the channel. Aggradation and net erosion of the flood plain evidently are separated by a threshold, the elevation of the flood plain. This type of geomorphic threshold separating aggradation and erosion was described by Bull (1979). It is also an intrinsic geomorphic threshold that, when crossed, will cause a complex response (Schumm, 1977) in the channel as the system attempts to reach a new equilibrium.

The existence of a threshold indicates that sedimentation, when considered over a long time span, perhaps several centuries, is not totally allocyclic. The stratigraphic record probably will consist of some combination of allocyclic and autocyclic deposits. Of the two basic alluvial environments, channel and flood-plain, the latter is likely to be allocyclic. The flood plain, as suggested by Leopold and others (1964), is a constructional land form, the upbuilding of which is related to discharge and climate. Flood-plain deposits in the stratigraphic column probably record a period when the river adjusted to a new discharge regimen by vertical flood-plain accretion. After the elevation of the flood plain reaches the height of the larger floods, it can become a terrace without a fluctuation in discharge. Further addition of flood-plain deposits to the stratigraphic column requires some additional or continuing climatic change that causes the river to adjust through further sedimentation. Channel deposits, on the other hand, are erosional forms that are not necessarily allocyclic; moreover, channel sediment is frequently reworked, so that only the most recent events are recorded by its stratigraphy. A sequence of channel deposits may be the result of lateral accretion that is caused by exceeding the threshold of flood-plain height or by climatic factors, such as during a period of frequent large floods.

In summary, the recent alluvial history of the Little Colorado suggests that flood-plain aggradation occurred allocyclically as the river adjusted to a change in hydrology. As Lewin (1978) noted, aggradation is not necessarily a response to autocyclic processes but may reflect an adjustment by the stream to external factors. To what extent allocyclicity applies to other ephemeral streams in the region is uncertain. It is unlikely that all streams in the diverse terrain of the arid and semiarid western United States are sensitive to changes in climate. Work by Baker (1977) suggests that the morphology and geology of the drainage basin determine the degree to which climate is capable of controlling fluvial processes. Some streams, such as the Little Colorado, thus may respond to even short-term climatic fluctuations, whereas others cannot because of basin size or other morphologic and geologic factors.

CLIMATE AND TWENTIETH-CENTURY STREAM PROCESSES ALONG THE LITTLE COLORADO RIVER

The recent changes in the morphology of the Little Colorado River channel were concurrent with variations in the discharge regimen of the river. It is reasonable, therefore, to expect that climate may have been a causative factor and to search for correlations between climatic history and the history of stream processes. Analysis of climatic records from the drainage basin suggests that variations in climate (as measured by mean annual precipitation, temperature, and tree ring–width indices) were occuring at about the same time as were the variations in annual discharge.

The precipitation and temperature data used in this analysis are from the weather stations at Flagstaff, Winslow, and Springerville (160 km southeast of Winslow), which have the longest continuous records in the Little Colorado River basin. The source of the data was the monthly tabulations in Green and Sellers (1964), Sellers and Hill (1974), and current U.S. Department of Commerce publications. Missing values are not a problem with these stations after 1915, with the exception of Springerville, which has an incomplete temperature record. The records were combined into a three-station average of total annual precipitation and a two-station (Flagstaff and Winslow) average of annual mean monthly

TABLE 2. FLOOD-PLAIN ALLUVIUM, FLOODS, AND LOCAL AND REGIONAL WEATHER

Alluvial unit	Flood	Weather
Older	1952 (January 21)	Unusually wet winter producing largest runoff on Arizona rivers since 1941. Annual discharge of the Colorado River among sixth largest since 1912
Intermediate	1969 or 1970 (September 7 or 11) 1971 (October 3)	Extensive flooding in 1970 in southwest from Pacific storm Norma (Roeske and others, 1978). 1969 to 1971 characterized by deficient precipitation in Arizona, New Mexico, and Texas (Bark, 1978)
	1972 (October 7) 1972 (October 19)	Wettest October in 92 yr at Flagstaff. Anomalous global weather patterns in 1972 and 1973 (Kukla and Kukla, 1974)
	1973 (April 18)	Heavy snowfall in winter of 1973. In the southwest, temperatures were abnormally low, snowfall especially abundant (American Meteorological Society Bulletin, 1973)
Younger	1978 (December 21)	Precipitation over northern Arizona began in November and continued into December. 1978 to 1980 characterized by wet winters and dry summers in northern Arizona. Precipitation in 1978 was heavy and floods were common over most of western United States (American Meteorological Society Bulletin, 1979)

TABLE 3. STATISTICS OF THREE PERIODS OF ANNUAL DISCHARGE, MEAN ANNUAL PRECIPITATION, AND MEAN ANNUAL TEMPERATURE COMPARED WITH STATISTICS OF ENTIRE SERIES

	Period	$\bar{x}$	95% confidence interval $\bar{x}$	Standard deviation
Discharge (m^3/s)	1927–1942	9.32	7.90 to 10.8	2.68
	1943–1961	5.28	4.75 to 5.81	1.10
	1962–1980	7.67	6.07 to 9.28	3.33
Entire series	1927–1980	7.32	6.50 to 8.14	2.99
Precipitation (mm)	1916–1942	348	335 to 361	32.5
	1943–1956	300	291 to 308	15.4
	1957–1980	346	334 to 359	29.5
Entire series	1916–1980	337	328 to 345	34.4
Temperature (°C)	1916–1932	9.83	9.61 to 10.1	0.436
	1933–1961	10.5	10.4 to 10.6	0.255
	1962–1980	10.0	9.88 to 10.2	0.306
Entire series	1016–1980	10.2	10.1 to 10.3	0.423

temperature from 1915 to 1981. The stations have been moved several times (Green and Sellers, 1964; Sellers and Hill, 1974), but analysis of the individual records did not reveal changes in the mean and variance that could be attributed to relocation. It is assumed that the precipitation and temperature time series derived from these stations reflect climatic patterns that affected the entire drainage basin, although spatial variability was not accounted for.

Time series of average ring-width indices determined from conifers at several widely spaced sites in the drainage basin also were examined. The data set is a seven-station average from 1800 to 1972 of the ring-width indices of stations 5, 6, 8, 9, 13, 14, and 25 in Dean and Robinson (1978). At semiarid sites such as these, ring width is directly related to precipitation and soil moisture and inversely related to temperature (Fritts, 1976).

The data were treated with a low-pass filter to remove high-frequency oscillations and to emphasize low-frequency variations. The filter consisted of a weighted 3-yr moving average with weights 0.25, 0.5, and 0.25. The filter removes oscillations of less than a three-year duration from the series without shifting the position of peaks and troughs. A three-year moving average is justified because conditions leading to changes in the channel occurred over several decades, as indicated by geomorphic (Table 2) and stratigraphic evidence (Fig. 13); consequently, the year-to-year variability is of little practical significance. In addition, stepwise multiple regression was performed on the filtered data with dis-

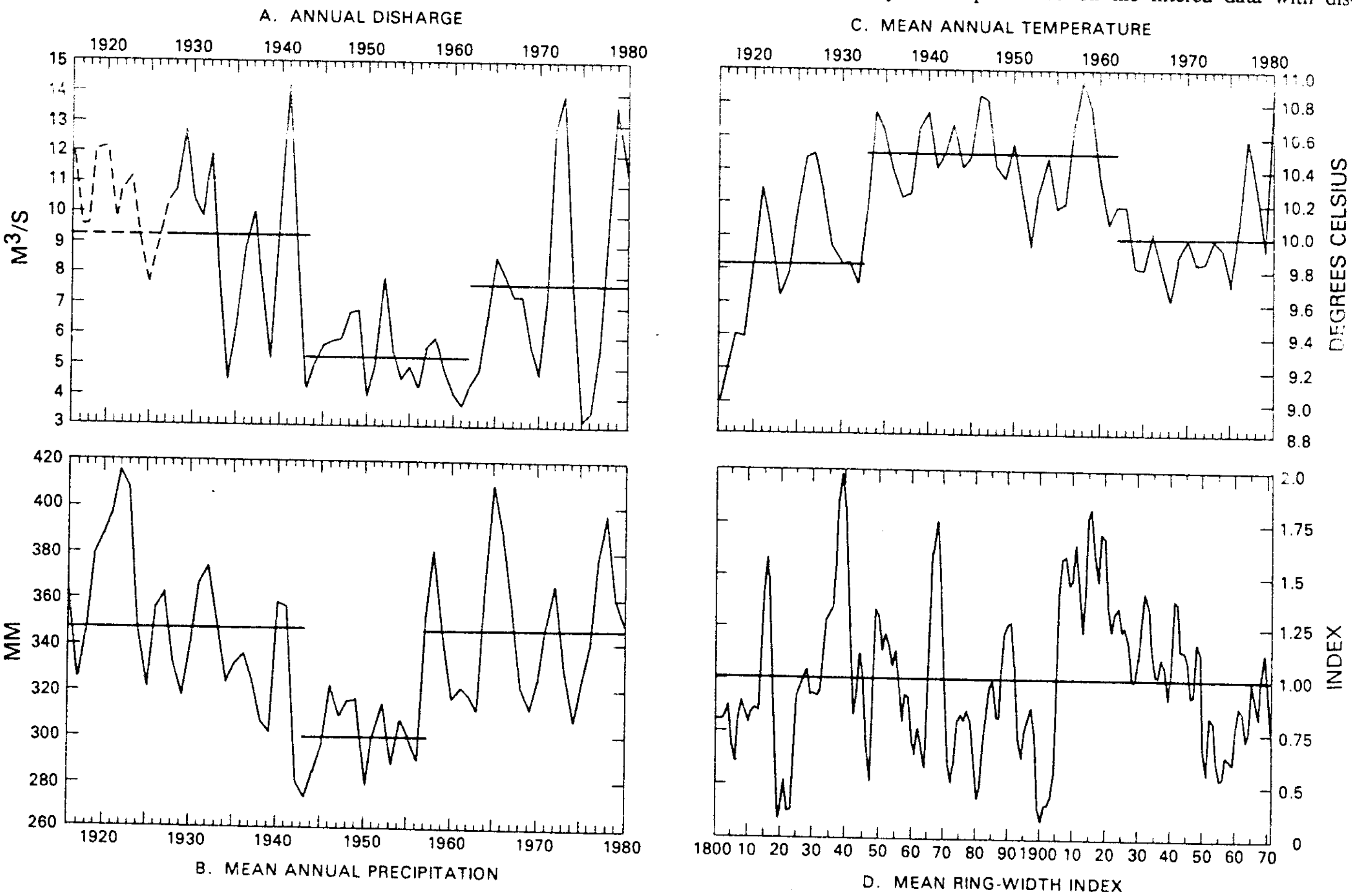

Figure 15. Time series of climatic variables in the Little Colorado River drainage basin. Filter (described in the text) results in loss of first and last years of data set. Horizontal lines in A, B, and C are the averages of the subseries in Table 3. A. Annual discharge, 1916 to 1980; dashed line was estimated from the precipitation and ring-width regression. B. Mean annual precipitation, 1916 to 1980. C. Mean annual temperature, 1916 to 1980. D. Mean ring-width indices, 1801 to 1971; horizontal line is the average.

charge as the dependent variable and precipitation, ring width, and temperature as the independent variables. All of the variables except temperature were significant in the regression, and 52% of the variability in discharge is explained by the precipitation and ring-width relationship. The regression was used to estimate discharge from 1916 to 1926.

One technique used to detect climatic variations is comparison of the over-all mean of the series with the means of several parts of the record (Mitchell and others, 1966; Potter, 1976). The discharge, precipitation, and temperature time series were divided into three subseries that isolated the central part of each record that appeared in time-series plots to have a different average. Table 3 compares the statistics of the three subseries with the statistics of the entire series on the assumption that the subseries are statistically independent. The 95% confidence interval for the average of the middle subseries does not overlap with the average for the entire series or with adjoining subseries, whereas the confidence intervals of the first and last segments overlap with each other and with the mean of the entire series.

This analysis indicates that the change in climate and discharge was caused by an abrupt shift in the central tendency of the generating process. Another explanation, which the analysis cannot determine, is that the change was gradual, indicating autocorrelation or interdependence of the process over a number of years. Historic time series such as these commonly show long-term interdependence that can be modeled either by a step function or by stochastic models that incorporate some form of pseudoperiodic behavior (Mandlebrot and Wallis, 1969; Potter, 1976). It is assumed that the step-function model adequately explains the shift in average values (Table 3), although the transitions may have been gradual.

The 1940s and 1950s constituted a climatically distinctive period of below-average discharge and precipitation and above-average temperature (Fig. 15; Table 5). The pattern of change in the ring-width indices (Fig. 15D) seems clear and statistical analysis is probably unnecessary. This pattern indicates that beginning in 1900 there was an abrupt rise in precipitation, and probably in discharge as well, that was unlike other fluctuations in the preceding 100 yr. From about 1910 until 1956, there was an intermittently broken but progressive decline in precipitation and soil moisture that reversed after 1956. Stockton (1975) found that the early 1900s was an unusually wet period in his reconstruction of Colorado River discharge using tree rings.

There seems to be an association between climatic history and stream processes along the Little Colorado River. The main stream processes modifying the geomorphology of the valley were (1) erosion resulting in incision of the cottonwood terrace and deposition of the older channel alluvium; (2) flood-plain development, that is, a reduction in channel width and the spread of riparian vegetation into the channel, beginning after 1937; and (3) flood-plain aggradation between 1952 and 1978. Erosion evidently was initiated by the increased precipitation and discharge of the early 1900s, and erosion persisted through most of an episode of intermittently declining moisture (Fig. 15D). Flood-plain development was approximately coincident with the decline in precipitation and discharge that began in 1942 (Figs. 2B, 15A, and 15B). Most of the aggradation of the flood-plain occurred after precipitation and discharge returned to nearly normal or average conditions in the late 1950s and early 1960s, which also coincided with a rise in soil moisture (Figs. 15A, 15B, and 15D).

The shift to less precipitation in 1942 and the variation in temperature (Figs. 15B, 15C) were not peculiar to the Little Colorado drainage basin. The period 1942–1956 was one of general, although intermittently broken, meteorologic and hydrologic drought that by 1956 encompassed the Rocky Mountain states from Arizona to Montana and the Great Plains region from South Dakota to Texas (Gatewood and others, 1963; Felch, 1978). This period was also noteworthy because of a reversal in the general warming trend that began in the Northern Hemisphere in the early part of this century (Kalnicky, 1974, Fig. 1). Since at least 1951, the middle and northern latitudes have undergone progressive oscillatory cooling (Kukla and others, 1977; Douglas and others, 1982). The causes of the recent global climatic changes are many and complex, but most of them probably are related to a shift in atmospheric circulation (Kalnicky, 1974) that changed surface temperature, precipitation patterns, and, ultimately, discharge. These global changes in climate are similar to those recorded in the Little Colorado basin (Table 3; Fig. 15), and their effect on the discharge of the river probably was responsible for the historic changes in the morphology of the river channel.

CONCLUSION

The geomorphic and stratigraphic history of the Little Colorado River in this century is recorded by three deposits. From oldest to youngest, these are the deposits beneath the cottonwood terrace, the older channel alluvium, and the flood-plain alluvium. The deposits forming the cottonwood terrace were incised sometime after 1880–1905, during a period of frequent large floods and high annual discharge. This discharge regimen persisted until the early 1940s and deposited the older channel alluvium. During the 1940s and 1950s, precipitation and discharge decreased for a number of years, and riparian vegetation, mainly nonnative salt cedar, spread into the higher, frequently dry parts of the channel. The flood plain began to aggrade at about this time, whenever large floods caused overbank deposition.

Climatic fluctuations probably caused most of the changes in channel morphology. The early erosive phase was associated with changes in land use but also occurred during a period of warming temperature affecting most of the Northern Hemisphere. Ring-width indices suggest that the erosive phase was initiated by an abrupt rise in moisture beginning in 1900. Flood-plain development probably began in 1942 and occurred during a climatically distinctive period of reduced precipitation, high temperature, and low discharge. On the other hand, aggradation of the flood plain was associated with a return to nearly average or normal conditions of precipitation and discharge.

The stratigraphy of the flood-plain alluvium indicates that it was formed by at least seven depositional events. Deposition occurred in 1952, 1969 or 1970–1973, and 1978. Sedimentation was caused by the largest floods in the period 1952–1978. Until the flood of 1980, the flood plain evolved by vertical accretion, but beginning in 1980 sedimentation shifted to lateral accretion; at present, the flood plain is being eroded. Although continued observation is needed to confirm this, it appears that the flood plain has become a terrace.

This recent change in geomorphic process from flood-plain aggradation to erosion, and in sedimentary style from vertical to lateral accretion, occurred without a fluctuation in discharge. It is likely that the flood plain has reached a critical height, beyond which further accretion is unlikely unless the channel aggrades or the height of future floods changes. The conclusion is that flood-plain development is initiated and sustained through hydrologic fluctuations, but that continued aggradation beyond a critical height is limited by geomorphic processes unrelated to the hydrology of the river.

The stratigraphic column of the Little Colorado, and perhaps of other ephemeral streams in the southwest United States as well, may record several episodes of flood-plain deposition within an interval of several centuries. Such deposits probably accumulated during periods of adjustment to new hydrologic conditions. The stratigraphic succession of flood-plain deposits from ephemeral streams such as the Little Colorado may represent repeated adjustments by the stream to changes in hydrology.

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REFERENCES CITED

Allen, J.R.L., 1965, A review of the origin and characteristics of recent alluvial sediments: Sedimentology, v. 5, p. 89–191.
——— 1970, Physical processes of sedimentation: New York, American Elsevier Company, Inc., 248 p.
American Meteorological Society Bulletin, 1973, NSSFC winter wrap-up '72–73: v. 54, p. 713–714.
——— 1977, Basin-by-basin summer drought outlook: v. 58, p. 534–535.
——— 1979, 1978 flooding casualties and damage: v. 60, p. 480–481.
Baker, V. R., 1977, Stream-channel response to floods, with examples from central Texas: Geological Society of America Bulletin, v. 88, p. 1057–1071.
Baker, V. R., and Penteado-Orellano, M. M., 1977, Adjustment to Quaternary climatic change by the Colorado River in central Texas: Journal of Geology, v. 85, p. 395–422.
Bark, L. D., 1978, History of American droughts, *in* Rosenberg, N. J., ed., North American droughts: American Association for the Advancement of Science Selected Symposium 15, p. 9–23.
Bigelow, J. M., 1856, General description of the botanical character of the country, *in* Whipple, A. A., Report of explorations and surveys to ascertain the practicable and economical route for a railroad from the Mississippi River to the Pacific Ocean, v. 4, part 5, Report on the botany of the expedition: U.S. 33rd Congress, 2nd session Senate Executive Document 78, p. 1–16.
Bryson, R. A., 1974, A perspective on climatic change: Science, v. 184, p. 753–760.
Bull, W. B., 1979, Threshold of critical power in streams: Geological Society of America Bulletin, v. 90, p. 453–464.
Christensen, E. M., 1962, The rate of naturalization of *Tamarix* in Utah: American Midland Naturalist, v. 68, p. 51–57.
Colton, H. S., 1937, Some notes on the original condition of the Little Colorado River: A side light on the problems of erosion: Museum of Northern Arizona Museum Notes, v. 10, p. 17–20.
Cooke, R. V., and Reeves, R. W., 1976, Arroyos and environmental change in the American Southwest: Oxford, England, Oxford University Press, 213 p.
Davey, J. R., 1980, Energy flow dynamics of a desert grassland ecosystem [Ph.D dissert.]: Flagstaff, Arizona, Northern Arizona University, 147 p.
Dean, J. S., and Robinson, W. J., 1978, Expanded tree-ring chronology for the southwestern United States: University of Arizona Laboratory of Tree-Ring Research Chronology Series 3, 58 p.
Douglas, A. V., Cayan, D. R., and Namias, Jerome, 1982, Large-scale changes in North Pacific and North American weather patterns in recent decades: Monthly Weather Review, v. 110, p. 1851–1862.
Emmett, W. W., 1974, Channel aggradation in western United States as indicated by observations at Vigil Network sites: Zeitschrift für Geomorphologie, Supplement Band 21, v. 2, p. 52–62.
Euler, R. C., Gummerman, G. J., Karlstrom, T.N.V., Dean, J. S., and Hevly, R. H., 1979, The Colorado plateaus: Cultural dynamics and paleoenvironment: Science, v. 205, p. 1089–1101.
Everitt, B. L., 1968, Use of the cottonwood in an investigation of the recent history of a flood plain: American Journal of Science, v. 266, p. 417–439.
——— 1979, Fluvial adjustments to the spread of tamarisk in the Colorado plateau region, discussion: Geological Society of America Bulletin, Part I, v. 90, p. 1183.
——— 1980, Ecology of saltcedar—A plea for research: Environmental Geology, v. 3, p. 77–84.
Felch, R. E., 1978, Drought: Characteristics and assessment, *in* Rosenberg, N. J., ed., North American droughts: American Association for the Advancement of Science Selected Symposium 15, p. 25–42.
Fischer, G. A., 1947, Little Colorado River, Arizona, from Tolchico Dam site to Lyman Reservoir: U.S. Department of the Interior Geological Survey Sheet 1-2, scale 1:31,680.
Fritts, H. C., 1976, Tree rings and climate: London, Academic Press, 567 p.
Gatewood, J. S., Wilson, Alfonso, Thomas, H. E., and Kister, L. R., 1963, General effects of drought on water resources of the Southwest 1942–56: U.S. Geological Survey Professional Paper 372-B, p. B1–B55.
Gifford, G. F., and Hawkins, R. H., 1978, Hydrologic impact of grazing on infiltration: A critical review: Water Resources Research, v. 14, p. 305–313.
Graf, W. L., 1978, Fluvial adjustments to the spread of tamarisk in the Colorado Plateau region: Geological Society of America Bulletin, v. 89, p. 1491–1501.
——— 1979, Fluvial adjustments to the spread of tamarisk in the Colorado Plateau region, reply: Geological Society of America Bulletin, v. 90, p. 1183–1184.
——— 1982, Spatial variation of fluvial processes in semi-arid lands, *in* Thorn, C. E., ed., Space and time in geomorphology: London, George Allen and Unwin, p. 193–217.
——— 1983, The arroyo problem—Paleohydrology and paleohydraulics in the short term, *in* Gregory, K. J., ed., [illegible] paleohydrology: London, John Wiley and Sons, p. 279–302.
Green, C. R., and Sellers, W. D., 1964, Arizona climate: Tucson, Arizona, University of Arizona Press, 503 p.
Gregory, H. E., 1917, Geology of the Navajo country: U.S. Geological Survey Professional Paper 93, 161 p.
Hadley, R. F., 1961, Influence of riparian vegetation on channel shape, northeastern Arizona: U.S. Geological Survey Professional Paper 424-C, p. C30–C31.
Harms, J. C., and Fahnestock, R. K., 1965, Stratification, bedforms and flow phenomena (with examples from the Rio Grande), *in* Primary sedimentary structures and their hydrodynamic interpretation: Society of Economic Paleontologists and Mineralogists Special Publication 12, p. 84–115.
Hereford, Richard, 1979, Preliminary geologic map of the Little Colorado River valley between Cameron and Winslow, Arizona: U.S. Geological Survey Open-File Report 79-1574.
Kalnicky, R. A., 1974, Climatic change since 1950: Annals of the Association of American Geographers, v. 64, p. 100–112.
Kukla, G. J., and Kukla, H. J., 1974, Increased surface albedo in the Northern Hemisphere: Science, v. 183, p. 709–714.
Kukla, G. J., and others, 1977, New data on climatic trends: Nature, v. 270, p. 573–580.
Leopold, L. B., 1951, Rainfall frequency: An aspect of climatic variation: American Geophysical Union Transactions, v. 32, p. 347–357.
——— 1976, Reversal of erosion cycle and climatic change: Quaternary Research, v. 6, p. 557–562.
Leopold, L. B., Wolman, M. G., and Miller, J. P., 1964, Fluvial processes in geomorphology: San Francisco, W. H. Freeman and Co., 522 p.
Lewin, John, 1978, Floodplain geomorphology: Progress in Physical Geography, v. 2, p. 408–437.
Lewis, J. L., 1948, Plan and profile of Little Colorado River from mouth to Tolchico Dam Site, Arizona: U.S. Department of the Interior Geological Survey Sheets A–E, scale 1:31,680.
Lockett, H. C., and Snow, Milton, 1939, Along the Beale Trail—A photographic account of wasted range land based on the diary of Lieutenant Edward F. Beale, 1857: Washington, D.C., U.S. Office of Indian Affairs, 56 p.
Love, D. W., 1979, Quaternary fluvial geomorphic adjustments in Chaco Canyon, New Mexico, *in* Rhodes, D. D., and Williams, G. P., eds., Adjustments of the fluvial system: Dubuque, Iowa, Kendall Hall, p. 277–308.
Mandelbrot, B. B., and Wallis, J. R., 1969, Some long-run properties of geophysical records: Water Resources Research, v. 5, p. 321–340.
Matthai, H. F., 1979, Hydrologic and human aspects of the 1976–77 drought: U.S. Geological Survey Professional Paper 1130, 84 p.
McDowell, P. F., 1983, Evidence of stream response to Holocene climatic change in a small Wisconsin watershed: Quaternary Research, v. 19, p. 100–116.
McKee, E. D., 1966, Significance of climbing-ripple structure: U.S. Geological Survey Professional Paper 550-D, p. D94–D103.
Miall, A. D., 1980, Cyclicity and the facies model concept in fluvial deposits: Bulletin of Canadian Petroleum Geology, v. 28, p. 59–80.
Mitchell, J. M., Jr., Dzerdzeevskii, H., Flohn, H., Hofmeyer, W. L., Lamb, H. H., Rao, K. N., and Wallen, C. C., 1966, Climatic change: Geneva, World Meteorological Organization Technical Note 79, 79 p.
Patton, P. C., and Schumm, S. A., 1981, Ephemeral-stream processes: Implications for studies of Quaternary valley fills: Quaternary Research, v. 15, p. 24–43.
Picard, M. D., and High, L. R., Jr., 1973, Sedimentary structures of ephemeral streams: New York, Elsevier Scientific Publishing Company, 223 p.
Potter, K. W., 1976, Evidence for nonstationarity as a physical explanation of the Hurst phenomenon: Water Resources Research, v. 12, p. 1047–1052.
Ray, P. K., 1976, Structure and sedimentologic history of the overbank deposits of a Mississippi River point bar: Journal of Sedimentary Petrology, v. 46, p. 788–801.
Reno, Philip, 1981, Navajo resources and economic development: Albuquerque, New Mexico, University of New Mexico Press, 183 p.
Robinson, T. W., 1965, Introduction, spread and areal extent of saltcedar *(Tamarix)* in the western States: U.S. Geological Survey Professional Paper 491-A, p. A1–A12.
Roeske, R. H., Cooley, M. E., and Aldrige, B. N., 1978, Floods of September 1970 in Arizona, Utah, Colorado, and New Mexico: U.S. Geological Survey Water-Supply Paper 2052, 135 p.
Schumm, S. A., 1977, The fluvial system: New York, John Wiley and Sons, 338 p.
Schumm, S. A., and Lichty, R. W., 1963, Channel widening and floodplain construction along Cimarron River in southwestern Kansas: U.S. Geological Survey Professional Paper 352-D, p. 71–88.
Sellers, W. D., and Hill, R. H., 1974, Arizona climate 1931–1972: Tucson, Arizona, University of Arizona Press, 616 p.
Sigafoos, R. S., 1964, Botanical evidence of floods and flood-plain deposition: U.S. Geological Survey Professional Paper 485-A, p. A1–A35.
Stevens, M. A., Simmons, D. B., and Richardson, E. V., 1975, Nonequilibrium river form: American Society of Civil Engineers, Proceedings, Journal of the Hydraulics Division, HY 101, no. [illegible]
Stockton, C. W., 1975, Long-term streamflow records reconstructed from tree rings—Papers of the Laboratory of Tree-Ring Research no. 5: Tucson, Arizona, University of Arizona Press, [illegible] p.
Stokes, M. A., and Smiley, T. L., 1968, An introduction to tree-ring dating: Chicago, Illinois, University of Chicago Press, 73 p.
Strahorn, A. T., Baldwin, Mark, and Carpenter, E. J., 1924, Soil survey of the Winslow area, Arizona: U.S. Bureau of Soils Field Operations for 1921, p. 155–188.
Thomas, H. E., 1963, General summary of effects of the drought in the Southwest: U.S. Geological Survey Professional Paper 373-H, p. H1–H22.
Thornthwaite, C. W., Sharpe, C.T.S., and Dosch, E. F., 1942, Climate and accelerated erosion in the arid and semi-arid southwest with special reference to the Polacca Wash drainage basin, Arizona: U.S. Department of Agriculture Technical Bulletin 808, 134 p.
Turner, R. M., 1974, Quantitative and historical evidence of vegetation changes along the upper Gila River, Arizona: U.S. Geological Survey Professional Paper 655-H, p. H1–H20.
Turner, R. M., and Karpiscak, M. M., 1980, Recent vegetation changes along the Colorado River between Glen Canyon Dam and Lake Mead, Arizona: U.S. Geological Survey Professional Paper 1132, 125 p.
Underwood, J. R., Jr., and Lambert, Wayne, 1974, Centriclinal cross strata, a distinctive sedimentary structure: Journal of Sedimentary Petrology, v. 44, p. 1111–1113.
U.S. Department of Agriculture, 1981, Summary report, Little Colorado River basin, Arizona–New Mexico: Washington, D.C., U.S. Government Printing Office, 41 p.
Williams, G. P., 1978, The case of the shrinking channels—The North Platte and Platte Rivers in Nebraska: U.S. Geological Survey Circular 781, 48 p.
Wolman, M. G., 1971, Evaluating alternative techniques of floodplain mapping: Water Resources Research, v. 7, p. 1383–1392.
Womack, W. R., and Schumm, S. A., 1977, Terraces of Douglas Creek, northwestern Colorado: An example of episodic erosion: Geology, v. 5, p. 72–76.

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