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The disparity between extreme rainfall events and rare floods — with emphasis on the semi-arid American West

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Abstract:

Research beginning 40 years ago suggested that semi-arid lands of the USA have higher unit discharges for a given recurrence interval than occur in other areas. Convincing documentation and arguments for this suspicion, however, were not presented. Thus, records of measured rainfall intensities for specified durations and recurrence intervals, and theoretical depths of probable maximum precipitation for specified recurrence intervals and areal scales are considered here for comparing extreme rainfalls of semi-arid areas with those of other climatic areas. Runoff from semi-arid lands, as peaks of rare floods, is compared with that of other areas using various published records. Relative to humid areas, semi-arid parts of the conterminous USA have lower 100-year, 6-h rainfall intensities and smaller depths of 100-year probable maximum precipitation for 26-km^2 areas. Nonetheless, maximum flood peaks, flash-flood potentials, and runoff potentials are generally larger in semi-arid areas than in more humid parts of the nation. Causes of this disparity between rainfall and runoff appear to be results of soil and vegetation that in humid areas absorb and intercept rainfall and attenuate runoff, but in semi-arid areas limit infiltration and enhance runoff from bare, crusted surfaces. These differences in soil and vegetation conditions are indicated by the relatively high curve numbers and drainage densities that are typical of semi-arid areas. Owing to soil and vegetation conditions, rare floods in semi-arid areas are more likely to cause landform change than are floods of similar magnitude elsewhere.

KEY WORDS rare floods; probable maximum precipitation; extreme events; semi-arid; vegetation; surficial geology; curve number; drainage density

INTRODUCTION

A long-standing perception that intense rainfall–runoff events are more common in semi-arid lands than elsewhere persists despite a lack of documentation. A series of papers, in response to an analysis of geomorphological work accomplished by streamflows of varying frequency (Wolman and Miller, 1960), included considerations of whether high-magnitude floods occur atypically often in semi-arid environments, but failed to provide data demonstrating that rainfall–runoff relationships differ significantly from those of other climates. Papers that questioned and elaborated on the geomorphological-work model of Wolman and Miller (1960) debated the extent to which rare floods of different climates cause landscape change, but did not specify whether observed change may be due largely to regional variations in runoff potential or due to climate-related differences in landform resistance to change.

Among the reactions to Wolman and Miller (1960) were papers by Baker (1977), who, within a context of landscape change by rare floods, implied that weathering processes and vegetation cover of the eastern USA are more conducive to infiltration and the attenuation of runoff from intense rainfall than occur in drier

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areas, and Wolman and Gerson (1978), who, in discussing landform response to infrequent storms, maintained that whereas 'rainfall intensities and maximum discharges are quite similar in different regions, their effectiveness as formative agents on channels and hillslopes is not'. Subsequent papers analysed the hydraulics of selected historical natural floods and dam-failure floods (Baker and Costa, 1987), and proposed a conceptual stream-power model based on total energy expenditure during a flood (Costa and O'Connor, 1995). All of these papers, whether explicitly or implicitly through tabulations of flood data, demonstrated a tendency for rare floods to be most common in semi-arid areas without identifying regional variations in rainfall–runoff relations.

In this paper the question considered is whether extreme rainfalls and rare floods, as indicated by return period for a given watershed size, are greater in semi-arid areas than in other climatic zones of North America. For the purposes of this discussion, an extreme rainfall is one of short duration, such as 6 h, with an average return period measured in 100 years. A rare flood is one that occurs through normal (repeatable) processes of intense precipitation, runoff and streamflow, that lends itself to standard techniques of flood-frequency analysis, and that has a return period estimated in 100 to 1000 years. The term catastrophic flood, as used by Baker (1977) and Costa (1987), among others, implies significant geomorphic change by infrequent repeatable events as well as by non-recurring natural flow events or multiple events such as glacial-outburst floods, failure of landslide dams, or rapid drainage of glacial-meltwater lakes, or failure of human-made impoundments. Although conditions common in semi-arid areas may make bottomlands especially vulnerable to catastrophic geomorphological change by rare floods, emphasis here is on comparisons of extreme rainfalls and rare, repeatable runoff events in semi-arid and other climatic environments.

The term semi-arid refers in this discussion to the climate of areas receiving from 250 to 500 mm mean annual precipitation, recognizing that many drainage basins of western North America are bounded by uplands receiving greater precipitation and have valley floors receiving precipitation less than those limits. Except where mountains collect orographic precipitation, semi-arid areas of the USA extend westward from the 100°W meridian generally through the Great Basin, basins of the Colorado and Snake Rivers, and the western Great Plains; parts of central and southern California are included also.

EXTREME RAINFALL

Estimates of intense-precipitation magnitudes, as rainfall depths, are available for the USA in atlases by Hershfield (1961), and more recently for the western USA in Miller (1973). These atlases were developed directly from weather-station data and support the contention that storm intensities in semi-arid regions of the USA do not differ notably from those in most other parts of the country. The magnitudes of 100-year, 6-h rainfall events (the rainfall depths for storms with a 6-h duration and a return period of 100 years), for example, are greatest, about 200 mm, along the Gulf Coast of the USA and decrease steadily to the east, north and west to about 100 mm in New England, Iowa and Minnesota, and eastern sections of New Mexico and Colorado (Figure 1).

West of the 100°W meridian, estimates of 100-year, 6-h rainfalls are locally greatest in mountainous parts of California, Oregon and Washington, but otherwise generally do not exceed the 100-mm rainfalls of southern and central Arizona. The greatest depths of 100-year, 6-h rainfalls in semi-arid areas of the USA have occurred in southern Arizona, southern Texas and parts of the south-eastern Great Plains (Figure 1). Estimates, based on extensive precipitation data (Osborn, 1981), of 100-year, 6-h rainfall events at Alamogordo Creek Experimental Watershed, eastern New Mexico (85 mm), and Walnut Gulch Experimental Watershed, south-eastern Arizona (74 mm), are consistent with the earlier maps of Hershfield (1961). In the two watersheds studied, Osborn (1981) found little change in depths of intense rainfalls with increasing elevation.

Probable maximum precipitation (PMP) is an estimated precipitation depth, the greatest that is considered physically possible for specified conditions of storm size, location and time of year (Hansen *et al.*, 1994). Thus, PMP is a potential, or theoretical, rainfall, unlikely ever to be observed, that is computed from data on

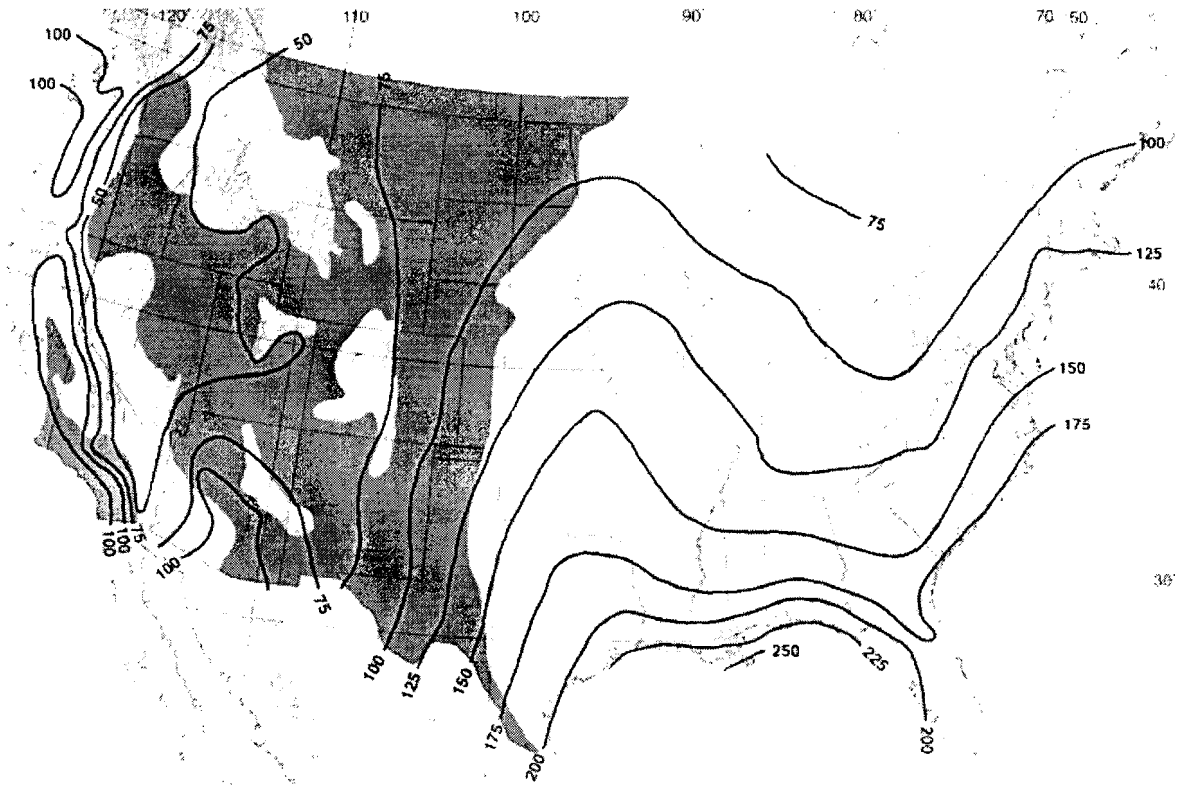

Figure 1. Map of the conterminous USA showing contours of depths of 100-year, 6-h precipitation events, in millimetres (simplified from Hershfield, 1961). Shaded (semi-arid) areas (modified from US Geological Survey, 1984) receive 250 to 500 mm mean annual precipitation)

potential free atmospheric precipitation, location relative to sources of moisture, an orographic factor, a storm-intensity factor and observed precipitation-frequency values (Hansen *et al.*, 1988). A series of hydrometeorological reports, by various authors of several US government agencies, provides detailed interpretations, by region, of PMP for the conterminous USA and parts of Canada. These interpretations, unlike maps of precipitation frequency that are generated from observed rainfalls but generally lack local detail, include considerations of orographic effects that are not necessarily validated by site-specific weather records. Interpreted values for 6-h PMP (for 26-km^2 areas) accordingly are as great as 800 mm or more and are ordinarily two to seven times greater than are those suggested by maps of precipitation frequency for 100-year, 6-h rainfall for numerous high-relief areas of the western USA (Hansen *et al.*, 1988, 1994).

The regional variation in 6-h PMP for 26-km^2 drainage basins (Figure 2) is similar to the pattern of 100-year, 6-h precipitation (Figure 1). The highest values occur in the areas closest to the source of hurricanes and warm moist air in the Gulf of Mexico and the southern Atlantic Ocean. The humid areas east of the 100°W meridian generally have higher PMP than either the semi-arid or the humid areas to the west. The principal difference between the two presentations (Figures 1 and 2) is the higher sensitivity to elevation of PMP. Elevation has a non-linear influence on extreme precipitation. The first movement of a warm, moist airmass forced upslope against landforms intensifies precipitation through orographic lifting (Maddox *et al.*, 1978; Smith *et al*, 1996); as an airmass rises still higher, PMP declines because the amount of moisture available in the air column declines (Hansen *et al.*, 1988). As a result, the upslope areas traversed first by a moist airmass, such as the High Plains and eastern foothills of the Rocky Mountains, have relatively high

PMP, whereas PMP in the Rocky Mountains themselves is low. The PMP in the western Great Plains and west Texas is only slightly below that of most of the eastern USA. At a finer scale, because the western USA generally has higher relief than does the eastern USA, the opportunities for orographic lifting are more abundant.

RUNOFF POTENTIAL

Runoff depends directly on storm intensity and duration but is also affected by topographic slope, soil conditions and other climate-related variables such as vegetation and drainage density. For storms of equal intensity and duration, flood magnitudes vary with the amount of precipitation that results in direct runoff, and with the time required for the runoff to move from uplands to channels. Semi-arid areas typically lack the vegetation cover, mature soil profile, and humic layer that generally intercept and store precipitation, enhance infiltration and attenuate storm runoff in areas of moist climate. Compared with those of wetter climates, sparsely vegetated soils of semi-arid areas commonly are thin or absent (i), are deficient in weathering products and organic matter (ii), may exhibit near-surface deposits of relatively impermeable calcrete (iii), and exhibit low rates of infiltration of rainfall due to soil-particle compaction by raindrop impact (iv).

A surficial-geology map (Hunt, 1979) characterizes the conterminous USA east of the 100°W meridian as being generally dominated by saprolitic residuum and colluvium south of the Ohio and Missouri Rivers

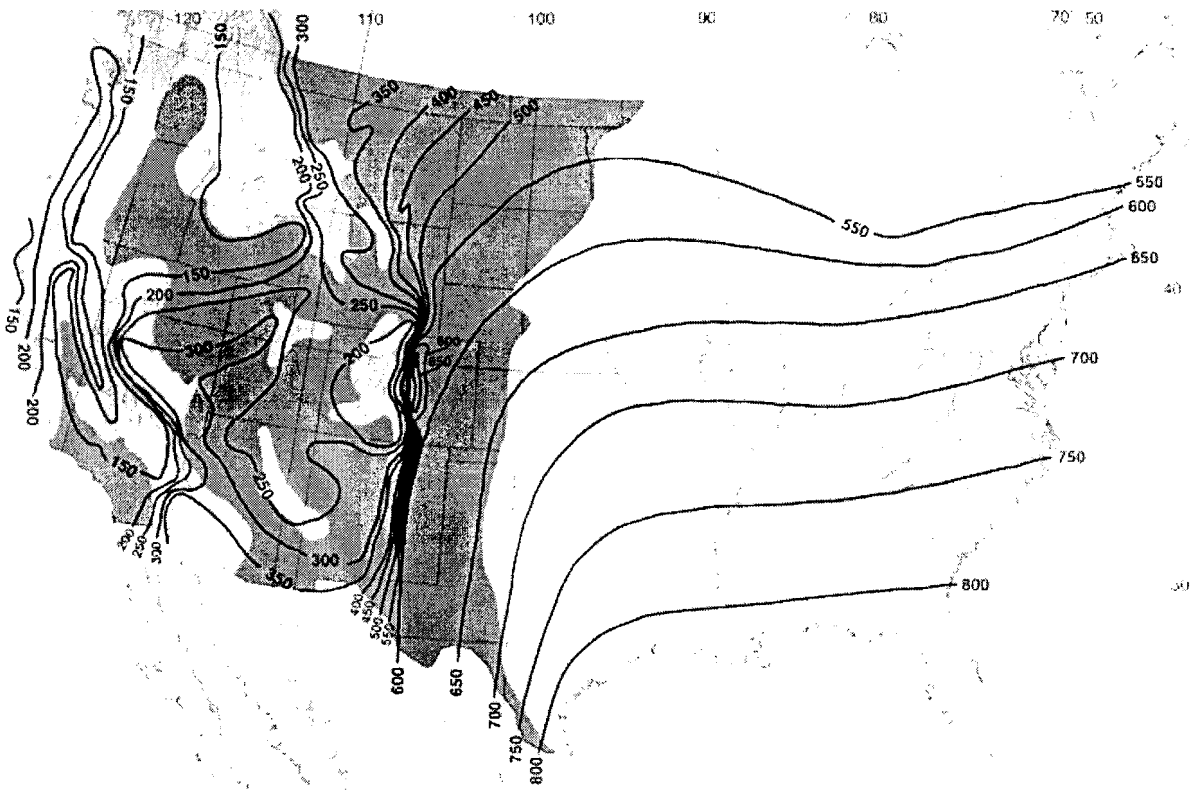

Figure 2. Map of the conterminous USA showing contours, in millimetres, of estimated depths of 6-h, 26-km^2 probable maximum precipitation (simplified from Hansen *et al.*, 1977, 1988, 1994; Schreiner and Riedel, 1978; Corrigan *et al.*, 1998). Shaded (semi-arid) areas (modified from US Geological Survey, 1984) receive 250 to 500 mm mean annual precipitation

(Figure 3; map areas A and C) and by glacial drift and loess north of the rivers (Figure 3; map area B). These materials, in particular saprolite and weathered glacial deposits, are typically clay-rich, stable and resistant to flood erosion. In contrast, surficial material west of the 100°W meridian is dominated by carbonate-rich valley fill and fan deposits, colluvium, sandy stream alluvium, and thin residuum or bedrock (Figure 3; map area C). Where calcrete is poorly developed, such as many bottomland areas, these deposits are generally poorly resistant to erosion by infrequent runoff events but readily yield runoff from typically bare, compacted soil surfaces. Runoff from upland areas underlain by extensive pedogenic calcrete, however, is often accentuated owing to induration and reduction of permeability by calcrete (Reeves, 1970); formation of well-developed pedogenic calcrete is nearly restricted to areas of semi-arid climate, where infiltrating rainfall is adequate to dissolve near-surface carbonates but inadequate to transport most dissolved solids to depth (D. H. Yaalon, in Reeves, 1976, p. 85).

A measure of expected rainfall runoff is provided by the curve-number concept (Soil Conservation Service, 1985; Ponce and Hawkins, 1996), which is a functional expression of the rainfall–runoff equation (Soil Conservation Service, 1985) but is largely a description of interception and infiltration of rainfall. Curve numbers are non-dimensional and have a possible range from 0 (no runoff) to 100 (total runoff). Values of curve number have been measured during numerous field investigations, leading to the development of nomographs and tables from which curve number can be estimated using required input data of soil type,

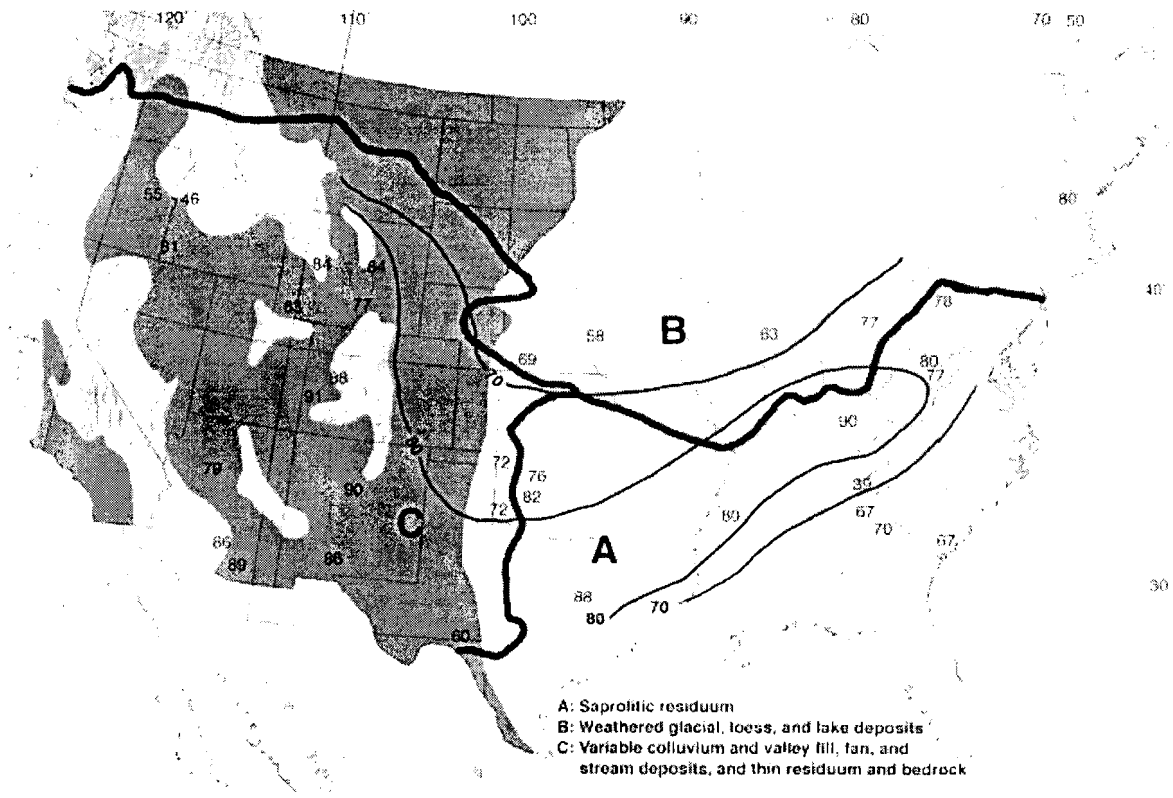

Figure 3. Map of the conterminous USA showing regional patterns of surficial geology (simplified from Hunt, 1979), measured curve numbers for natural conditions (in some cases adjusted) at specific study sites (compiled from various sources by DeAnne Rietz, University of Arizona, personal communication, 1999), and contours of curve-number variation, in selected 10-unit intervals, suggesting possible zones of curve-number variation in non-mountainous areas. Shaded (semi-arid) areas (modified from US Geological Survey, 1984) receive 250 to 500 mm mean annual precipitation

vegetation type and density, land use, and soil-moisture condition (Hawkins, 1980). High curve-number values, over about 80, are indicative of substantial storm runoff, whereas values less than 50 suggest that most storm precipitation infiltrates or replenishes soil moisture.

Figure 3 includes interpretations of curve-number variation for undisturbed watersheds across parts of the conterminous USA. The interpretations are based on measured curve-number data compiled by DeAnne Rietz (University of Arizona, personal communication, 1999) for 34 study sites and watersheds of the USA (Table I); in some instances, multiple measured values of curve number from the same site were averaged and others were adjusted to account for land use. Interpolations resulting in interpreted curve-number contours are based partially on surficial geology and distribution of soil types. Curve numbers are typically less than 60 in forested areas that effectively intercept rainfall and may store much of it in surface litter, duff and organic soil horizons. Curve numbers from watersheds of the semi-arid American West are variable but generally may be 85 and higher in sparsely vegetated, small- to medium-sized basins underlain by bedrock or calcrete-rich soil (Hawkins, 1993).

Another landscape parameter that is, in part, a function of infiltration capacity of a surface is drainage density — the ratio of the combined stream lengths of a drainage basin to the basin area (km/km^2).

Table I. List of measured curve numbers from study sites of the USA

Site	State	Curve number	Comments
Beaver Creek	Arizona	79	Pine–juniper forest
Tombstone	Arizona	89	Grassland
Tucson	Arizona	86	Desert brush
Boco Mountain	Colorado	88	Sagebrush
Badger Wash	Colorado	91	Desert shrub
Watkinsville	Georgia	70	Pasture; adjusted to natural cover
Whitehall	Georgia	67	Pine–hardwood forest
Treynor	Iowa	58	Pasture; adjusted to natural cover
Reynolds	Idaho	81	Grazed; adjusted to natural cover
Silver Creek	Idaho	46	Forest; average of several sites
Lafayette	Indiana	63	Meadow; average of two sites
Berea	Kentucky	90	Oak–hickory forest
Oxford	Mississippi	80	Cropland; adjusted to natural cover
Coweeta	North Carolina	39	Forest
Hastings	Nebraska	69	Meadow
Hubbard Brook	New Hampshire	80	Northern hardwood forest
Albuquerque	New Mexico	90	Range; average of two sites
Dona Ana County	New Mexico	86	Range; average of four sites
Coshocton	Ohio	77	Meadow; average of seven sites
Chickashaw	Oklahoma	72	Pasture; adjusted to natural cover
Guthrie	Oklahoma	82	Meadow; average of three sites
Cherokee	Oklahoma	72	Cropland; adjusted to natural cover
Stillwater	Oklahoma	76	Pasture; adjusted to natural cover
Big Creek	Oregon	55	Forest
Leading Ridge	Pennsylvania	78	Oak–hickory forest
Charleston	South Carolina	67	Pine–hardwood forest
Riesel	Texas	88	Meadow
Sonora	Texas	60	Grazed; adjusted to natural cover
Moorefield	West Virginia	77	Pasture; adjusted to natural cover
Fernow	West Virginia	80	Northern hardwood forest
Eastern	Wyoming	84	Grazed; adjusted to natural cover
Southeastern	Wyoming	77	Grazed; adjusted to natural cover
Central	Wyoming	84	Grazed; adjusted to natural cover
Southwestern	Wyoming	63	Grazed; adjusted to natural cover

Drainage density (based here on mapped channel lengths, regardless of streamflow frequency) is indicative of the facility by which excess rainfall moves from an upland surface; high values of drainage density are suggestive of high transfer efficiency and, therefore, of low infiltration and transmission loss and high peak discharge (Schumm, 1977). More specifically, numeric values of drainage density reflect, as do values of erosion, sediment discharge and curve number, the availability of precipitation to enter the stream network, and thus are functions of land-surface slope, soil development (and geology), and other climate-related variables of topography and vegetation (Gregory and Gardiner, 1975). The highest values of drainage density occur in semi-arid areas of well dissected, fine-grained rocks 'where surface runoff rates are high owing to high rainfall intensities over surfaces with sparse vegetation cover and limited soil development.' (Knighton, 1984, p. 20).

Gregory and Gardiner (1975) compiled previously published data of drainage density and climate of 13 areas world-wide, and grouped the data according to a logarithmic index relating basin areas and cumulative channel lengths. Values from arid climates, as defined here, were unavailable, but are typically regarded as low (less than 4 km/km^2) owing to limited amounts of rainfall available for runoff (Knighton, 1984). Values of drainage density in areas of semi-arid climate, 250 to 500 mm, averaged about 10 km/km^2, indicative of maximum event runoff, short distance to drainageways, and a high potential for high-magnitude discharge (Knighton, 1984). With increasing mean precipitation, average values of the drainage-density data declined from the peak values of semi-arid areas to about 7 km/km^2 in subtropical areas and 2 km/km^2 in humid extratropical areas (Gregory and Gardiner, 1975). These decreased drainage-density values appear to be results of increased infiltration, storage of water as soil moisture, and attenuation of maximum runoff to drainageways by vegetation and other biological processes (Gregory and Gardiner, 1975; Knighton, 1984). Variation in drainage density with mean annual precipitation resembles equivalent relationships between sediment yield and precipitation (Langbein and Schumm, 1958; Wilson, 1973); combined, the two relationships indicate that the likelihood for rapid movement of water and sediment is best developed from landscapes of semi-arid areas.

The effects of soil-particle packing, or crusting, by direct raindrop impact where vegetation is lacking has been shown to reduce infiltration rates into silt-loam soils by up to 95% (Mahamad, 1985). Rainsplash impact on bare soil surfaces not only results in particle detachment, erosion and sediment transport, but also causes formation of a relatively impermeable crust-seal of about 0·1 mm thickness (Duley, 1939; Morin *et al.*, 1989). In arid and semi-arid areas of deficient plant cover, and dependent on soil type, chemistry and condition (such as soil disturbance by bioturbation), and the duration, intensity and kinetic energy of the rainfall, infiltration rates into rain-compacted soils are normally 3 to 20% of those expected for an uncompacted condition (Duley, 1939; Morin and Benyamini, 1977; Morin *et al.*, 1989). The combined effects of generally poorly developed silty soils, sparse vegetation and decay products, widespread occurrence of calcrete and soil-surface compaction by raindrop impact explain the high values of curve number and drainage density observed in uplands of semi-arid areas.

FLOODS AND FLOOD POTENTIAL

Flood potential, based on a flash-flood magnitude index, was evaluated for the USA by Beard (1975). The index, which was developed using streamflow records of at least 20 years from about 2900 gauge sites on unregulated streams, is expressed as a ratio between logarithms of unusual and common flood discharges adjusted for the effects that zero flows and low flows have on flood-frequency analyses. Thus, the tendency for outlier data of high flood peaks and low flows to yield high index values was minimized. Beard (1975) assumed that almost all flash flooding is the result of intense, short-duration rainfalls and asserted that 'flash floods are most common in the arid and semi-arid regions of the west and southwest.' (p. 1) owing to meteorological and physiographic conditions that are conducive to the development of intense thunderstorms.

A composite (Figure 4) of regional maps presented by Beard (1975) represents flash-flood potential for basins of the conterminous USA less than 2600 km^2 in area. The map of flash-flood magnitude index indicates that east of central Texas and generally the 100°W meridian, few areas are subject to significant flash-flood hazard. West of the 100°W meridian, the potential for flash flood is high from central Texas through to coastal parts of southern California and in the plains of Colorado and Kansas northward through eastern Montana and North Dakota into Saskatchewan, Canada (Figure 4). In general, index values of flash-flood magnitude range from 0·5 to 0·8 in semi-arid areas near moisture sources and from 0·3 to 0·6 in semi-arid areas more removed from moisture sources; parts of the USA receiving more than 500 mm mean annual precipitation mostly have values of 0·2 to 0·5 (Figure 4).

Envelope curves for potential maximum floodflows in basin areas up to 26 000 km^2 were compiled by Crippen and Bue (1977) for 17 flood regions of the conterminous USA. The curves, developed from thousands of gauge-site records of the US Geological Survey, suggest that the greatest floodflow potential for basins of 1000 km^2 occurs in Basin and Range channels of south-eastern California, Nevada, western Utah, southern Arizona and western New Mexico (nearly 20 $m^3/s/km^2$), and in streams of Texas draining directly to the Gulf of Mexico (about 19 $m^3/s/km^2$). Elsewhere, the potential maximum floodflows identified by Crippen and Bue (1977) for basins of 1000 km^2 range from about 1·4 $m^3/s/km^2$ in the parts of the Colorado Plateau of northern Arizona and Utah, to 4 to 9 $m^3/s/km^2$ east of the Mississippi River, to 10 to 12 $m^3/s/km^2$ in much of California, Oregon and Washington, and the plains of eastern New Mexico, Colorado, Kansas and Nebraska.

At an areal scale of 26 km^2, similar to areas of convective thunderstorms common in the semi-arid American West, maximum unit floodflows range from 9 $m^3/s/km^2$ in the Colorado Plateau to nearly 60 $m^3/s/km^2$ in the Great Basin and southern parts of California, Arizona and New Mexico. High values of maximum floodflow, 50 to 55 $m^3/s/km^2$, also occur in the southern Appalachian Mountains, the southern

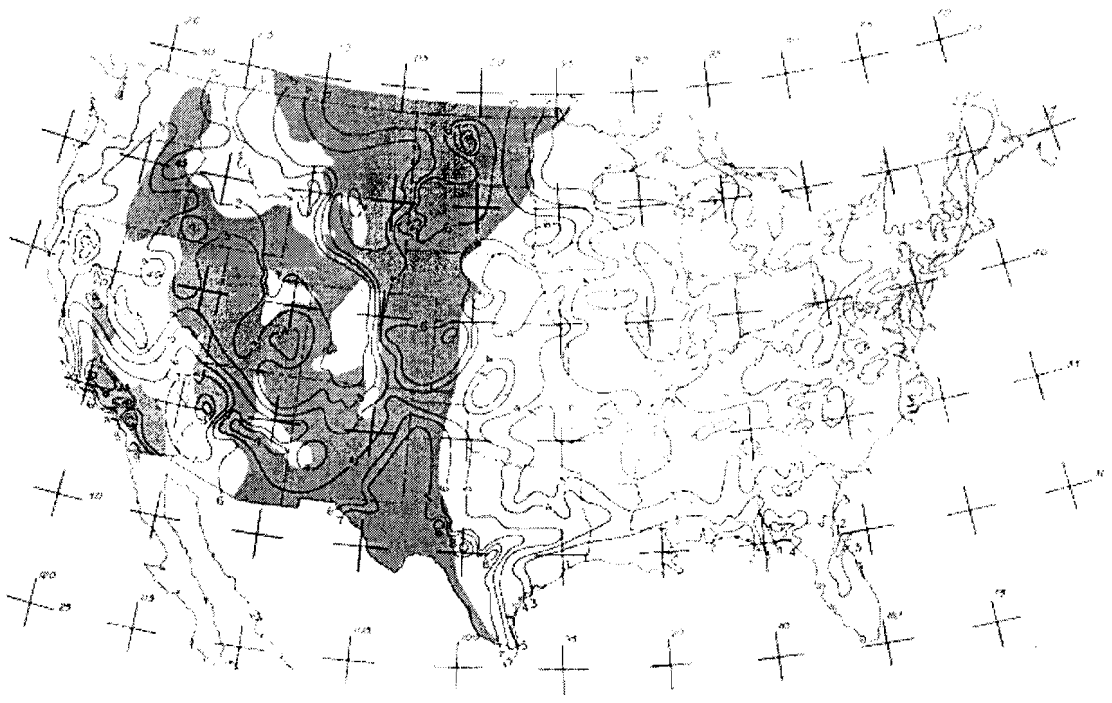

Figure 4. Map of the conterminous USA showing regional variation of flash-flood magnitude index (modified from Beard, 1975)

and central Great Plains, the Gulf Coast of Texas, and portions of the Pacific Northwest (Crippen and Bue, 1977). Maximum (estimated) floodflows from 26-km^2 basins of other parts of the USA generally vary from 20 to 40 $m^3/s/km^2$. Although this pattern of areal variation for floodflows from 26-km^2 basins does not duplicate the variation in PMP values mapped by Hansen *et al.* (1977, 1988, 1994), Schreiner and Riedel (1978) and Corrigan *et al.* (1998), the highest values predicted for flood discharge and PMP mostly are in the same areas, have little or no correlation with mean precipitation, and are most typical of semi-arid climates.

The analysis of Beard (1975) also indicates that the greatest potential for infrequent and rare flooding in the USA does not occur in areas of greatest rainfall but from south-eastern Texas across southern New Mexico and Arizona into south-eastern California and much of southern Nevada. Combined with information on soil characteristics, especially as indicated by curve numbers, the maps and curves of Beard (1975) and Crippen and Bue (1977) seem expectable.

Additional confirmation of the tendency for high flash-flood potential in semi-arid areas was provided by Costa (1987), who compiled a global list, with emphasis on the USA and China, of the largest known rainfall–runoff floods for drainage-basin areas up to about 5×10^6 km^2. The largest measured floods, worldwide, in basins greater than 1000 km^2 occurred mostly in areas of average annual basin precipitation over 1000 mm; for very large basins, such as those of the Mississippi and Amazon Rivers, average basin precipitation exceeded 1500 mm per year. Although weather data of China may be unreliable, rare floods in drainage basins of less than about 200 km^2 generally have occurred in areas of less than 700 mm mean annual precipitation, whereas extreme flooding in basins larger than 200 km^2 has been mostly in areas of 1000 mm or more mean annual precipitation.

Table II lists 30 sites, with drainage-basin areas and approximate values of mean annual precipitation, of rare floods of the conterminous USA compiled by Crippen and Bue (1977) and Costa (1987). Of the 23 basins less than 1000 km^2 in area, nine have a semi-arid climate of 250 to 500 mm mean annual precipitation and 17 have less than 750 mm annual rainfall. The other seven flood sites are in basins larger than 1000 km^2 and have climates ranging from subhumid to humid. The computed peak flow at one site, Bronco Creek near Wickieup, AZ (Table II), may be invalid owing to unreliable field data.

DISCUSSION

Floods in any natural environment are recurrent events that help shape bottomlands and are necessary stimulants for a vigorous and diverse biotic system. Perennial streams of most regions typically overflow on to adjacent areas of floodplain once every year or two, the basis upon which the floodplain is generally defined. Rare, or highly infrequent, floods may inundate most or all of the bottomland features, including normally dry alluvial terraces. In regions of deficient precipitation, ephemeral streamflow may spill on to adjacent bottomlands much less frequently. Regardless of climate, however, rare runoff events may be agents of extreme and nearly instantaneous landscape change, and that change may have extensive effects on biota, especially vegetation.

The implications of the relationships and possible disparity between extreme rainfalls and rare floods are relevant to the concept of geomorphological effectiveness. Relationships proposed by Wolman and Miller (1960) included the concept that the mean annual flood, relative to flow frequency and sediment discharge, is the flow that most effectively accomplishes geomorphological work. Subsequent papers demonstrated that the frequency of a dominant discharge may be influenced by climatic and geomorphological characteristics other than the mean annual flood. Thus, attention turned to arid and semi-arid areas, where floods and measurable landscape change may occur only during extreme rainfall–runoff events.

Wolman and Gerson (1978) argued that whereas mean precipitation and mean discharge for a given area are lower in semi-arid than in humid regions, the intensity of extreme precipitation and the magnitude of extreme unit discharges are independent of aridity. Although they discussed the decrease in precipitation and unit discharge with increasing basin area, they did not discuss local or regional variation in maximum precipitation or unit discharge. Subsequent investigators of both semi-arid and humid regions have stressed

Table II. List of extreme-flood sites used by Crippen and Bue (1977) to develop a nationwide envelope curve[1] and by Costa (1987) to identify the largest rainfall-runoff floods of the (conterminous) United States[2]

Flood site	Area (km^2) (mm)	Mean annual precipitation (mm)
Praire Creek tributary near Tyler, TX[2]	0·0065	1270
Unnamed tributary near Riesel, TX[2]	0·0124	619
Brawley Wash tributary near Tucson, AZ[2]	0·021	290
Castle Creek tributary 2 near Rockford, SD[2]	0·05	480
Cimarron Creek near Cimarron, NM[2]	0·13	325
Wenatchee River tributary near Monitor, WA[2]	0·39	690
Lahonton Reservoir tributary near Silver Springs, NV[2]	0·57	122
Lake Pinto Creek tributary near Newcastle, UT[1]	0·78	263
Boney Draw at Rock Port, MO[1]	1·97	855
Humboldt River tributary near Rye Ranch, NV[2]	2·30	205
Big Creek near Waynesville, NC[1]	3·42	1190
Stratton Creek near Washta, IA[1]	4·92	735
Lane Canyon near Nolin, OR[2]	13·1	935
El Rancho Arroyo near Pojoaque, NM[1]	17·4	350
Meyers Canyon near Mitchell, OR[1,2]	32·9	290
Bronco Creek near Wickieup, AZ[1,2]	49·2	263
Eldorado Canyon at Nelson Landing, NV[2]	59·3	205
North Fork Hubbard Creek near Albany, TX[2]	102	706
Jimmy Camp Creek near Fountain, CO[1,2]	141	380
Wilson Creek near Adako, NC[1]	168	1200
Maintrail Creek near Loma Alta, TX[2]	195	659
Seco Creek 18 km above D'Hanis, TX[1,2]	368	655
East Bijou Creek at Deer Trail, CO[1]	782	361
West Nueces River near Kickapoo Springs, TX[1,2]	1040	650
West Nueces River near Brackettville, TX[1]	1810	516
Bijou Creek near Wiggins, CO[1]	3400	570
Eel River at Scotia, CA[1,2]	8060	1220
Susquehanna River at Conowingo, MD[2]	70 200	1150
Ohio River at Metropolis, IL[2]	526 000	1220
Mississippi River near Arkansas City, AR[2]	2 930 000	880

this variability and its underlying causes, including proximity to a moisture source, mesoscale circulation patterns, and orographic lifting (Maddox *et al.*, 1980; Hirschboeck, 1988; Smith *et al.*, 1996).

The variables assumed responsible for most rare and catastrophic floods can be simplified as (i) weather and climate, including particularly the intensity and duration of precipitation events, (ii) climatic controls on upland soils and vegetation and hence runoff characteristics, and (iii) climatic controls on bottomland alluvium and vegetation. Elements of climate are frequencies of intense storms and magnitudes of extreme precipitation events. The effects of upland soils and vegetation on flood potential are suggested by infiltration capacity and overland flow, whereas alluvium and vegetation of bottomlands are measures of resistance to landscape change by intense flooding. These bulk variables suggest flood classes and permit comparison of infrequent floods in semi-arid areas with those of other climatic zones.

Floods that cause significant geomorphological change are those 'that generate high values of peak stream power per unit area, and have moderate to long duration' (Costa and O'Connor, 1995, p. 54). Long-duration precipitation of meso-scale storms, common in the moist east and north-west, but less typical of the drier parts of western North America, may yield substantial runoff of long duration but not high values of stream power. Conversely, hurricanes, cyclones and convective storms that generate intense precipitation for a sustained period in drainage basins generally less than 1000 km^2 in area may occur in a wide range of climatic zones of temperate North America.

Bottomland vegetation affects flood magnitude, flood duration and the onset of landform change by flooding in at least two primary ways:

(1) by providing root cohesiveness to otherwise unconsolidated alluvium, thereby impeding near-surface disturbance;
(2) by providing resistance to flow that generally reduces velocity and unit stream power, and increases flood duration.

Where bottomland vegetation is lush and dominated by grasses, forbs and shrubs, resistance to flood erosion owing to the stabilizing effect of stems and roots may be high. Large, closely spaced trees in bottomlands of humid areas with shallow groundwater also provide root-protection to near-surface alluvium and, combined with dead stems and branches that become caught against living trees, may significantly increase resistance to flow during floods (Arcement and Schneider, 1989). Channel and riparian-zone vegetation of semi-arid areas often is dominated by phreatophytes, such as salt cedar, willow and cottonwood, where streamflow is perennial or intermittent, and by xerophytes, such as mesquite and paloverde, where streamflow is ephemeral. Many phreatophytes have flexible stems that bend with the flow during floods and thus do not increase flow resistance dramatically; stems of riparian-zone forests of mesquite and paloverde often are not flexible enough to withstand large floods and may contribute to landform damage if toppled during infrequent floods (Phillips *et al.*, 1998).

The energy of floods, given these conditions, may be largely dissipated owing to overbank flow and low unit stream power. The greatest amounts of geomorphological work accomplished occur in watersheds large enough that sediment is stored as bottomland alluvium and small enough that single storms may provide runoff from much or all of the basin. Thus, the potential for intense runoff and flood damage is greatest where vegetation offers little flow resistance and curve numbers are high, conditions most likely met along channels of semi-arid areas.

SUMMARY

Evidence of regional variation in extreme precipitation suggests that magnitudes of intense precipitation in some semi-arid areas of the western USA have been less than in much of the moist central and southern USA, but were generally similar to or greater than those of other sections of the north and west USA. Regardless of mean precipitation, intense rainfall in semi-arid areas appears more closely related to patterns of atmospheric circulation and sources of moisture than to large-scale variation in mean precipitation (Hansen *et al.*, 1988, 1994). Specifically, magnitudes of 100-year, 6-h rainfall events and computed magnitudes of PMP for 26-km^2 watersheds indicate that extreme precipitation in semi-arid areas such as western Texas and central and southern Arizona is generally lower than that in humid areas; the contrast is stronger for large watersheds. Local variation in extreme rainfall is great in both humid and semi-arid regions. The orographic, or topographic, influence on PMP is substantial in most high-relief areas of temperate North America (Hansen *et al.*, 1994), but is not everywhere observed in the semi-arid USA (Osborn, 1981).

Although extreme rainfall in semi-arid areas tends to be lower than in humid areas, peaks of infrequent floods are typically greater in semi-arid areas. Evidence includes records of maximum floodflows, as assembled and graphed for the USA by Crippen and Bue (1977), flash-flood potentials generated by Beard (1975), and a compilation of the largest well documented rainfall–runoff floods of the USA (Costa, 1987). This evidence shows that many of the greatest recorded unit floodflows have occurred in drainage basins of less than 1000 km^2 in semi-arid areas.

The disparity between rainfall–runoff patterns for extreme events in semi-arid and other areas is explained in large part by differences in climatic and geological influences on vegetation and soil characteristics. Near-surface calcrete, which generally forms near the surface in semi-arid areas, indurates sediment of the upper soil profile and thereby increases runoff by reducing infiltration (Reeves, 1970). Vegetation cover in semi-arid

areas is generally incomplete, resulting in limited interception of rainfall and little or no absorption of rainfall and runoff by surface and soil organic material. Where vegetation is absent, rainsplash impact typically results in soil compaction, reduced infiltration and thus relatively high runoff (Duley, 1939; Mahamad, 1985; Morin *et al.*, 1989).

Generally higher values of curve number and drainage density in semi-arid areas than elsewhere indicate that runoff potential is enhanced where soils are poorly developed and vegetation is sparse. Although variable with local soil conditions, values of curve number tend to be highest, indicating high runoff, in semi-arid areas (Hawkins, 1993). Drainage density, a measure of runoff proportion and high peak discharge, is generally highest in semi-arid areas (Gregory and Gardiner, 1975; Schumm, 1977; Knighton, 1984).

Extreme flooding is most likely in areas close to a source of warm moist air and that have (i) enough relief to increase precipitation orographically, (ii) thin or relatively impermeable soils, (iii) sparse vegetation, and (iv) high land-surface slopes. Although moisture patterns are more conducive to extreme rainfall in humid areas, geology, soils, vegetation and drainage density enhance runoff in semi-arid areas. As a result, rare floods and accompanying geomorphological change are more likely in small drainage basins of semi-arid areas than elsewhere, although the variation within climate zones may be greater than that between them.

Data show also that rare floods in drainage basins smaller than 1000 km^2 are more common in semi-arid areas than elsewhere. Owing to soil and vegetation conditions, erosion and thus geomorphological work by rare floods of semi-arid areas typically are greater, if not much greater, than by floods of similar magnitude and duration in wetter areas (Baker, 1977; Costa and O'Connor, 1995). The concept of geomorphological work and its relation to a dominant discharge, which were applied first to work accomplished by mean annual floods of humid landscapes, in semi-arid regions appear to relate to increasingly greater and infrequent rainfall–runoff events as measures of mean precipitation decrease.

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