Geomorphology, 2 (1989) 257-284
Elsevier Science Publishers B.V., Amsterdam — Printed in The Netherlands

The Role of Catastrophic Geomorphic Events in Central Appalachian Landscape Evolution

ROBERT B. JACOBSON[1], ANDREW J. MILLER[2] and JAMES A. SMITH[3]

[1]*U.S. Geological Survey, Mailstop 926, Reston, VA 22092 (U.S.A.)*
[2]*Department of Geography, University of Maryland, Baltimore County, Baltimore, MD 21228 (U.S.A.)*
[3]*Hydrologic Research Lab, National Weather Service, Silver Spring, MD 20910 (U.S.A.)*

(Received February 20, 1989; accepted after revision June 7, 1989)

Abstract

Jacobson, R.B., Miller, A.J. and Smith, J.A., 1989. The role of catastrophic geomorphic events in central Appalachian landscape evolution. In: T.W. Gardner and W.D. Sevon (Editors), Appalachian Geomorphology. Geomorphology, 2: 257-284.

Catastrophic geomorphic events are taken as those that are large, sudden, and rare on human timescales. In the nonglaciated, low-seismicity central Appalachians, these are dominantly floods and landslides. Evaluation of the role of catastrophic events in landscape evolution includes assessment of their contributions to denudation and formation of prominent landscape features, and how they vary through space and time.

Tropical storm paths and topographic barriers at the Blue Ridge and Allegheny Front create significant climatic variability across the Appalachians. For moderate floods, the influence of basin geology is apparent in modifying severity of flooding, but for the most extreme events, flood discharges relate mainly to rainfall characteristics such as intensity, duration, storm size, and location. Landslide susceptibility relates more directly to geologic controls that determine what intensity and duration of rainfall will trigger slope instability.

Large floods and landslides are not necessarily effective in producing prominent geomorphic features. Large historic floods in the Piedmont have been minimally effective in producing prominent and persistent geomorphic features. In contrast, smaller floods in the Valley and Ridge produced erosional and depositional features that probably will require thousands of years to efface. Scars and deposits of debris slide-avalanches triggered on sandstone ridges recover slowly and persist much longer than scars and deposits of smaller landslides triggered on finer-grained regolith, even though the smaller landslides may have eroded greater aggregate volume.

The surficial stratigraphic record can be used to extend the spatial and temporal limits of our knowledge of catastrophic events. Many prominent alluvial and colluvial landforms in the central Appalachians are composed of sediments that were deposited by processes similar to those observed in historic catastrophic events. Available stratigraphic evidence shows two scales of temporal variation: one related to Quaternary climate changes and a more-recent, higher-frequency variation due to rare events during the Holocene. In much of the central Appalachians, landforms related to Quaternary climate changes persist as the most prominent features, despite the modifying effects of late-Holocene catastrophic events.

Introduction

Hack and Goodlett (1960) hypothesized that rare, large, geomorphic events may be the dominant sculptors of the central Appalachian landscape. At nearly the same time, Wolman and Miller (1960) developed the concept that moderate-sized events occurring with moderate frequency were responsible for transporting most of the sediment load on some landscapes. Both of these concepts, and Wolman and Gerson's (1978) elaboration of the magnitude/fre-

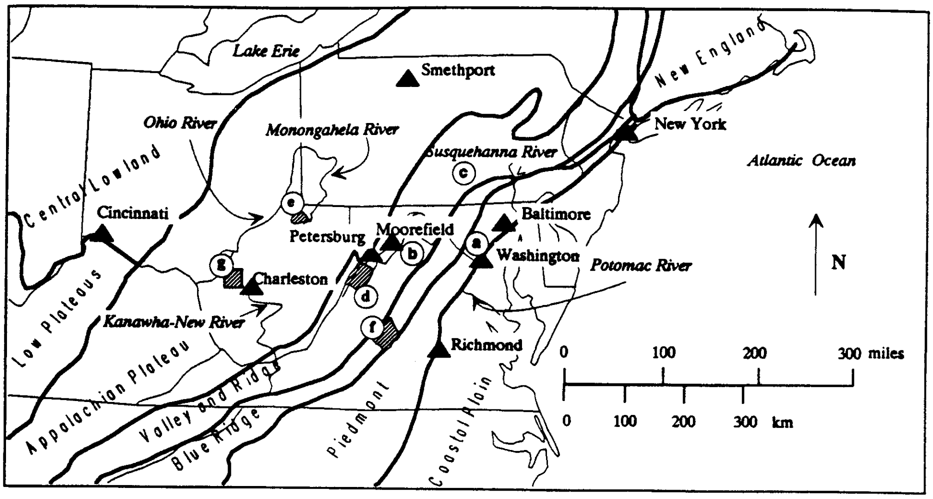

Fig. 1. Physiographic provinces and locations in the central Appalachians. a = Seneca Creek, b = Passage Creek, c = Little Lehigh Creek, d = Pendleton County, W. Va., e = Buffalo Creek basin, f = Nelson County, Va., g = St. Albans archeological site.

quency argument, were substantiated with data and examples from the central Appalachians, an area with great physiographic variety. Our goal in this paper is to examine the current state of knowledge of the role of catastrophic events in evolution of the landscapes of the central Appalachians. In this paper, the central Appalachians are considered to include the nonglaciated portions of the Piedmont, Blue Ridge, Valley and Ridge, and Plateau physiographic provinces of Pennsylvania, Maryland, Virginia, and West Virginia (Fig. 1).

Almost any geomorphic event could be considered catastrophic, depending on one's reference frame in time and space. From social and economic perspectives, however, there is little debate over the catastrophic effects of a few dozen flood and landslide events that have affected parts of the central Appalachians during historic time. Therefore, to provide a reference frame for discussion, and to emphasize applications in hazards mitigation, human time and spatial scales are adopted here for definition of the term "catastrophic". This paper considers events to be catastrophic that are rare on human timescales (one or two per generation to several during the Holocene) and sufficiently intense and (or) extensive as to threaten lives and property. In the central Appalachians these are usually storm-induced landslides and floods. An understanding of the role of these events in shaping the Appalachian landscape will allow better prediction of where and how often catastrophic flood and landslide hazards occur.

Assessment of the role of catastrophic geomorphic events in shaping the landscape should be cast in terms of magnitude, frequency, and effectiveness of geomorphic processes (Wolman and Miller, 1960; Wolman and Gerson, 1978). In these terms, two separable questions are apparent: are catastrophic events important in total geomorphic work and denudation? and, are catastrophic events important in producing prominent landforms? The answers to these questions vary with the highly variable geology and climate of the central Appalachians. Lack of sediment transport data precludes quantitative consideration of the first question, but observations of the causes and effects of extreme geomorphic events across the region apply to the second question. In the first part of this paper we consider some examples of historic catastrophic flood and landslide events to generalize about the spatial and temporal controls on present-day catastrophic events. In the second part of the paper we consider selected examples of the surficial stratigraphic record to evaluate the long-term effect of catastrophic events on the Appalachian landscape.

Controls on the spatial and temporal distributions of catastrophic geomorphic events

Variations in structure, lithology, and regolith have produced four distinctly different physiographic provinces in the central Appalachians (Fig. 1). The geologic framework and variable meteorological settings determine a wide range of factors controlling hydrologic and hillslope processes across the region.

Geologic framework

The Piedmont physiographic province is dominated by gently rolling topography underlain by thick saprolite regolith formed from schists, gneisses, and other metamorphic rocks. Small areas underlain by carbonates and mafic rocks have thinner regolith and the Mesozoic sedimentary rocks have thin soils and low relief.

In the Blue Ridge province, regolith thickness and relief vary markedly with bedrock lithology and texture (Pavich, 1986). Generally, quartzites, mafic rocks, carbonates, and massive siliceous igneous rocks have thin residuum and underlie a steep landscape. Metasedimentary and metaigneous rocks have thick saprolite and underlie a gently rolling landscape similar to the Piedmont. Thick colluvium occurs on some hillslopes and alluvium and debris are common in fans along the Blue Ridge. Debris is defined as poorly sorted to diamictic hillslope sediment transported in part by debris flow with variable fluvial reworking (Jacobson, 1988a).

In the folded and faulted Valley and Ridge, thin regolith is dominant on shaley lithologies and on tops of steep, quartzite ridges. Thicker colluvium and debris have collected at the bases of many slopes and may be tens of meters thick where deposited over dissolving limestone.

Some of the steepest land in the central Appalachians occurs in the Plateau province where major rivers have incised nearly flat lying Paleozoic sandstones; relief is more subdued in areas underlain by shales and limestones. Thin residual weathering profiles are dominant in the Plateau although thicker colluvium exists at high elevations, in hillslope hollows, and at the bases of slopes.

Meteorological patterns and hydrologic responses

Mean annual precipitation varies markedly across the four physiographic provinces (Fig. 2), from a maximum of over 1420 mm (56 in) in the Allegheny Mountains at the eastern edge of the Appalachian Plateau and along the crest of the Blue Ridge, to a minimum of less than 914 mm (36 in) at low elevations in the western Valley and Ridge. Approximately 125 mm (5 in) of the mean annual precipitation in the Appalachian Plateau and 30 mm (1.2 in) in the Blue Ridge is from snow.

The role of topography in determining precipitation characteristics is complex. More important than total altitude are the effects of the two major topographic boundaries, the Blue Ridge and Allegheny Front. The mechanisms by which the Blue Ridge and Allegheny Front influence precipitation are quite different. The Blue Ridge produces an elongate area with high values of mean annual precipitation which varies with that of the two neighboring provinces, as shown by high correlations of annual precipitation among three representative rain gages in the Piedmont, Blue Ridge, and Valley and Ridge, Table 1. In contrast, the Allegheny Front separates much more distinctly the climate of the Appalachian Plateau from the eastern physiographic provinces. Mean annual precipitation increases more than 500 mm over a distance of less than 48 km and, more significantly, annual precipitation in the eastern Appala-

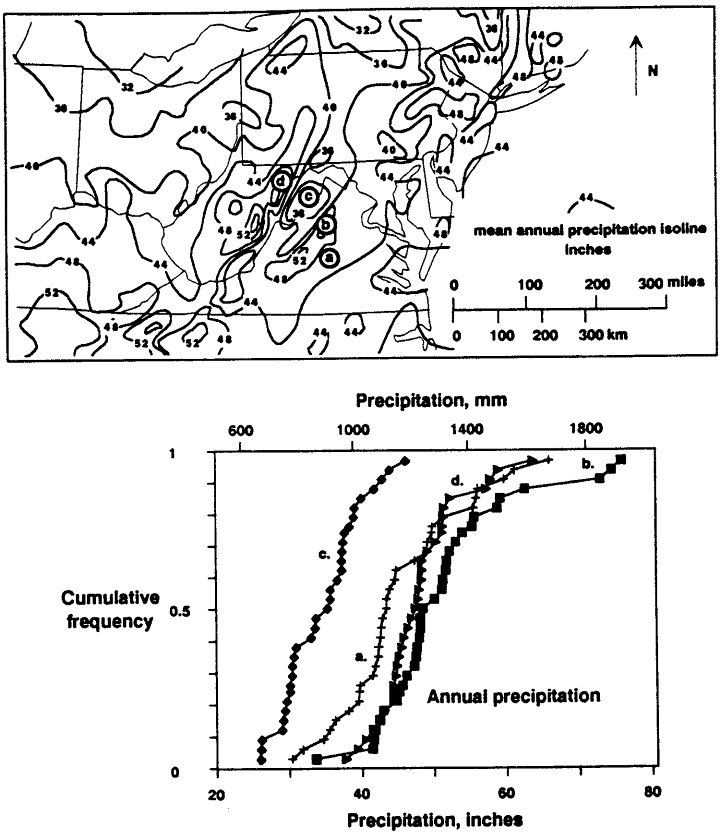

Fig. 2. Map of mean annual precipitation of the central Appalachians (modified from Environmental Data Service, 1968) and cumulative frequency distributions of annual rainfall at four representative raingage stations. a = Piedmont, Charlottesville, Va, b = Blue Ridge, Big Meadows, c = Valley and Ridge, Wardensville, W. Va., d = Appalachian Plateau Bayard, W. Va. Rainfall isohyets in inches can be converted to mm by multiplying by 25.4.

TABLE 1

Cross correlation of annual rainfall for four stations in the central Appalachians

	Big Meadows (Blue Ridge)	Charlottesville (Piedmont)	Wardensville (Valley and Ridge)	Bayard (Plateau)
Big Meadows	1			
Charlottesville	0.64	1		
Wardensville	0.75	0.70	1	
Bayard	0.27	0.29	0.31	1

chian Plateau is very poorly correlated with gages in the other provinces despite small distances between gages (Table 1, Fig. 2). The Appalachian Plateau experiences a significantly different precipitation regime on the annual time scale from the other provinces.

Values of extreme, short-duration rainfall generally decline in magnitude east to west across the central Appalachians, increasing locally along the Blue Ridge topographic boundary but not at the Allegheny Front (Fig. 3). Contrasts in extreme rainfall between the east-

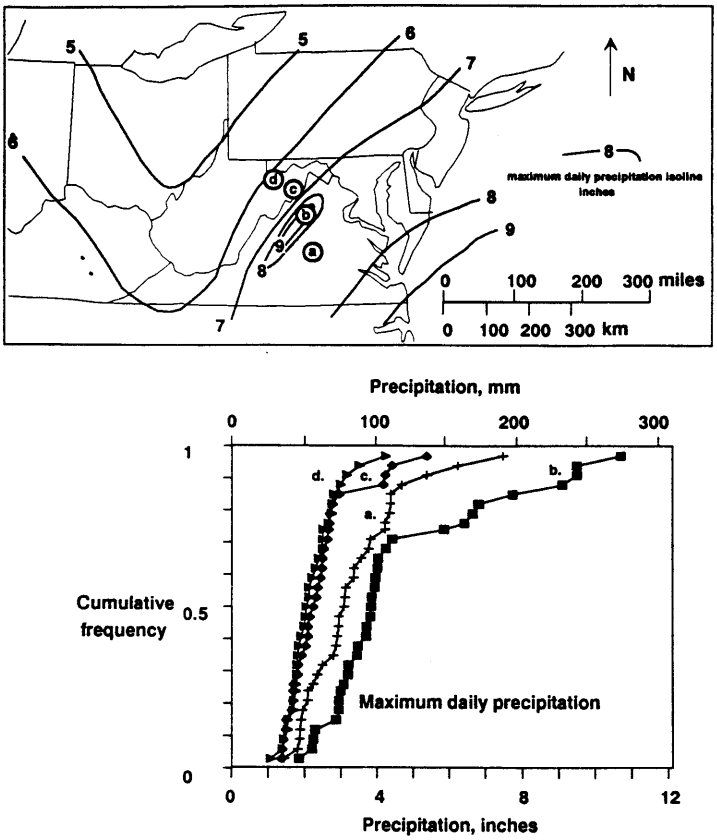

Fig. 3. Map of maximum daily rainfall with 100 yr recurrence interval (modified from Hershfield, 1961) and cumulative frequency distributions of annual maximum daily precipitation at four representative raingage stations: a = Piedmont, Charlottesville, Va, b = Blue Ridge, Big Meadows, c = Valley and Ridge, Wardensville, W. Va., d = Appalachian Plateau, Bayard, W. Va. Rainfall isohyets in inches can be converted to mm by multiplying by 25.4.

ern and western central Appalachians can be explained in large part by the typical paths of tropical storms along the coast (Neumann et al., 1981), that impose an east to west gradient in mean rainfall patterns.

To assess geomorphic effects of extreme storm rainfall at the basin scale, spatial variability of storms must also be considered. A statistical model of storm rainfall for the Valley and Ridge portion of the central Appalachians (based on storm data 1949–1981) depicts storms as consisting of intense cells representative of convective complexes (Smith and Karr, in press). Applying this model to Valley and Ridge storm data yields the following statistical description of storms: Storms with rainfall exceeding 125 mm over three days (measured at a point) occur with a frequency of 1 per 2.2 years over the 10,000 km^2 area; each storm consists of a mean density of 3 cells/10,000 km^2. The average rainfall at cell centers is 64 mm and the radial distance over which cell rainfall decreases by one half is 28 km. Virtually all of the rainfall of a single cell falls within a 50 km radius, corresponding to the meteorological small mesoscale (50–1000 km^2). Using the model, and assuming a stationary climate over the Holocene, we obtained the probability distribution of maximum 3-day totals at a point for the 10,000 km^2 area (Fig. 4). These results show that extrapolation of historic rainfall data over 1000–10,000 years predicts high probabilities

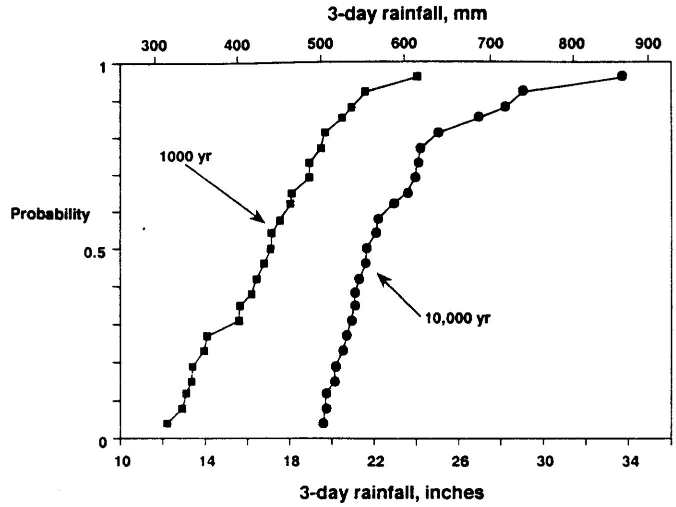

Fig. 4. Calculated probability distributions of maximum point rainfall for the central Appalachians over periods of 1000 and 10,000 yr.

that each point in the central Appalachians will be (and has been) subjected to extreme rainfall events on the order of 305-760 mm (12–30 in) over a 3-day period.

The tendency for extreme rainfall to occur in cells of 50–1000 km^2 area strongly determines the basin scales at which rainfall produces geomorphically significant floods. Decrease in maximum unit discharge with increasing basin size is controlled in part by organization of extreme rainfall at the small mesoscale (Fig. 5a). Note that maximum unit discharge and variability in maximum unit discharge decrease sharply beyond approximately 400 km^2; basin areas of maximum unit discharge are generally less than corresponding storm cell areas because storm cells usually distribute rainfall to two or more basins.

The role of spatial inhomogeneities of extreme rainfall in determining extreme flood characteristics is even more pronounced when contrasting only Appalachian Plateau and Piedmont basins (Fig. 5b). The contrast in values of extreme, short-duration rainfall between the Plateau and the Piedmont (Fig. 3) are clearly reflected in extreme floods. Unit discharges at small drainage areas are much larger in the Piedmont than in the Plateau. Except for the anomalous 1985 storm, which resulted from complex interactions of a tropical cyclone with two low-pressure systems (Jacobson et al., 1989), big floods in the Appalachian Plateau are much smaller than big floods in the Piedmont. For all the Piedmont basins included in Fig. 5b, the flood of record is associated with a tropical storm. In contrast, the floods of record for the Plateau basins are associated with a variety of meteorological events, including convective thunderstorms, frontal rainfall, and snowmelt.

In a discussion of the relative magnitudes and geomorphic effectiveness of several historic floods in the Potomac River Basin and adjacent watersheds, Miller (in press) describes basin-scale hydrologic response as a function of the intensity, duration, and area of precipation. Three discharge–drainage area envelope curves enclosing the maximum recorded values of discharge for each of three great floods on the Potomac River illustrate differences in extreme floods attributable to storm differences (Fig. 6, Table 2). Each flood has a steep rising limb; peak discharge increases downstream 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 in-

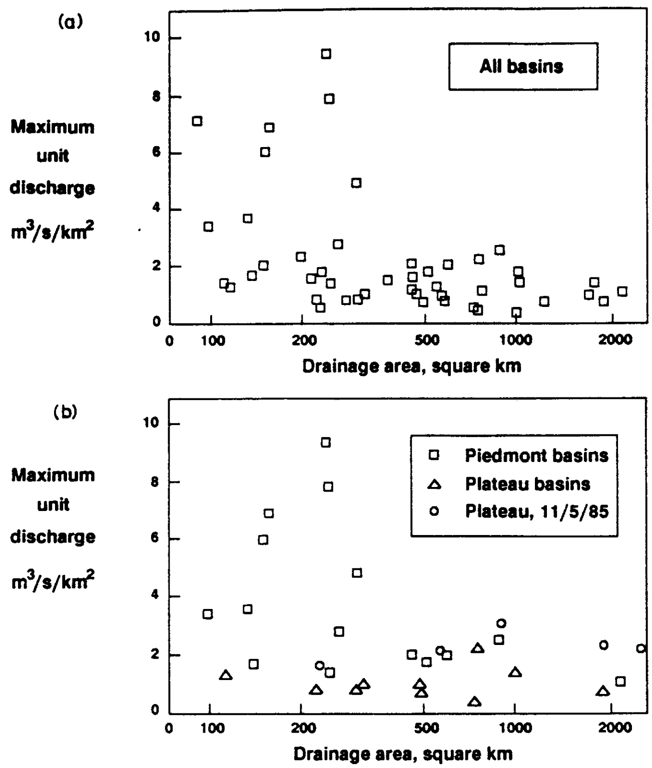

Fig. 5. (a) Plot of maximum unit discharge versus drainage area for 70 streamgage stations in the central Appalachians. (b) Plot of maximum unit discharge versus drainage area showing contrasts between Piedmont and Plateau floods. Modified from Smith (1988).

crease in peak discharge is observed. The flat limb of each curve represents translation and possibly some attenuation of the flood wave, with only modest contributions from tributary basins outside the areas of greatest precipitation depth.

The three floods were generated by three different kinds of precipitation events. The greatest intensity and smallest area of precipitation were observed in the 1949 storm, a conventional complex astride the Shenandoah and South Branch Potomac River basins (Hack and Goodlett, 1960), that affected basin areas less than 200 km^2. This storm triggered more than 100 debris avalanches and caused reworking of valley floors in small basins (Stringfield and Smith, 1956; Hack and Goodlett, 1960). The lowest intensity and largest area affected were observed in the 1936 flood produced by a series of winter–spring frontal storms, partly on snow and frozen ground (Grover, 1937). Despite widespread inundation of river valleys and extensive property damage, there are no reports indicating major geomorphic changes on slopes or on valley floors resulted from this event. The intermediate event, the 1985 storm, was produced by a tropical cyclone interacting with two extratropical low-pressure systems. Peak discharge of this storm occured at drainage areas between those of the 1936 and 1949 storms; valleys with drainage areas between 500 and 2000 km^2 incurred extreme channel widening and flood-plain erosion (Miller, 1987).

Analysis of other storms in the central Appalachians shows that flood response is not simply related to the meterological context of the storm, because storms of similar genesis can produce very different floods depending on in-

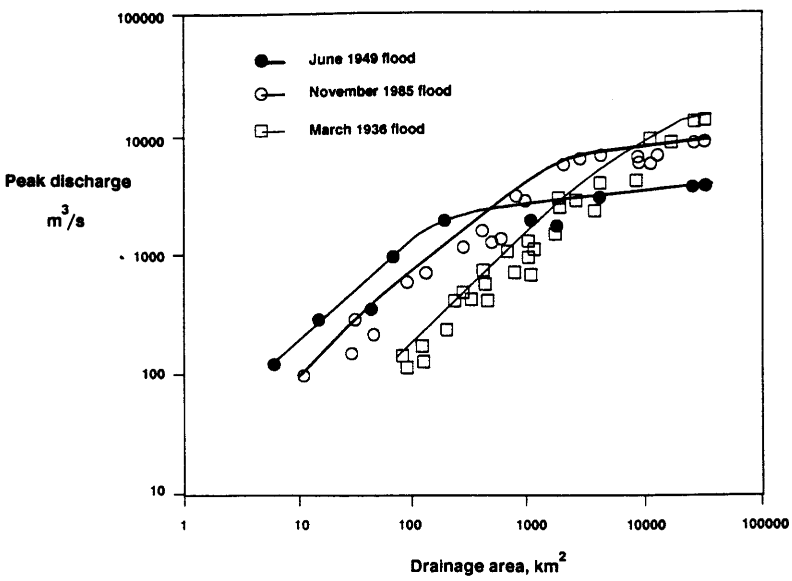

Fig. 6. Plot of peak discharge versus drainage area for three extreme floods on the Potomac River. Modified from Miller (in prep).

TABLE 2

Peak two-day precipitation totals for selected storms

Storm	Date	Official peak (mm)	Unofficial peak (mm)
Potomac, James, and upper Ohio floods (Grover, 1937)	March 1936	179	no data
Smethport, Pa. (Eisenlohr, 1952)	July 1942	209	909
Petersburg, W. Va. and North River basin, Va. (Stringfield and Smith, 1956)	June 1949	305	413
Johnstown, Pa. et al (Hoxit, 1982)	July 1977	305	no data
Hurricane Camille, Nelson County, Va. (Camp and Miller, 1970)	August 1969	278	711
Tropical storm Agnes, Md., Pa., and Va., (Bailey, 1975)	June 1972	442	406
November 1985 storm (Lescinsky, 1987)	November 1985	310	no data

tensity, duration, and area of the storm (Fig. 7). The Smethport storm (Fig. 1, July 1942), like the June 1949 storm, was a short-lived convectional system producing high precipitation intensity in a limited area. Camille and Agnes were both tropical cyclones, but Camille moved through the Appalachians very quickly and delivered all of its precipitation in a small area in less than 12 h, whereas Agnes, like the November 1985 storm, combined with an extratropical low-pressure system and, stalled by a high-pressure system off the east coast, delivered

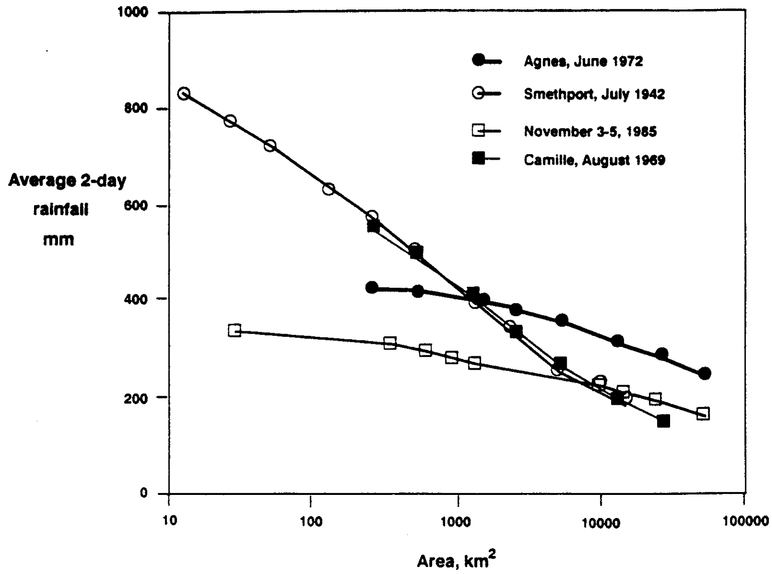

Fig. 7. Plot of rainfall depth (over two-day-duration) and area for four extreme storms in the central Appalachians. Modified from Miller (in press).

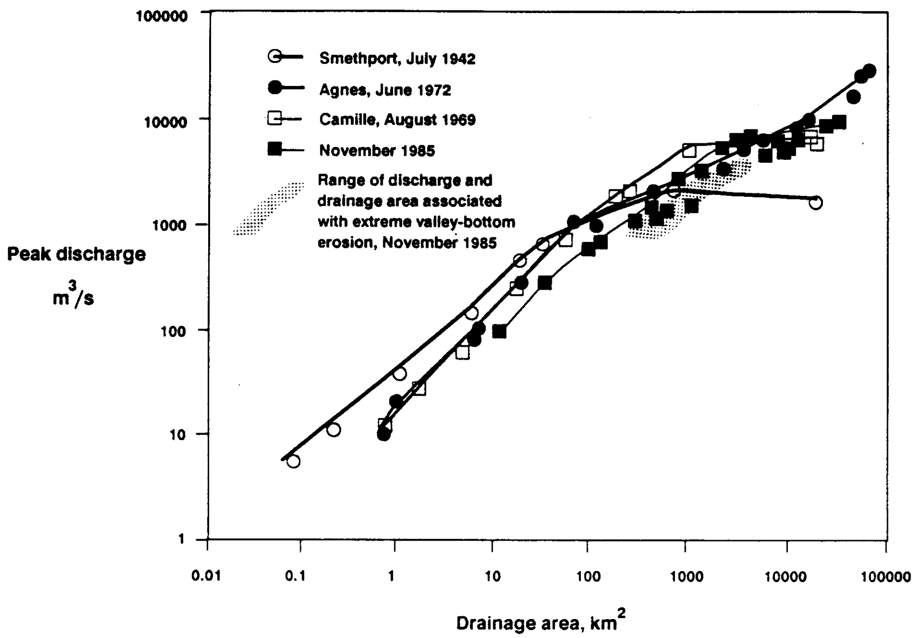

Fig. 8. Plot of peak discharge versus drainage area for four extreme floods in the Central Appalachians. Modified from Miller (in press).

precipitation over a broad region for several days.

Although the four storms form two distinct classes of rainfall depth–area curves, discharge–drainage area envelope curves for the resulting floods (Fig. 8) show greater diversity in flood responses, with peak discharges occuring over a range of 700–70,000 km^2. The differences in discharge–drainage area curves are presumably related to different combinations

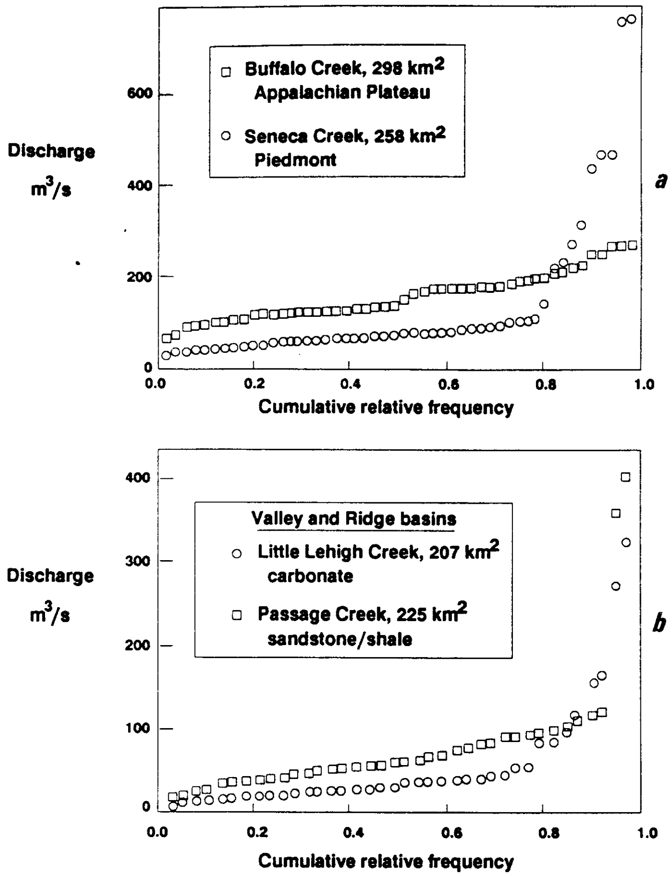

Fig. 9. (a) Plot of discharge versus probability that the discharge is equal to or less than the given value for two similar size drainage basins in the Piedmont and Plateau. (b) Plot of discharge versus probability that the discharge is equal to or less than the given value for two similar size drainage basins in the Valley and Ridge on different bedrock types. Modified from Smith (1988).

of rainfall intensity and duration, and differing landscape hydrologic responses.

Geologic controls are capable of modifying flood responses to moderate and high-magnitude storms but are ineffective in mediating the most extreme events. In a comparison of the annual peak flow distributions for two streamgage stations of similar drainage area, Seneca Creek in the Piedmont, and Buffalo Creek in the Appalachian Plateau, annual peak quantiles for Seneca Creek are lower than Buffalo Creek for discharges less than the 80th percentile (Fig. 9a). This reflects the higher storage capacity of thick saprolite in the Piedmont compared to the thin discontinuous, colluvial and residual soils of the Plateau. In the upper tail of the annual peak distribution, however, Seneca Creek discharges are much higher, reflecting the dominant effect of extreme rainfall magnitudes associated with tropical storms.

In a comparison between two Valley and Ridge basins of similar size, Little Lehigh Creek, draining an area dominated by carbonates, and Passage Creek, draining an area dominated by sandstone and shale, the effects of geologic control are apparent for annual discharges up to the 80th percentile but are less important for the extreme, infrequent storms (Fig. 9b). For the small and moderate-size events, the annual discharges from Passage Creek are larger than

Little Lehigh Creek, evidently reflecting the higher relief and lower storage available in the non-carbonate, Passage Creek basin. Similar dramatic effects of karst features on the central portion of annual flood peak distributions have been described by White (1976). For extreme floods, however, the discharge differences between the carbonate and sandstone/shale basins are negligible.

Landslide susceptibility and meteorological triggers

Landslides are often catastrophic geomorphic events. When landslides are triggered during a storm that also produces catastrophic flooding, the sediment supplied to the flood can increase its potential for geomorphic change.

As with hydrologic processes, landslide processes vary with physiography and climate across the central Appalachians. Factors influencing landslide susceptibility are interdependent with one another and rainfall characteristics. Intrinsic factors operating at a slope site include regolith shear strength, regolith thickness, regolith permeability, slope morphology, and shear strength provided by roots. The intrinsic factors interact with extrinsic factors, such as antecedent soil moisture levels, and intensity and duration of rainfall, to determine landslide susceptibility. At the slope scale, interaction of these factors make prediction of landslide susceptibility at a site extremely difficult during any particular storm (Mills et al., 1987; Jacobson et al., 1989). At a regional scale, where vegetation influences can be averaged out and rock types can be assumed to have characteristic slope morphologies, landslide susceptibility can be described in terms of characteristic regolith infiltration and drainage rates that determine the accumulation of soil pore-water pressures under given rainfall intensity and duration.

Landslides are uncommon on natural slopes of the Piedmont. Even the prolonged, heavy rainfall of Hurricane Agnes did not trigger noticeable slope instability in the Piedmont of Maryland (J.E. Costa, unpubl. data; Clark, 1987). In contrast to the Piedmont, the steeper slopes of the Blue Ridge have been subjected repeatedly to landsliding, especially in the southern Blue Ridge of North Carolina and Tennessee (Clark, 1987). One of the most dramatic cases of landsliding in the central Blue Ridge was the regionally extensive destabilization caused by intense rainfall from Hurricane Camille in Nelson County, Va., in 1969. Landslides were dominantly debris slide-avalanches that delivered large amounts of coarse sediment to streams and fans (Williams and Guy, 1973; Johnson, 1983; Kochel, 1987; Gryta and Bartholomew, 1987).

Landslide processes and spatial distributions in the Valley and Ridge have been related to rock and regolith, and variations in their hydrologic response to different types of storms (Jacobson, 1988b; Carr and Jacobson, 1988). For example, in November of 1985, 200–240 mm rainfall over 48 h in Pendleton County, W. Va. produced thousands of shallow landslides on thin colluvial and residual soils formed on shale, and eight debris slide-avalanches on quartzite ridges. In contrast, when the same landscape experienced much more intense, short-duration rainfall (approx. 356 mm in 24 h, Stringfield and Smith, 1956) from the 1949 convective thunderstorm, dozens of debris slide-avalanches were triggered on quartzite ridges. Presumably, long-duration/low-intensity rainfall is required to fill soil moisture reservoirs in fine-grained regolith, whereas, short-duration/high-intensity rainfall is required to fill soil-moisture reservoirs in quartzite regolith where infiltration and drainage rates are high to coarse materials and fractures in underlying bedrock.

At the slope scale, other factors appear to control landslide location, but these factors have variable influence depending on the storm characteristics. Locations of the 1949 debris avalanches were insensitive to bedrock dip (Clark, 1987) whereas most of the debris slide locations in 1985 were on dip slopes where bedding plane weaknesses and down-dip flow of groundwater may have been important contributors to slope instability (Jacobson et al.,1989). Smaller landslides triggered in 1985 were strongly biased toward cleared land but were fairly insensitive to hillslope topography (Jacobson et al., 1989).

The eastern edge of the Allegheny Plateau has steep slopes on Mississippian and Pennsylvanian sandstones along major river valleys. These slopes, like those documented by Pomeroy (1980), are subject to debris slide-avalanches triggered by intense rainfall. In contrast, shaley Pennsylvanian and Permian lithologies in the Plateau have gentler slopes characterized by shallow slips, slumps, and earth flows (Jacobson, 1985; Pomeroy, 1982). On interbedded sandstones and mudstones of the Permian Dunkard Group, ages of a population of landslides dated by tree ring growth correlated best with long, wet periods, mostly spring months with high values of runoff over 30-day durations, rather than with any specific storm event (Jacobson, 1985).

Geologic controls and meteorological variations determine a broad range of conditions that can trigger major landslide events. As indicated by Wieczorek (1987), slope instability is triggered by pore-water pressures that rise as a function of regolith thickness, infiltration rate, and drainage rate, plus convergence or divergence of shallow groundwater flow, as well as intensity and duration of rainfall. Because of the complex conditions that determine rainfall intensity and duration and how they trigger slope instability, flood-producing events and landslide-producing events are not necessarily the same. For example, Hurricane Agnes produced extreme amounts of rainfall over durations from 12 to 96 h, which resulted in extreme flooding (Fig. 8) but negligible landsliding (J.E. Costa, unpubl. data; Clark, 1987) because it mainly affected the Piedmont (Fig. 10, lines i, j). Much lower storm totals at comparable durations (Fig. 10, lines a–h) produced thousands of small landslides (Jacobson et al., 1989) and

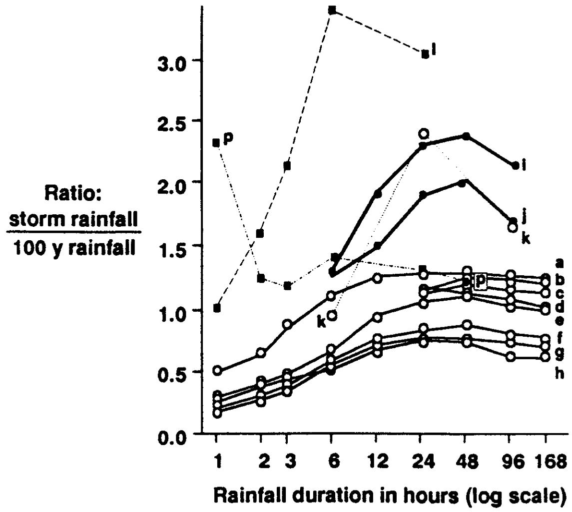

Fig. 10. Plot of ratio of storm rainfall to estimate of 100 y recurrence rainfall (from Hershfield, 1961) versus storm duration for extreme storms in the central Appalachians. a–h = Nov. 1985 storm; i–j = tropical Storm Agnes, June 1972; k = June 1949 storm; l = Hurricane Camille, August 1969; p = July 1942 storm (Smethport, Pa.).

extreme flooding (Figs, 6, 8) in the Valley and Ridge in November, 1985. In the same area, smaller storm totals at higher intensity produced dozens of large debris slide-avalanches (Stringfield and Smith, 1956) with less-extreme flooding (Fig. 6) in June, 1949 (Fig. 10, line k). The extremely intense rainfall recorded during the Smethport storm (Fig. 10, line p) evidently produced some small "blowout" landslides but was not noted for triggering extensive slope instability (Eisenlohr, 1952) probably because rainfall intensities were too great and duration too short to allow deep infiltration. Extremely high-intensity/moderate-duration rainfall from Hurricane Camille (Fig.10, line 1) was very effective in triggering landslides in Nelson County, Va. and in producing extreme flooding at small drainage areas (Figs. 7, 8).

History of landsliding in an area also influences the response to extreme rainfall. After a hillslope site has failed and regolith has been removed, some time interval is necessary to replenish regolith by bedrock weathering or accumulation of colluvium in the landslide scar to critical thickness for subsequent failure (c.f. Reneau and Dietrich, 1987). Lack of hillslope sites with critical regolith thicknesses may have prevented extensive slope instability in Nelson County, Va. in 1985 when 325 mm of rain fell over 48 h. Alternatively, rainfall intensity and duration may not have been adequate to trigger landslides during the 1985 storm.

Geomorphic work and effectiveness

Wolman and Miller (1960) concluded that most of the geomorphic work accomplished by rivers, as measured by the sediment loads they transport, is performed by flood discharges with frequencies of one or two per year in many areas. However, their data did not extend to consider catastrophic events with frequencies of 1 in 100 years and smaller. Data for assessing the work performed for these events are rare and estimated recurrence intervals for threshold of transport are tenuous. This is especially true for rare events in steep mountainous terrain where debris flows are common and flood transport is dominated by coarse bedload.

Existing data on sediment transport by catastrophic floods indicate that they transport a very large percentage of total annual suspended sediment. Floods from Hurricane Agnes transported 135–435% of the estimated annual sediment load from three small basins (<2 km^2) in Pennsylvania (Reed, 1980) and 300% of the annual sediment load of the Susquehanna River (Gross et al., 1978). No comparable data are available for steep basins in the central Appalachians, although rough estimates of debris-flow denudation rates in Central Virginia (Osterkamp and Costa, 1986) are close to long-term denudation rates, implying that infrequent debris flows are the dominant denudational process in that area. However, debris-flow denudation rates are also ten times greater than present-day erosion rates based on river sediment loads, indicating disequilibrium between slopes and rivers (Osterkamp and Costa, 1986).

While available data are insufficient to adequately characterize and compare geomorphic work performed by catastrophic events in the central Appalachians, we can characterize their geomorphic effectiveness. Geomorphic effectiveness was defined by Wolman and Gerson (1978) as the ability of events to form or shape the landscape, regardless of how those events rank in denudational importance. Effectiveness also is defined relative to the time required to restore the landscape to the form it had before the event, that is, a recovery time. A geomorphic event is effective if the resulting erosional and (or) depositional landforms are large, cover a broad area, and last a long time. For most of the historic, catastrophic events discussed here, absolute erosion attributable to the event, and event frequency, are poorly known, but thicknesses of flood deposits, channel widening, and depths of landslide scars attest that the events produced many times the mean annual erosion.

Effectiveness in the fluvial regime

Within any one basin or physiographic province, geomorphic effectiveness of a storm event on river channels and flood plains may vary with intensity, duration, and spatial extent of precipation. For example, no major geomorphic changes were reported from the flood of largest contributing area and lowest rainfall intensity in the Potomac River Basin (1936 flood, Fig. 6). The smaller, more intense storm of June 1949 caused channel widening and valley-floor erosion at small drainage areas (<50 km^2) or at sites where debris-avalanche deposits blocked river channels in larger valleys (A.J. Miller, unpubl. data; Hack and Goodlett, 1960; Stringfield and Smith, 1956). In contrast, size and intensity of the November 1985 storm produced flood peaks capable of widespread valley-floor erosion at drainage areas of several hundred km^2. Flood features documented by Miller (1987), Clark et al. (1987), McKoy (1988), and Miller (in press) include cobble bars, cobble levees, sand splays, stripped flood-plain surfaces, extremely widened channels, and chutes cut across flood plains. Sites of extreme erosion and stripping of flood-plain surfaces are slowly recovering since the 1985 flood by incremental deposition of clay- to sand-size material from moderate floods. Depositional features, like cobble bars and levees, however, probably will persist much longer as prominent features.

The contrast of geomorphic effects of the 1949 and 1985 floods is typified by the sequence of events occuring at a location along the South Branch Potomac River in the Smoke Hole, a canyon reach near Petersburg, W.Va. (Fig. 1). This site was located in the area of most intense precipitation during the 1949 storm, and debris avalanches occurred in most of the local headwater valleys. Debris avalanches in the Redman Run basin delivered a large volume of coarse sediment directly into the channel of the South Branch Potomac River; discharge on the South Branch was incapable of removing this material, and the channel was diverted to the right, cutting into the alluvial bottomland on the opposite bank. Comparison of aerial photographs from 1945 (Fig. 11a) and 1952 (Fig. 11b) shows the resulting change in configuration of the channel and valley floor. During the subsequent 36-year period the only notable change was the growth of vegetation on the surface created in the 1949 flood (Fig. 11c shows the site in 1980). In November 1985, only a few debris avalanches occurred in the South Branch Potomac River basin and all of them came to rest well upstream of tributary confluences with the main trunk river. At the site illustrated in Fig. 11, some reworking of tributary valley deposits was observed but the volume of new sediment delivered to the South Branch was relatively small. Peak discharge on the South Branch was much larger than in 1949, however, and resulted in complete removal of the surface created in the 1949 storm and formation of a large gravel bar along the right side of the valley (Fig. 11d).

Although the comparison of storms and flood effectiveness cited above is valid for that part of the Valley and Ridge, storms of comparable intensity and size occurring in different geomorphic settings may not be equally effective. The range of peak discharges measured at stations in the Potomac River basin in areas that experienced severe erosion in 1985 is indicated by the shaded area in Fig. 8. The other envelope curves included in the figure indicate that Agnes, Camille, and the Smethport storm all produced discharge peaks falling within or above the shaded area. Geomorphic impacts of Camille and of the Smethport storm have been documented (Eisenlohr, 1952; Williams and Guy, 1973; Johnson, 1983) and are characterized chiefly by slope failures and reworking of valley floors in small drainage basins that are directly affected by slope failures. Fluvial effects in larger basins affected by Camille were concentrated in the vicinity of bridges and roads (Williams and Guy, 1973) and were not comparable in magnitude to the fluvial impacts observed in the Potomac and Cheat River basins at drainage areas of 250–2000 km^2 in 1985. Similarly, geomorphic effects of Tropical Storm Agnes were also meagre. Most rivers recovered fairly quickly from any channel widening occurring during the flood (Costa, 1974; Gupta and Fox, 1974; Moss and Kochel, 1978). Gravel and sand were deposited only at scattered sites on flood plains, and extensive erosion of the flood plain was not documented.

Differences in geomorphic effectiveness appear to result in part from the controlling influence of valley physiography on flood hydraulics in the Valley and Ridge province (Miller, in press). The three forks of the South Branch Potomac River are narrow and steep, with average gradients between 0.002 and 0.009 and average width of the valley floor varying between 200 and 400 m for drainage areas up to about 800 km^2. Structural and lithologic controls effectively prevent any systematic increase in valley width as drainage area increases. Furthermore, these valleys also have many local expansions and contractions of val-

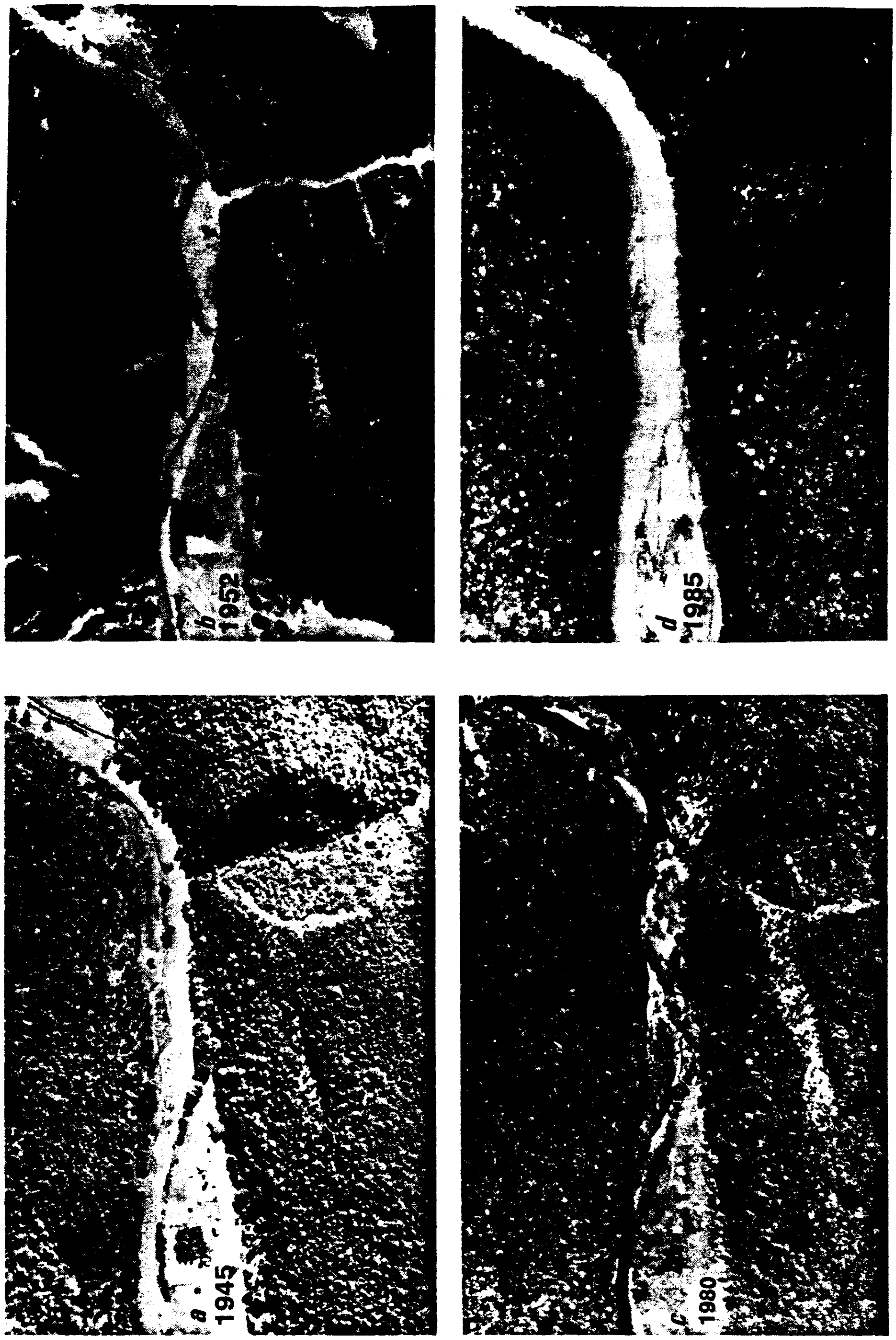

Fig. 11. Sequence of four aerial photographs of site along the South Branch River upstream of Petersburg, W. Va. described in text. The valley width is approximately 50 m; flow is to left.

ley width that create backwaters upstream from the contractions and severe erosion of the valley floor by jet-shaped flow patterns downstream from the valley expansions. Preliminary investigations suggest that unit stream power, which is a function of both slope and discharge and varies inversely with width, may be more useful than discharge alone as an indicator of geomorphic effectiveness.

Effectiveness in the hillslope regime

Geomorphic effectiveness of landsliding is at a minimum in the Piedmont where, apparently, thick, well-drained saprolite rarely accumulates pore-water pressures sufficient for failure. In other areas where a variety of types and sizes of landslides occur, relative effectiveness must be measured in terms of the size of the resulting landforms and their recovery time.

For landslide scars, recovery time is the time required to re-establish critical thicknesses of colluvium and (or) residuum on a hillslope site. Re-establishment of vegetation is not considered recovery of a landslide scar. Failure susceptibility generally increases with the time interval between triggering events as failure thresholds are steadily lowered.

The time required for recovery of landslide scars is variable depending on the rock type involved and the capability of high-frequency, low-magnitude processes to operate on filling the scar. For example, shallow landslides triggered in 1985 on steep pastured slopes underlain by thin, shaley residual and colluvial soils in Pendleton County, W. Va., have substantial natural revegetation after only three years. Ravel and wash from steep scarps around slip scars have filled portions of scars with 2–5 cm of fine colluvium. As scarp steepness declines, however, rates of filling are expected to decrease and total recovery to prefailure thickness may require many years. On shaley rocks in the Buffalo Creek basin (Fig. 1), diffusional colluvial transport by bioturbation and creep, and weathering of bedrock into failure-susceptible residuum were calculated to require several thousand years for a landslide scar to recover to mean failure conditions, although head and toe scarps disappear much more rapidly (Jacobson, 1985). Scars from large debris slide-avalanches in Nelson County, Va. (1969) and Grant County, W. Va. (1949) persist as prominent landscape features. In both cases, landslides stripped colluvium and weathered bedrock from hillslope sites; refilling with colluvium derived from side slopes has been minimal and weathering of bedrock negligible.

Recovery of landslide deposits is a fundamentally different process from recovery of landslide scars. Many landslides produce sediment that falls or flows and accumulates in debris cones, colluvial benches, or debris fans. Unlike scars, these deposits can have much longer residence times if they are on low-angle, inactive slopes and (or) if the deposit itself is resistant to erosion. Some clayey mudflows triggered in the November 1985 storm in Pendleton County, W. Va. have been in corporated into pre-existing colluvial cones and benches. After three years, only very subtle turf levees remain; however, the colluvial landforms may integrate many such events over very long time intervals, and thus establish aggregate effectiveness. In contrast, individual bouldery debris deposits from 1949 debris slide-avalanches still retain primary depositional features like lobes, levees, and fans. Older, prehistoric deposits also are evident. The most dramatic examples of Appalachian landslide features with long residence times are the extremely large rock-block slides and debris avalanche features identified by Schultz (1986) and Southworth (1987). The size (several million to 1 billion cubic meters) and resistant materials that comprise these features, assure that the landforms will remain prominent for thousands of years.

The surficial geologic record of catastrophic events

Surficial deposits contain a discontinuous, transient record of depositional geomorphic events. The history contained in these deposits is essential for understanding the spatial distribution and long-term recurrence of rare catastrophic events. An example of the utility of such data is the formal statistical procedures that have been developed recently to incorporate paleoflood information directly into frequency analyses of rare floods (e.g., Hupp, 1988). In other cases, however, progress would be made if we could simply count the number of large events over a time interval and assign heuristic recurrence intervals.

As the time interval of interest is extended backward into the middle Holocene, it is increasingly tenuous to assume stationary climatic conditions. One of the most important goals of surficial stratigraphic research is to determine what portions of the episodic, surficial stratigraphic record are attributable to extreme events under the present-day climate as opposed to major Quaternary climatic changes. This section reviews selected surficial stratigraphic studies in the central Appalachians to consider the role of catastrophic floods and landslides in landscape evolution during the Quaternary.

Valley bottoms and alluvial terraces

Interpretation of catastrophic flood histories from valley-bottom alluvium and alluvial terraces requires comparisons to effects of historic catastrophic flooding. In the Piedmont, the Hurricane Agnes flood in 1972 significantly widened channels but produced only minor flood-plain deposition in rare areas that were hydraulically favorable (Gupta and Fox, 1974; Costa, 1974; Moss and Kochel, 1978). Reasoning from the Agnes example, we would not expect isolated extreme flood events like the Agnes flood to be extensively recorded in the Piedmont valley-bottom stratigraphy.

Few studies exist on prehistoric alluvial stratigraphy of Piedmont valley bottoms. Wolman and Leopold (1957) used a conceptual model that flood plains form from floods of 1–2 year recurrence and radiocarbon dates from Pennsylvania and North Carolina to suggest that Piedmont flood plains are young (less than 2000 yr) and adjust rapidly to frequent flood events. In view of paleobotanical evidence that the eastern U.S. has been dominated by woodlands for most of the Holocene (Knox, 1983; Delcourt and Delcourt, 1986), and the historic insensitivity of Piedmont streams to extreme events, it is unlikely that Holocene climate changes or catastrophic floods have been recorded unambiguously in flood plain strata of stream basins located entirely within the Piedmont. In contrast, agriculturally induced upland erosion produced a distinct, broadly distributed layer of silty flood-plain sediment throughout the Piedmont of Maryland (Jacobson and Coleman, 1986), Georgia (Trimble, 1974), and presumably elsewhere in the Piedmont. In the Maryland Piedmont, agricultural sediments are much more extensive and distinct that any produced by major floods, including the Tropical Storm Agnes flood. Hence, it would appear that a widespread, intense, and long-lasting disturbance equivalent to agricultural disturbance is necessary to produce an unambiguous stratigraphic record in the Piedmont.

In contrast to the Piedmont, stream valleys of the Blue Ridge, Valley and Ridge, and Plateau, and streams draining from these areas across the Piedmont, exhibit more evidence of episodic deposition. Historical observations, again, help to calibrate the surficial record. Hurricane Camille in 1969 produced dramatic widening of stream channels with small drainage areas and extensive deposition of alluvium and debris-flow in channels and fans (Williams and Guy, 1973; Johnson, 1983). Many of these features are distinct today. In the Valley and Ridge, the 1949 storm around Harrisonburg, Virginia produced channel widening, flood-plain incision, and deposition of levees and flood-plain bars (Hack and Goodlett, 1960). Similar features were created by the 1985 storm in the same area (Kochel et al., 1987); the effects of the 1985 storm on the South Branch Potomac River basin have been described above. All these storms occurred in mountainous areas with thin soils, steep stream gradients, and coarse sediment supply. The 1949 and 1969 events also were characterized by extensive delivery of coarse landslide debris to the floods. These conditions appear to be the most favorable for creation and preservation of a stratigraphic record of extreme events.

There have been only a few alluvial stratigraphic studies to document the prehistoric record of mountainous parts of the central Appalachians. Most of these studies have attributed flood-plain strata or terraces to climate change rather than individual catastrophic events but the number of observations are insufficient to generalize over the region.

Multiple fans and terraces have been studied by Kite et al. (1986) along the west flank of the Blue Ridge and in the Valley and Ridge of Virginia. They observed post settlement alluvium, an indistinct Holocene record, and extensive alluvial terraces and fans of undetermined age, some of which are extensively weathered and are probably pre-Holocene. Terraces along the Shenandoah River in the vicinity of the Thunderbird archeological site also have been attributed to major Pleistocene climate changes (Segovia and Franco, 1977; Foss, 1977).

Alluvium of the South Branch Potomac River valley near Moorefield and Petersburg, W. Va. appears typical of the Valley and Ridge. A complex series of cut and fill episodes is represented by strata in the valley bottom (Fig. 12). The oldest valley bottom unit is presumed to be early Holocene to Late Wisconsin based on relative weathering criteria. A younger aggradational sequence has yielded basal dates of 7060 +/− 230 yr BP (W-6077) and 2170 +/− 180 yr BP (W-6076) at two separate locations (R.B. Jacobson, unpubl. data). Without additional radiocarbon dates, episodic deposition within the valley bottom sediments cannot be attributed reliably to either climatically induced periods of increased flooding or individual catastrophic floods. Similarity of the cut and fill scour features in valley bottom alluvium to flood-plain chutes produced by the 1985 flood, however, suggest that they may have been incised by a catastrophic flood event and subsequently filled by moderate flood events. At least three higher, undated, highly weathered terraces flank the valley bottom. We interpret the major terraces to reflect aggradation under Pleistocene climates when the capacity of streams to transport sediment eroded from slopes was proportionally less compared to present conditions.

Valley-bottom alluvium in the Buffalo Creek drainage basin (Fig. 1) may be typical of small drainage basins in the Appalachian Plateau. Alluvium there is composed of uniform silty and sandy units, lacking buried weathering profiles, and dominated by late-Holocene radiocarbon ages of less than 4000 yr. (Jacobson, 1985). This was interpreted as evidence that the area had experienced no major climate changes over this time interval and that extreme storms, if they had affected the area, had not left distinct stratigraphic evidence.

In another alluvial stratigraphy study in the Plateau, an archeological excavation along the Kanawha River (26,700 km^2 drainage area, Kanawha County, W. Va.) exposed 5.5 m of stratified sediment with seventeen occupation zones spanning the years 9850–8160 yr BP (Broyles, 1971). If each occupation zone corresponds to a depositional hiatus between flood events, the average recurrence interval is approximately 100 years. Variations in the amount of sediment between occupation zones suggest that flooding conditions were not uniform during the time interval and the entire sequence is thought to reflect a climatically controlled period of aggradation unlike the late Holocene (Olafson, 1971; Barlow, 1971).

Older and higher alluvial terraces along ma-

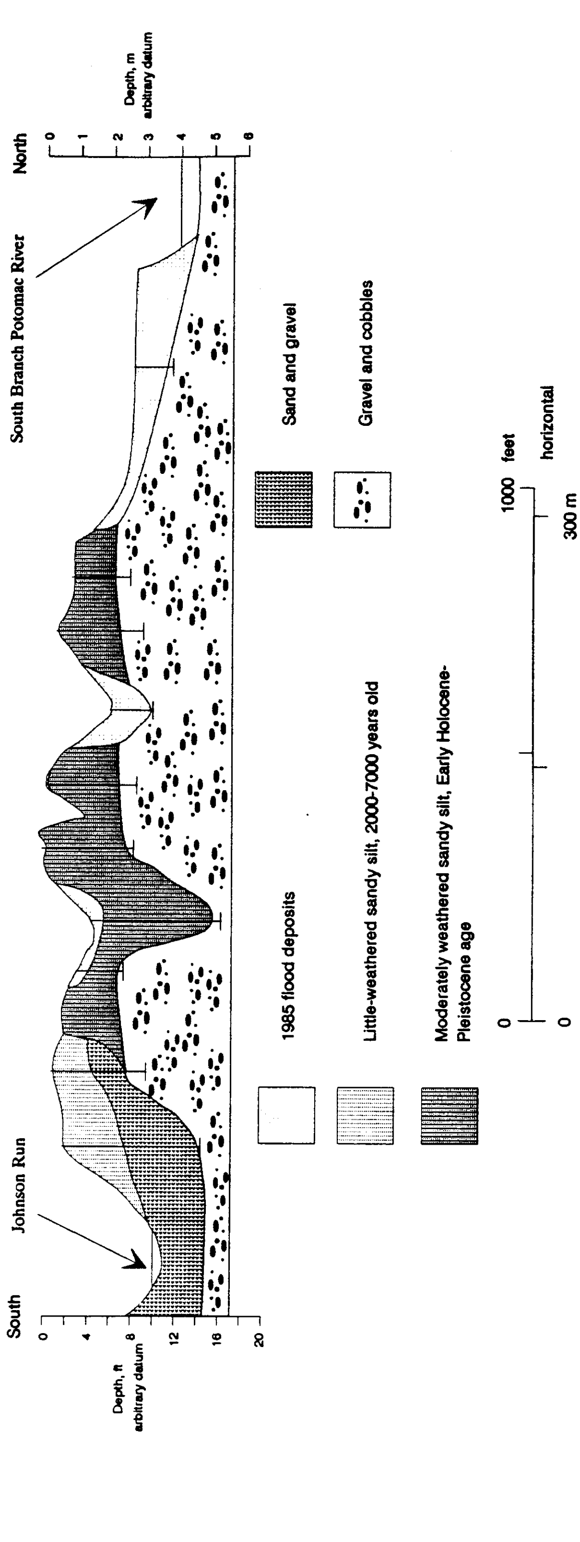

Fig. 12. Cross section of valley-bottom alluvium near Petersburg, West Virginia, showing 1985 flood deposits and erosional features compared to underlying deposits.

jor rivers in the central Appalachians are considered generally to be Pleistocene and pre-Pleistocene in age. These include terraces along the New River (Mills and Wagner, 1985), Monongahela River (Jacobson et al., 1988), Rappahannock River (Dunford-Jackson, 1978), Rappahannock Estuary (Newell, 1985), and Potomac River (Reed, 1981). Episodic terrace formation for these rivers over the late Cenozoic has been ascribed variably to climate change and tectonism.

In summary, the alluvial stratigraphic record in the central Appalachians has two contrasting components. The high-level terrace remnants that comprise some of the most dramatic elements of the landscape were deposited during the Pleistocene or earlier, as the result of climatically induced changes in sediment supply and (or) hydrology, or tectonism. Meagre studies of alluvium in the valley flats suggest that aggradational episodes have been associated with both climatically induced periods of more frequent flooding and with catastrophic floods. Based on historic flood effects and preserved flood-plain morphologic features and deposits, it is apparent that catastrophic flooding is more effective in sculpting flood plains in steep basins of the Valley and Ridge and Blue Ridge, than in the Piedmont and Plateau.

Hillslope sediments

The surficial stratigraphic record of colluvium, debris, and related slope deposits is the most direct available evidence for the sediment supply side of the geomorphic system. As such, hillslope sediments may comprise a more sensitive record of extreme storms and climatic events than alluvial strata. Additionally, colluvium and debris can have longer residence times, especially in cases where colluvium and debris are composed of coarse materials that resist erosion.

In the Piedmont, scattered, thick, colluvial deposits have been attributed to episodic depostion, but lack of datable materials or a reliable stratigraphic framework has precluded a definite paleoclimatic interpretation (Whittecar, 1985), although thick, colluvial deposits in North and South Carolina have been attributed to cool-wet paleoclimates (Whitehead and Barghoorn, 1962; Eargle, 1977).

Debris and alluvial fans along Blue Ridge of Virginia have been described by Kochel and Johnson (1984), Kochel and Simmons (1986), and Kochel (1987). Radiocarbon dates on debris flow deposits were used to distinguish three or four debris-flow events in the last 11,000 years, for an average recurrence interval of 3000–4000 yr for the area upslope from the fans. Kochel (1987) interpreted the fan stratigraphy to reflect increased debris-flow activity due to the influence of tropical air masses following the northward migration of the Polar Front at the end of the Holocene. In contrast to the relatively active debris fans east of the Blue Ridge, Kochel (1987) describes inactive, coalescent alluvial fans along the western flank that are interpreted to be pre-Holocene age based on weathering criteria.

In the Valley and Ridge of West Virginia, toeslope colluvial deposits with a source area in fine shale bedrock have revealed an episode of colluviation radiocarbon dated from 12,000–10,000 yr BP (Fig. 13). These dates may indicate a series of landsliding events at this time, or alternatively, they may mark the transition from a time of general hillslope instability during the late Wisconsin, to a period of less frequent landsliding during the Holocene. Overlying the thick, late Wisconsin colluvium are at least two debris flows separated by a distinct soil horizon: the top one is from 1985, the lower was radiocarbon dated as 315 yr BP. On adjacent slopes situated below quartzite ridges, extensive debris terraces composed of mixed matrix-supported diamicton and clast-supported gravel, ascend in steps to the tops of spur ridges at elevations up to 50 m above modern channels (Fig. 14). Although undated, elevations of these terraces above present-day stream channels and lack of weathering suggest the higher terraces

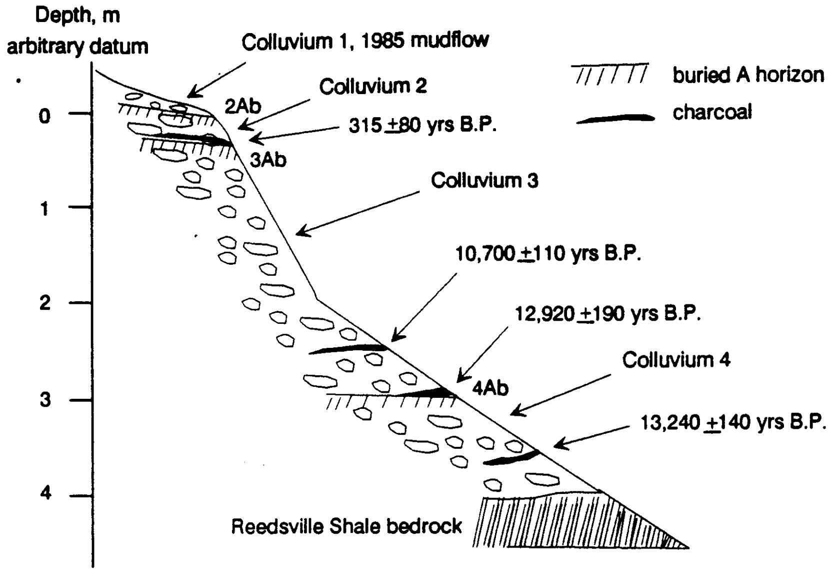

Fig. 13. Stratigraphic section of colluvium in Penleton County, W. Va. showing episodic debris-flow deposition. Radiocarbon identification numbers are: AA-4350 (315 yr BP), AA-4351 (10,700 yr BP), AA-4352 (12,900 yr BP), and AA-4353 (13,240 yr BP).

may be as old as middle Pleistocene or older. Lack of weathering of the lowest, youngest unit suggests it is probably Holocene age. Significantly, all but the youngest debris terrace are much more extensive in width than large debris flows triggered nearby in the same landscape setting in 1949 and 1985. Hence, the older remnants were either deposited by much larger single events, or they were produced by lateral aggradation of numerous smaller events. Stratigraphic exposures are not sufficient to chose between these alternatives. However, the large apparent time intervals between periods of debris terrace aggradation suggest episodicity with a frequency comparable to Pleistocene climate change.

Similar debris deposits have been noted in southwestern Virginia (Mills et al., 1987; Mills, 1989). Mills (1989) subdivided debris deposits into those produced by debris flow and those produced by gelifluction. Based on relative weathering characteristics, Mills (1989) also concluded that younger deposits tended to be emplaced by debris flows and older deposits showed more influence of periglacial processes.

Studies of colluvium in the Valley and Ridge of Pennsylvania also have documented episodic colluvial deposition (Ciolkosz et al., 1979; Sevon, 1985; Hoover and Ciolkosz, 1988). Although no absolute dates exist, regional climatic models and the existence of periglacial sedimentary features suggest that these units were deposited during Pleistocene cold periods.

Similarly, thick weathered colluvial deposits in the Buffalo Creek drainage basin have been attributed to accelerated colluviation under early Wisconsin paleoclimatic conditions (Jacobson, 1985). These thick diamicton colluvial deposits were then entrenched and eroded materials were incorporated in younger debris fans that grade to late Wisconsin (?) to late Holocene age alluvial units. The younger deposits have a higher percentage of clast-supported units compared to the early Wisconsin deposits, thus indicating a change in dominant transport processes from the early Wisconsin to the late Wisconsin. Holocene fans have multiple units but individual dates for events could not be obtained. Because the associated alluvial stratigraphic record did not disclose regional

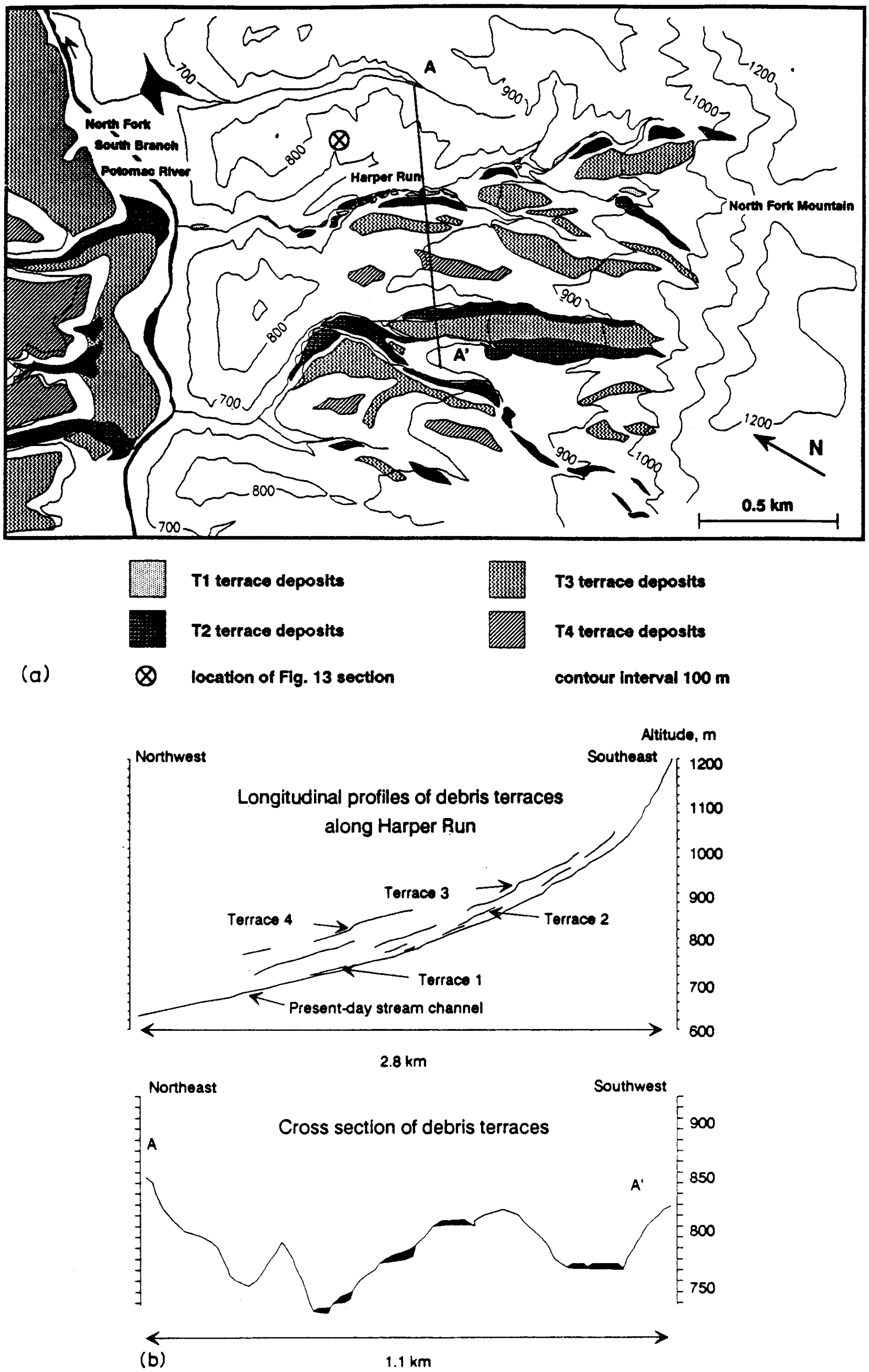

Fig. 14. Map of debris terraces in Pendleton County, W. Va., with cross section and longitudinal profile.

Fig. 15. Aerial photograph of muddy debris flows triggered in Pendleton County, W. Va. in November, 1985. Bulk of debris-flow volume in central part of photo was intercepted and stored on a Pleistocene debris terrace, Q_{dt}.

depositional events, Jacobson (1985) interpreted the fan stratigraphy to reflect asynchronous destabilization by small landslides scattered in time and space. Relict prehistoric debris flows were also recognized in southwestern Pennsylvania by Pomeroy (1983), who interpreted them as the product of the Late Pleistocene climate. Pleistocene climates are also invoked to explain thick colluvial deposits on the eastern edge of the Appalachian Plateau (Jordan et al., 1987).

Under present-day conditions Pleistocene deposits continue to influence moderate and extreme geomorphic events. The debris terraces described above (Fig. 14) act as long-term sediment storage sites that intercept sediment produced by landslides on adjacent slopes, thus mitigating the effects of extreme events (Fig. 15) Miller (1987) noted that valley constrictions formed by debris fans were important hydraulic controls that influenced erosional and depositional effects of 1985 flood waters on the South Branch Potomac River.

In summary, the surficial stratigraphic record of hillslope processes in the central Appalachians is dominated by deposits that owe their origin to climate change. In the Piedmont, where slopes presently are insensitive to catastrophic storms, the stratigraphic record shows old, episodically deposited sediments. In the Plateau, scattered occurrences of landslides under present-day climatic conditions have produced thin, scattered deposits and landforms while most of the colluvium present is Pleistocene age (see also Braun, this volume). In the Valley and Ridge and Blue Ridge provinces, where modern-day events have produced distinct and dramatic deposits, colluvial and debris deposits record large Holocene events, but even these are much less extensive than those produced during the Pleistocene.

Discussion and conclusions

Most historic catastrophic geomorphic events in the central Appalachians have been triggered by extreme rainfall events from either tropical storms or from small mesoscale features associated with frontal systems and extra-tropical cyclones. Our model of spatially varying rainfall indicates high probability that all points within the Valley and Ridge province have been subjected during the Holocene to rainfall amounts that historically have been sufficient to trigger catastrophic geomorphic events (Fig. 4). Although extreme rainfall amounts will vary across the region with climate and physiography, this conclusion suggests that all parts of the central Appalachians have been affected by one or more extreme rainfall events during the Holocene.

Tropical storm tracks and topographic barriers of the Blue Ridge and Allegheny Front act to influence the spatial distribution of the extreme rainfall. These topographic and climatic controls are superimposed on the highly variable physiography of the Central Appalachians to determine variable hydrologic responses and landslide susceptibilities. Geologic and physiographic controls strongly modify flood discharges and determine landslide locations when floods and landslides are triggered by moderate storm events, but intrinsic controls are less important for truly rare, catastrophic events when storm characteristics dominate. Size of storm, track and location of storm elements with respect to basin boundaries, and intensity and duration of rainfall combine to determine peak discharges occurring at different drainage areas. Locations and spatial densities of landslides are primarily determined by rainfall intensity and duration relative to the infiltration and drainage rates of regolith.

In contrast to the direct dependency of extreme flood discharges on storm characteristics, the geomorphic effectiveness of central Appalachian floods are significantly modified by basin geology. Extreme discharges in the Piedmont have left little lasting geomorphic record because the deposits can be reworked by subsequent moderate flows and sufficient fine sediment is available to fill in widened channels. In contrast, smaller discharges in steep areas of the Blue Ridge and Valley and Ridge have been effective in producing geomorphic change because of the direct sources of coarse sediments supplied by hillslopes and high stream powers produced in steep channels. Furthermore, geologic constraints like the water gaps in Valley and Ridge impose hydraulic controls that increase stream power relative to similar discharges elsewhere in the Appalachians.

Catastrophic landslides are effective geomorphic events when they are triggered in hard rocks on slopes, producing prominent scars and deposits with long residence times. Despite relatively common occurrence of long-duration, high-intensity rainfall from tropical storms, the Piedmont landscape is minimally susceptible to landsliding because the thick and highly permeable saprolite there allows infiltration and drainage of rainfall without allowing soil moisture to accumulate to critical levels. The same storms can trigger many landslides on steep slopes of the Blue Ridge. In the Valley and Ridge, shaley regolith is more likely to be destabilized by long-duration, low-intensity rainfall whereas the steeper, better-drained sites on sandstone ridges are more likely to destabilized by short-duration, high-intensity rainfall. However, scars and deposits from the small landslides on shale recover quickly whereas those from the sandstone ridges remain prominent for much longer. Thin, residual and colluvial soils on shaley lithologies of the Plateau have not been susceptible to catastrophic landsliding because extreme rainfall totals are relatively rare in the region; instead, most hillslopes are adjusted to more frequent movement triggered during long wet periods in the spring.

The ultimate geomorphic effect of a storm results from the combination of landslides and floods. The importance of sediment supplied directly from landslides to flood waters was obvious in Camille (1969) and the 1949 storm when sediment from debris slide-avalanches was entrained in the flood and was deposited downstream on fans and flood plains (Johnson, 1983; Schleiff and Kite, 1989). Less directly, correlations of the locations of thousands of landslides triggered by the November 1985 storm with areas of flood erosion, suggests that sediment introduced from many small landslides can contribute in aggregate fluvial erosion, probably by increasing flow density (Harper et al., 1987). Thus, the most catastrophic of central Appalachian geomorphic events will be those in which conditions simultaneously promote landslides and high flood discharges.

The answer to the question of whether catastrophic events or smaller, more-frequent events are the most important in modifying the central Appalachian landscape under present-day conditions cannot be answered reliably until more data are gathered on meteorology, hydrology, and sediment transport. Without these data, recurrence intervals and geomorphic work cannot be evaluated. However, the foregoing discussion of spatial variations of intrinsic and extrinsic geomorphic controls across the central Appalachians implies that the answer will vary with location. Our preliminary conclusion is that catastrophic events are currently less important in shaping the Piedmont and Plateau than the Blue Ridge and Valley and Ridge. This results primarily from the interplay of decreasing importance of tropical storms east to west and increasing sensitivity to catastrophic events in areas of hard rocks and high relief.

Because the historic record is too short for adequate assessment of the long-term importance of catastrophic events, the surficial stratigraphic record becomes an important source for additional data. Most alluvial stratigraphic studies attribute terrace formation to aggradation triggered by changing climates during the Holocene and Pleistocene. However, the detailed mechanisms of such aggradational events have not been explored. In particular, we do not know how changing meteorological patterns and seasonal climate changes have interacted to determine relative changes in sediment supply, hydrologic response, and dominant transport processes. Periods of increased flooding during the Holocene have been documented at other sites in the eastern U.S. (Knox, 1983, 1988; Brakenridge, 1980) and should be expected in the central Appalachians. Conditions for preservation of stratigraphic evidence vary widely across the Appalachians because of variable hydrology and hydraulic settings. Conditions for preservation of a long stratigraphic record are more favorable on low-energy rivers like the Kanawha (Barlow, 1971) than high-energy rivers like the South Branch Potomac River or smaller streams like Buffalo Creek.

Evidence for individual extreme floods has not been distinguished stratigraphically from climatically induced periods of aggradation in the central Appalachians. Detailed stratigraphic studies like those of Knox (1987) have been successful in documenting extreme and moderate floods in the midwest U.S. and could be applied to the central Appalachians. Dendrochronological paleoflood analyses have been used successfully to date individual extreme floods (Hupp, 1988) but slackwater methods appear less promising, because few hydraulically suitable sites are available (Kite and Linton, in prep).

The surficial stratigraphic record of hillslope sedimentation is dominated by discrete time intervals of aggradation. Weathering of the most extensive deposits suggests that they formed during the Pleistocene, probably due to an increase in the rate of regolith production on hillslopes relative to the transport capability of streams. The surficial record also shows that Holocene episodes have been smaller in aggregate than Pleistocene episodes. Furthermore, sedimentological data (Mills, 1989) and observations of remnant periglacial features (Clark and Ciolkosz, 1988) show that in some parts of the central Appalachians, hillslope transport was at least in part by periglacial processes, presumably during Pleistocene cold periods. Therefore, aggradational episodes probably are not solely due to increased magnitude and frequency of catastrophic events relative to the present.

In general, the surficial stratigraphic record of the central Appalachians indicates two timescales of variation. Most of the region exhibits multiple, prominent hillslope and alluvial terrace deposits indicative of climate changes with durations and recurrence frequencies of tens of thousands of years or more. Historic and late Holocene catastrophic events create, add to, or erode these features, but their effects are generally smaller. Thus, the events that have been the most effective in sculpting the Appalachian landscape are largely the result of events associated with what appears to have been a very different geomorphic climate in the Pleistocene. Present-day catastrophic events, although effective in modifying the landscape in some parts of the central Appalachians, have not yet erased much of the legacy of previous climates.

Acknowledgements

We are indebted to M. Rubin, G. Burr, and D. Donohue for providing radiocarbon dates.

References

Bauley, J.F., Patterson, J.S. and Paulhus, J.L.H., 1975. Hurricane Agnes rainfall and floods, June–July 1972. U.S. Geol. Surv. Prof. Pap., 924: 403 pp.

Barlow, J.A., 1971. Recent geological historical of the St. Albans site. In: B.J. Broyles, (Editor). The St. Albans Site, Kanawha County, West Virginia. W. Va. Geol. Econ. Surv., Rep. Archeol. Invest., 3: 80–92.

Brakenridge, G.R., 1980. Widespread episodes of stream erosion during the Holocene and their climatic cause. Nature, 283: 655–656.

Braun, D.D., 1989. Glacial and periglacial erosion of the Appalachians. In: T.W. Gardner and W.D. Sevon (Editors), Appalachian Geomorphology. Geomorphology, 2: 233–256 (this volume).

Broyles, B.J., 1971, The St. Albans site, Kanawha County, West Virginia. W. Va. Geol. Econ. Surv. Rep. Archeol. Invest., 3: 1–76.

Camp, J.D., and Miller, E.M., 1970, Flood of August 1969 in Virginia. U. S. Geol. Surv. Open File Rep., 120 pp.

Carr, C.E. and Jacobson, R.B., 1988. Physical properties of regoliths and the distribution of slope movements triggered by the November 3–5, 1985, storm, Pendleton County, West Virginia. EOS, Am. Geophys. Union, 69: p. 348 (abstr.).

Ciolkosz, E.J., Petersen, G.W., Cunningham, R.L. and Matelski, R.P., 1979. Soils developed from colluvium in the ridge and valley area of Pennsylvania. Soil Sci., 128: 153–162.

Clark, G.M., 1987. Debris slide and debris flow historical events in the Appalachians south of the glacial border. In: J.E. Costa and G.F. Wieczorek, (Editors), Debris flows/avalanches: process, recognition, and mitigation. Geol. Soc. Am., Rev. Eng. Geol., VII: 125–138.

Clark, G.M. and Ciolkosz, E.J., 1988. Periglacial geomorphology of the Appalachian highlands and interior highlands south of the glacial border - a review. Geomorphology, 1: 191–220.

Clark, G.M., Jacobson, R.B., Kite, J.S. and Linton, R.C., 1987. Storm-induced catastrophic flooding and related phenomena in Virginia and West Virginia, November, 1985. In: L. Mayer and D. Nash (Editors), Catastrophic Flooding. Allen and Uwin, Boston, pp. 355–379.

Costa, J.E., 1974. Response and recovery of a Piedmont watershed from tropical storm Agnes, June 1972. Water Resour. Res., 10: 106–112.

Delcourt, H.R. and Delcourt, P.A., 1986. Late Quaternary vegetation change in the central Atlantic states. In: J.N. McDonald and S.O. Bird (Editors), The Quaternary of Virginia. Va. Div. Miner. Resour., Publ., 75: 23–35.

Dunford-Jackson, C.S., 1978. The geomorphic evolution of the Rappahannock River basin. Unpubl. M.S. thesis, University of Virginia, Charlottesville, Va., 92 pp.

Eargle, D.H., 1977. Piedmont Pleistocene soils of the Spartanburg area, South Carolina. Geol. Notes, S.C. Div. Geol., 21: 57–74.

Eisenlohr, Jr., W.S., 1952. Floods of July 18, 1942 in north-central Pennsylvania. U.S. Geol. Surv., Water Supply Pap., 1134B: 98 pp.

Environmental Data Service, 1968. Climate atlas of the United States. U.S. Dep. Commer., Environ. Sci. Serv. Admin., 43 pp.

Foss, J.E., 1977. The pedological record at several paleoindian sites in the northeast. New York Acad. Sci., Ann., 228P: 234–244.

Gross, M.G., Karweit, M., Cronin, W.B. and Schubel, J.R., 1978. Suspended sediment discharge of the Susquehanna River to northern Chesapeake Bay, 1966 to 1976. Estuaries, 1: 106–110.

Grover, N.C., 1937. The floods of March 1936 - 3 Potomac, James, and upper Ohio Rivers. U. S. Geol. Surv., Water Supply Pap., 800: 351 pp.

Gryta, J.J. and Bartholomew, M.J., 1987. Frequency and susceptibility of debris avalanching induced by Hurricane Camille in central Virginia. In: A.P. Schultz and C.S. Southworth (Editors), Landslides of eastern North America. U. S. Geol. Surv., Circ., 1008: 16–18.

Gupta, A. and Fox, H., 1974. Effects of high-magnitude floods on channel form. A case study in the Maryland Piedmont. Water Resour. Res., 10: 499–509.

Hack, J.T. and Goodlett, J.C., 1960. Geomorphology and forest ecology of a mountain region in the central Appalachians. U. S. Geol. Surv., Prof. Pap., 347: 66 pp.

Harper, J.M., Jacobson, R.B., Howard, A.D. and McGeehin, J.P., 1987. Relationship between landslides, and tributary channel changes from an extreme flood, North Fork, South Branch Potomac River basin, West Virginia. Program Abstr., Catastrophic Flooding Symp., Oxford Ohio, 1987, 32 pp. (abstr.)

Hershfield, D.M.; 1961. Rainfall frequency atlas of the United States. U. S. Weather Bur., Tech. Pap., 40: 114 pp.

Hoover, M.T. and Ciolkosz, E.J., 1988. Colluvial soil parent material relationship in te ridge and valley physiographic province of Pennsylvania. Soil Sci., 145: 163-172.

Hoxit, L.R., Maddox, R.A., Chappell, C.F. and Brua, S.A., 1982. Johnstown-western Pennsylvania storm and floods of July 19-20, 1977. U. S. Geol. Surv., Prof. Pap., 1211: 68 pp.

Hupp, C.R., 1988. Plant ecological aspects of flood geomorphology and paleoflood history. In: V.R. Baker, R.C. Kochel and P.C. Patton (Editors), Flood Geomorphology. Wiley, New York, N.Y., pp. 335-356.

Jacobson, R.B., 1985. Spatial and temporal distributions of slope processes in the upper Buffalo Creek drainage basin, Marion County, West Virginia. Unpubl. Ph. D. thesis, The Johns Hopkins University, Baltimore Md., 485 pp.

Jacobson, R.B., 1988a. Debris deposits and Quaternary drainage basin evolution in the central Appalachian Mountains of West Virginia. EOS, Am. Geophys. Union, 69: p. 1225 (abstr.).

Jacobson, R.B., 1988b. Rainfall and slope movements in the central Appalachian Mountains of West Virginia. EOS, Am. Geophys. Union, 69: p. 348 (abstr.).

Jacobson, R.B. and Coleman, D.J., 1986. Stratigraphy and recent evolution of Maryland Piedmont flood plains. Am. J. Sci., 286: 617-637.

Jacobson, R.B., Elston, D.P. and Heaton, J.W., 1988. Stratigraphy and magnetic polarity of the high terrace remnants in the upper Ohio and Monogahela Rivers in West Virginia, Pennsylvania, and Ohio. Quat. Res., 29: 216-232.

Jacobson, R.B., Cron, E.D. and McGeehin, J.P., 1989. Slope movements triggered by heavy rainfall, November 3-5, 1985, in Virginia and West Virginia, In: A.P. Schultz and R.W. Jibson (Editors), Landslides of Eastern North America. Geol. Soc. Am., Spec. Pap., 236: 1-13.

Johnson, R.A., 1983. Stream channel response to extreme rainfall events. the Hurricane Camille storm in central Nelson County, Virginia. Unpubl. Ma. Thesis, University of Virginia, Charlottesville, Va., 109 pp.

Jordan, M.K., Behling, R.E. and Kite, J.S., 1987. Characterization and late Quaternary history of the alluvial and colluvial deposits of the Pendleton Creek valley, Tucker County, WV. Geol. Soc. Am., Abstr. Programs, 19:p. 92 (abstr.).

Kite, J.S. and Linton, R.C., in prep. Depositional aspects of the November 1985 flood on Cheat River and Black Fork, West Virginia. U.S. Geol. Surv., Bull.

Kite, J.S., Bell, A. and Allamong, M., 1986. Preliminary observations on floodplain, terrace, and fan deposits in the upper Potomac River basin of Virginia and West Virginia, In: J.N. McDonald and S.O. Bird (Editors), The Quaternary of Virginia. Va. Div. Miner. Resour., Publ., 75: 121-122.

Kochel, R.C., 1987. Holocene debris flows in central Virginia. In: J.E. Costa and G.F. Wieczorek (Editors), Debris Flow/Avalanches: Process, Recognition, and Mitigation. Geol. Soc. Am., Rev. Eng. Geol., VII: 139-4155.

Kochel, R.C. and Johnson, R.A., 1984. Geomorphology and sedimentology of humid-termperate alluvials fans, central Virginia. In: E.H. Koster and R.J. Steele (Editors), Sedimentology of Gravels ans Conglomerates. Can. Soc. Pet. Geol., Mem., 10: 109-122.

Kochel, R.C. and Simmons, D.W., 1986. Quaternary alluvial fans in central Virginia. In: J.N. McDonald and S.O. Bird (Editors), The Quaternary of Virginia. Va. Div. Miner. Resour., Publ., 75: 123-125.

Kochel, R.C., Ritter, D.F. and Miller, J., 1987. Role of tree dams on the construction of pseudo-terraces and variable geomorphic response to floods in the Litter River valley, Virginia. Geology, 15: 718-721.

Knox, J.C., 1983. Responses of river systems to Holocene climates. In: H.E. Wright (Editor), Late-Quaternary Environments of the United States - Volume 2, The Holocene. University of Minnesota Press, Minneapolis, Minn., pp. 26-41.

Knox, J.C., 1987. Stratigraphic evidence of large floods in the upper Mississippi Valley. In: L. Mayer and D. Nash (Editors), Catastrophic Flooding. Allen and Unwin, Boston, Mass., pp. 155-180.

Knox, J.C., 1988. Climatic influence on upper Mississippi valley floods. In: V.R. Baker, R.C. Kochel and P.C. Patton (Editors), Flood Geomorphology. Wiley, New York, N.Y., pp. 279-300.

Lescinsky, J.B., 1987. Flood of November 1985 in West Virginia, Pennsylvania, Maryland, and Virginia. U.S. Geol. Surv., Open File Rep., 86-486: 33 pp.

McKoy, M.L., 1988. The November 1985 storm and flood and resulting geomorphic effects on South Branch Basin of the Potomac River, West Virginia. W. Va. Geol. Econ. Surv., Open File Rep. OF8803: 143 pp.

Miller, A.J., 1987. What does it take to make a geomorphically effective flood? Some lessons from the November 1985 flood in West Virginia. In: J.S. Kite (Editor), Research on the Late Cenozoic of the Potomac Highlands. W. Va. Open File Rep., OF8801: Geol. Econ. Surr., 3-30.

Miller, A.J., in press. Flood hydrology and geomorphic effectiveness in the central Appalachians. Earth Surf. Process. Landforms.

Mills, H.H., 1989. Surficial geology and geomorphology of the Mountain Lake Area, Giles County, Virginia, including sedimentological studies of colluvium and boulder streams. U.S. Geol. Surv., Prof. Pap., 1469: 57 pp.

Mills, H.H. and Wagner, J.R., 1985. Long-term change in the regime of New River indicated by vertical variation in extent and weathering intensity of alluvium. J. Geol., 93: 131-142.

Mills, H.H., Brakenridge, G.R., Jacobson, R.B., Newell, W.L., Pavich, M.J. and Pomeroy, J.S., 1987. Appalachian mountains and plateaus. In: W.L. Graf (Editor), Geomorphic Systems of North America. Geol. Soc. Am., Cent. Spec. Vol., 2: 5-50.

Moss, J.H. and Kochel, R.C., 1978. Unexpected geomorphic effects of the Hurricane Agnes storm and flood, Conestoga drainage basin, south-eastern Pennsylvania. J. Geol., 86: 1-11.

Neumann, C.J., Cry, G.W., Caso, E.L. and Jarvinen, B.R., 1981. Tropical cyclones of the North Atlantic Ocean, 1871-1980. National Climatic Data Center, 152 pp.

Newell, W.L., 1985. Architecture of the Rappahannock estuary - neotectonics in Virginia. In: M. Morisawa and J.T. Hack (Editors), Tectonic Geomorphology. Allen and Uwin, Boston, Mass., pp. 321-342.

Olafson, S., 1971. Late Pleistocene climate and the St. Albans site. In: B.J. Broyles (Editor), The St. Albans site, Kanawha County, West Virginia. W. Va. Geol. Econ. Surv., Rep. Archeol. Invest., 3: 77-80.

Osterkamp, W.R. and Costa, J.E., 1986. Denudation rates in selected debris-flow basins. Fourth Fed. Interagency Sediment. Conf., Proc. 1: 4-91-4-99.

Pavich, M.J., 1986. Processes and rates of saprolite production and erosion on a foliated granitic rock of the Virginia Piedmont. In: S.M. Colman and D.P. Dethier (Editors), Rates of Chemical Weathering of Rocks and Minerals. Academic Press, Orlando, Fl., pp. 552-590.

Pomeroy, J.S., 1980. Storm-induced debris avalanching and related phenomena in the Johnstown area, Pennsylvania, with reference to other studies in the Appalachians. U.S. Geol. Surv., Prof. Pap., 1191: 23 pp.

Pomeroy, J.S., 1982. Mass movement in two selected areas of Washington County, Pennsylvania. U.S. Geol. Surv., Prof. Pap., 1170-B: 17 pp.

Pomeroy, J.S., 1983. Relict debris flows in northwestern Pennsylvania. Northeast. Geol., 5: 1-7.

Reed, Jr., J.C., 1981. Disequilibrium profile of the Potomac River near Washington, D.C. - A result of lowered base level or Quaternary tectonics along the Fall Line? Geology, 9: 445-450.

Reed, L.A., 1980. Suspended-sediment discharge in five streams near Harrisburg, Pennsylvania, before, during, and after highway construction. U.S. Geol. Surv., Water Supply Pap., 2072: 37 pp.

Reneau, S.L. and Dietrich, W.E., 1987. Size and location of colluvial landslides in a steep forested landscape. Erosion and Sedimentation in the Pacific Rim (Proceddings of the Corvallis Symposium, August, 1987). Int. Assoc. Hydrol. Sci., Publ., 165: 39-48.

Schleiff, K.A. and Kite, J.S., 1989. Geomorphology of debris slides, debris flows, and related phenomena developed during the June 1949 flood in eastern West Virginia. Geological Society of America, Abstracts with Programs, 21: p. 56 (abstr.).

Schultz, A.P., 1986. Ancient, giant rockslides, Sinking Creek Mountain, Southern Appalachians, Virginia. Geology, 14: 11-14.

Segovia, A.V. and Franco, L, 1977. Archaeologic implications of the geomorphologic evolution of the lower Shenandoah Valley, Virginia. New York Acad. Sci., Annal., 228: 189-193.

Sevon, W.D., 1985. Pennsylvania's polygenetic landscape. Harrisburg Geol. Soc., 4th Annual. Field Trips Guidebook., 55 pp.

Smith, J.A., 1988. Regional flood frequency analysis using upper order statistics of the annual peak record. Water Resour. Res., 25: 311-317.

Smith, J.A. and Karr, A.F., in press. A statistical model of extreme storm rainfall. J. Geophys. Res.

Southworth, C.S., 1987. Prehistoric giant rockslides of the Appalachian valley and ridge province. types and characteristics. In: A.P. Schultz and C.S. Southworth (Editors), Landslides of Eastern North America. U.S. Geol. Surv., Circ., 1008: 24-32.

Stringfield, V.T. and Smith, R.C., 1956. Relation of geology to drainage, floods, and landslides in the Petersburg area, West Virginia. W. Va. Geol. Econ. Surv., Rep. Invest., 13: 19 pp.

Trimble, S.W., 1974. Man-induced soil erosion of the southern Piedmont, 1700-1970. Soil Conserv. Soc. Am., Ankeny, Iowa, 180 pp.

Wieczorek, G.F., 1987. Effect of rainfall intensity and duration on debris flows in central Santa Cruz Mountains, California. In: J.E. Costa and G.F. Wieczorek (Editors), Debris flows/avalanches: process, recognition, and mitigation. Geol. Soc. Am., Rev. Eng. Geol., VII: 93-104.

Williams, G.P. and Guy, H.P., 1973. Erosional and depositional aspects of Hurricane Camille in Virginia, 1969. U.S. Geol. Surv., Prof. Pap., 804: 80 pp.

White, E., 1976. Role of carbonate rocks in modifying flood flow behavior. Water Resour. Bull., 12: 351-170.

Whitehead, D.R. and Barghoorn, E.S., 1962. Pollen analytical investigations of Pleistocene deposits from western North Carolina and South Carolina. Ecol. Monogr., 32: 347-369.

Whittecar, G.R., 1985. Stratigraphy and soil development in upland alluvium and colluvium north-central Virginia Piedmont. Southeast. Geol., 26: 117-129.

Wolman, M.G. and Gerson, Ran, 1978. Relative scales of time and effectiveness of climate in watershed geomorphology. Earth Surf. Process., 3: 189-208.

Wolman, M.G. and Leopold, L.B., 1957. River flood plains. some observations on their formation. U.S. Geol. Surv., Prof. Pap., 282-C: 87-109.

Wolman, M.G. and Miller, J.P., 1960, Magnitude and frequency of forces in geomorphic processes. J. Geol., 68: 54-74.