Geomorphic response to seasonal variations in rainfall in the Southwest United States

Devin Etheredge[†]
David S. Gutzler[‡]
Department of Earth and Planetary Sciences, University of New Mexico, Albuquerque, New Mexico 87131, USA
Frank J. Pazzaglia
Department of Earth and Environmental Sciences, Lehigh University, 31 Williams, Bethlehem, Pennsylvania 18015, USA

ABSTRACT

The interaction of the North American Monsoon with watershed hydrology and landscape response is evaluated by observing geomorphic characteristics of hillslopes, hydrology, and stream channels in two mountain ranges with contrasting intensity of precipitation. The study compares watersheds in the Hualapai and Santa Catalina Mountains in Arizona, which are similar in lithology, elevation, tectonic setting, vegetation, and annual precipitation, but differ in the proportion of precipitation received in summer thunderstorms. The Hualapai Mountains receive most of their precipitation in winter, whereas rainfall in the Santa Catalina Mountains occurs mostly in summer. Drainages in the Santa Catalinas are more variable in local relief and exhibit much more exposed bedrock and higher drainage density. The trunk channel of a major drainage in the Santa Catalina Mountains has measured discharges several orders of magnitude greater than a channel draining a comparable area in the Hualapai Mountains. The alluvial segment of the channel in the Santa Catalina piedmont additionally displays greater concavity, smaller width-to-depth ratios, and a larger caliber of bed material. In contrast to conventional interpretation, summer monsoonal precipitation is not primarily responsible for generating most of the discharges that modify channels in large-scale drainages. In the monsoonal climate regime of the Santa Catalinas, small basins flood most often in summer, whereas larger drainages exhibit peak discharges in response to low-intensity winter precipitation. We attribute the paradoxical discharge response of larger drainages to the small spatial scale of summer thunderstorms, which fail to deliver enough precipitation to generate floods in these larger basins, but do prime hillslopes by stripping colluvium and lowering hillslope-infiltration rates, making the large drainages more responsive to areally extensive winter storms.

Keywords: monsoons, precipitation, Southwest United States, seasonal variations, hydrology, landscape evolution.

INTRODUCTION

Geomorphic studies of arid and semiarid environments have placed considerable emphasis on the role of climate in landscape evolution (Gerson and Yair, 1975; Budel, 1982; Wells et al., 1987; Bull, 1991). In the Southwest United States (Arizona and New Mexico), the relationship between temporal variation in the frequency and intensity of rainfall over the region and the cycles of erosion and deposition that can be observed in Quaternary landforms is commonly a subject of speculation. Seasonal, high-intensity precipitation from summer thunderstorms associated with the North American Monsoon is traditionally cited as the principal agent in accelerating landscape denudation via increased rates of sediment erosion and transport (Leopold, 1951; Faulkner, 1986; Hereford and Webb, 1992; Hereford, 1993; Reneau et al., 1996). The assumption that high-intensity rainfall generates flashy, high-volume discharge from hillslopes and streams and thus increases rates of sediment erosion and transport rests on reasonably sound principles of physical hydrology and sedimentology, but field studies focused on the observed response of the landscape to forcing by different intensities of rainfall remain conspicuously rare.

The objective of this study is to provide a firmer observational foundation for arid-region climate and landscape interactions through the intensive study of a limited area. We evaluate the influence of seasonal variation in rainfall intensity on two watersheds by using a comparative description of basic landscape metrics and processes in locales that differ in the frequency of intense rainfall that they receive. Our study is pertinent to the accepted models of regional Quaternary geologic history; it may also provide some insight into future environmental changes that can be expected in the Southwest if predictions of global climate alteration prove to be correct. The general strategy of this study is to compare observations of hydrologic, hillslope, and fluvial metrics of the landscape system in a strongly monsoonal watershed and a semi-monsoonal setting in different parts of the Southwest United States.

SETTING AND METHODS

In this comparative study we have been cautious to ensure that rainfall rate is properly isolated from the major independent variables affecting landscape evolution: total annual precipitation, rainfall duration, rock type, elevation, relief, tectonics, and anthropogenic effects. The sites selected for intensive study are the Hualapai Mountains south of Kingman, Arizona, and the Santa Catalina Mountains north of Tucson, Arizona (Fig. 1). Both of these ranges are primarily the result of middle to late Tertiary extensional tectonics. Movement along the normal faults bounding

[†]E-mail: euler-pole@msn.com.
[‡]E-mail: gutzler@unm.edu.

GSA Bulletin; May/June 2004; v. 116; no. 5/6; p. 606–618; doi: 10.1130/B22103.1; 14 figures; 2 tables.

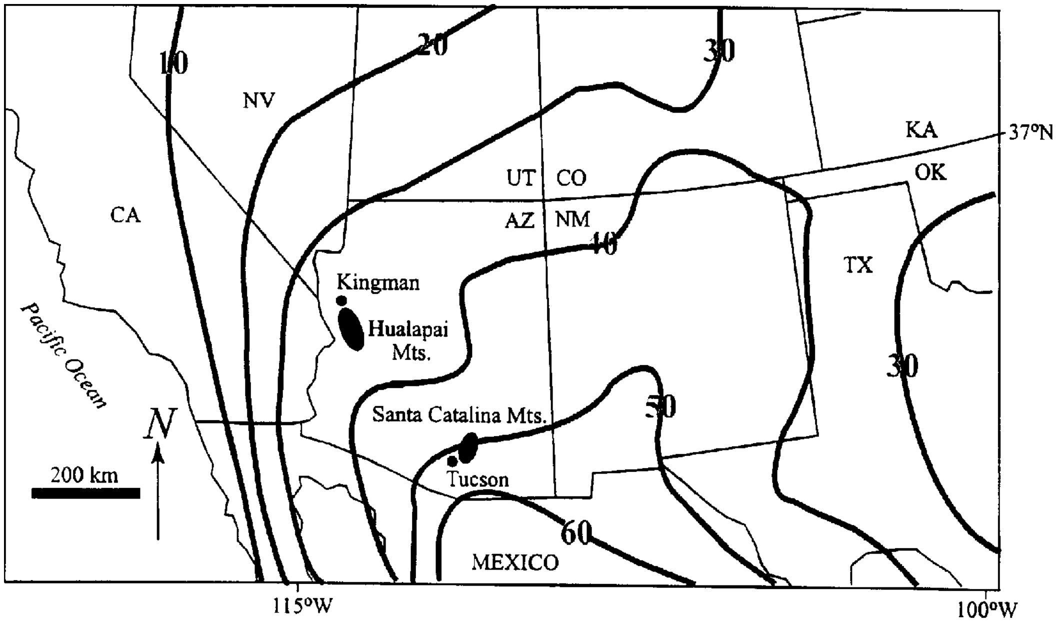

Figure 1. The percentage contribution of the North American Monsoon (months of July, August, and September) to the annual total precipitation. The locations of the Santa Catalina and Hualapai Mountains are also indicated. Adapted from Douglas et al. (1993).

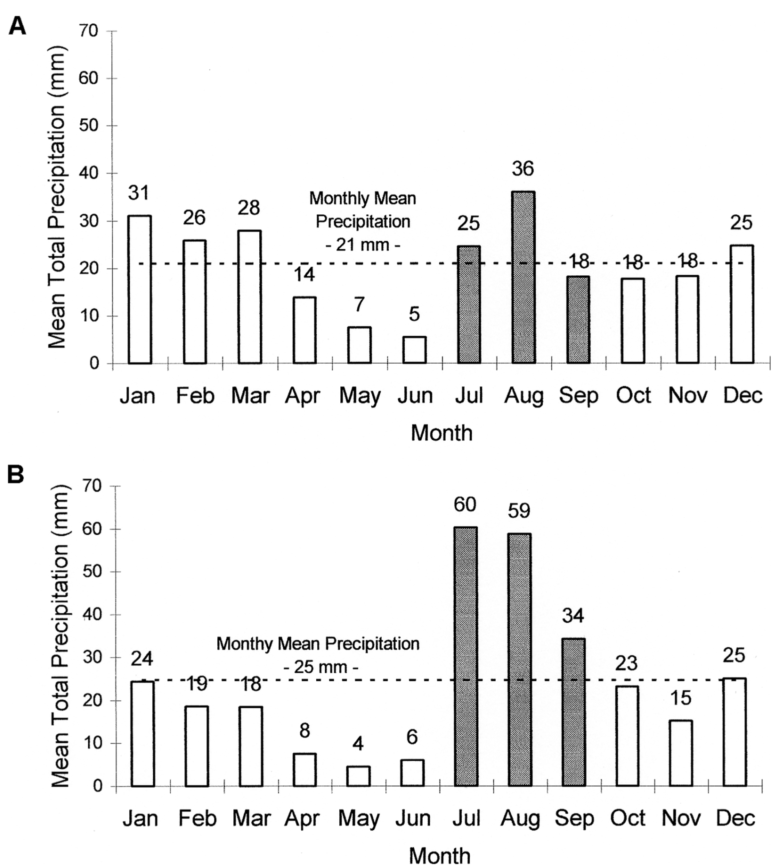

Figure 2. Mean monthly precipitation (1948–1993) at the study locations, (A) Kingman and (B) Tucson International Airport. The monsoon months of July, August, and September are shaded.

these ranges is thought to have ceased by the early Pliocene (Menges and Pearthree, 1989). The Hualapai Mountains are composed mostly of Precambrian granites and gneiss as well as small areas of exposed quartzite and schist (Eldred and Moore, 1959). The lithology of the Santa Catalina Mountains is composed almost entirely of Tertiary gneiss and granite (Eldred et al., 1960).

The Hualapai Mountains cover an area of ~2000 km² and rise to their maximum elevation at Hualapai Peak, 2496 m above sea level. Mean elevation for the entire range is 1258 m, and total relief is ~1950 m. The Santa Catalina Mountains are a smaller range, 660 km² in area. Mount Lemmon, the highest peak in this range, has an elevation of 2743 m. The mean elevation of the Santa Catalina Mountains is 1552 m, and total relief is ~1975 m.

Mean annual precipitation in Kingman, 19 km north of the Hualapai Mountains study area, is 250 mm; ~30% of that total falls during the summer monsoon months of July, August, and September (Sellers and Hill, 1974). The mean annual precipitation at Tucson International Airport, 25 km south of the Santa Catalina mountain front, is <300 mm, >50% of which arrives with the monsoon (Fig. 1). Below 1200–1500 m elevation the vegetation at both study sites can generally be described as Upper Sonoran type (Shreve and Wiggins, 1964). White bur sage (*Ambrosia dumosa*) and creosote bush (*Larrea tridentate*) dominate the basin floor, whereas saguaro (*Cereus giganteus*), ocotillo (*Fouquieria splendens*), and several varieties of cholla (*Opuntia*) are more common on hillslopes. Between 1500 and 2000 m, oak and juniper woodland becomes increasingly dense. Forests of pine, spruce, and fir dominate elevations above 2000 m. Both mountain ranges are sparsely populated and have undergone relatively minimal human impact.

Hourly precipitation recordings, obtained from the National Oceanic and Atmospheric Administration (NOAA) National Climatic Data Center in Asheville, North Carolina, were used to characterize rainfall intensities at the study sites (Table 1). Care was taken to use data from stations that were located in similar topographic settings in order to minimize variation caused by local orographic effects on precipitation. The recording stations at Tucson International Airport and Kingman are the sole providers of hourly precipitation data in the vicinity of the mountain study areas. Daily precipitation records from Sabino Canyon and Palisade Ranger Station in the Santa Catalina Mountains were used to provide additional information on the variation in total precipitation with elevation.

Records used in the statistical analysis of the data were taken from the period between October 1948 and September 1993. This period ensured that the Tucson and Kingman data sets overlapped temporally while at the same time providing a long (46 yr) and complete record for study. Precipitation data were separated by month to isolate the seasonal cycle of rainfall.

Characterization of the hydrologic response of the landscape to monsoonal climate forcing comes primarily from analysis of stream-flow statistics, in the form of peak annual discharge data obtained from stream-flow gages in both the Hualapai and Santa Catalina mountain ranges. The U.S. Geological Survey (USGS) maintains a series of gages in these ranges as part of the National Water Information Service (NWIS; Table 2). The stream gages report the calendar date and maximum flow rate of all annual discharge peaks that occurred during the interval of time that the gage was in operation. All of the streams studied are ephemeral, being dry channels at least two months of the year. Likewise, all of these streams flow south or southwest from their respective ranges, and all are underlain by a gneiss/granite watershed (Eldred and Moore, 1959; Eldred et al., 1960).

The comparative part of the peak-discharge analysis focuses on Walnut Creek in the Hualapai Mountains and Bear Creek in the Santa

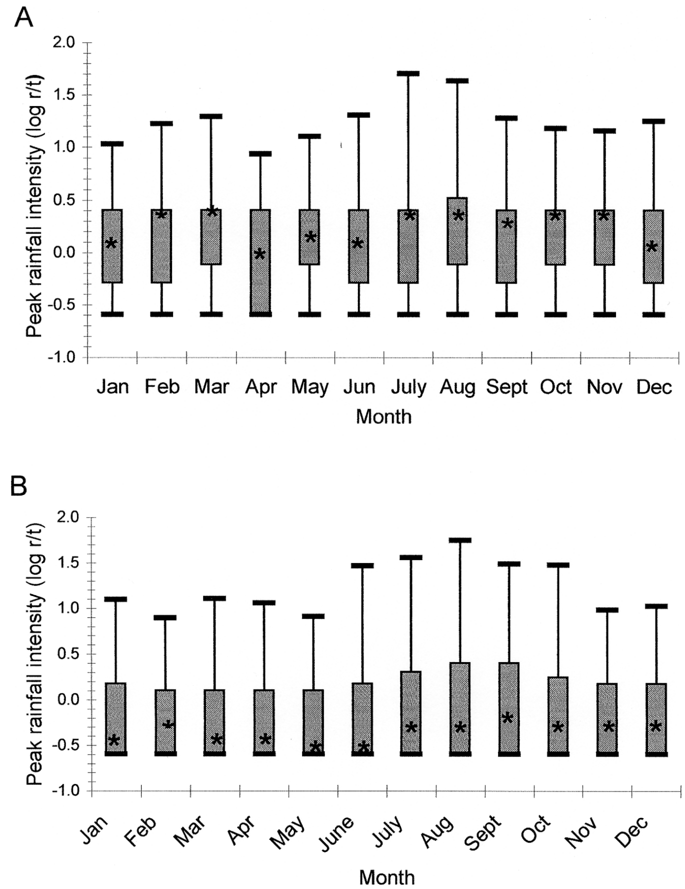

Figure 3. Whisker plot of peak rainfall intensity (log *r/t*, where *r* is rainfall in millimeters and *t* is time in hours) recorded in storms at (A) Kingman and (B) Tucson International Airport for the period 1948–1993. Stars indicate the median intensity for each month. The range of the middle 50% is shown by the shaded boxes, and the maximum and minimum intensities in the period of record are represented by the bars. Minimum intensity recorded is 0.01 in./h, which is equivalent to log $r/t = -0.6$, where r/t is in millimeters per hour.

TABLE 1. PRECIPITATION RECORDING STATIONS USED IN THIS STUDY

Station name	Co-op number	Location	Elevation (m)	Duration of record†
Tucson International Airport	028820	32°8′N, 110°57′W	777	7/1/48–12/24/97
Sabino Canyon	027355	32°30′N, 110°49′W	804	1/1/61–12/31/90
Palisade Ranger Station	026202	32°25′N, 110°43′W	2425	1/1/65–10/1/81
Kingman	024639	35°11′N, 114°3′W	1025	7/1/48–8/22/67
Kingman no. 2	024645	35°12′N, 114°1′W	1079	8/22/67–10/31/93

†The date format is month/day/year.

TABLE 2. NWIS STREAM-FLOW RECORDING STATIONS

Drainage	Station number	Location	Basin size (km^2)	Duration of record
Craycroft Wash	09484530	32°18′05″N, 110°52′13″W	0.1	1982–1990
Agua Caliente Wash	09483200	32°16′07″N, 110°44′15″W	5.3	1965–1980
Geronimo Wash	09485950	32°19′56″N, 110°56′37″W	5.6	1964–1981
Pima Wash	09485900	32°20′15″N, 110°57′35″W	12.8	1964–1984
Ventana Wash	09484510	32°15′35″N, 110°50′20″W	16.7	1965–1981
Bear Creek	09484200	32°18′35″N, 110°48′03″W	42.2	1959–1978
Sabino Creek	09484000	32°19′01″N, 110°48′36″W	92.0	1948–1996
Canada Del Oro	09486100	32°33′54″N, 110°50′48″W	109.6	1984–1991
Walnut Creek	09423780	35°02′00″N, 114°01′05″W	81.1	1965–1976

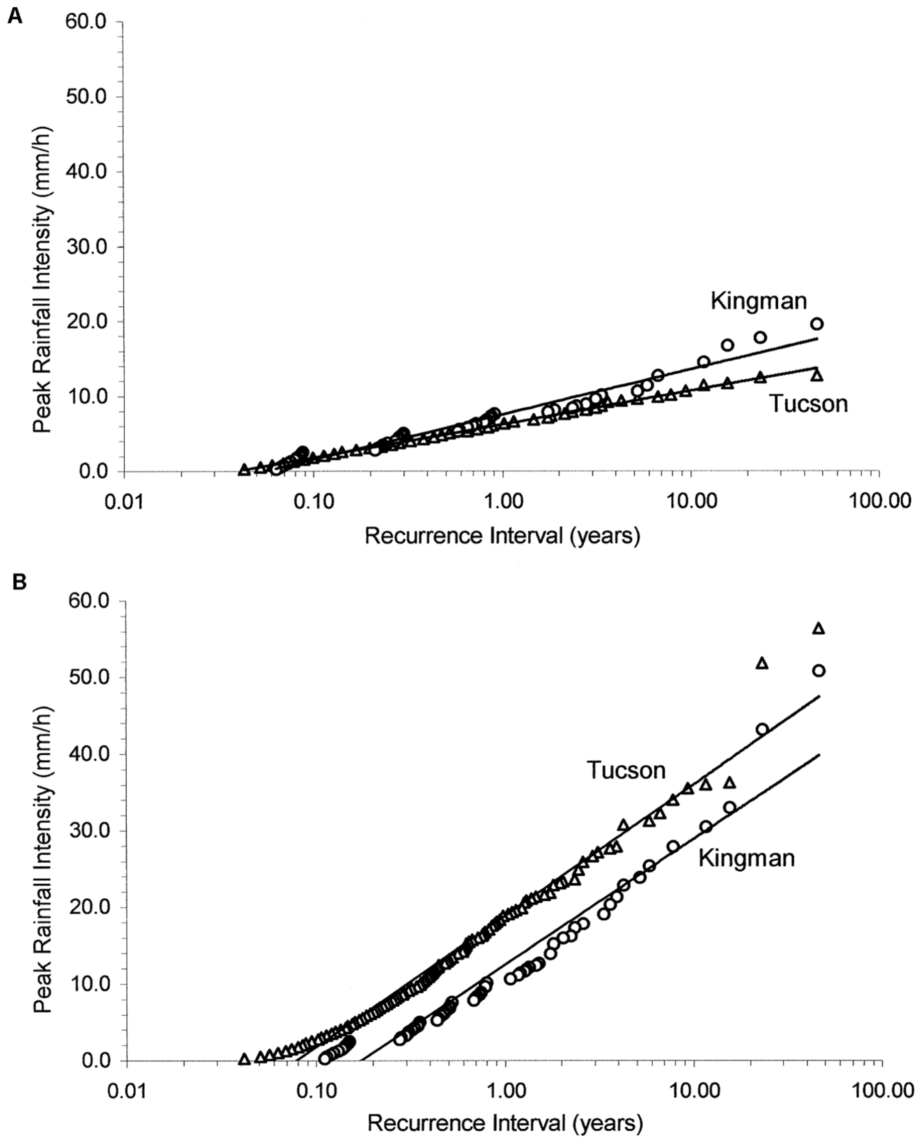

Figure 4. Frequency and magnitude of storms based on partial-duration series of peak rainfall intensities recorded from 1948 to 1993. (A) Recurrence interval of winter mid-latitude cyclones (November–May). (B) Recurrence interval of summer monsoon storms (July–September). Triangles are data from Tucson International Airport; circles are data from Kingman.

Catalina Mountains. The data sets from these streams overlap for the period from 1965 to 1976, providing a decade-long record of discharge for comparison. The records of peak-discharge rates for these streams were analyzed as a partial-duration series, which allowed their flood-magnitude and flood-frequency behavior to be contrasted.

Spatial studies of extensive properties such as local relief and drainage density were conducted by utilizing a geographic information system (ARC/INFO GIS). Raw data for these analyses were obtained in the form of 1:250,000 digital elevation models (DEMs) from the USGS. The 1:250,000 DEMs are based on square-raster cells ~90 m on a side. Finer resolution is possible with 1:24,000 DEMs, but coverage at this scale was unavailable for a large part of the Hualapai Mountains so the coarser 1:250,000 scale data were used.

Aerial photographs obtained from the National Aerial Photography Program, sponsored by the USGS, were utilized to inventory the extent of exposed bedrock. These black and white photographs were taken at the scale of 1:40,000, which provides enough resolution to map bedrock exposures as small as 20–30 m across. Outcrops of bedrock were identified from the photographs, verified by field scouting, and then digitized into ArcView GIS for analysis.

Field collection of data was targeted at both hillslope studies of bedrock exposures and the survey of alluvial-stream channels. Rock slopes were examined in order to determine their rock-mass strength and to evaluate any potential difference in the erodibility of Precambrian (Hualapai Mountains) and Tertiary (Santa Catalina Mountains) intrusive rocks. Field measurements of rock elasticity with a Schmidt hammer as well as joint spacing, width, and orientation recordings were used to calculate a numerical value for rock-mass strength according to the method of Selby (1980).

Field surveys were conducted on the piedmont alluvial segments of Walnut and Bear Creeks. The stream channels in this setting are incised into predominantly coarse-grained alluvial-fan and/or terrace deposits. Starting immediately downstream of the stream gages, the channel's hydraulic geometry parameters (width, depth, gradient) were recorded at 90–100 m intervals until 1 km of the channel had been surveyed for each stream. At each station, a topographic profile of the channel was drawn, and its gradient was measured by using a handheld survey level and stadia rod. Both channels were dry when surveyed, so an estimate was made of the depth of the most recent, bankfull-representative flow on the basis of vegetation lines and flood debris.

At each station the caliber of the bedload sediment was sampled. The difficulty of accurately sampling sediments in coarse bedload channels is well documented (Mosley and Tindale, 1985; Bathurst, 1987; Church et al., 1987). For this reason it was decided to utilize a sampling method similar to that of Baker (1973) and Costa (1983), in which five of the largest exposed clasts in the channel bed were measured at each stop. The mean measurement for the intermediate axis was then used to constrain the maximum caliber (D_{max}) of sediment that the stream is capable of transporting.

PRECIPITATION AT THE STUDY SITES

The North American Monsoon is a distinctive climatic feature of the American Southwest. It has probably existed more or less in its present manifestation but with variable strength since the beginning of the Holocene (Anderson et al., 2000). This system is alternatively referred to as the Arizona (Smith and

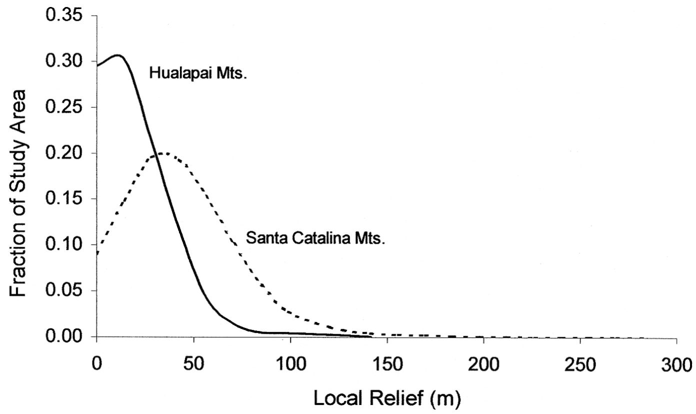

Figure 5. Distribution of local relief across the study areas, calculated for a circular window of ~1 km diameter moved across a 90-m-resolution DEM. The Santa Catalina Mountains (dashed line) are greater in both mean local relief and in variability of relief compared to the Hualapai Mountains (solid line).

Gall, 1989), Mexican (Douglas et al., 1993), or North American (Adams and Comrie, 1997; Higgins et al., 1997) Monsoon. Its northern extension into the United States takes place approximately from July to September, when heating across the high topography of the Rocky Mountains and Colorado Plateau destabilizes the atmosphere. A stationary low-pressure trough develops over the Rocky Mountain–Colorado Plateau region, supporting a southerly lower tropospheric flow that disrupts the northwesterly middle to lower tropospheric springtime airflow over the Southwest. Moisture is advected into the region in a near-surface airflow from the Gulf of California (Hales, 1972) and at higher levels from the Gulf of Mexico (Schmitz and Mullen, 1996). The combination of high humidity and surface temperature leads to convective instability resulting in thunderstorms.

The spatial extent and amplitude of the monsoon can be defined by the percentage of total annual rainfall that locations receive in the months of July, August, and September or JAS (Fig. 1; Douglas et al., 1993). In Arizona this percentage ranges from ~60% along its southeastern border with Mexico to 30% in the northwestern corner of the state. Monsoon onset can be defined in terms of abrupt increases in either rainfall (Higgins et al., 1997) or humidity (Adams and Comrie, 1997); either way, monsoon onset across the Southwest United States occurs on average early in July. Early monsoon onsets have been observed to occur in June in some years (Higgins et al., 1997), but for simplicity we here consider the three months of July, August, and September (JAS) to define the monsoon season in Arizona.

Tucson is situated in the region of the United States where precipitation is most predominantly monsoonal. During the period from 1948 to 1993, Tucson International Airport received a mean annual total of 297 mm of precipitation, of which 154 mm (52%) fell in JAS. Kingman, by contrast, is located near the edge of the Sonoran and Mojave Deserts where the monsoon's influence is much more marginal. Kingman receives a mean annual precipitation total of 251 mm, ~15% less than KTUS. The mean rainfall total for JAS at Kingman is 79 mm (just 31% of the annual total), about half the amount accumulated on average at Tucson in the same summer months.

The effect of orography on rainfall is illustrated by observations of precipitation at different elevations in the Tucson–Santa Catalina Mountains region. The annual precipitation total rises from 297 mm at Tucson on the basin floor to 351 mm at the southern front of the Santa Catalina Mountains (Sabino Canyon) and eventually reaches 753 mm near the summit of this mountain range (Palisade Ranger Station). The fraction of annual rainfall received in the summer months drops off with increased elevation, suggesting that orographic enhancement of precipitation is greater in winter. The mean total rainfall for JAS is 157 mm at the mountain front and 319 mm at Palisade Ranger Station. Thus, JAS rainfall accounts for 52% of the annual total precipitation on the basin floor but just 45% at Sabino Canyon and 42% at Palisade Ranger Station.

The relative importance of the different types of storm conditions at the study sites is illustrated by the monthly mean total precipitation (Fig. 2). Winter mid-latitude cyclones are clearly the dominant rain-producing storm type at Kingman. Winter precipitation (November–May) accounts for 149 mm or 59% of the mean annual total. Monthly precipitation peaks in December and remains at above-average levels until March. Frontal systems bring less rain to the area in April and have nearly ceased by May. June is a transition month for the atmosphere; for this reason June is the driest month of the year, recording a scant 5 mm of rain on average. Rainfall totals increase back to winter levels in July as thunderstorm activity increases. Average rainfall in August is 36 mm, the highest monthly precipitation total for the year. The summer precipitation regime at Kingman terminates quickly, and average September rainfall (18 mm) is well below the annual average monthly mean.

The monsoon onset at Tucson is spectacular by comparison with that at Kingman. July is the rainiest month at Tucson, recording 60 mm of total rainfall on average, following a three-month dry season during which <20 mm of rainfall occurs on average. Rainfall amounts in both July and August are more than twice the annual average monthly mean. The monsoon abates in September, but that month's average rainfall still exceeds the annual average monthly mean. Rainfall totals for the rest of the year remain at or below the annual average monthly mean, again highlighting the predominance of the North American Monsoon at Tucson. Winter rainfall peaks in the months of December and January at 25 and 24 mm, respectively, just reaching annual monthly mean levels. As a whole, mid-latitude cyclones provide only 38% of the mean annual total to Tucson. The contribution of tropical cyclone remnants, which can be crudely estimated by October rainfall totals, is <10% of the mean annual totals at Tucson and Kingman.

Rainfall intensity is largely a function of the atmospheric conditions that produced the rain, so it is instructive to look at the peak rainfall intensities observed in different storms at different times of the year. Figure 3 shows a graph of the range of peak hourly rainfall intensity recorded for each storm in each month over the 46 yr of hourly observations. From November through May the highest rainfall intensity shows very little variability at both study sites. The mean maximum rainfall intensity for the winter at Kingman is 13.7 mm/h and ranges from a March high of 19.6 mm/h to a low of 8.6 mm/h in April. At Tucson the maximum winter rainfall intensities are slightly weaker, averaging 10.4 mm/h and ranging from 12.7 mm/h in March to 7.9 mm/h in February. These data suggest that the upward limit on rainfall intensity from a frontal system associated with a mid-latitude cy-

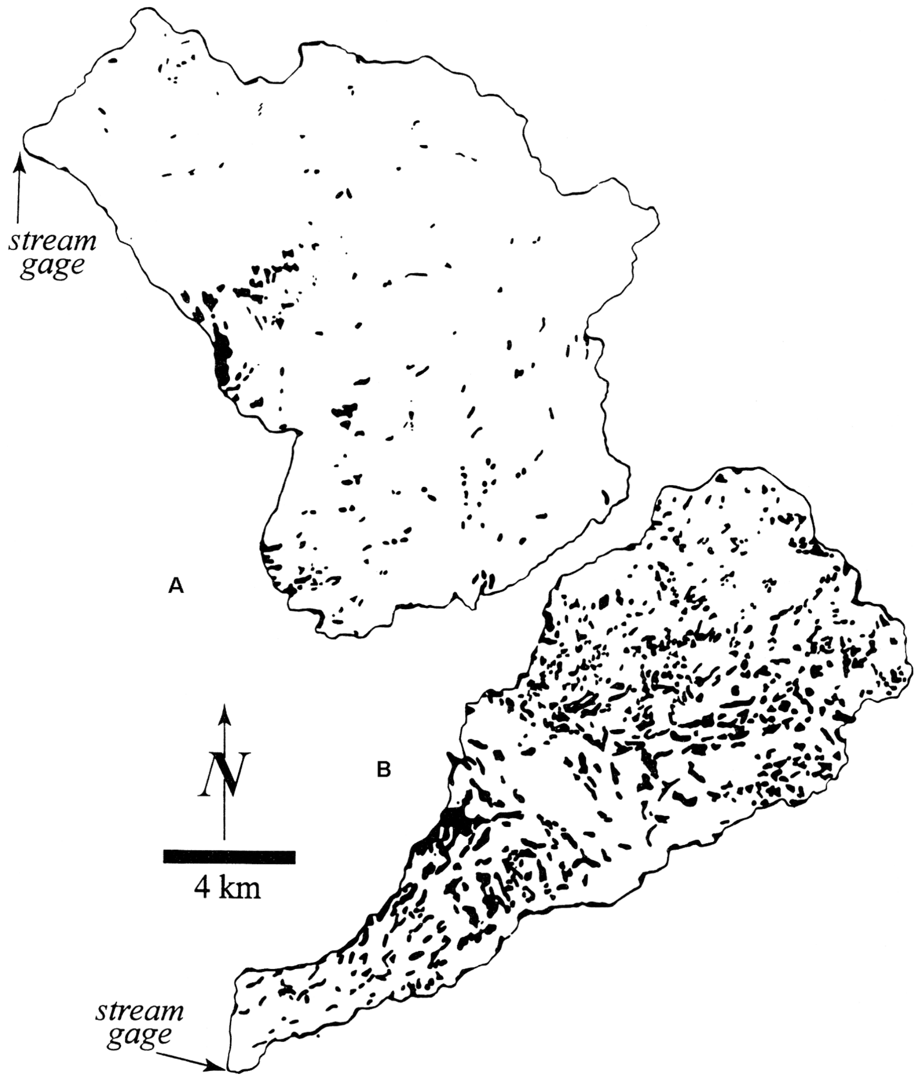

Figure 6. Extent of bedrock exposures (black areas) across (A) Walnut Creek and (B) Bear Creek drainage basins. Note that bedrock coverage is considerably denser at Bear Creek in the Santa Catalina Mountains.

clone lies in the 10–20 mm/h range at both locations.

Because intensity of summer thunderstorms depends on the instability of the atmosphere, it is no surprise that the maximum summer rainfall intensities occur during the month when surface temperatures reach their annual maximum. The additional importance of the monsoon as a moisture source to fuel summer thunderstorms is highlighted by the maximum rates recorded for the months of June and September. In June, prior to monsoon onset, the maximum rainfall intensity rises abruptly above 20 mm/h but never rises to the intensities observed during the subsequent core monsoon months. In September the lower temperatures and diminished monsoon moisture source act to hold maximum rainfall intensities between 20 and 30 mm/h.

Rainfall intensities for the monsoon season at both Kingman and Tucson are much more variable than for the winter. The maximum intensity recorded at both locations in JAS is just above 50 mm/h. At Tucson the peak rainfall intensity occurs in August at 56.4 mm/h, whereas at Kingman the greatest intensity peak is in July at 50.8 mm/h. The mean maximum rate at Tucson is 41.1 mm/h, only slightly higher than the 37.7 mm/h average at Kingman. Thus, the maximum rainfall intensity possible for a summer storm is approximately the same at Kingman and Tucson. Therefore the principal difference in the monsoon influence at these locations must be the frequency of high-intensity rainstorms, as can be shown directly by plotting the recurrence frequency of peak rainfall intensities at the two sites.

During winter (Fig. 4A), Kingman and Tucson exhibit similar peak rainfall intensities for short (subannual) recurrence intervals. Less-frequent, higher-intensity storms occur more often at Kingman than at Tucson. The modest slope of the regression line, compared to the summer data in Figure 4B, suggests that the peak intensities of very infrequent winter storms are not much larger than more common rainfall events. The difference in infrequent, high-intensity winter precipitation between the sites increases at longer recurrence intervals, although such estimates contain very large uncertainties as they are barely resolvable with our data sets.

In contrast, summertime peak rainfall intensity of a given magnitude occurs much more frequently at Tucson (about two and a half times more frequently) than at Kingman for all recurrence intervals (Fig. 4B). Rainfall intensities are much larger in summer than in winter for recurrence intervals of 1 yr or longer at both sites but especially at Tucson. On average, Tucson receives a summer thunderstorm with a peak rainfall intensity of 18.9 mm/h or greater at least once a year, whereas the corresponding peak intensity at Kingman is just 12.5 mm/h. The absolute difference of 6–7 mm/h between the sites remains nearly constant for longer recurrence intervals.

The seasonal cycle of rainfall duration per event is similar at the two sites. Average duration is longest (3–5 h) in winter months (between November and February) and shortest (1–2 h) in summer months (between May and August). These values are consistent with the different lifting mechanisms and precipitation types in the two seasons: relatively long duration frontal lifting in the winter and shorter-lived but intense thunderstorms in the summer.

LANDSCAPE METRICS AND THE HILLSLOPE SYSTEM

The total relief of the Hualapai and Santa Catalina mountain ranges is very similar at just under 2000 m from basin floor to their highest peaks; however, there are significant differences between these two ranges when local relief is considered. Taken over circular unit cells of $\sim$0.03 km^2 in area, the mean local relief for the entire Hualapai Mountains is 22.8 m. Despite possessing a similar rock type, the mean local relief of the Santa Catalina Mountains is twice as great at 45.2 m.

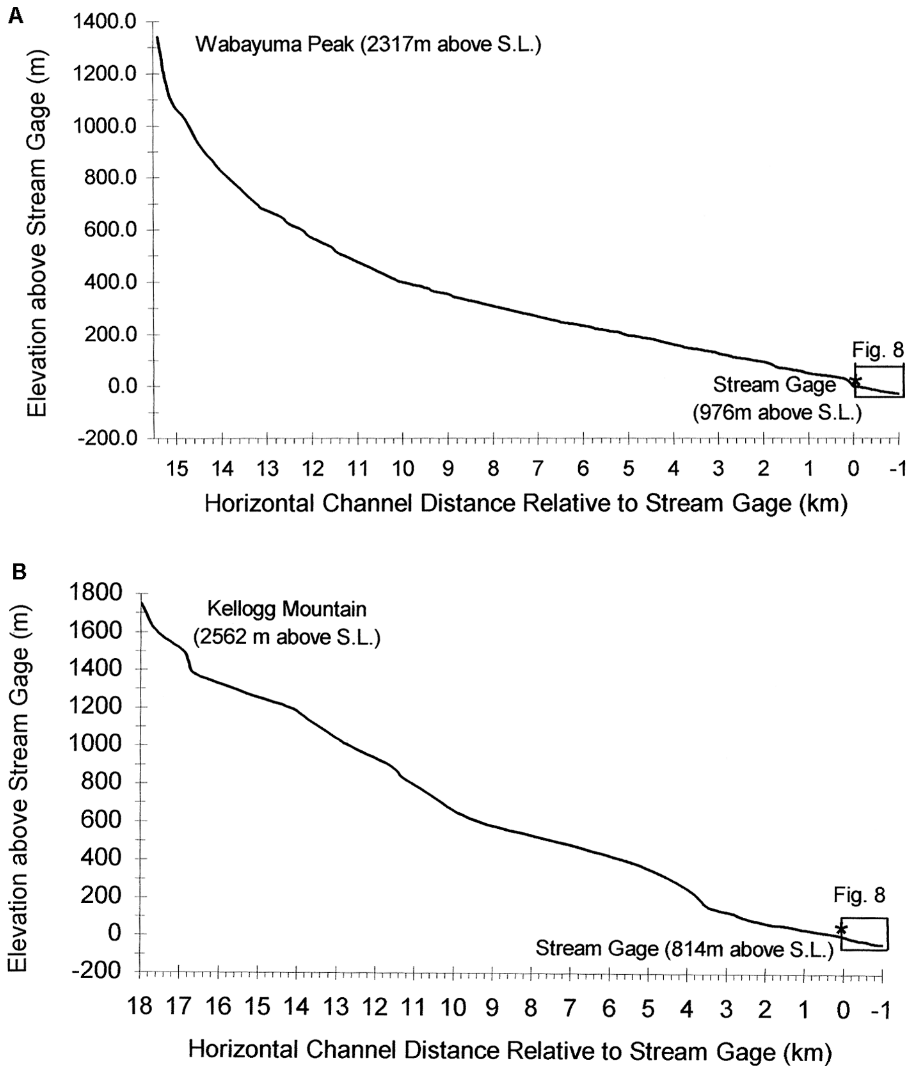

Figure 7. The longitudinal stream profiles of (A) Walnut Creek and (B) Bear Creek. At both drainages, the USGS gage station is located at the mountain front/piedmont transition. Upstream of the gages the channels are predominantly bedrock type; downstream from the gages, the channels are alluvial.

The Santa Catalina Mountains have a broader distribution of relief as well (Fig. 5). The standard deviation of local relief is 19.4 m for the Hualapai Mountains, whereas the maximum relief is 146 m. The standard deviation of local relief is 29.9 m in the Santa Catalina Mountains, and their maximum relief is 282 m.

Drainage density—defined as the total length of stream channels (identified from DEM data) divided by the area of the drainage basin—is sensitive to the balance between infiltration, overland flow, and erosion in the watershed. The drainage density over the area of the Hualapai Mountains measures 72.6 km/km^2. The corresponding value in the Santa Catalina Mountains is 29% denser, 94.0 km/km^2. The higher drainage density in the Santa Catalina Mountains results in overall shorter hillslope reaches between channels in this watershed compared to the Hualapai Mountains watershed. Stated another way, in comparison to the Hualapai Mountains, more of the precipitation input to the Santa Catalina Mountains encounters short-length, steep hillslopes and is ultimately routed out of local watersheds via overland flow.

Even though the watersheds are underlain by the same rock type compositionally, a vital premise of the research design is that the rock type has a similar resistance to erosion for both study areas (Bull and Schick, 1979). Measurements of rock-mass strength in the Hualapai Mountains ranged between 85 and 63 (on a scale from 1 to 100) and had a mean value of 75.0. In the Santa Catalina Mountains, rock-mass strength measurements ranged between 85 and 67 and had a mean value of 76.5. We conclude that rock-mass strength is essentially the same at both study areas.

The extent and distribution of exposed bedrock for Walnut Creek in the Hualapai Mountains and for Bear Creek in the Santa Catalina Mountains illustrates important differences (Fig. 6). Exposed bedrock covers ~2.6 km^2 of the Walnut Creek drainage basin or 3.2% of its total surface area. In contrast, total exposed bedrock in the Bear Creek catchment covers 12.5 km^2 or 29.7% of its total area. If these two drainages are representative, then the regional extent of exposed bedrock is nearly ten times greater in the Santa Catalina Mountains compared to that in the Hualapai Mountains. Bedrock in the Walnut Creek watershed is more concentrated near the divide where it underlies short, steep hillslopes somewhat isolated from major channels. In contrast, there is a more uniform distribution of bedrock in the Bear Creek watershed where bedrock underlies elongate narrow hillslopes with good continuity downslope to major channels. The relative percentage of exposed bedrock is consistent with the difference in drainage density of the respective catchments.

HYDROLOGY AND THE FLUVIAL SYSTEM

Walnut Creek in the Hualapai Mountains flows west and southwest for 15.4 km between the highest point in the basin and the USGS stream gage draining an area of 81.1 km^2. Dividing the approximate diameter of the basin by the length of its main channel yields an elongation ratio of 0.92, making this basin fairly circular in shape. Circular basins like the Walnut Creek watershed tend to focus runoff to the trunk channel, resulting in significant increases in the peak discharge downstream. Total relief for the basin is 1341 m from the stream gage (976 m above sea level) to Wabayuma Peak, the highest point of the basin at 2317 m above sea level. The mean channel gradient for this segment of Walnut Creek is 0.087.

Bear Creek in the Santa Catalina Mountains flows southwest a distance of 18.0 km between its highest point (Kellogg Mountain) and the USGS gage. The area of this drainage basin is 42.2 km^2. Bear Creek basin has an elongation ratio of 0.49, making this basin markedly more elongate in shape than Walnut Creek. Elongate basins like the Bear Creek watershed tend to slow the delivery of runoff

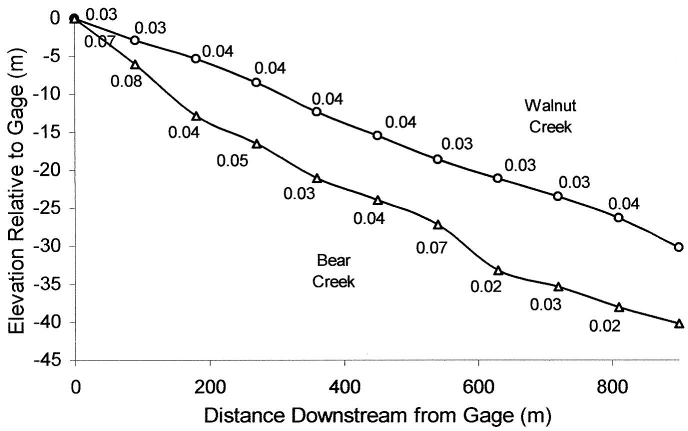

Figure 8. Longitudinal stream profiles for Walnut Creek (circles) and Bear Creek (triangles) immediately below the USGS gage stations. The measured channel gradient is displayed for select points along each profile.

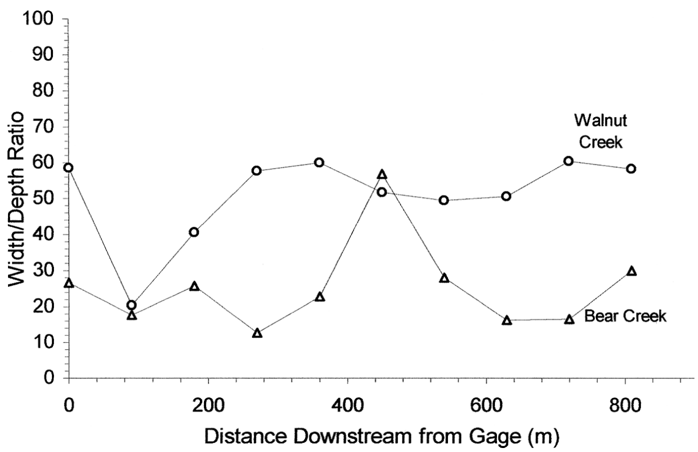

Figure 9. Channel width/depth ratios measured along Walnut Creek (circles) and Bear Creek (triangles) just below the USGS stream gage stations. Walnut Creek has a shallow channel that increases in width downstream, typical of alluvial-stream channels, whereas Bear Creek is more deeply incised.

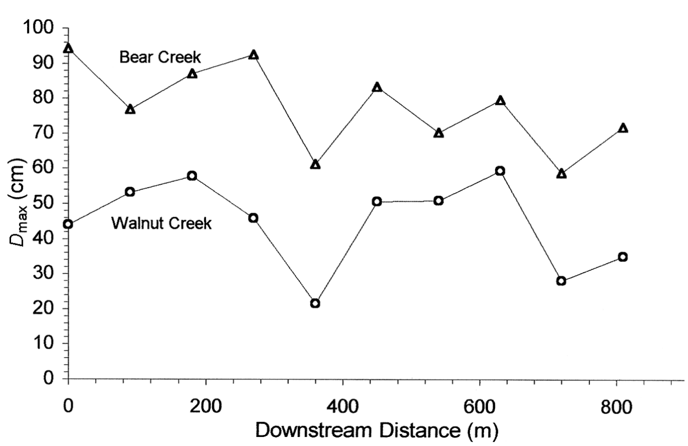

Figure 10. Sediment caliber for Walnut Creek (circles) and Bear Creek (triangles) based on the five largest clasts sampled.

to the trunk channel, resulting in a peak discharge that does not increase dramatically downstream. The total basin relief is 1748 m between the 2562 m summit of Kellogg Mountain and the stream gage (elevation, 814 m). Thus, the elevation and overall basin relief of Bear Creek are very close to those of Walnut Creek. The mean channel gradient for Bear Creek is 0.097 over this length.

The longitudinal profiles for Walnut and Bear Creeks above the USGS gage stations (Fig. 7) exhibit different concavities. Long-profile concavity is a complicated manifestation of several mutually dependent factors. The principal factors pertinent to our setting include the decrease in sediment caliber, increase in peak annual discharge, and increase in channel width downstream (Tucker and Bras, 2000). Walnut Creek has a relatively smooth, noticeably concave profile, typical of graded alluvial channels. In contrast, Bear Creek possesses a relatively straight, irregular profile with numerous knickpoints. These are commonly produced by vertical displacement along faults, changes in base level, streambed lithology changes, or the addition of large amounts of material to the channel by mass wasting (Knighton, 1987, 1998).

The differences between the Bear Creek and Walnut Creek profiles may be attributed to the different hydrology of the respective drainages. The smooth, graded profile of Walnut Creek is consistent with an alluvial channel undergoing more frequent sustained discharges that steadily increase downstream. In contrast, the Bear Creek profile is more consistent with an incised bedrock channel undergoing frequent flashy discharges.

The channels continue downstream from the USGS stream gages into their respective piedmonts (Fig. 8) where they incise into predominantly coarse-grained alluvium. The mean gradient for the entire mapped segment of Bear Creek is 0.040, slightly greater than the 0.033 mean gradient of Walnut Creek. Channel gradient remains fairly constant downstream at Walnut Creek, whereas at Bear Creek there is a dramatic decrease in channel slope from values of 0.06–0.08 down to 0.02–0.03, making the profile overall more concave in contrast to the longitudinal profile of the bedrock reach.

The widths of the channel cross sections measured at the alluvial stretch of Walnut Creek ranged from 12.2 to 29.3 m and had a mean span of 19 m. Measurements of channel width at Bear Creek were similar to those at Walnut Creek, ranging from 13 to 24 m and having a mean width of 18.9 m. Channels were about twice as deep overall at Bear

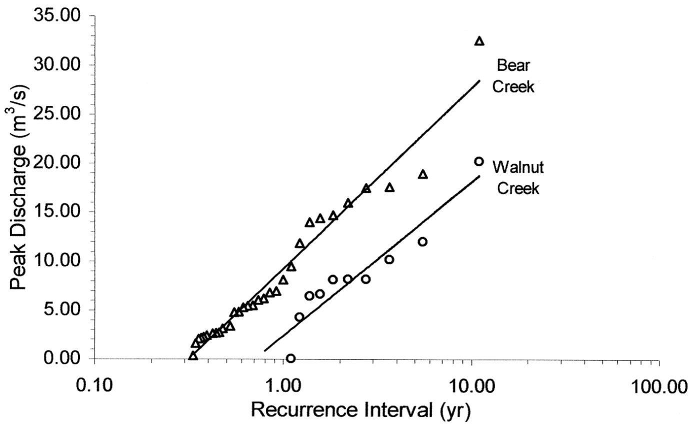

Figure 11. A comparison between magnitude and frequency of floods in Walnut Creek (circles) and Bear Creek (triangles) based on partial-duration series of stream discharge records from 1965 to 1974.

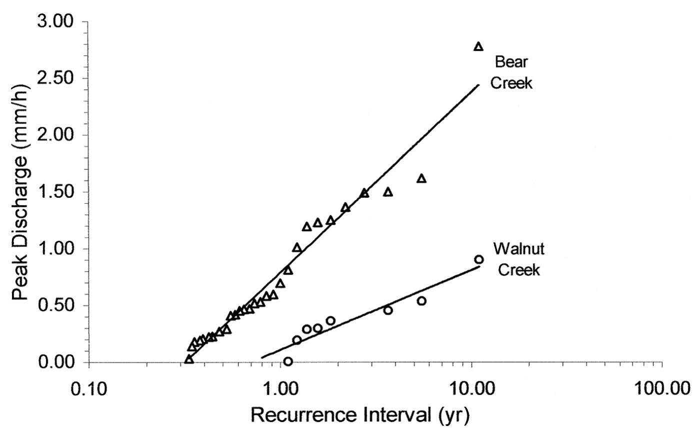

Figure 12. Partial-duration series, normalized by drainage area, for Walnut Creek (circles) and Bear Creek (triangles). Units are mm/h (see text for explanation).

Creek compared to Walnut Creek. The mean channel depth was 0.85 m at Bear Creek compared to a mean depth of 0.39 m at Walnut Creek. Channel depth at Bear Creek also displayed more variability, ranging from 0.4 to 1.5 m, whereas depths at Walnut Creek were between 0.3 and 0.6 m.

The channel width/depth ratio at set intervals along the length of the stream illustrates two trends (Fig. 9). First, the width/depth ratios are roughly twice as great for Walnut Creek as they are for Bear Creek. Second, Walnut Creek widens downstream but Bear Creek maintains a relatively constant ratio.

The caliber of channel sediments is a general proxy for the stream's ability to do geomorphic work, assuming that the Santa Catalina and Hualapai hillslopes do not produce boulders of vastly different size. Measurements of D_{max} in the streambeds of Bear and Walnut Creek (Fig. 10) show that Bear Creek possesses a much larger caliber of bed material than Walnut Creek. The mean D_{max} of the clasts sampled along the length of Bear Creek is 77.6 cm compared to a D_{max} of 43.6 cm for clasts at Walnut Creek. Bear Creek has been able to entrain and transport clasts roughly twice the size of those at Walnut Creek.

These grain-size data are consistent with partial-duration series of discharge frequency and magnitude for Walnut and Bear Creeks. The mean annual flood for the period 1965–1974 (Fig. 11) was 2.4 m^3/s at Walnut Creek and 9.3 m^3/s at Bear Creek. Thus, the Bear Creek drainage produces a mean annual discharge peak almost four times larger than that of Walnut Creek despite being only half its area. The 10 yr flood in the series is equivalent to 18.1 m^3/s at Walnut Creek and 27.7 m^3/s at Bear Creek. Thus, the data suggest that the larger area of the Walnut Creek catchment begins to play a greater role in modulating the size of the discharge as the return period of the flood (and by analogy the magnitude of the rainfall forcing signal) increases.

By normalizing the gage flow rates for basin area, it is possible to exclude the influence of drainage size on the magnitude of discharge. The normalized channel discharge (in mm/h) then represents the equivalent depth of water that is being shed from the entire basin surface that forms the discharge peak. As such, it provides a focused interpretation of runoff and sediment transport operating over the entire watershed. The mean annual flood at Walnut Creek produces a normalized peak discharge of 0.1 mm/h, whereas Bear Creek produces a flow rate of 0.8 mm/h (Fig. 12). The peak discharge in the 10 yr flood measures 0.8 mm/h at Walnut Creek and 2.4 mm/h at Bear Creek. A projection of the 100 yr flood yields a flow rate of 1.5 mm/h for Walnut Creek and 3.9 mm/h for Bear Creek. It is therefore evident that although the disparity in discharge rates decreases as recurrence interval increases, the magnitude of discharge from the monsoon-influenced catchment remains significantly greater per unit area than that of the nonmonsoonal basin.

Each of the landscape metrics we have examined—channel depth, sediment caliber, and discharge frequency—involves large uncertainties involving scouring, time since most recent intense flood, data record, etc., but taken together the data present a consistent and compelling picture of the basic geomorphic expression of these two fluvial systems. In both the drainage basin and on the piedmont of Walnut Creek, longer sustained discharges in the winter foster the development of alluvial channels with smooth, concave, long profiles, higher width/depth ratios, and smaller D_{max} values. In contrast, Bear Creek is characterized by a more irregular incised bedrock profile, lower width/depth ratios on the piedmont, and larger D_{max} values associated with frequent high-intensity summer precipitation.

The percentage of high-magnitude discharge events (defined here as discharge peaks greater in magnitude than the mean annual flood) that are generated in response to storms of a given type for several drainages in the Santa Catalina Mountains (Fig. 13) shows variability in the discharge response of streams that seem to depend on catchment area. Drainage basins $<\sim 20$ km^2 in area flood most often in response to high-intensity summer rainfall events. However, in drainages $>\sim 40$ km^2, low-intensity, long-duration rainfall from cyclonic storms is most often responsible for generating floods. For drainages of <20 km^2 area, there is a definite inverse relationship be-

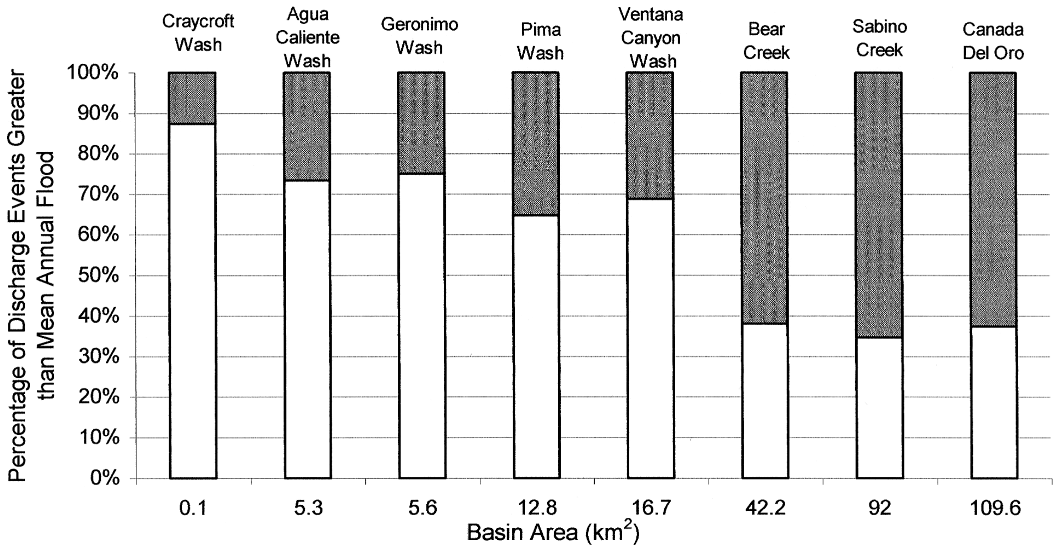

Figure 13. Percentage of discharge events greater than the mean annual flood resulting from different storm types in drainages in the Santa Catalina Mountains (see Table 2). Discharges preceded by monsoon thunderstorms are unshaded; discharges due to winter cyclonic storms are shaded.

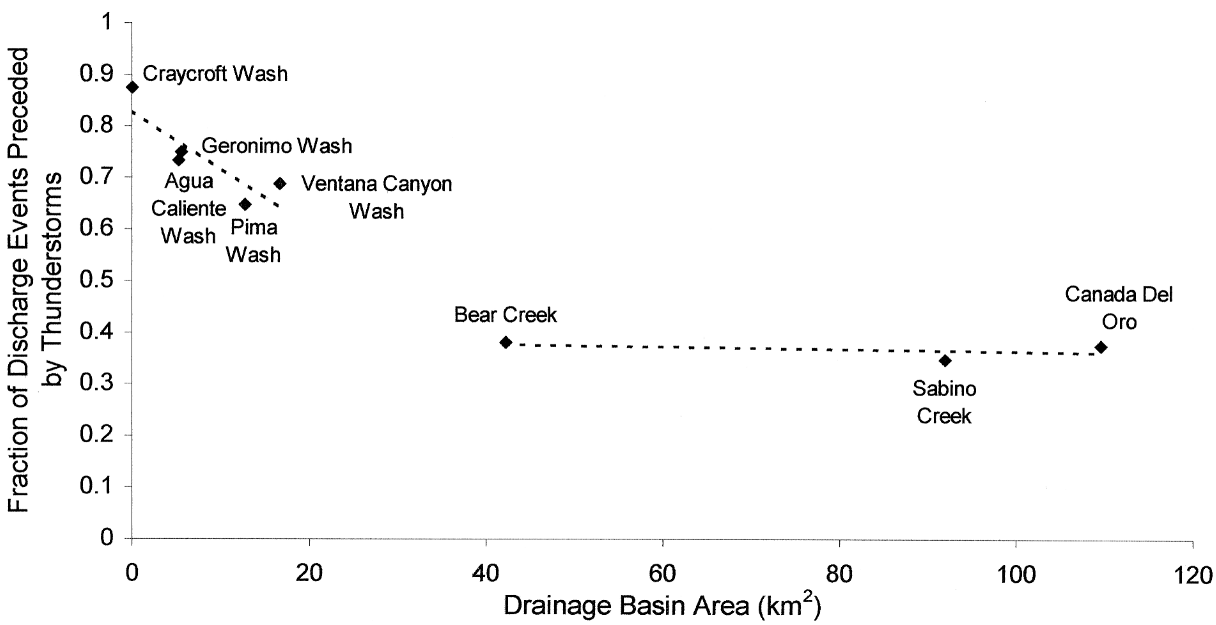

Figure 14. The relationship between basin size and responsiveness to monsoon thunderstorms. As basin size increases, fewer discharge events greater than the mean annual flood are generated from thunderstorms. In basins larger than 40 km², thunderstorms are responsible for less than 40% of the events larger than the mean annual flood.

tween basin area and sensitivity to high-intensity rainfall (Fig. 14). In contrast, drainages >40 km² in area are uniform in their low responsiveness to thunderstorms.

DISCUSSION

The comparison of two drainage systems carried out in this study supports the assertion that there is a measurable geomorphic expression of runoff generation and hillslope response to variations in high-intensity rainfall associated with the North American Monsoon. We develop this argument by first exploring the relationship between watershed geomorphology and hydrology. Hillslopes in the Santa Catalina Mountains are clearly more sparsely mantled by regolith and are shorter and steeper, compared to slopes in the Hualapai Mountains. All of these characteristics are consistent with lower infiltration rates, greater downslope flow continuity, and the resulting generation of overland flow delivered to a channel (Yair, 1992). The assertion that overland flow is a more prevalent process influencing hillslopes in the Santa Catalina Mountains is supported by the increased drainage density of this region compared to the Hualapai Mountains and by the greater peak discharges typically observed in the streams of the Santa Catalina Mountains. A likely consequence of increased density of drainages and magnitudes of stream discharge would be higher rates of vertical channel incision and concomitant increases in local relief, both of which are consistent with our data.

The lithology of the study sites should offer similar resistance to erosion as attested to by the field measurements of rock-mass strength. Likewise the total regional relief values of the Santa Catalina and Hualapai Mountains are similar. Given similar lithology, relief, and annual precipitation, it is possible to conclude that the increased runoff generation in the Bear Creek watershed has accelerated rates of hillslope processes and thus increased drainage density and variability in local relief while decreasing the extent of sediment-mantled hillslopes.

The effects of decreased infiltration and increased runoff generation are also expressed in the trunk-channel hydraulic geometry and hydrographs. The mean annual flood at Bear Creek is roughly eight times larger than that at Walnut Creek (when normalized for basin area), and the magnitude of the discharge peak increases at a much more rapid rate with decreasing return times in the monsoonal catchment.

The increased discharge levels of Bear Creek come in spite of the fact that this basin is much more elongate in shape than Walnut Creek. The greater elongation of Bear Creek is most likely a result of a strong structural lineation that is present in the Santa Catalina Gneiss, which trends toward the southwest. Channel erosion is increased along the axis of lineation so that many of the major drainages of the Santa Catalinas appear to follow this structural trend. The assumed effect of increased basin elongation would be to weaken the magnitude of the discharge peak recorded in the main trunk channel of the catchment. Thus, solely based on basin shape, the more circular watershed at Walnut Creek should undergo greater peaks in stream discharge, which is clearly not the case. The smooth longitudinal profile of Walnut Creek must result primarily from some other factor such as a rapid downstream fining of sediment, a possibility consistent with the primarily transport-limited nature of this watershed.

The higher peak discharge of Bear Creek fosters incision, expressed as an irregular bedrock profile in the drainages and a concave alluvial profile with a relatively low width/depth ratio on the piedmont. We attribute this to the high runoff generation on hillslopes above Bear Creek. Alternatively the deeper channels at Bear Creek may result from the better consolidation of bank materials there compared to Walnut Creek (Gregory and Walling, 1973; Grissinger, 1982) because the channel banks at Bear Creek appear to be more heavily vegetated.

The relationship between runoff characteristics and geomorphology in these two basins leads us next to consider the role that precipitation plays in shaping the landscape. Rainfall variability has been associated with higher drainage density in both field (Chorley, 1957; Chorley and Morgan, 1962) and modeling studies (Tucker and Bras, 1998, 2000). Rainfall variability is also thought to influence sediment transport along trunk streams, especially those with coarse bedloads (Tucker and Bras, 2000). Both of these observations are consistent with our results.

It was expected that high-magnitude discharge events in the drainages of the Santa Catalina Mountains would occur most often during the monsoon season because it was assumed that such events occur in direct response to high-intensity rainfall, but this was not always the case. In smaller drainages (<17 km^2), peak-discharge events in excess of the mean annual flood occur most often in the monsoon season, just as expected. But in larger drainages (>40 km^2), $\sim$60% of the discharge peaks exceeding the mean annual flood occur in the winter months. Thus, the hydrologic response of stream channels in the Santa Catalina Mountains appears to be dependent on both the seasonality of rainfall and drainage-basin scale.

The disparity in the hydrologic response of different sized basins to summer thunderstorms is probably due to the limited areal coverage of rainfall at any instant in these storms, a concept that has been demonstrated by several researchers using different techniques (Turnage and Mallory, 1941; Sharon, 1972; Ying et al., 1996). As a result, the rainfall coverage over a large basin is usually fractional. The volume of precipitation delivered to the basin in this case is small relative to the capacity of the stream channels and thus results in a low-magnitude discharge. Rainfall coverage from a thunderstorm is more likely to be complete over a small basin, which allows the high-intensity rainfall of the thunderstorm to generate a high-magnitude discharge.

In contrast, rainfall from a winter or late summer cyclonic storm is much more areally extensive, so rainfall coverage over the large basin is more complete, and a volume of precipitation sufficient to generate a large discharge is captured. As demonstrated above, the monsoon acts to decrease infiltration rates on hillslopes. The reduction of hillslope infiltration rates in turn allows even low-intensity winter rainfall to generate significant rates of overland flow. In this way summer rainfall primes the large basin for response to winter cyclonic storms. In the small basin (<17 km^2), where the channel system has been formed in equilibrium with the high-volume discharges generated by summer storms, low-intensity winter rainfall usually fails to produce floods, but the integrated effect of many smaller basins is to generate maximum peak discharge in the drainages covering >40 km^2.

The monsoonal impacts on (1) the hillslope itself (increased sediment erosion inferred from exposed bedrock and increased local relief and drainage density), and (2) the hydrological components of the landscape system (increased peak-discharge rates) should have an extensive effect on the behavior of piedmont channels given similar base-level conditions. Bull (1991) has noted that on Holocene time scales, even large-scale regional base-level changes are incapable of influencing the evolution of channels such as Bear and Walnut Creeks, which are several hundred kilometers upstream from their ultimate base levels, and thus 10^3–10^4 yr piedmont incision and aggradation must be in response to upstream hydrology rather than base-level fluctuation.

The data on which this discussion is based are limited with respect to spatial coverage and temporal length. We have restricted the analysis to two limited study areas in order to control for factors other than precipitation variability. It would be highly desirable to have longer stream-gage records at comparable sites, hourly precipitation data coincident with stream gages, long-term records of duration and spatial extent of precipitation events in the watersheds, and many more study sites, in order to properly generalize the results we have obtained. Despite these important limitations, we offer the following interpretive discussion as a working hypothesis for future studies.

THE MODEL FOR SOUTHWEST LANDSCAPE EVOLUTION

The detailed observations made in this study of two drainage systems in contrasting precipitation regimes support the general models proposed by Gerson and Yair (1975) and Bull (1991) for watershed response to climate change. Beginning at ca. 10,000 to 8000 yr B.P., an increase in summer rainfall marked the initiation of the North American Monsoon as a climate phenomenon prevalent in the Southwest (Van Devender and Spaulding, 1979; Davis and Shafer, 1992; Davis, 1994; Peterson, 1994; Anderson et al., 2000). High-intensity rainfall stripped away regolith in regions where hillslopes were previously mantled with colluvium, dramatically increasing sediment yields. A greater frequency of debris flows, valley-floor aggradation, and acceleration of alluvial-fan head deposition (fan steepening) would have been consequences of the initial monsoon signature in this type of setting. Discharge from hillslopes and stream channels increased as a direct result of increased high-intensity rainfall and indirectly as hillslope-infiltration rates were lowered in response to the erosion of colluvium.

As hillslopes became increasingly weathering-limited, sediment yield began to decline. Debris flows became less frequent, and stream channels began a phase of incision into alluvial surfaces that left behind terraces and dissected fans. Discharge rates remained high, whereas sediment yields achieved an equilibrium rate dictated by the ability of rock hillslopes to supply material to the stream channel. In the piedmont, older alluvial surfaces continued to be incised and in some cases may have been completely obliterated.

A significant new dimension to this model is suggested by our data on the variability of hydrological response to rainfall with storm type and basin size; there are implications for future hydrological changes as well. Climate predictions based on global warming scenarios include the possibility of an increased frequency and magnitude of warm El Niño/Southern Oscillation (ENSO) events (Timmermann et al., 1999). Warm ENSO events are associated with higher than average amounts of winter precipitation in the Southwest United States (Ropelewski and Halpert, 1986). The effect of ENSO events on summer precipitation is more problematic. Gutzler and Preston (1997) and Gutzler (2000) have shown evidence for a negative correlation between spring snowpack in the Rocky Mountains and the abundance of summer precipitation in the Southwest in recent years. They have hypothesized that anomalously extensive snowpack in the southern Rocky Mountains, such as typically occurs during warm ENSO events, acts to suppress the development of the stationary low over the Rockies responsible for the subsequent monsoon. A possible increase in frequency of warm ENSO events as predicted in global warming scenarios could therefore lead to a major shift in the seasonal pattern of rainfall over the Southwest toward increased winter precipitation and decreased summer rainfall.

A shift in proportion from summer to winter precipitation would subsequently reduce the amount of high-intensity rainfall currently received by the region and cause significant landscape repercussions. The most immediate effect would be a decrease in high-magnitude discharge events from the small basins that show the greatest sensitivity to summer rainfall. Unfortunately, at the same time, increased winter precipitation would increase the chance of major floods in larger basins.

CONCLUSION

Better understanding of the effect of high-intensity rainfall on rates of geomorphic processes is vital for the interpretation of Holocene landscape evolution and for predicting the consequences of climate change on future environments. The basic assumption that the high-intensity rainfall of the North American Monsoon fosters increased rates of hillslope processes, leading to higher rates of stream discharge, sediment yield, and regional denudation, is largely validated by the results of this study. The monsoon season is characterized by an increased frequency of high-intensity rainstorms, which in turn translates into a higher frequency of major overland-flow events. Overland flow increases erosion of sediment from hillslopes, which shows up in the landscape as increased bedrock exposure, drainage density, and local relief. Monsoon rainfall leads to higher rates of peak discharge in stream channels, both directly by generating high rates of overland flow and indirectly by stripping hillslopes of sediment and thus lowering infiltration capacity.

The area of drainage basins appears to play a fundamental role in modulating the hydrologic response of the catchment to different regimes of precipitation. Basins in the Santa Catalina Mountains of <17 km^2 in area tend to undergo most of their high discharge events in the summer monsoon season, whereas basins of >40 km^2 in area generate peak discharges most often after winter cyclonic storms. The existence of a threshold basin size between 20 and 40 km^2 is suggested by our study but cannot be constrained by available data. The robustness of this result should be examined in other monsoonal regions in the Southwest.

Evidence of the increased peak-discharge rates of streams in the monsoon environment is manifested in alluvial-stream channels. Streams fed by the monsoon are capable of entraining and transporting coarser bedload and have an increased tendency to incise into the alluvial piedmont. These effects do not occur principally in summer, however. Instead, stream flow is affected by the monsoon indirectly. Summer precipitation primes the slopes of small watersheds, enhancing the effects of large-scale winter cyclonic storms that do most of the geomorphic work of sculpting the landscape.

ACKNOWLEDGMENTS

Les McFadden provided useful feedback and encouragement. Jessica Preston assisted in the field. Critical reviews by P. Pearthree, A. Yair, and P. Knuepfer led to significant improvements in the manuscript. This research has been supported by a grant from the National Science Foundation (award 97-31089).

REFERENCES CITED

Adams, D.K., and Comrie, A.C., 1997, The North American Monsoon: Bulletin of the American Meteorological Society, v. 78, p. 2197–2213.

Anderson, R.S., Betancourt, J.L., Mead, J.I., Hevly, R.H., and Adam, D.P., 2000, Middle- and late-Wisconsin paleobotanic and paleoclimatic records from the southern Colorado Plateau, USA, *in* Kershaw, A.P., and Whitlock, C., eds., Paleoecological records of the last glacial/interglacial cycle: Patterns and causes of change: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 155, p. 31–57.

Baker, V.R., 1973, Paleohydrology and sedimentology of the Lake Missoula flooding in eastern Washington: Geological Society of America Special Paper 144, 79 p.

Bathurst, J.C., 1987, Measuring and modeling bedload transport in channels with coarse bed materials, *in* Richards, K., ed., River channels: Environment and process: Institute of British Geographers Special Publication, p. 272–294.

Budel, J., 1982, Climatic geomorphology: Princeton, New Jersey, Princeton University Press, 443 p.

Bull, W.B., 1991, Geomorphic response to climate change: New York, Oxford Press, 326 p.

Bull, W.B., and Schick, A.P., 1979, Impact of climatic change on an arid watershed: Nahal Yael, southern Israel: Quaternary Research, v. 11, p. 153–171.

Chorley, R.J., 1957, Climate and morphometry: Journal of Geology, v. 65, p. 628–638.

Chorley, R.J., and Morgan, M.A., 1962, Comparison of morphometric features, Unaka Mountains, Tennessee and North Carolina, Dartmoor, England: Geological Society of America Bulletin, v. 73, p. 17–34.

Church, M.A., McLean, D.G., and Wolcott, J.F., 1987, River bed gravels: Sampling and analysis, *in* Thorne, C.R., Bathurst, J.C., and Hey, R.D., eds., Sediment transport in gravel-bed rivers: Chichester, John Wiley & Sons, p. 43–88.

Costa, J.E., 1983, Paleohydraulic reconstruction of flash-flood peaks from boulder deposits in the Colorado Front Range: Geological Society of America Bulletin, v. 94, p. 986–1004.

Davis, O.K., 1994, The correlation of summer precipitation in the southwestern USA with isotopic records of solar activity during the Medieval Warm Period: Climate Change, v. 26, p. 271–287.

Davis, O.K., and Shafer, D.S., 1992, A Holocene climate record for the Sonoran Desert from pollen analysis of Montezuma Well, Arizona, USA: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 92, p. 107–119.

Douglas, M.W., Maddox, R.A., and Howard, K., 1993, The Mexican Monsoon: Journal of Climate, v. 6, p. 1665–1677.

Eldred, D.W., and Moore, R.T., 1959, Geologic map of Mo-

have County, Arizona: Arizona Bureau of Mines, scale 1:250,000.
Eldred, D.W., Moore, R., and O'Haire, R.T., 1960, Geologic map of Pima and Santa Cruz Counties, Arizona: Arizona Bureau of Mines, scale 1:375,000.
Faulkner, H., 1986, Gully evolution in response to both snowmelt and flash flood erosion, western Colorado, *in* Gardiner, V., ed., International geomorphology, Part 1: Chichester, John Wiley & Sons, p. 947–969.
Gerson, R., and Yair, A., 1975, Geomorphic evolution of some small desert watersheds and certain paleoclimatic implications, Santa Catherine area, southern Sinai: Zeitschrift für Geomorphologie, v. 19, p. 66–82.
Gregory, K.J., and Walling, D.E., 1973, Drainage basin form and process: A geomorphological approach: London, Edward Arnold, 456 p.
Grissinger, E.H., 1982, Bank erosion of cohesive materials, *in* Thorne, C.R., Bathurst, J.C., and Hey, R.D., eds., Gravel-bed rivers: Chichester, John Wiley & Sons, p. 273–287.
Gutzler, D.S., 2000, Covariability of spring snow pack and summer rainfall across the Southwest United States: Journal of Climate, v. 13, p. 4018–4027.
Gutzler, D.S., and Preston, J.W., 1997, Evidence for a relationship between spring snow cover and summer rainfall in New Mexico: Geophysical Research Letters, v. 24, p. 2207–2210.
Hales, J.E., 1972, Surges of maritime tropical air northward over the Gulf of California: Monthly Weather Review, v. 100, p. 298–306.
Hereford, R., 1993, Entrenchment and widening of the upper San Pedro River, Arizona: U.S. Geological Survey Professional Paper 282, 46 p.
Hereford, R., and Webb, R.H., 1992, Historic variation of warm-season rainfall, southern Colorado Plateau, southwestern USA: Climate Change, v. 22, p. 239–256.
Higgins, R.W., Yao, Y., and Wang, X.-L., 1997, Influence of the North American Monsoon system on the U.S. summer precipitation regime: Journal of Climate, v. 10, p. 2600–2622.
Knighton, A.D., 1987, River channel adjustment—The downstream dimension, *in* Richards, K., ed., River channels: Environment and process: Institute of British Geographers Special Publication, p. 95–128.
Knighton, A.D., 1998, Fluvial forms and processes, a new perspective: New York, Toronto, Wiley and Sons, 381 p.
Leopold, L.B., 1951, Rainfall frequency: An aspect of climate variation: American Geophysical Union Transactions, v. 32, p. 347–357.
Menges, C.M., and Pearthree, P.A., 1989, Late Cenozoic tectonism in Arizona and its impact on regional landscape, *in* Jenney, J.P., and Reynolds, S.J., eds., Geologic evolution of Arizona: Arizona Geological Society Digest, v. 17, p. 649–680.
Mosley, M.P., and Tindale, D.S., 1985, Sediment variability and bed material sampling in gravel-bed rivers: Earth Surface Processes and Landforms, v. 10, p. 465–482.
Petersen, K.L., 1994, A warm and wet Little Climatic Optimum and a cold dry Little Ice Age in the Southern Rocky Mountains, USA: Climate Change, v. 26, p. 243–269.
Reneau, S.L., McDonald, E.V., Gardner, J.N., Kolbe, T.R., Carney, J.S., Watt, P.M., and Longmire, P.A., 1996, Erosion and deposition on the Pajarito Plateau, NM, and implications for geomorphic responses to late Quaternary climate changes, *in* New Mexico Geological Society Guidebook, 47th Conference, Jemez Mountains region: Socorro, New Mexico Geological Society, p. 391–397.
Ropelewski, C.F., and Halpert, M.S., 1986, North American precipitation and temperature patterns associated with the El Niño/Southern Oscillation (ENSO): Monthly Weather Review, v. 114, p. 2352–2362.
Schmitz, J.T., and Mullen, S.L., 1996, Water vapor transport associated with the summertime North American Monsoon as depicted by ECMWF analyses: Journal of Climate, v. 9, p. 1621–1634.
Selby, M.J., 1980, A rock mass strength classification for geomorphic purposes: With tests from Antarctica and New Zealand: Zeitschrift für Geomorphologie, v. 24, p. 31–51.
Sellers, W.D., and Hill, R.H., 1974, Arizona climate, 1931–1972: Tucson, University of Arizona Press, 616 p.
Sharon, D., 1972, The spottiness of rainfall in a desert area: Journal of Hydrology, v. 17, p. 161–175.
Shreve, F., and Wiggins, I.L., 1964, Vegetation and flora of the Sonoran Desert, Volume 1: Palo Alto, California, Stanford Press, 1740 p.
Smith, W.P., and Gall, R.L., 1989, Tropical squall lines of the Arizona Monsoon: Monthly Weather Review, v. 117, p. 1553–1569.
Timmermann, A., Oberhuber, J., Bacher, A., Esch, M., Latif, M., and Roeckner, E., 1999, Increased El Niño frequency in a climate model forced by future greenhouse warming: Nature, v. 398, p. 694–697.
Tucker, G.E., and Bras, R.L., 1998, Hillslope processes, drainage density, and landscape morphology: Water Resources Research, v. 34, p. 2751–2764.
Tucker, G.E., and Bras, R.L., 2000, A stochastic approach to modeling the role of rainfall variability in drainage basin evolution: Water Resources Research, v. 36, p. 1953–1964.
Turnage, W.V., and Mallory, T.D., 1941, An analysis of rainfall in the Sonoran Desert and adjacent territory: New York, Carnegie Institute, 45 p.
Van Devender, T.R., and Spaulding, W.G., 1979, Development of vegetation and climate in the southwestern United States: Science, v. 204, p. 701–710.
Wells, S.G., McFadden, L.D., and Dohrenwend, J.C., 1987, Influence of late Quaternary climatic changes on geomorphic and pedogenic processes on a desert piedmont, eastern Mojave Desert, California: Quaternary Research, v. 27, p. 130–146.
Yair, A., 1992, The control of headwater area on channel runoff in a small arid watershed, *in* Parsons, A.J., and Abrahams, A.D., eds., Overland flow: Hydraulics and erosion mechanics: Chapman and Hall, p. 53–68.
Ying, F., Wood, E.E., Baeck, M.L., and Smith, J.A., 1996, Fractional coverage of rainfall over a grid: Analysis of NEXRAD data over the southern Plains: Water Resources Research, v. 32, p. 2787–2802.

Manuscript Received by the Society 24 July 2001
Revised Manuscript Received 25 November 2002
Manuscript Accepted 12 December 2002

Printed in the USA