To Gordon Grant - Best wishes - Andy

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FLOOD HYDROLOGY AND GEOMORPHIC EFFECTIVENESS IN THE CENTRAL APPALACHIANS

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ABSTRACT

This paper compares hydrologic records and geomorphic effects of several historic floods in the central Appalachian region of the eastern United States. The most recent of these, occurring in November 1985, was the largest ever recorded in West Virginia, with peak discharges exceeding the estimated 500-year discharge at eight of eleven stations in the South Branch Potomac River and Cheat River basins. Geomorphic effects on valley floors included some of the most severe and widespread floodplain erosion ever documented and exceeded anything seen in previous floods, even though comparable or greater rainfall and unit discharge have been observed several times in the region over the past 50 years. Comparison of discharge–drainage area plots suggests that the intensity and spatial scale of the November 1985 flood were optimal for erosion of valley floors along the three forks of the South Branch Potomac River. However, when a larger geographic area is considered, rainfall totals and discharge–drainage area relationships are insufficient predictors of geomorphic effectiveness for valley floors at drainage areas of 250 to 2500 km^2.

Unit stream power was calculated for the largest recorded flood discharge at 46 stations in the central Appalachians. Maximum values of unit stream power are developed in bedrock canyons, where the boundaries are resistant to erosion and the flow cross-section cannot adjust its width to accommodate extreme discharges. The largest value was 2570 $W m^{-2}$; record discharge at most stations was associated with unit stream power values less than 300 $W m^{-2}$, but more stations exceeded this value in the November 1985 flood than in the other floods that were analysed. Unit stream power at indirect discharge measurement sites near areas experiencing severe erosion in this and other central Appalachian floods generally exceeded 300 $W m^{-2}$; reach-average values of 200–500 $W m^{-2}$ were calculated for valleys where erosion damage was most widespread. Despite these general trends, unit stream power is not a reliable predictor of geomorphic change for individual sites. Improved understanding of flood impacts will require more detailed investigation of interactions between local site characteristics and patterns of flood flow over the valley floor.

KEY WORDS Floods Erosion Geomorphic effectiveness Floodplains

INTRODUCTION

Increasing attention has been focused over the past three decades on the role of rare, large-magnitude floods as geomorphic agents. Wolman and Miller (1960) evaluated the relative importance of such floods, using cumulative sediment transport over time as a measure of geomorphic work, and concluded that their significance was outweighed by frequently recurring events of lower magnitude. The alternative concept of geomorphic effectiveness, as described by Wolman and Gerson (1978) and reviewed by Brunsden and Thornes (1979), Newson (1980), and Kochel (1988), refers to the ability of an event to alter landforms and to the relative persistence of the altered landforms under the influence of processes tending to restore the landscape to its previous condition. Some of the larger floods documented in the literature have had little lasting impact on the landscape (Wolman and Eiler, 1958; Costa, 1974; Gupta and Fox, 1974; Gardner, 1977; Moss and Kochel, 1978), but the published record is replete with examples of modern floods that have

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modified river valleys in important ways (Hack and Goodlett, 1960; Stewart and LaMarche, 1967; Williams and Guy, 1973; Anderson and Calver, 1977; Baker, 1977; Costa, 1978; Shroba *et al.*, 1979; Newson, 1980; Johnson, 1983; Sullivan, 1983; Gupta, 1983; Nanson, 1986; Baker and Pickup, 1987; Calvet, 1987; Clark *et al.*, 1987; Kochel *et al.*, 1987; Osterkamp and Costa, 1987; Webb, 1987; Hickin and Sichingabula, 1988; Kochel, 1988; Kresan, 1988; Ritter, 1988). There is little doubt that, in some environments, large, rare floods leave a lasting, and perhaps even a dominant, imprint on the landscape.

The literature allows some generalization regarding factors that influence geomorphic effectiveness: (1) The effects of large floods tend to persist for longer periods in semiarid and arid environments than in most humid environments (Wolman and Gerson, 1978). Streams in these environments that have sand beds and sandy, noncohesive sediments making up the river banks and floodplain are particularly susceptible to periodic episodes of catastrophic channel widening (Schumm and Lichty, 1963; Burkham, 1972; Osterkamp and Costa, 1987). (2) Although channel recovery processes are more active in humid environments, catastrophic floods still may have a dominant influence on channel and valley form if they recur frequently enough. This is particularly true in tropical climates (Gupta, 1983, 1988). (3) Regardless of climate, the effectiveness of large floods is greater in steep, narrow, rock-bound, or partially confined valleys of mountainous regions than in regions characterized by broad, flat alluvial valleys. Nanson (1986) characterized floodplains along mountain rivers affected by alternating episodes of aggradation and stripping as disequilibrium landforms. Availability of coarse bedload to be transported is also an important determinant of effectiveness (Baker, 1977, 1984; Kochel, 1988); the deposits left behind in river channels and on floodplains often are too coarse to be moved by subsequent events of lesser magnitude.

Despite the abundance of literature describing geomorphically effective floods, there are relatively few studies (Wolman and Eiler, 1958; Nanson and Hean, 1985; Baker and Costa, 1987) that attempt to provide quantitative information on what threshold conditions, if any, distinguish floods that are effective from those that are not. Because the flood history of the central Appalachian region of the eastern U.S. includes several well-documented events encompassing a broad range of geomorphic impacts, comparative analysis of floods occurring in this region allows some progress toward this goal. The most recent major flood in the region occurred in November 1985 and caused fluvial erosion and reworking of channel and floodplain sediments on a scale not previously observed (Miller and Parkinson, in preparation). The purpose of this paper is to explore the question of how and why the geomorphic impacts of this event differed from those of previous events. In the following pages I briefly describe the 1985 flood, compare it with other large floods that have occurred in the region, and examine the available evidence to determine whether threshold relations between flood characteristics and geomorphic impacts can be identified. In the absence of evidence about long-term recovery processes, an event is regarded here as effective if it causes substantial alteration of existing landforms in ways that cannot be accomplished by events of lesser magnitude, and if those alterations cannot be reversed by the steady operation of normal geologic processes over a period of at least several years. The primary focus is on modification of valley floors; slope processes are considered in greater detail by Jacobson *et al.* (1989a) and by Jacobson *et al.* (1989b).

THE NOVEMBER 1985 FLOOD

The storm of November 3–5, 1985 resulted from the confluence of the remnants of Tropical Storm Juan, which made landfall over the Gulf Coast of the U.S., with two other low-pressure cells moving northward over the southeastern U.S. A blocking high-pressure system off the Atlantic coast pumped moisture into the system and slowed the migration of the storm as it moved northeast through the central Appalachians of West Virginia and Virginia. This was a long-period, moderate-intensity meteorological event with multiple precipitation cells covering a broad area; maximum recorded 48-hour precipitation was 310 mm, and the maximum recorded intensity was 38·3 mm hr^{-1} (Jacobson *et al.*, 1989a; Clark *et al.*, 1987).

The resulting flood of November 4–6, 1985 affected portions of West Virginia, Pennsylvania, Maryland, and Virginia. In the South Branch Potomac River and Cheat River basins of West Virginia (Figure 1) this was the largest flood ever recorded; the ratio of peak discharge to the previous record discharge at eleven U.S. Geological Survey gauging stations in these two basins ranged from 0·98 to 3·84, with an average value of 2·23.

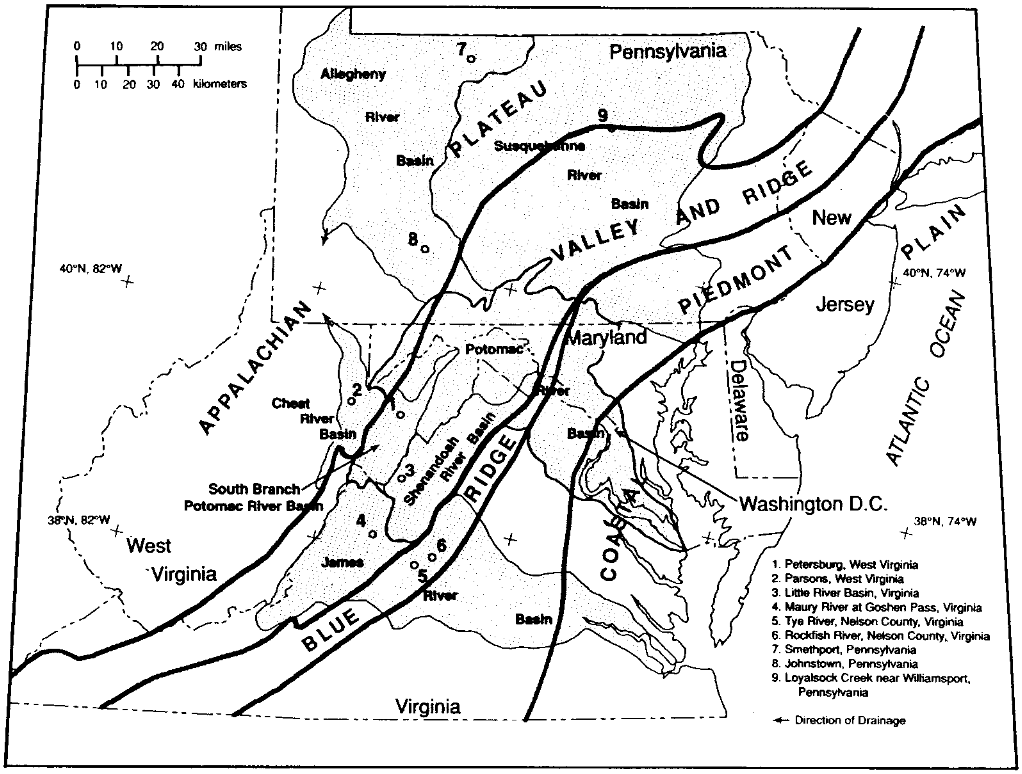

Figure 1. Location map for the central Appalachians showing physiographic provinces, river basins and specific locations referred to in the text

All but one of these stations had record lengths exceeding 40 years, with historical flood information extending back at least to 1877. Recurrence intervals calculated following the guidelines in U.S. Water Resources Council Bulletin 17B (Interagency Advisory Committee on Water Data, 1982) exceed 100 years at all eleven gauge sites and also exceed 500 years at eight of these sites.

Slope failures were widespread in the South Branch Potomac River drainage (Jacobson *et al.*, 1989a) and flood power was sufficient to cause stripping and dissection of alluvial bottomlands at many sites along several hundred kilometres of channel in the South Branch basin and at many sites in the Cheat River basin. Erosion and deposition features were documented in aerial photographs taken for the West Virginia Department of Highways during the week after the flood. Bottomland erosion features observed in the study area include longitudinal grooves; scour marks; floodplain stripping and channel widening; floodplain chutes with maximum dimensions up to 800 m long, 50 m wide, and 2–3 m deep; dissection by networks of anastomosing channels; and jet-shaped erosion forms up to 200 m wide (Figure 2). An extended discussion of the morphology and distribution of flood-generated landforms may be found in Miller and Parkinson (in preparation). Additional documentation of the geomorphic effects of this flood in the Potomac and adjacent river basins is provided by Scatena (1986), Jacobson *et al.* (1989a), Kochel *et al.* (1987), and Kochel (1988).

COMPARISON WITH OTHER FLOODS

Although some of the peak discharge values estimated for the November 1985 flood are among the highest ever recorded for comparable drainage areas in the central Appalachians, they are considerably lower than the values defining the envelope curve published by Costa (1987) for the largest recorded rainfall–runoff floods in the United States (Figure 3). Furthermore November 1985 flood peaks are not unprecedented in the central Appalachian region, as the following analysis indicates.

Figure 2. Photograph showing catastrophic erosion of the valley floor along the North Fork South Branch Potomac River at Cabins, West Virginia (upstream of Petersburg, location indicated in Figure 1). The preflood river channel is in the foreground; note that trees along the river bank are intact and a bridge crossing the channel survived the flood. The eroded area at the centre of the photo has a maximum width of more than 200 m; no channel was present in this location before November 1985. Peak discharge in the constriction upstream of this site was 2550 $m^3 s^{-1}$ and unit stream power was 550 W m^{-2}. Flow is from right to left. Photograph provided courtesy of the Moorefield Examiner

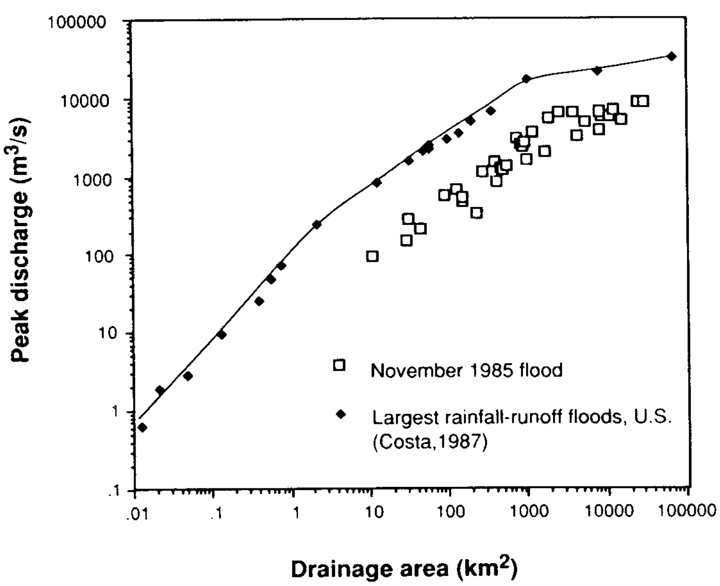

Figure 3. Plot of peak discharge against drainage area for selected stations affected by the November 1985 flood, compared with the envelope curve representing maximum values recorded for rainfall–runoff floods in the United States

Figure 4 is a plot of peak discharge against drainage area with envelope curves projected for three floods: the November 1985 flood; the June 1949 flood, which was the previous flood of record in the South Branch Potomac River basin and at some stations in the Shenandoah River basin; and the March 1936 flood, which remains the flood of record on the main stem Potomac River near Washington. The envelope curves enclose the extreme or limiting values in the discharge–drainage area relationship for each event. Meteorological and hydrologic characteristics and geomorphic impacts are compared in Table I.

Despite the differences between these three events, what is most interesting about the envelope curves shown in Figure 4 is their similarity. In each case there is a steep rising limb; peak discharge increases with increasing size of area contributing runoff, and the rising limbs of all three curves have similar slopes. On each curve there is a break in slope at a drainage area above which little or no increase in peak discharge is observed; the flat limb of the curve represents translation and possibly some attenuation of the flood wave, with only modest contributions from tributary basins outside the area of greatest precipitation. Differences between the three curves are determined by intensity and spatial scale of the runoff-generating event: the rising limb of the storm event with the greatest precipitation intensities is located farthest to the left, whereas the slope break on the curve of the storm event with the largest contributing area is located farthest to the right.

The June 1949 flood was caused by an intense convectional storm. It had the highest recorded precipitation intensity and the smallest contributing area among the three events described in Figure 4; its discharge envelope curve lies above the curves for the other two storms at small drainage areas but begins to level off at drainage areas above 200 km^2. Cloudburst floods of this type typically exhibit this pattern, probably as a result of the relatively small spatial scales of the cells of highest precipitation intensity. The location of the slope break on the envelope curve is controlled not only by the size of the area of intense precipitation but by its position with respect to basin boundaries. In the case of the June 1949 flood, the cell of most intense precipitation was split by the divide between the Shenandoah and the South Branch Potomac River basins (Hack and Goodlett, 1960, Figure 23, p. 42). If all of this precipitation had fallen within a single basin, more extreme discharges would have been observed at larger drainage areas. A similar observation was made by Eisenlohr (1952) with respect to the Smethport storm of July 1942 in north–central Pennsylvania.

Geomorphic impacts of the 1949 storm were documented by Hack and Goodlett (1960) for the Little River basin of Virginia and by Stringfield and Smith (1956) for the area around Petersburg, West Virginia (see

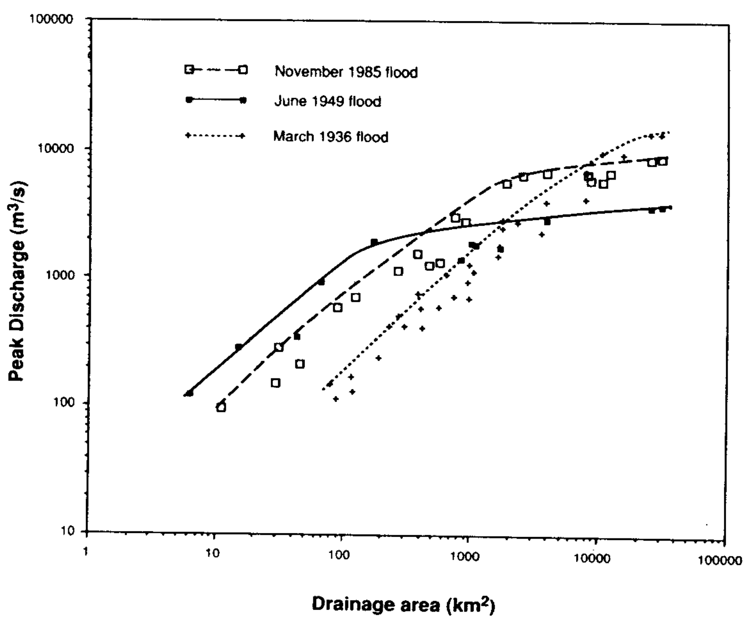

Figure 4. Envelope curves illustrating trends in the relationship between peak discharge and drainage area for three great floods in the Potomac River basin and adjacent drainage basins. Data from Lescinsky (1987), Grover (1937), and Mussey (1950)

Table I. Hydrologic and geomorphic characteristics of historic floods in the central Appalachians

Maximum recorded precipitation for selected duration	Hydrologic characteristics of storm and flood	Geomorphic impacts
March 9–22, 1936—Potomac, James, and upper Ohio River basins (Grover, 1937)		
127 mm in 24 hours 285 mm in 14 days	Low-intensity, long-duration storm; large contributing area, enhanced by frozen ground; record-breaking discharge at large drainage areas	None reported
July 18, 1942—Susquehanna and Allegheny River Basins, centered around Smethport, Pa. (Eisenlohr, 1952)		
909 mm in 12 hours	High-intensity, short-duration storm; extremely high discharge at small drainage areas; limited contributing area	Scattered hillside 'blowouts' and debris avalanches; incision of new channels on forested hillslopes and farm fields drained by low-order tributaries; some tributaries scoured to bedrock; local reworking of larger channels at drainage areas of several hundred km^2
June 17–18, 1949—North River Basin, Va. and South Branch Potomac River Basin, W. Va. (Stringfield and Smith, 1956; Hack and Goodlett, 1960)		
413 mm in 24 hours	High-intensity, short-duration storm; extremely high discharge in small basins; limited contributing area	Hundreds of debris avalanches in two small areas; reworking of river channels and erosion of floodplains in small drainage basins; local diversion of river channels by debris-avalanche deposits in narrow valleys at drainage areas up to 1000 km^2
August 19–20, 1969—Hurricane Camille, Tye and Rockfish River Basins, Nelson County, Va. (Camp and Miller, 1970; Williams and Guy, 1973)		
711 mm in <12 hours	High-intensity, short-duration precipitation delivered by tropical cyclone; small contributing areas, but with enough water to generate record-breaking peaks at drainage areas up to 1500 km^2	Hundreds of debris avalanches associated with reworking of river channels and erosion of floodplains in small headwater basins; localized dissection of floodplains at drainage areas up to 250 km^2; some channel widening at drainage areas up to 1500 km^2
June 21–24, 1972—Susquehanna, Potomac, James, and other river basins draining to eastern seaboard of U.S. (Bailey et al., 1975)		
269 mm in 12 hours 442 mm in 48 hours	Moderate to high-intensity, long-duration precipitation caused by remnants of tropical cyclone combined with extratropical cyclone; extreme discharge peaks in small basins and in large basins; large contributing area	Local widening of stream channels in Piedmont; scattered local examples of channel and floodplain erosion in Valley and Ridge
July 19–20, 1977—Conemaugh River Basin, Pennsylvania		
305 mm in 12 hours	High-intensity, short-duration precipitation; discharge characteristics similar to 1942, 1949 floods	Debris avalanches, local reworking of small valleys (most dramatic impacts downstream of breached dams)
November 3–6, 1985—South Branch Potomac and Cheat River Basins, W. Va. and James River Basin, Va. (Lescinsky, 1987)		
310 mm in 48 hours	Moderate-intensity, long-duration storm generated by combination of tropical and extratropical cyclones stalled by a high-pressure system off the east coast	Thousands of small landslides and 8 debris avalanches in headwater basins tributary to the South Branch Potomac River; widespread reworking of river channels and erosion of floodplains in small drainage basins; catastrophic erosion of river channels and floodplains at drainage areas up to several thousand km^2

Figure 1 for locations). More than a hundred debris avalanches occurred on steep slopes underlain by resistant sandstones in each of these areas. Evidently their spatial distribution was partly a function of local precipitation gradients. Documented fluvial impacts included extensive reworking of valley floors in the Little River basin (drainage area $=65\ km^2$) and along tributaries to the South Branch and North Fork South Branch Potomac Rivers. Erosion of valley floors at drainage areas of several hundred km^2 occurred in the South Branch Potomac River basin primarily at sites where avalanches descended from tributary valleys and blocked flow along the South Branch or the North Fork South Branch Potomac River.

The March 1936 flood (Grover, 1937) was characterized by multiple hydrograph peaks over a two-week period at many stations along the major rivers. Runoff was enhanced by snowmelt and by the presence of frozen ground over much of the area affected; record-breaking discharges at the largest drainage areas were attributable more to the size of the contributing area than to the local intensity of precipitation or magnitude of unit discharge in any one subbasin. In this respect the 1936 flood resembled other major winter–spring floods, which occur under conditions of high antecedent soil moisture or reduced infiltration capacity and are not generally associated with rainfall intensities as high as those observed in connection with summer thunderstorms or tropical depressions.

Despite widespread inundation of river valleys and extensive property damage (Grover, 1937), we have no record to indicate the occurrence of major geomorphic impacts resulting from the 1936 flood. Precipitation rates probably were insufficient to cause significant slope failures, and peak discharges in small and intermediate-size basins apparently were not powerful enough to cause significant channel widening and floodplain erosion. The record-breaking discharge on the main stem Potomac River at Washington, D.C. was accompanied by one of the highest velocities ever measured directly ($6{\cdot}7\ m\ s^{-1}$; Leopold *et al.*, 1964), but this measurement was made at Chain Bridge, a narrow, deep, rock-bound channel section. Elsewhere along the main stem Potomac, most cross-sections are wider and velocities would have been much lower, particularly as overbank flow spread over the floodplain and low terraces flanking the channel. For this reason the March 1936 event and other events of its type are not likely candidates for causing major geomorphic change in the basins discussed here, although such storms do tend to carry very large suspended sediment loads. The downstream effects are primarily depositional and have a negligible effect on the morphology of the channel–floodplain system.

The November 1985 event falls between the other two floods discussed above, both in scale and in intensity. Peak unit discharges at small drainage areas probably were smaller than for the 1949 storm, but the size of the area experiencing extreme values of unit discharge was larger. The envelope curve for the 1985 storm exceeds the curves for both other storms shown in Figure 4 for drainage areas between about 500 km^2 and 7500 km^2.

Geomorphic impacts were in some respects less dramatic and in other respects more dramatic than in 1949. As Jacobson *et al.* (1989a) indicate, peak rainfall intensities in the South Branch Potomac River basin were close to the thresholds required to trigger debris avalanches. Whereas debris avalanches occurring in 1949 delivered large volumes of coarse sediment directly into the main river valleys, all of those occurring in 1985 came to rest well upstream of tributary confluences with the main trunk river. Unlike the June 1949 event, the 1985 storm also generated thousands of smaller slope failures, including shallow slip-flows in regolith and some rotational slumps. The spatial distribution of these other failures was a function of bedrock lithology and precipitation depth (Jacobson *et al.*, 1989a).

Impacts on valley floors were more intense over a larger area than in 1949. Although many headwater and tributary valleys experienced virtually total reworking of the channel and riparian corridor, some of the most impressive cases of valley-floor modification were observed along the major rivers at drainage areas of 250 to 2500 km^2 (Figure 2; also see Miller and Parkinson, in preparation). Much of the coarse material mobilized from tributary valleys by debris avalanche and delivered to the main valley of the South Branch Potomac River in 1949 was left in place by that flood but was removed by the 1985 flood.

Examination of Figure 4 in the context of what is known about the three floods leads to the suggestion that the intensity and spatial scale of the November 1985 event were optimal for the initiation of floodplain erosion in the major valleys. If we adopt the terminology of Newson (1980), the 1985 event may be characterized as both a 'channel' and a 'slope' flood, but was predominantly a 'channel' flood in that the most spectacular fluvial impacts occurred irregardless of sediment inputs from slope failures; whereas the 1949 event was primarily a 'slope' flood. Newson's assessment of the conditions typically associated with these two types of events is not applicable here: in his study area the more intense, shorter-duration storm produced greater channel impacts and modest slope impacts, whereas the moderate-intensity, longer-duration storm with abundant antecedent precipitation produced modest channel changes and had a greater impact on slopes. In the central Appalachians, debris avalanches originating in massive sandstones seem to be favoured by high-intensity precipitation, whereas shallow failures in regolith overlying shale are more likely when pore-water pressures are allowed to build over the course of a long-duration, moderate-intensity precipitation event. The likelihood of debris-avalanche activity also may depend on the amount of time elapsed since the last episode of significant mass wasting at the same location, as the critical conditions for failure can be attained only when a sufficient depth of weathered material has formed in place or accumulated as colluvium in a local depression in the bedrock surface (Newson, 1980; Reneau and Dietrich, 1987; Kochel, 1988; Jacobson *et al.*, 1989b). High-intensity precipitation events usually are smaller in spatial scale and therefore may be less effective than longer-duration moderate-intensity events in modifying channels and floodplains at large drainage areas. The storm of lowest intensity and greatest duration (March 1936) produced the largest absolute discharges but left no lasting imprint on channels or slopes.

Despite the differences among the floods described in Figure 4 and their geomorphic impacts in the Potomac River basin, the discharge–drainage area curve cannot by itself define the threshold conditions for extensive bottomland erosion. This becomes more readily apparent if we compare the 1985 storm and flood with extreme events that have affected other basins in the central Appalachians and the mid-Atlantic states (Table I). Depth–area curves for maximum 2-day precipitation in four historic events (Figure 5) show that the 1985 storm delivered less total precipitation with lower intensity than the other three storms for areas up to several thousand km^2. Two of these other storms (Tropical Storm Agnes in June 1972 and Hurricane Camille in August 1969) were tropical cyclones and one (the Smethport storm in July 1942) was caused by a series of intense convection cells. Camille and the Smethport storm both were characterized by intensities approaching the probable maximum precipitation over small areas, and their depth–area curves show rapidly decreasing average precipitation depth with increasing area; the two curves are almost perfectly aligned and both dip steeply to the right. Precipitation intensity recorded during Agnes was lower, and the storm was characterized by a broad area of heavy precipitation; the slope of its depth–area curve is relatively gentle. Although the 1985

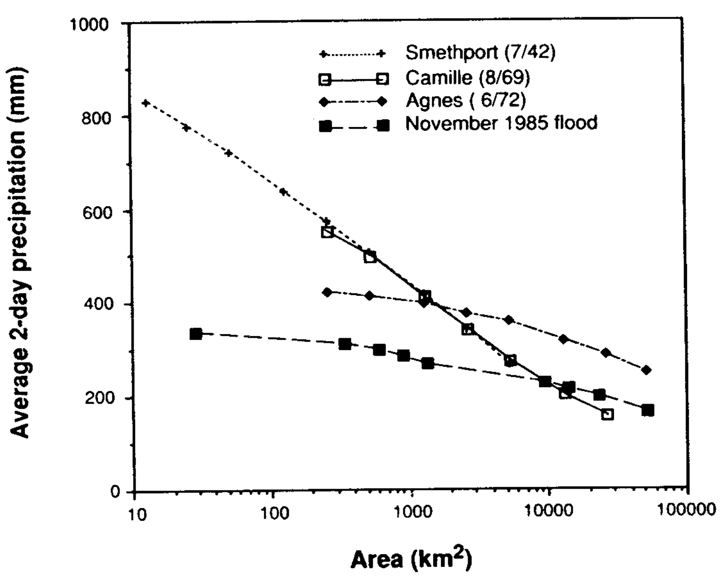

Figure 5. Depth–area curves illustrating maximum values for average two-day precipitation occurring in four great storms in the central Appalachian and mid-Atlantic regions of the eastern United States. Data from Eisenlohr (1952) and Bailey *et al.* (1975); depth–area data for the November 1985 storm were compiled using an isohyet map prepared by R. B. Jacobson of the U.S. Geological Survey

curve and the Agnes curve have comparable slopes, precipitation depths delivered by Agnes were greater across the entire range of areas.

Envelope curves for peak discharge–drainage area relationships in these four storms (Figure 6) reflect the trends shown in the precipitation data. The curve representing the Smethport storm has its rising limb farthest to the left, and peak discharges occurring during Hurricane Camille exceed those plotted for other storms at drainage areas between about 200 and 1500 km^2. The curves representing Agnes and the November 1985 flood attain the highest discharges at the largest drainage areas. Agnes has larger discharge values than the 1985 flood for drainage areas below 500 km^2 and above 10000 km^2: at the larger end of the spectrum, unprecedented peak discharges were probably caused by the enormous area contributing runoff, whereas the maximum discharges associated with smaller drainage areas are attributable to small, high-intensity precipitation cells embedded within the larger storm (Bailey *et al.*, 1975, Figure 34).

Two other central Appalachian storms with extreme values of unit discharge are not shown in order to preserve the legibility of Figure 6. A storm that inundated Johnstown, Pa. in July 1977 (Hoxit *et al.*, 1982; see Figure 1 for location) produced discharges falling in the range between the Smethport and Agnes envelope curves for drainage areas up to 25 km^2 and exceeding values measured in the November 1985 flood for drainage areas up to about 160 km^2. A storm that occurred in October 1942 produced the second highest discharge ever recorded on the Potomac River at Washington, D.C.; its envelope curve closely resembles the one plotted in Figure 4 for the March 1936 storm, exceeding discharges measured in November 1985 at drainage areas above 12000 km^2.

Although the envelope curves in Figures 5 and 6 reveal nothing unique about the November 1985 storm, its erosive impacts on bottomlands in the South Branch Potomac River and Cheat River basins clearly exceeded those of the other storms described here. The range of peak discharges measured at stations in areas that experienced severe erosion in the 1985 flood is indicated by the shaded area in Figure 6. Almost all of the great floods occurring in the central Appalachians within the last 50 years have produced discharge peaks falling within or above the shaded area, but none were accompanied by damage to valley floors of comparable intensity and spatial extent (Table I). During Hurricane Camille, fluvial erosion features comparable to those

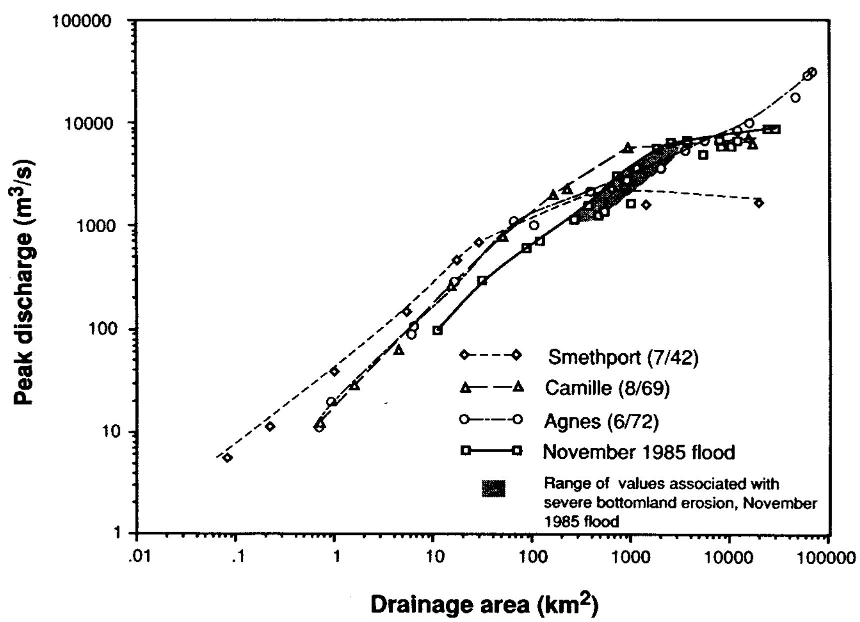

Figure 6. Envelope curves illustrating trends in the relationship between peak discharge and drainage area for four great floods in the central Appalachian and mid-Atlantic regions of the eastern United States. Data from Eisenlohr (1952), Camp and Miller (1970), Bailey *et al.* (1975), and Lescinsky (1987)

observed in the South Branch Potomac River basin in 1985 were formed in the valleys of the Tye and Rockfish Rivers (see Figure 1 for location) at drainage areas up to about 250 km^2, but geomorphic impacts at larger drainage areas were limited primarily to bank erosion and scour along channel margins and were concentrated most heavily in the vicinity of bridges and roads. In the aftermath of Tropical Storm Agnes, local examples of severe bottomland erosion were observed along the West Branch Susquehanna River and its tributaries (W. A. Turnbaugh, University of Rhode Island, unpublished archaeological site notes, 1972; Turnbaugh, 1978); these included a floodplain chute adjacent to Loyalsock Creek near Williamsport, Pa. (drainage area about 1150 km^2) that was 610 m long and 4·5 m deep. Scour holes scattered along the floodplains of the West Branch Susquehanna River and some of its other tributaries in the same area were 1·5 to 4 m deep and 6 to 23 m across, but similar features were not reported from other areas; nor were there reports of catastrophic stripping of alluvial deposits from the valley floor, as occurred in the November 1985 flood.

The evidence examined thus far indicates that events causing significant geomorphic change on steep slopes and in headwater valleys are more common than those causing significant change in larger valleys; Newson (1980) reached a similar conclusion. Furthermore, rainfall and discharge data are insufficient predictors of geomorphic effectiveness of fluvial processes in the larger valleys. As flood hydraulics are strongly influenced by boundary conditions as well as by the amount of discharge at a particular site, lithologic and structural controls on the gradient, cross-section and plan form of a valley must be considered in any attempt to explain flood impacts. Thus, as Kochel (1988) and Jacobson *et al.* (1989b) point out, observations indicating that Agnes was not a geomorphically effective flood were made in moderate-relief areas of the Piedmont physiographic province. Major geomorphic effects of Camille, the June 1949 flood, the Johnstown flood, and the November 1985 flood (as well as those sites where Agnes was geomorphically effective) occurred in the Blue Ridge, Valley and Ridge, or Plateau provinces of the central Appalachians (see Figure 1), where stream gradients are steeper, bedload is coarser, and slopes are less stable.

UNIT STREAM POWER AND FLOOD DAMAGE

Steep, narrow valleys are likely to develop large values of power per unit area of bed (or unit stream power) during an extreme flood. This quantity, which can be calculated as the product of discharge, unit weight of water, and energy slope divided by flow width:

$$\omega = \gamma Q s / w \tag{1}$$

has been recommended as a useful parameter for evaluating the ability of rivers to erode and transport sediment during high-magnitude flow events (Baker and Costa, 1987). The largest values of unit stream power and boundary shear stress are developed in bedrock canyons, where the boundaries are resistant to erosion and the flow cross-section cannot adjust its width to accommodate extreme discharges. Observations of the aftermath of the November 1985 flood indicate that where bedrock gorges and constrictions alternate with slightly wider valley reaches exhibiting alluvial bottomland development, conditions are optimal for application of a highly erosive flow to bottomland materials that are vulnerable to erosion.

Slope-area measurements of peak flood discharge were obtained from the files of U.S. Geological Survey district offices for 44 gauge locations that experienced record discharges in the June 1949 flood, Camille, Agnes, or the November 1985 flood. For two additional locations where a slope-area measurement could not be obtained, energy gradient and width of the valley floor were estimated from topographic maps. Calculated values of unit stream power are plotted in Figure 7 together with values tabulated by Baker and Costa (1987) in their discussion of the most extreme recorded floods.

The values reported here are not extraordinary by comparison with the values reported by Baker and Costa (1987); of the 35 events they cited, only four have values of unit stream power less than 1000 W m^{-2}, and the maximum value is 18 582 W m^{-2}. The additional points plotted in Figure 7 indicate that the upper limit of recorded values of unit stream power in the central Appalachians for discharges at drainage areas greater than 50 km^2 is about 2600 W m^{-2}. The largest recorded discharges at most stations are associated with values of unit stream power an order of magnitude lower; the median value is 224 W m^{-2}.

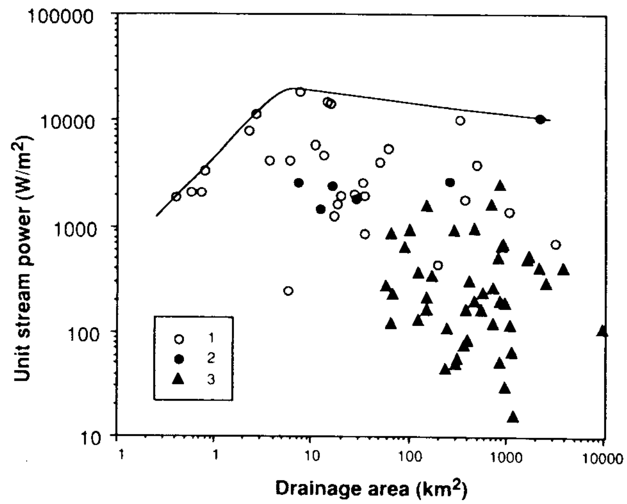

Figure 7. Plot of unit stream power against drainage area for rainfall–runoff floods and dam-failure floods; envelope curve adapted from Baker and Costa, 1987. 1. Values tabulated for rainfall–runoff floods by Baker and Costa, 1987. 2. Values tabulated for dam-break floods by Baker and Costa, 1987. 3. Values tabulated for record floods at 46 stations in the central Appalachians, from unpublished records of the U.S. Geological Survey

Additional plots of unit stream power vs. drainage area for the same set of central Appalachian stations are provided in Figure 8, with symbols identifying the values associated with individual events (Figure 8a) and identifying values calculated near sites where severe channel or floodplain erosion were observed (Figure 8b). Although the November 1985 flood was not the largest among those discussed here with respect to discharge per unit of drainage area, there were more stations with values of unit stream power exceeding 300 W m^{-2} in this flood than in the other floods described by Figure 8a. This observation helps to explain why the November 1985 flood caused so much damage. Of the 13 points associated with severe erosion in Figure 8b (marked by open triangles), 11 have unit stream power values exceeding 300 W m^{-2}; of the 33 remaining points, 25 have unit stream power values less than 300 W m^{-2}. These results suggest a trend, but they do not establish a clear threshold suitable for predictive purposes: unit stream power values exceeding 1000 W m^{-2} were calculated for sites where geomorphic impacts were negligible, and severe erosion was observed near sites with calculated values as low as 45 W m^{-2}. The highest value (2570 W m^{-2}) was calculated from a slope-area measurement made on the Maury River in Goshen Pass (located in the James River basin; see Figure 1), where a discharge of about 2500 m^3 s^{-1} passed through a bedrock canyon with a gradient of nearly 0·01 and a minimum width of less than 100 m. Boulders with intermediate axes of 1·5 m were moved at this site and some larger boulders may have moved; in addition some bedrock scour occurred along the canyon walls. Although the river channel was heavily reworked and a bridge downstream of Goshen Pass was destroyed, impacts on the valley floor below the mouth of Goshen Pass appear to have been minor.

The results indicate that factors other than unit stream power need to be considered. The most obvious is erosion resistance: the highest values of unit power are attained in bedrock canyons precisely because their boundaries are not readily eroded. The orientation of the river channel within the valley also is important. Severe erosion downstream of a bedrock constriction is most commonly observed where the channel bends to the left or the right and flood waters continue straight downvalley across the floodplain; where the channel and valley are parallel through the reach below a constriction, comparatively little erosion occurs. Another factor is the lateral variation of hydraulic conditions within a single cross-section. Even if a valley inundated by flood waters is relatively wide, much of the flow may be concentrated within a small portion of the cross-section. Under these conditions a value of unit stream power calculated for the cross-section as a whole would seriously underestimate the rate of energy expenditure per unit area of bed in that part of the valley cross-section where most of the flow was concentrated. Finally, longitudinal variations in energy gradient and

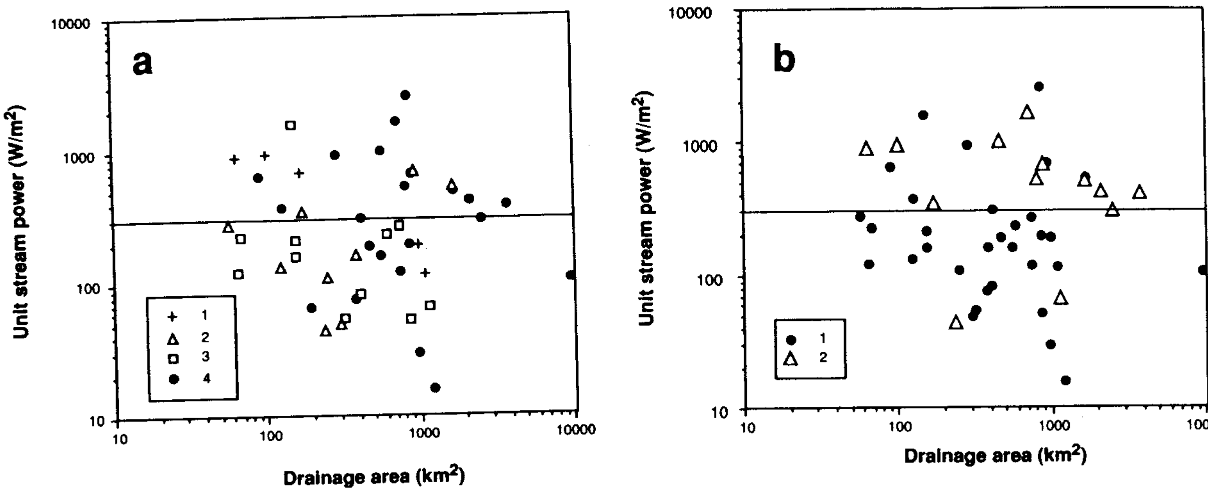

Figure 8. Plot of unit stream power against drainage area for record floods at 46 stations in the central Appalachians. (a) 1. June 1949 flood. 2. Hurricane Camille, August 1969. 3. Tropical Storm Agnes, June 1972. 4. November 1985 flood. (b) 1. Little or no reported damage to valley floor. 2. Evidence of severe channel and floodplain erosion reported from nearby location

valley width between the slope-area sites and the actual erosion sites add some scatter to the relationship between unit stream power and severe erosion.

Instead of relying on point measurements of unit stream power and identifying simple presence or absence of severe valley-floor erosion, Miller and Parkinson (in preparation) mapped the spatial distribution of erosion along the three forks of the South Branch Potomac River above their confluence, annotating $7\frac{1}{2}$-minute topographic quadrangles with information derived from manual interpretation of post-flood aerial photographs. Mapping covered a cumulative channel distance of 290 km. A simple hierarchical classification scheme consisting of 4 classes was ranked in order of increasing severity of erosion. The classes are described as 'no erosion', 'incipient to moderate erosion', 'severe erosion', and 'catastrophic erosion'; classification criteria are outlined in Miller and Parkinson (in preparation). After mapping the distribution of erosion classes, Miller and Parkinson divided each of the three forks of the South Branch Potomac River into several long reaches and calculated the percentage of each reach assigned to each of the four erosion classes. The cumulative percentage of each reach assigned to the 'severe' and 'catastrophic' classes was used as an index of the severity of floodplain erosion and compared with other parameters in an effort to explain the spatial patterns observed. The same procedure has subsequently been carried out for portions of the Tye and Rockfish Rivers, both of which are tributary to the James River and were affected by extreme flood discharges resulting from Hurricane Camille in August 1969.

The narrowest valley reaches are dominantly bedrock gorges with limited bottomland development. As was discussed above, these reaches experience relatively little change in even the largest floods. Thus one might expect that the erosion index would reach a peak for values in some intermediate range of valley widths. This does in fact appear to be the case, as Figure 9 indicates. All valley reaches with more than 50 per cent of their length mapped in either the 'severe' or the 'catastrophic' erosion class had mean widths between 230 and 370 metres; the sole outlier approaching 50 per cent is the South Branch Potomac River in the vicinity of Petersburg, West Virginia, which is a broad reach punctuated by two narrow bedrock constrictions.

It would be desirable to plot unit stream power against percentage of reach length in the severe and catastrophic erosion classes, but field measurements needed to calculate unit stream power were collected at only a few sites in the South Branch Potomac River basin following the 1985 flood and at only a few sites in the Tye and Rockfish River basins following Hurricane Camille in 1969. A surrogate parameter, here described as reach-average unit stream power, is based on average values of channel gradient, valley width, and peak discharge for each reach where erosion statistics are available. Gradient and valley width were measured from topographic maps and discharge for each reach was estimated by interpolation from flood

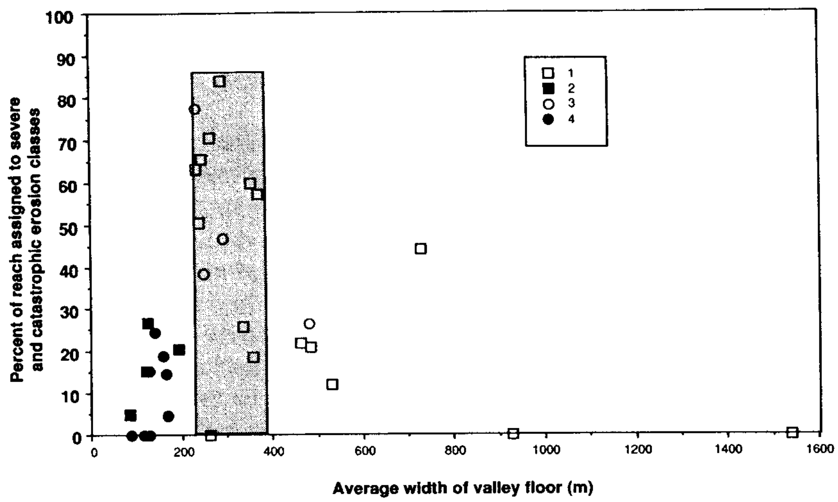

Figure 9. Plot of erosion severity against average valley width for valley reaches along the three forks of the South Branch Potomac River (November 1985) and along the Tye and Rockfish Rivers (August 1969). 1. South Branch Potomac River basin, average valley width >200 m. 2. South Branch Potomac River basin average valley width <200 m. 3. Tye and Rockfish rivers, average valley width >200 m. 4. Tye and Rockfish Rivers, average valley width <200 m. Modified from Miller and Parkinson (in preparation). The shaded area indicates that the highest values of the erosion index fall within a narrow range of valley widths

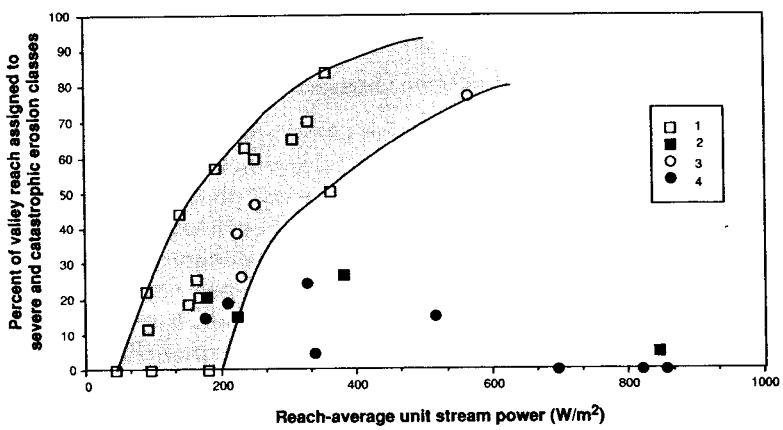

Figure 10. Plot of erosion severity against reach-average unit stream power. See caption of Figure 9 for additional information. Modified from Miller and Parkinson (in preparation).

peaks measured at known locations in the basin. The plot of per cent reach length in the severe and catastrophic erosion classes against reach-average unit stream power defines a curved belt enclosing values calculated for both the November 1985 flood and Hurricane Camille (Figure 10). Intensity of erosion increases sharply with increasing unit power, and percentages of more than 50 per cent are associated with values in the range 200–600 W m^{-2}. Assuming that valley reaches with average widths of 200 m or less are often constrained by bedrock and provide fewer opportunities for erosion of the valley floor by flood flows, the plot can be separated into two different populations. The narrow valley reaches are relatively insensitive to variations in unit stream power (Figure 10).

DISCUSSION AND CONCLUSIONS

The analysis in the preceding paragraphs represents an initial attempt at providing quantitative guidelines in relating geomorphic effectiveness to hydraulic parameters. Studies to date indicate that a flood need not approach world-record values of precipitation, unit discharge, or unit stream power in order to cause catastrophic change on the valley floor. Conversely, it is also possible for a flood to have extremely high values of precipitation, unit discharge, or unit stream power without causing change of the magnitude observed in the November 1985 flood.

For a flood to be geomorphically effective with respect to erosion of valley floors at drainage areas comparable to those discussed here requires the coincidence of sufficiently large peak flows with a physiographic setting where large values of unit stream power can be applied to valley reaches with erodible alluvial bottomlands. For valleys wider than about 200 m, evidence discussed in this paper clearly shows a trend toward increasing severity of erosion with increasing values of unit stream power. Narrower valley reaches are less sensitive to unit stream power, in part because they are more likely to have resistant boundaries and in part because the channel more often runs parallel to the valley margins and there are fewer opportunities for the central core of the flow to cross from the channel onto the adjacent valley floor.

To the extent that a threshold value of unit stream power can be associated with severe channel and floodplain erosion, 300 $\mathrm{W\,m^{-2}}$ appears to be a reasonable mininum estimate of that threshold. However, there are so many complicating factors that this parameter is a relatively poor predictor of the likelihood of extensive erosion at individual locations. Erosion features seen in the field and in aerial photographs are associated with specific configurations of valley width and slope, channel pattern, spatial arrangements of roughness elements, and local flow obstructions; a single value of unit stream power calculated for a valley cross-section cannot unravel the complexities of the flow pattern. More detailed analysis will require the use of both physical models and two- or three-dimensional numerical flow models to elucidate the relationship between channel and valley form, flow pattern, and observed patterns of erosion and deposition in an extreme flood.

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