Available online at www.sciencedirect.com
ScienceDirect
Forest Ecology and Management 242 (2007) 77–98
Forest Ecology and Management
www.elsevier.com/locate/foreco

Review

Forest operations, extreme flooding events, and considerations for hydrologic modeling in the Appalachians—A review

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Received 17 May 2006; received in revised form 30 August 2006; accepted 25 January 2007

Abstract

The connection between forests and water resources is well established, but the relationships among controlling factors are only partly understood. Concern over the effects of forestry operations, particularly harvesting, on extreme flooding events is a recurrent issue in forest and watershed management. Due to the complexity of the system, and the cost of installing large-scale hydrologic studies, data are usually limited. Therefore, hydrologic models are employed to evaluate specific land use issues during extreme conditions. Our objectives were to review literature regarding: (1) relevant forest hydrology concepts, (2) the effects of silviculture and forest operations on peak discharges and flood yields, and (3) the suitability of existing modeling approaches for assessing these effects on extreme peak discharges. Numerous studies have shown that the effects of forest operations on streamflow vary, and that the influence of vegetation, soils, and land use on streamflow generation diminishes as larger volumes of water are introduced to the system. The most significant impact forest operations might have on extreme flows is by routing via poorly located and designed road networks. Extreme events appear to have different hydrologic controls than lower-flow events, and that sharp thresholds may exist between these paradigms. There are a large number of hydrologic models currently available that have been developed for a wide variety of applications. Issues such as uncertainty, overparameterization, extrapolation of flood data, and logistic issues limit the use of hydrologic models for evaluating the specific controls and outcome of land-use change on extreme peak discharges.

Keywords: Best management practices; Forest operations; Flooding; Watershed management; Hydrologic modeling

Contents

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0378-1127/$ – see front matter
doi:10.1016/j.foreco.2007.01.051

1. Introduction

The impact of forestry practices, particularly harvesting, on flooding and site hydrology is a recurrent scientific, social, and political theme in watershed management (Lull and Reinhart, 1972; McCulloch and Robinson, 1993; Andreassian, 2004; Calder, 2006). Floods caused an estimated 90 billion dollars in damage in the United States during the 1990s (Pielke et al., 2002). The premise that forests and related land-use affect watershed hydrology is ancient (Lull and Reinhart, 1972; Keller, 1988; Andreassian, 2004). It seems that throughout history, people have perceived that floods were occurring with increasing frequency and devastation. That perception remains today, and may be true from a purely economic standpoint (Harr et al., 1975; Pielke et al., 2002; Yeo, 2002; DeWalle, 2003; FAO and CIFOR, 2005).

The impact of floods could be as much the result of exposure due to population pressures as changes in climate or the environmental impact of human activity. There is some evidence that the frequency of severe flooding may be increasing due to climate change and permanent large-scale changes in land use (Macklin and Lewin, 2003). In addition, there has been a slight upward trend in 1-day rainfalls greater than 50 mm in the US since the 1930s (Kunkel et al., 1999). Some attribute perceived increases in flood damage to the increased development within flood-prone areas, and conclude that there is little evidence of a connection between forest conversion and large-scale, extreme flooding (FAO and CIFOR, 2005; Calder, 2006). Therefore, the magnitude of the influence of land use on flooding, and its specific mechanisms, remains the focus of much research and debate throughout the world.

Regardless, there is little doubt that forests influence the storage and movement of water in watersheds. The removal of trees through harvesting, or conversion to other land uses, generally reduces water demand and will affect water yield particularly during the growing season. Soil disturbance, skid trails, and road systems may alter hillslope hydrology and flow routing to rivers and streams (Megahan, 1972; Wemple et al., 1996; Sidle and Onda, 2004; Ziegler et al., 2004; Sidle et al., 2006b). Changing the timing and magnitudes of hillslope runoff in response to storms may in turn increase the frequency and magnitude of local and regional floods. Despite centuries of scientific observations and research inspired by significant flooding events, many aspects about the relationship between land-use and flooding, in particular extreme flooding, remain unresolved.

Hydrologists commonly define floods as any flow event that exceeds the normal banks of a river or stream (Jarvis, 1936; Leopold and Maddock, 1954). Floods are also defined by their return period or relative frequency as the maximum event for a given year in the long term (Barrows, 1948). Hydrologic research has traditionally been directed at studying frequent, minor to moderate flood events. In contrast, the public perception of floods is often restricted to more extreme events that result in loss of life or property. Significant public and political pressure to prevent future events often follows large, damaging floods and forest protection is regularly a centerpiece of any action plan (Miller, 1997; FATT, 2002; Brzozowski, 2004; FAO and CIFOR, 2005; Calder, 2006). The perceived importance of forests as a primary mechanism for comprehensive flood protection is at the core of new litigation against forest landowners claiming that harvesting activities increase risks for major floods (Mortimer and Visser, 2004).

A general misunderstanding regarding the nature of forest hydrologic function has been propagated by over-simplifications of the water cycle (Keller, 1988; Miller, 1997; FAO and CIFOR, 2005; Calder, 2006). This is coupled with an innate political need to take action in response to natural disasters. Unfortunately, most of what is known about the functional connection between forests and flooding is restricted to information based on non-extreme events. It is possible, if not probable, that information garnered from lesser events may apply to extreme events, which may be subject to entirely different hydrologic controls (Hawkins, 1993; Gaume et al., 2003, 2004; Lavigne et al., 2004). The infrequency of extreme floods, and the effort required to properly instrument watershed studies, has severely limited reliable scientific information about extreme events. As a result, hydrologic modeling has been used to estimate flood characteristics, but results have not always been satisfactory.

The objectives of this review paper are to explore (1) relevant forest hydrology concepts, (2) the effects of silviculture and forest land uses on flooding, and (3) to evaluate the suitability of existing models and modeling approaches for assessing the effects of forest practices on flooding, and in particular extreme peak discharges (return periods of 50–500 years). Although general concepts and a variety of results will be included, some emphasis will be placed on flooding issues in the United States and particularly the Appalachian region.

The Appalachian region is broadly defined as the chain of mountain ranges and foothills along the eastern seaboard of North America extending from Alabama in the south, and Newfoundland, Canada in the north (Williams, 1978). The region may be more finely divided into a diverse set of provinces and ranges each with distinct geologic origins and characteristics. Individual peaks average around 1000 m above mean sea level, but the average elevation is approximately 600 m and ranges between 100 and 2000 m. Drainage topography is either dendridic, rectangular, or karstic (Hewlett, 1982; White and White, 1989). The Appalachian region receives between 800 and 1300 mm of precipitation annually (Ward and Trimble, 2004). It has historically been prone to flooding caused by late winter and early spring rains (and snowmelt), summer cloudbursts, and remnants of tropical systems (Perry and Combs, 1998; Perry et al., 2001a; Hicks et al., 2005). The most extensive forest cover types include oak-pine, oak-hickory, maple-beech-birch, and spruce-fir (Eyre, 1980). Less prevalent cover types include, loblolly shortleaf pine, white-red-jack pine, and elm-ash-cottonwood.

2. Contemporary forest hydrology concepts

2.1. Watershed and hillslope hydrology

Watersheds have been called the basic ecological unit because they are relatively self-contained (Lotspeich, 1980). They are useful for hydrologic work because the boundary conditions, inputs, and outputs are more easily described than at smaller scales (Hooper, 2001). Watersheds function to collect, store, and discharge water as well as alter water chemistry and modify sediment export. Streamflow is the integrated product of the unique character of climate, geology, vegetation, and soils of individual watersheds.

The simplest conception of the hydrologic budget is that when precipitation is deposited on a watershed, water moves toward the stream, is retained by above or belowground storage components, or is lost via evapotranspiration (ET) (Fig. 1)

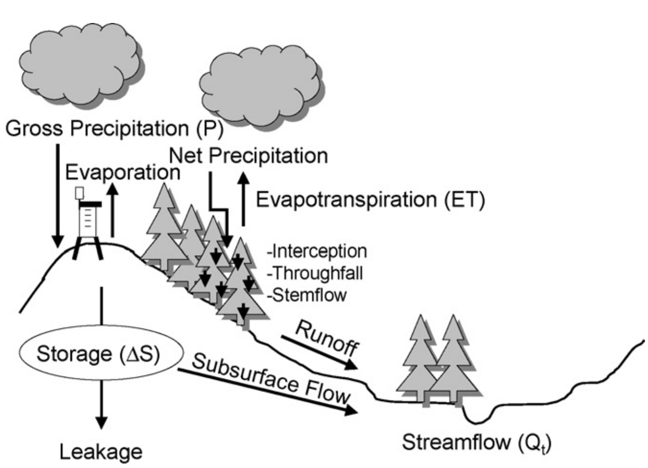

Fig. 1. A simplified water budget.

(Hewlett, 1961). Forest soils are generally characterized as having very high infiltration rates, and Hursh (1936) was the first to note that overland flow was not the main source of storm runoff in forested watersheds. More than 90% of water that falls the soils of a forested watershed annually interacts with the soil in some way before returning to the atmosphere or contributing to streamflow (Hewlett, 1982; Kirkby, 1988). Canopy interception may result in losses of 15–30% of gross annual precipitation depending on climate (Chang, 2006). However canopy storage may amount to only a few millimeters; the remainder reaching the ground via throughfall or streamflow. Approximately one-third of the water that interacts with the soil over the course of a year contributes to streamflow as interflow (shallow groundwater) or baseflow (deeper groundwater); the rest leaves via evapotranspiration. Of the water that reaches the stream, only a quarter (8% of total) does so as a result of storm flow (Hewlett, 1982). Certainly, the specific quantity in each of these pathways varies considerably with scale.

Modeling these interactions is a difficult problem due to very complex interactions among site attributes (Hibbert, 1967; Phillips, 2004). Climate, geology, and watershed geometry are all essentially fixed (Benda et al., 2004). Soils, topography, and vegetation can all be influenced by management practices and land-use. Antecedent moisture conditions also greatly affect watershed responses (Findell and Eltahir, 1997). Thus, the hydrologic response of an individual watershed can be unique, especially in the Appalachian Region where geology and soils, and their response to treatments, can be so variable (Bonta et al., 1997; Phillips, 2004).

2.2. Components of stormflow

The volume discharged over the course of the year is known as annual water yield. Water yield is typically partitioned into two components: baseflow and stormflow (quickflow). Baseflow (groundwater flow, dry-weather flow) is rather obscurely defined by the US Geologic Survey as streamflow that results from precipitation infiltrating into the soil and eventually moves through the soil to the stream channel. Stormflow is water entering stream channels promptly after rainfall or snowmelt as direct runoff. While these definitions are mutually exclusive, they are problematic in cases such as headwater systems. Stormflow in headwater watersheds may substantially utilize rapid flow pathways that are technically defined as baseflow (Hammermeister et al., 1982; Mulholland, 1993; Sidle et al., 1995). Baseflow is often arbitrarily partitioned such that direct runoff is overestimated in generalized hydrographs (Freeze and Cherry, 1979; Hewlett, 1982; Brooks et al., 2003).

Relative to understanding flooding issues, a more convenient partitioning of the hydrograph may be based on the timing of streamflow component. Stormflow might be comprised of direct components (i.e. channels and contributing road drainage), rapid components (overland flow, preferential flow in the vadose zone), slow components (preferential and matrix flow in shallow groundwater), and baseflow components (water released gradually from deep and shallow groundwater storage). Each of these proposed components may be further partitioned into specific flow pathways, which may be affected by specific changes within a watershed. Some changes, such as the installation of a road network, may increase the volume of water transported by a specific pathway such that there is an increase in the peak discharge (and therefore stage). Other changes may desynchronize pathways so that, although there is an increase in total streamflow for the event, there is little change or even a decrease in peak discharge (Gomi et al., 2002; Brooks et al., 2003).

2.3. Spatial conceptions of stormflow generation

The variable source area concept (VSAC) and Hortonian flow (Winter, 1981; Pearce et al., 1986) are two classic conceptualizations of the spatial and temporal nature of streamflow generation as it relates to processes occurring upslope. Few examples demonstrate Hortonian flow in forested watersheds although it has been predicted (Walter et al., 2003). Hortonian flow occurs when the precipitation intensity exceeds the capacity of the water to infiltrate into the soil (infiltration excess), which is in contrast to saturation excess flow where water cannot enter the soil due to saturated conditions. Although forests have high infiltration rates and rainfall intensity is normally unlikely to affect stormflow and only have a minor affect on peakflow (Hewlett et al., 1984), particularly heavy precipitation events could cause Hortonian flow to manifest itself where infiltration is poor (e.g. shallow soils, fine-textured soils, saturated soils, rock outcrops, compacted road surfaces).

The VSAC provides a more widely accepted conceptualization of streamflow generation in forested watersheds. VASC and its variants hold that only portions of a watershed area are significantly contributing as a source of flow at any time (Tsukamoto, 1963; Hewlett and Hibbert, 1967; Kirkby and Chorley, 1967). During storms, the contributing area expands as the near-stream areas become saturated and can no longer store additional water. The degree of expansion is a function of antecedent soil moisture conditions and precipitation volume and intensity. Conceptualizations such as VASC or Hortonian flow work for a wide variety of applications; however, they do not include complex hydrological processes that may be driving the system at other scales (Anderson et al., 1997; Torres, 2002). Sidle et al. (2000) extend the VASC beyond an explanation driven primarily by antecedent moisture conditions using the ‘‘hydrogeomorphic conceptual model of stormflow generation’’ (HCMSG). HCMSG attempts to address the dynamic spatial and temporal nature of steep catchments by tying soil hydrology components and geomorphology. Key additions of the HCMSG include preferential flow, soil depth, and the behavior of ‘‘zero-order’’ sub-basins, which imply non-linear responses to increasing wetness. The influence of soil hydrologic properties and antecedent moisture conditions may be expected to have decreasing influence as floods become more significant (Wood et al., 1990).

The generation of stormflow is dependent on hillslope hydrology. Rainwater takes four major flowpaths from the surrounding hillslopes into streams: (1) direct input or throughfall, (2) surface runoff, (3) shallow groundwater pathways, and (4) deep groundwater pathways (Fig. 1) (Mulholland et al., 1990; Wilson et al., 1991; Brooks et al., 2003). First order streams tend to have more rapid response times (greater flashiness) because the hydrologic gradients are greater and there are fewer system components with which water must interact on its path to the stream. Higher order streams have longer response times as multiple lower-order watersheds react differently to a given rainfall event (Brooks et al., 2003). The response time is affected by many variables. Vegetation, soils, and topography delay the stream response as water interacts with the canopy, litter layer, microtopography, soil surface, shallow and deep groundwater, and channel storage (Hewlett, 1982; Sidle et al., 2000).

Residence times along flow pathways vary from minutes to years. Surface flow is intercepted and detained by vegetation, litter, and surface irregularities that allow infiltration into the soil profile. Infiltration and subsurface flow either follows preferential flowpaths, or is absorbed into the soil matrix. The rates of exchange of water between larger and smaller pores, and between the surface and deeper groundwater system, control flow and storage within the hillslope and also control streamflow generation (Luxmoore et al., 1990; Mulholland et al., 1990; Wilson et al., 1990; Guebert and Gardner, 2001; Jones and Connelly, 2002; Torres and Alexander, 2002).

Subsurface storage and flow of water in forested hillslopes are mainly affected by the size and distribution of pores within the soil matrix and organic matter content. Mineral soils generally contain between 40% and 60% pore space by volume, of that only 20–50% may contain water readily available to plants (Shuttleworth, 1993). Pores exist in an infinite array of shapes and sizes whose structure and arrangement can have profound affects on water storage and movement. In bulk soil, macropores (>1 mm) are the primary conduits for the movement of water within the soil profile (Wilson and Luxmoore, 1988). Mesopores, drainable pores between 0.01 and 1 mm, are also significant conduits for water during storm events when the profile becomes saturated. Micropores (<0.01 mm) are the primary repositories for water storage.

The amount of precipitation that is transmitted to the stream as stormflow for an individual event may range from 4% to nearly 100% (Hewlett, 1982; Gaume et al., 2003). The sources of flow may vary. On wetland soils in a glacial outwash in Ontario, 80–90% of the stream volume for a summer storm was comprised of ‘pre-event’ water, implying that the precipitation water was replacing water stored in the system (Hill and Waddington, 1993). However, in east Tennessee relatively rapid transport was observed along preferential flowpaths due to perched water tables (Wilson et al., 1990). Eighty to 90% of the stream discharge from a spring storm event was comprised of water quickly transmitted through unsaturated and saturated soil, rather than bedrock, with little surface runoff on the hillslopes (Mulholland, 1993).

3. Forestry effects on peak discharge and flooding

3.1. Summary of published research

The primary focus of forestry hydrology research in the US has been on water quality issues (Aust and Blinn, 2004), hillslope hydrology and streamflow generation (Bonell, 1998; Cammeraat, 2002), annual water yields (Stednick, 1996), and ecological function (Swank and Crossley, 1988). Historically, the more common goal of forest management has been for the purpose of increasing water yield rather than flood control (Hursh, 1951; Douglass and Swank, 1972; Douglass, 1983; Hibbert, 1983; Krutilla et al., 1983; Ponce and Meiman, 1983). The lack of emphasis on flooding is not due to a lack of interest, but rather opportunity and the complexity of the issue. Hydrologic responses of forest watersheds to silvicultural practices have been described as ‘‘highly variable, and for the most part unpredictable’’ (Hibbert, 1967) in the past, and remain enigmatic in the present. The complexity and scale of many hillslope processes are well beyond our ability to test extreme flood hypotheses rigorously in the field or satisfactorily simulate rainfall-runoff transformations (Hooper, 2001; Cloke et al., 2003; Uhlenbrook et al., 2003; Phillips, 2004). Thus, the study of extreme flooding events has generally been the realm of geomorphological research of specific events, which are frequently observational in nature (Hack and Goodlett, 1960; Phillips, 2002; Hicks et al., 2005).

McCulloch and Robinson (1993) categorized watershed studies into three groups: correlation studies, single catchment studies, and paired catchment studies. Correlation studies utilize geologically similar watersheds that vary based on vegetation or land use. Experimental replication is key to the reliability of this approach, but these studies may suffer from variance inflation due to autocorrelation. The main limitation of correlation studies is the assumption that the treatment differences account for the majority of any differences observed. Single catchment studies calibrate watershed responses to climate, which provides a means to compare watershed post-treatment responses to a predicted pretreatment response. The limitation of this approach is that post-treatment weather conditions may fall outside the bounds for which the study was calibrated. Paired watersheds provide a means to account for the issues associated with correlation and single catchment studies. Similar watersheds are studied for a calibration period, and after a time treatments are applied leaving one unmodified catchment as a reference. The first paired watershed study in the US for forestry was established in 1909 to evaluate the effects of timber harvesting on annual water yield at Wagon Wheel Gap, Colorado (Bates and Henry, 1928). Nested catchment designs represent a fourth category of watershed studies (Sidle et al., 1995, 2000; Casper et al., 2003; Tetzlaff and Uhlenbrook, 2005). The purpose of this approach is to improve the interpretation of flowpaths and processes at multiple scales on heterogeneous watersheds.

3.2. Forestry effects on annual water yield

The most substantial impact due to forest harvesting is on baseflow and annual water yield (Hornbeck et al., 1993; Stednick, 1996; Bonell, 1998). In general, there is an inverse relationship between vegetative cover and water yield (Hibbert, 1967; Lal, 1997). Calder (1993) calculated that runoff increased 3.26 mm for each percent area deforested ($r^2 = 0.50$) based on a world-wide database of hydrologic studies. In the US the increase was slightly less at 2.46 mm ($r^2 = 0.17$) (Stednick, 1996). Studies of harvesting effects across the Appalachian region have detected increases in annual water yield from 0 to more than 400 mm (Hornbeck et al., 1993; Stednick, 1996). On average, these studies suggest that each 10% increase in harvest area results in an annual increase in runoff of approximately 28 mm. However, the specific impacts due to silvicultural treatments can vary substantially within fairly localized areas (Cornish and Vertessy, 2001). The effects on annual water yield are usually found to last less than 10 years.

3.3. Forestry effects on peak discharge and stormflow volumes

In the eastern US, the water yield from forested headwater watersheds after a typical rainfall event (>25 mm) is 0.10–0.34 of gross precipitation; meaning 10–34% of the precipitation leaves a watershed as streamflow (Hewlett, 1982). Response values for specific watersheds may range from 0.04 to greater than 0.40. In general, half of precipitation has a residence time in the watershed measured in days or less, and about 1–2% of water that reaches the forest floor leaves watersheds without interacting with soils at all. The rate at which rivers and streams respond to precipitation events is dependent on the magnitude, intensity and duration of storms, and important watershed properties such as antecedent moisture conditions, soil physical properties, geology, vegetation, topography, geomorphology (Sopper and Lull, 1967; Hewlett, 1982; Black, 2004; Chang, 2006), and land use history (Brissette et al., 2003). Changes in hydrology are usually attributed to soil disturbance, erosion, road construction, channelization, and decreased evapotranspiration.

Caissie et al. (2002) suggests that peak discharge may be the most sensitive parameter to changes in watershed hydrology for clearcuts in excess of 20% of the total watershed area or basal area. Peak discharge responses to harvesting can range between 0% and 330%, mostly regarding small and moderate events (Jones and Grant, 1996; Guillemette et al., 2005). Increases in peak discharge following timber harvesting have been attributed to decreased evapotranspiration (ET), increased soil water content (Hibbert, 1967; Hornbeck, 1973a; Dietterick and Lynch, 1989), and soil disturbance (Lal, 1997) and road networks (Jones and Grant, 1996; LaMarche and Lettenmaier, 2001). However, broader reviews of the literature seem to indicate that there is at least no consistent connection between peak flows and logging practices (Thomas and Megahan, 1998).

Brooks et al. (2003) highlights six studies that reported changes in peak discharge (−22% to 200%) and total stormflow (0–200%) after the removal of forest cover, including several studies from the Appalachian region. At Coweeta (North Carolina) peak discharges were observed to increase by 14–15%, and total storm volumes by 10% after clearcutting (Swank et al., 2001). In 1964, Hewlett and Helvey (1970) observed a 22% increase in stormflow volume at Coweeta during two separate events that approached the 7-day, 100-year return period for rainfall, seemingly to imply that forestry affects larger events; however, they did not draw this conclusion. They were less certain about the effects of forest clearing on peak flow, except that the variability of larger peakflows increased. The residuals from the two events represented the largest positive and negative deviations from the regression line.

At Fernow (West Virginia), four-fold increases in peak discharges were observed after clearcutting; however, this result was observed predominantly in smaller peak discharges (Patric and Reinhart, 1971). At Hubbard Brook (New Hampshire), peak discharge increases up to 30% were observed for 4 years after clearcutting (Hornbeck, 1973a). Total stormflow was also three times higher during the growing season, although it was still only a third of dormant season volumes. In the third-order Catamaran Brook watershed (New Brunswick, Canada; 5200 ha) harvesting had a significant effect on stormflow at relatively low volumes (0.1–1 mm); however, there was a convergent pattern such that no differences were observed for storm flows greater than 10 mm (Caissie et al., 2002). Additionally, first-order drainages appeared to be more sensitive to forest harvesting than second-order drainages; however, the total proportion of the watershed area harvested was below 20% at the larger scale. Harr et al. (1975) similarly found that the effect of forest operations became smaller for larger storms in the Oregon Coast Range.

3.4. Other land-use effects on floods

In order to better understand the impact of forest operations on hydrology, other prevalent land uses (e.g. agriculture, mining, and urbanization) should be briefly discussed. As with forestry, changes in hydrology are attributed to increased soil disturbance, erosion, channelization, and decreased evapotranspiration, and include the creation of impervious surfaces and dense disturbance patterns. Based on simulations, the potential positive effect that afforestation has for decreasing water yield on very large areas is notable (Krause, 2002; Wegehenkle, 2002). Forested land use is already associated with good stream quality compared with other land practices (Thornton et al., 2000). However, the conversion of forests to other land uses constitutes a permanent or long-term change in the hydrologic cycle compared to the temporary impact of forest harvesting (Sidle et al., 2006a). Forest ecosystems in the Appalachians, for example, have been shown to be fairly resilient to disturbances associated with conventional management (Kochenderfer and Wendel, 1983).

Agriculture is often viewed as an analog to forestry. In many areas worldwide, there is concern that conversion to agriculture has major effects on flooding (FAO and CIFOR, 2005). The impact has reportedly varied, but increases are generally more pronounced than forest harvesting due to sustained decreases in ET. In addition, removal of organic matter may lead to increased Hortonian flow (Sidle et al., 2006b). Fitzpatrick and Knox (2000) used models to predict 200% increases in peakflow and 400% increases in sedimentation for heavy agricultural activity following clearcutting in Wisconsin. In Germany, conversion of forests to agriculture reduced ET by almost half (Robinson et al., 1991). Conversion of forests to agriculture in Nigeria resulted in 500% increases in discharge for rain events between 50 and 100 mm (Lal, 1997).

Surface mining operations have probably raised some of the greatest public concerns with regards to their impacts on watershed hydrology and their role in flooding. Paired-watershed studies of surface mining in the Appalachians date back to the late 1950s and early 1960s (Phillips, 2004). With regards to flooding, results have also been mixed. While mining operations appeared to increase the risk of flooding, the risk is dependent on reclamation efforts, changes to soils or substrate, modification of drainage patterns that either increase or detain runoff, or enhance baseflow. However, it is notable that modeling studies cited by Phillips (2004) and others (FATT, 2002) universally predict increased peak discharges due to mining in spite of the mixed results from field studies. Additionally, Bonta et al. (1997) found that modeled runoff potentials after mining are not well correlated to pre-disturbance conditions. Significant controversy in the Appalachian region surrounds the practice of mountaintop removal and valley fill (MTR/VF) practices related to coal mining. Phillips (2004) finally concluded: ‘‘It is difficult to state with confidence that MTR/VF does, or does not, increase or even tend to increase peak discharges downstream. The situation is analogous to studies of downstream impacts of dams, where it has been shown that many outcomes are possible, that these outcomes are highly contingent on local conditions, and that predictions are not feasible except on a case-by-case basis.’’

Urbanization and development in flood prone areas are also concerns both from the standpoint of their hydrologic impact and placing lives and property in harm's way (Boyle et al., 1998; Bhaduri et al., 2001). The effect of urbanization is related to the percentage of the area affected, but increases in annual yield were 2–3.5 times for urbanization rates greater than 20% (Bhaduri et al., 2001). In a 70 km^2 watershed in Indiana, 49% and 67% urban coverage in 1973 and 1991, contributed 86% and 95%, respectively, to the total annual runoff volume (Bhaduri et al., 2000).

4. Mechanisms of hydrologic effects

4.1. Vegetation removal

One of the primary ways timber harvesting is thought to affect flood yield is through the removal of vegetation. The site-specific consequences of vegetation removal are not easily predicted. The responses from individual studies are variable because of climate, topography, and cover types (Stednick, 1996). In a review of 52 studies, the maximum transpiration rates for individual tree species ranged between 10 and 1190 kg day^{-1} with a median of about 75 kg day^{-1} (Wullschleger et al., 1998); the range for tree species similar to those found in the Appalachian region were slightly above the median. A typical mature, fully stocked hardwood stand for an Appalachian forest might contain approximately 350 trees ha^{-1} with an average diameter of 30 cm, a basal area around 25 $m^2\,ha^{-1}$. Although there are problems with scaling up from individual trees, if it is assumed the stand consumes above the median the daily usage might only be 2.6 and 5.3 mm day^{-1} at most.

Trees can mitigate peakflow by maintaining soil moisture deficits through ET over days or weeks thereby resulting in increased potential for soil storage and infiltration capacity. In addition, vegetative cover intercepts and detains water within the canopy delaying or preventing some water from reaching the ground. Increased streamflow (baseflow) is usually assumed to be directly proportional to the intensity of the harvest (Patric, 1978). A linear relationship is often assumed between vegetation removal and decreases in ET is based on the presumption that water is not limiting on a particular site (Hibbert, 1983). While large clearcuts may negate water use by vegetation in the short-term (months), demand on water-limited sites should quickly return as vegetation reestablishes, and may even exceed the water demand of mature stands (Vertessy et al., 1993). Additionally, harvesting disturbance may diminish soil storage and infiltration as soils are subjected to higher surface temperatures and continue to drain via baseflow pathways.

Forest cover types also influence annual yield differently. Seasonal differences in water use and canopy interception between evergreen and deciduous forests will affect watershed response. Although hardwoods generally have higher individual water demand (Kramer and Kozlowski, 1979), coniferous forests have the greatest demand for soil water, followed by hardwoods and brush/grassland (Bosch and Hewlett, 1982; Jones and Post, 2004). Bosch and Hewlett (1982) reported that for each 10% change in cover, changes in annual runoff were 40 mm for conifer forests, 25 mm for hardwood forests, and 10 mm for brush/grassland.

During storm events, up to 1.5 mm of water may be stored in the canopy of dense coniferous forests, compared to 0.5 mm in pasture or farmland (Hewlett, 1982). Interception may detain half the water deposited for small storm events. However, during larger or extreme events, this component should become negligible. Thus, as rainfall amount and intensity increase the ability of vegetation to buffer streamflow should decrease.

Holtan and Creitz (1969) calculated that the effects of vegetation on flood flows are greatest on soils with higher storage, but they diminish or disappear during longer and larger storm events. On some mixed hardwood sites in Quebec, Canada, interception by stands decreased rapidly during storms for cumulative weekly rainfalls between 0 and 10 mm the first year after harvesting (Pothier et al., 2003). On unharvested stands, interception was only 20–30% for total rainfall $\geq$20 mm. Harvests where tree removal exceeded 50% reduced interception to 5–15% for cumulative rainfalls >20 mm. After 4 years, interception increased from 12% to 20% for a 20 mm cumulative weekly rainfall, compared to 30% for the unharvested stand. In Bavaria, the reduction of peak discharges was proportionally greater for smaller storm events; peak discharges were reduced significantly over the first 10 years of new forest establishment (Robinson et al., 1991).

The essential point is that even on unharvested sites the influence of vegetation on large rainfall events is not constant and is a fraction of the influence at lower flows (Calder, 1993). Although forest cover has been shown to influence runoff during small precipitation events during the growing season, the presence or absence of forest cover has less to do with flood generation. Large precipitation events have consistently been shown to generate floods (Perry and Combs, 1998; Lecce, 2000; Perry et al., 2001a). However, research has shown that forest management features that affect flow routing such as roads, culverts, ditches, and soil impacts are more likely to influence water yields and flooding.

4.2. *Soil and stream impacts*

While vegetation removal via harvesting may have a short-term effect, long-term effects (>10 years) on hydrology are related to soil disturbance caused by road construction and subsequent traffic (Hatchel et al., 1970; Greacen and Sands, 1980; Megahan et al., 1995; Lal, 1997; Megahan et al., 2001). Soil disturbances include erosion, compaction, rutting, litter displacement, and fire. Soil disturbance can reduce infiltration or induce water repellency via physical or chemical changes, removal of organic matter, or the creation of bare soil (Worrell and Hampson, 1997; Croke et al., 2001; Wahl et al., 2003). Soil compaction due to traffic can reduce water storage capacity by decreasing total porosity, but in particular macroporosity (Greacen and Sands, 1980; Kozlowski, 1999; Miwa et al., 2004). However, macropores do not retain water, and are more related to the hydraulic conductivity of the soil; therefore the influence of compaction on flooding may be difficult to discern. Fire, depending on intensity, can consume organic matter and create hydrophobic layers in the soil that inhibit infiltration (DeBano and Rice, 1973; Megahan, 1983; DeBano et al., 1998).

Soil disturbance resulting from forest management activities could contribute to flooding if significant erosion and sediment transport result in stream aggradation and decreased channel capacity. Elevated erosion may be indicative of higher levels of surface flow in sufficient volumes with enough energy to displace soil particles. Soil erosion rates in mature deciduous forests are generally less than 1 Mg ha^{-1} year^{-1} (Patric, 1978; Jackson et al., 2005). Although the Universal Soil Loss Equation may overestimate erosion rates at large scales, harvesting on some Appalachian forests in southwest Virginia were estimated to have erosion rates of 48 Mg ha^{-1} year^{-1} immediately after harvest, but decreased to less than 10 Mg ha^{-1} year^{-1} by the second year (Hood et al., 2002). However, by the second year there were no significant differences between harvested stands and controls, and the average erosion rate over 100 years was estimated to be only 2.2 Mg ha^{-1} year^{-1}. In contrast, erosion rates on agricultural land are nearly 10-times higher; roads nearly 50 times that of forest land (Jackson et al., 2005). Erosion rates of an undisturbed forest floor during rainfalls >125 mm day^{-1} should generally be minimal.

Landslides materials alter stream channel capacity or may create natural dams that can cause significant flooding (Sidle and Ochiai, 2006). Isolated landslides can be significant sedimentation process in steep terrain (Patric and Kidd, 1982; Neary et al., 1986; Neary and Swift, 1987). Catastrophic events may increase long-term sedimentation rates by an order of magnitude (Kirchner et al., 2001).

Channel restrictions caused by the aggradation of sediments and woody debris are frequently cited as leading to more intense floods (Stover and Montgomery, 2001; Zhang et al., 2006). Sediment and debris deposition can change stream morphology by increasing overbank sedimentation, inducing bank erosion, channel downcutting, and altering stream course (Wondzell and Swanson, 1999; Phillips, 2002; Hicks et al., 2005; Zhang et al., 2006). Minor flood flows are thought to be sufficient to prevent the aggradation of fine sediments (Phillips, 2002). Annual flows rates generally maintain the streambed at a relatively constant elevation, while the surrounding floodplain aggrades, which can actually lessen flood frequency and severity of minor floods (Ruhlman and Nutter, 1999; Knox, 2001; Faustini and Jones, 2003). Extreme events, alternatively, are capable of permanently altering stream morphology (Myers and Swanson, 1996; Hicks et al., 2005). Large amounts of sediments may be flushed, replaced, and redistributed during major flood events. However, bed material and channel morphology are typically poor predictors of hydrologic regimes (Phillips, 2002).

Large woody debris (LWD) may be comprised of both logging slash, and downed riparian vegetation and are common features in undisturbed late-successional streams and recently harvested catchments (Robison and Beschta, 1990; Hedman et al., 1996). LWD has commonly been identified as a contributor to flood hazards by creating jams and anthropogenic or natural channel constrictions, or by blocking drainage structures and causing road washouts (Wondzell and Swanson, 1999; Jeffries et al., 2003; Haehnel and Daly, 2004). As a result, many state BMPs recommend against the disposal of logging slash in streamside management zones and federal BMPs forbid such disposal in jurisdictional wetlands. However, other literature has indicated that LWD plays little role in the absolute level of an individual flood peak (Geyer et al., 2000; Miller and Benda, 2000). Locally, LWD can attenuate flood runoff by the formation of levees, or exacerbate floods when logjams fail or culverts are clogged. These materials affect water and sediment routing, dissipate stream energy, as well as define and serve as habitat (Harmon et al., 1986; Maser et al., 1988; Hedman et al., 1996). LWD has been shown to improve stream habitat and perform important ecosystem functions (Gippel, 1995; Boyd et al., 2005), and its introduction has been recommended for the purpose of improving habitat for many types of streams (Seehorn, 1985; Brooks et al., 2004). However, natural recruitment may be a better source of these materials and may function differently than anthropogenic sources (Gomi et al., 2002; Faustini and Jones, 2003).

4.3. Roads

Temporary and permanent road and skid trail systems are integral components of forest management, and impact mountain watersheds and streams. Although the effects of vegetation removal alone on annual water yields generally last less than 10 years, the effect on peak discharges may be more persistent when road systems are present (Harr et al., 1975; Jones and Grant, 1996). Skid trails and roads account for approximately 10% of the total area of a typical clearcut in the Central Appalachians, or about 7 km km^{-2} (Kochenderfer, 1977). In the western Cascades, road networks may reach densities of 2–3 km km^{-2} (Harr et al., 1975; Jones and Grant, 1996; Wemple et al., 1996). Road networks can increase the effective drainage density by 40–100%.

Floods from small to moderate events from small western watersheds were as large or larger on areas that were 25% harvested with road networks compared to areas that were clearcut without roads (Jones and Grant, 1996). Thomas and Megahan (1998) conducted a follow-up study to Jones and Grant (1996) and found that increases in peak flow were greatest on the smallest events, but that treatment effects decline with time. They also concluded that the effects of harvesting activities on large watersheds are difficult in the context of diverse land uses. Harr et al. (1975) found that roads and clearcutting changed summer stormflow volumes from slightly negative to >100% increase for 16–300 ha watersheds in the Oregon Coast Range. An increased frequency and magnitude of peak discharges for relatively frequent flood events were attributed to the effect of roads on water routing (Jones et al., 2000).

Roads can interfere with natural hillslope storage and flow functions by acting as a corridor, barrier, or sink for water, bypassing some natural drainage features while being concentrated into others. Unless water is redirected at topographic breaks and other permeable sites, roads concentrate flow so that it is less likely to interact with subsurface storage that may mediate peak discharges (Wemple et al., 1996; Jones et al., 2000; Tague and Band, 2001). Road cuts can also intercept subsurface water by severing natural pipeflow, or by creating artificial areas of water resurgence by disrupting flow nets (Megahan et al., 2001; Sidle and Onda, 2004; Negishi et al., 2006; Sidle et al., 2006a). The first 10 years after road installation are the most critical period as the road system stabilizes, after which it is difficult to discern the impact on larger events (Jones and Grant, 1996).

High sediment production is indicative of road systems with few drainage controls that move large volumes of water. Swift (1984) found that roads located at Coweeta generated an average of 6.7 Mg ha^{-1} month^{-1} over a 23-month period in 1976 and 1977. The most active source of sediment on these roads was the cutslopes immediately after harvest and during the winter months (Swift et al., 1988). In Western Oregon, Megahan et al. (2001) observed sediment yields of granitic soils around 75 Mg ha^{-1} year^{-1}. Properly graveled roads (70–100 mm of gravel) produce far less sediment than poorly graveled roads (40–50 mm of gravel), which are also difficult to maintain. Grayson et al. (1993) estimated that a 4 m $\times$ 100 m road section generated as much sediment as a 30 ha watershed. In 1979, a 1000-year rain event in West Virginia, generated >50 Mg ha^{-1} for roads on stony soils; however, erosion was negligible where logging roads were covered by litter, maintained limited grades, and transmitted minimal volumes of water (Patric and Kidd, 1982).

Ideally, roads systems should be carefully designed so that environmental impacts are minimized (Kochenderfer, 1977; Walbridge, 1997; Megahan et al., 2001; Keller and Sherar, 2003). Well-designed road systems minimize connectivity with streams and conform to the topography as much as possible (Sidle and Onda, 2004). The configuration, placement, construction, and maintenance of logging roads and skid trails, particularly relative to stream intersections, plays an important role in controlling flow and protecting streams. Water control structures such as water turnouts, water bars, and broad based dips are used to control the velocity and volume of water conveyed by a road. On more heavily utilized, primary and secondary roads, culverts and ditches may also be used to control drainage. Water control structures should be used with sufficient frequency to prevent water from gaining enough momentum to cause erosion and road damage. Requirements for the type and spacing of water control structures vary from state to state (Grace, 2002; VDOF, 2002; WVDOF, 2002).

5. Watershed hydrology and flood modeling

There are a large number of rainfall-runoff models that have been developed for evaluating the effects of land-use change on stream generation and a wide variety of other hydrologic applications (e.g. Todini, 1988; Singh and Frevert, 2002a,b; Singh and Woolhiser, 2002). Singh and Woolhiser (2002) provide a comprehensive review of mathematical modeling of watershed hydrology, and cite 69 hydrologic models developed since the 1960s. Lavigne et al. (2004) mention 14 models and integrated model systems used to simulate the impacts of forest conversion on hydrologic regimes. Model selection is thus a difficult process that requires trade-offs between model complexity, capabilities, and robustness. It is often driven by the preferences of the nation, region, scientific discipline, or individuals for the sake of politics, familiarity, consistency, or commercial availability. There are few comprehensive rules for selecting a specific model for a specific application, or for deciding which parameters will be key for describing a specific system or phenomenon (Martin and McCutcheon, 1999; Andreassian et al., 2004; Wagener et al., 2004). Unfortunately, it is unusual to have detailed access to the source code of models to evaluate the consequences of model configuration for specific applications (Cloke et al., 2003). One option is to utilize a suite of models (Georgakakos et al., 2004; Reed et al., 2004); however, this approach is unlikely a logistically viable option for land managers. A detailed discussion of the considerations for model selection may be found in Martin and McCutcheon (1999) and Singh and Frevert (2002a,b).

The variety of mathematical and statistical complexity used in contemporary hydrologic models make them difficult to neatly categorize (Singh and Frevert, 2002a). Modelers often self-describe their models using somewhat descriptive nomenclature, but they do not often provide insight how the model utilizes the data or other assumptions. Broadly, models may be based on a theoretical understanding of the hydrologic cycle (physically based), or they may depend on empirically derived fits of observed data (empirical). Spatially, models may be lumped, while others are distributed or semi-distributed (Singh and Frevert, 2002a). Computationally, some models are deterministic, while others are stochastic.

Many other classification schemes for hydrologic models have been used. Wheater et al. (1993) and Wagener et al. (2004) generically categorized models based on their computational approach: (1) mechanistic, (2) parametric, (3) metric, and (4) hybrids. Todini (1988) similarly based a classification scheme on numerical approach. Singh (1995) used a broad classification scheme that considered land-use, process description, spatial and temporal scales, and modeling approach. ASCE (1996) classified flood analysis models based on their application: (1) event based rainfall-runoff models, (2) continuous precipitation-runoff models, (3) steady flow-routing models, (4) unsteady flow flood routing models, (5) reservoir regulation models, and (6) flood frequency analysis models (Singh and Woolhiser, 2002). Lavigne et al. (2004) classified models based on data inputs: (1) global data, (2) spatial data, (3) physical data, and (4) spatial data with physical bases. Because the nature of the available data drives most flood analyses, categorizing based on the models' computational approach as described by Wheater et al. (1993) is probably the most useful starting point for the purpose of outlining specific pros and cons of modeling approaches.

Mechanistic (also physically based) models use mathematical representations of physical laws to abstract the actual processes that control water movement (Wheater et al., 1993; Wagener et al., 2004). These are the most detailed and precise abstractions of watersheds behavior. However they are limited by their requirement for large amounts of site-specific data and computing power. The most serious issue concerning the use of mechanistic models is overparameterization leading to model bias (FATT, 2002; Cloke et al., 2003; Wagener et al., 2004). Hydrologic models are commonly considered parsimonious with as few as six components; however, they can become overparameterized with as few as two or three (Jakeman and Hornberger, 1993).

Metric (also data-based, empirical, black box, curve number) and parametric (also conceptual, or explicit soil moisture accounting) models represent two kinds of empirically driven models. They rely on time-series data gathered from the watershed of interest in order to calibrate the final model.

Parametric hydrologic models are a common approach used in practical applications of watershed simulation (Wheater et al., 1993; Wagener et al., 2004). Parametric models are distinguished from metric models in that their structure is predefined in the form of major storage components (e.g. atmosphere, soil water, groundwater, vegetation, streams). Fluxes are simulated for these components via coefficient driven transfer processes (e.g. rainfall, infiltration, percolation, evapotranspiration, runoff, drainage). They are generally the most intuitive for users because they are organized based on common understandings of the hydrologic system.

Parametric models are subject to several limitations. Like mechanistic models, over-parameterization is an issue. Second, they utilize data from the system being simulated to fit parameter values using complicated calibration procedures

(Boyle et al., 2001). The dependence on flow data makes them particularly impractical for ungauged catchments (Wagener et al., 2004). Third, processes and parameters may be aggregated into a single component that cannot be derived directly from field observation. Fourth, when these models are applied spatially they are generally not fully distributed (i.e. independent parameter values are usually not available for all locations) and are often lumped in semi-distributed models (Singh and Woolhiser, 2002; Chappell et al., 2006). For small watershed or hillslope-scale models, where reliable parameter estimates do not exist at the equivalent scale, or where relevant soil properties are known only for a limited number of points, semi-distributed models are nearly compulsory (Jakeman and Hornberger, 1993; Chappell et al., 1998).

Metric hydrologic is distinguished from parametric models because they rely purely on data collected from the watershed in question. Unlike parametric models metric models generally depend only on fitting data to a predetermined curve rather than constructed based on our understanding of watershed processes. They usually assume a linear, proportional, or constant response to increased rainfall, and are based on concepts such as the unit hydrograph theory or the rational formula (Newson, 1975; Pilgrim and Cordery, 1993; Wheater et al., 1993; Bonta et al., 1997; Wagener et al., 2004). Their primary advantage is that they require minimal data and are easily applied to ungauged watersheds. They can provide a sound initial assessment of a watershed response (Phillips, 2004).

A key disadvantage of curve number and similar approaches, particularly with regards to extreme floods, is that they assume linear or curvilinear decay (proportional) responses between rainfall and runoff. These models are not ideal for forested watersheds (Hawkins, 1993), provide highly variable predictions (Bonta et al., 1997), and have been shown to overestimate larger events (Svoboda, 1991). However, as strongly implied by Gaume et al. (2003, 2004) and Hawkins (1993), extreme events are probably not subject to the same hydrologic paradigms for which these models have been developed. Hydrologic responses may change for extreme events and therefore conclusions drawn from results outside of the model's frame of inference would be suspect (Chiu and Huang, 1970).

The final class of models consists of hybrids and modeling systems. Metric-parametric and data-based mechanistic models obscure the distinctions between mechanistic, metric, and parametric models, which can lead to some confusion when comparing models (Bonell, 1998; FATT, 2002; Singh and Woolhiser, 2002; Wagener et al., 2004). Hybrids utilize observational data following the metric paradigm, but seek to include parametric or mechanistic elements to evaluate hypotheses regarding model components relative to watershed function (Wagener et al., 2004). A secondary purpose is to expand the application of metric models spatially.

Similarly, model systems are usually an amalgam of several model packages for specific areas tailored to address spatial systems (e.g. Boss International HEC-HMS, Gestion Intégrée des Bassins versants à l'aide d'un Système Informatisé—GIBSI). The use of hybrids and systems is attractive for the purpose of evaluating treatment effects on hydrology and a variety of related issues because of their versatility and detail. However, as with the root model types, overparameterization and extrapolation errors remain concerns.

6. General issues related to models and the nature of extreme floods

Hydrologists and planners face several challenges for predicting extreme flooding events or evaluating watershed responses to disturbances. These include (1) issues associated with testing hypotheses, specifically the hypotheses regarding disturbance effects on hydrologic processes; (2) model errors associated with uncertainty, heterogeneity, and scale; (3) issues associated with hydrologic thresholds and extrapolation; (4) issues associated with stream channel morphology and sedimentation; (5) limits associated with flood and rainfall frequency at applicable scales.

6.1. Hypothesis testing

Although modeling is a useful tool in hydrologic investigations of normal watershed conditions, there are concerns regarding the application of simulation models to extreme floods. Controlled manipulations are difficult to perform in the field at relevant scales. Therefore, as computing power and availability of intensive spatial data have increased, models have been increasingly used to test hypotheses regarding the relationships among model components, land-use, and hydrologic function (Bormann and Diekkruger, 2003; Wagener et al., 2004). However, models generally perform poorly at extremes that are beyond the range of conditions for which they were developed (Kuczera et al., 1993), and the dangers of extrapolation are fundamental to data analysis. Model evaluation is often only concerned with goodness of fit, rather than structural analyses which is more useful when treating the model as a hypothesis (Hooper, 2001).

There is a basic assumption that once a model that has been fit to match the observed hydrograph that all other inputs and output is therefore correct (Hooper, 2001; Singh and Woolhiser, 2002). This assumption is fundamentally opposed to the scientific method because there is no means for rejecting a model's structure once it has been selected. Model performance is highly sensitive to structure, and alternatives may just as easily explain an observed phenomenon and structural errors are almost always present (Jakeman and Hornberger, 1993; Young et al., 1996; Cloke et al., 2003; Butts et al., 2004). Calibrated models that utilize lumped data or conceptual frameworks may perform better, but they have less power as tools for testing hypotheses (Hooper, 2001; Bormann and Diekkruger, 2003; Reed et al., 2004; Chappell et al., 2006). Fully distributed, mechanistic models, although uncalibrated and subject to overparameterization, are potentially more useful for isolating factors that control hydrologic change. However, the magnitude of their results may be uncertain and they may do a poor job of emulating hydrologic processes at specific points within a mixed watershed. Utilizing suites of models may provide the most information, and the most powerful evaluations of model performance (Georgakakos et al., 2004).

6.2. Uncertainty, scale, and heterogeneity

Issues of heterogeneity and scale are among the most difficult problems associated with studying natural systems (Levin, 1992; Waring and Running, 1998; Uhlenbrook, 2003). Uncertainty (i.e. the amount an observed or calculated estimate differs from the true value in nature) is a major issue for hydrologic modeling (Butts et al., 2004; Georgakakos et al., 2004), and in particular the modeling of extreme floods (Jakeman et al., 1990; Herbst and Diekkruger, 2003). There are four types of uncertainty: (1) random and systematic error in model inputs or boundary conditions, (2) random or systematic errors in outputs, (3) sub-optimal parameterization, or (4) model bias (Butts et al., 2004). In addition, measurement error can greatly affect model accuracy, particularly in ungauged catchments. Results can be highly sensitive to parameters that are difficult to quantify, and small errors in our assumptions can result in errant conclusions. Measurement errors can become greatly magnified for very large events, particularly if monitoring equipment becomes damaged. Eyewitness accounts are often relied upon to reconstruct the conditions and timing of extreme flow events, but they are generally of minimal use for describing anything with precision except channel conditions (FATT, 2002; Gaume et al., 2003, 2004; Hicks et al., 2005).

Individual models are often applied with little discussion of uncertainty or parameter sensitivity (FATT, 2002; Walter et al., 2003). A prime example is the reliance of hydrology models on digital elevation models (DEMs) to provide topographic information. Commonly available DEMs (10–30 m resolution) do not capture microtopography, and more importantly do not address recent or vestigial road networks that may greatly affect surface hydrology (Walker and Wilgoose, 1999; Duke et al., 2003). A second example is the lumping of soils parameters such as saturated hydraulic conductivity, porosities because these values are difficult to quantify in the field. Efforts such as the distributed model intercomparison project (DMIP) (Smith et al., 2004) allow uncertainties with model structure and model capabilities to be to be evaluated and compared (Reed et al., 2004). The DMIP clearly illustrates that the performance of models varies, and that no model is superior in all circumstances.

The most capable models have multiple levels of organization and utilize components across multiple scales (Ryan et al., 2000). Hoosbeck and Bryant describe this on a three-axis diagram that includes the degree of complexity (empirical to mechanistic), degree of computation (qualitative to quantitative), and scale hierarchy (molecular to planetary). Hydrologic models used to evaluate land-use effects on floods are frequently developed for use at regional scales (Singh and Woolhiser, 2002), rather than the scale of forest operations (<1 km^2). As spatial and temporal scale decrease the number of parameters required to describe the system tends to increase (Wagener et al., 2004), along with uncertainty (Reed et al., 2004). Unfortunately, there is a fundamental lack of data for extreme events, particularly on smaller headwater watersheds, and particularly for hillslope hydrologic parameters controlling streamflow generation (Hooper, 2001; Gaume et al., 2004). Metric and parametric models often rely on parameters that cannot be measured directly and must be fit to the data. Consequently there is a lack of insight in the processes that control extreme floods. The problem is exacerbated in mountainous terrain because steep slopes and shallow soils are most sensitive to parameter selection due to these heterogeneity and scale issues (Cloke et al., 2003).

Heterogeneity is also an important issue with regards to scaling. From an ecological standpoint, heterogeneity is a description of the complexity of a system; however, it is not strictly a quantitative property, but also a qualitative one (Li and Reynolds, 1995; Weins, 2000). Therefore it is difficult to describe in terms that are easily recognizable, or neatly quantified. Hydrologic modelers address scale and heterogeneity either through lumping, regionalization, or complex geostatistical methods (Herbst and Diekkruger, 2002, 2003; Giannoni et al., 2003) methods (Boyle et al., 2001; Herbst and Diekkruger, 2002, 2003; Giannoni et al., 2003; Uhlenbrook, 2003; Wagener et al., 2004). However, as results are scaled up or down, heterogeneity will manifest itself differently because environmental gradients are manifested as patchy mosaics and processes in nature (Addicott et al., 1987; Ehrenfeld et al., 1997; Sidle et al., 2001).

6.3. Hydrologic thresholds

Two studies (Hawkins, 1993; Gaume et al., 2003) provide us with specific insights into extreme events. Hawkins (1993) observed that runoff responses of watersheds to specific rainfall amounts follow either a standard form, or two deviant forms (Fig. 2). Watersheds exhibiting standard behavior conform to the widely applied and assumed rainfall/runoff scenario observed in many watersheds; that is, the runoff coefficient declines with increasing storm magnitude to approach a constant value. This coefficient, or curve number, is widely reported and used in rainfall/runoff modeling, particularly ungauged catchments. The curve number for a complacent watershed also declines with increasing rainfall and are

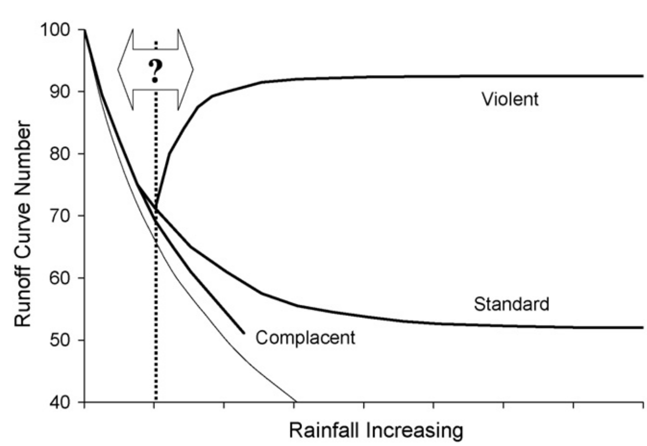

Fig. 2. Three variations in watershed runoff responses to specific rainfall depths (adapted from Hawkins, 1993).

indicative of a partial source area condition (source area between 0.1% and 5%). On the Berea Watershed (forested), located in Kentucky, he identified a violent response curve that initially appears similar to standard or complacent responses, but changes suddenly and asymptotically to approach a high curve number (90–100). Hawkins (1993) hypothesized that the violent watershed responses might be triggered by some critical rainfall depth and intensity, although in the case-study presented the violent response was due to a rainfall depth of only 25 mm. Water repellant soils caused by fires may be subjected lower thresholds for violent behavior (Neary et al., 2005).

The Avene watershed in the French Mediterranean covers an area of 57 km^2 with mountainous headwaters, a diverse geology (sandstone, limestone, and schist), and mixed land-use (agriculture, forests) (Gaume et al., 2003). The October 1997 flood was caused by a rain event with intensities as high as 300 mm over 6 h. The specific discharge for the event was 10–15 $m^3 s^{-1} km^2$ with less than 2 $m^3 s^{-1} km^2$ being generated in the upper portions of the watershed. Eyewitnesses to the event recalled sudden (half-hour) rises in the water levels late in the storm, along with a recession to near normal conditions within 36 h. Similar to Hawkins (1993) investigators believed that runoff coefficients must have been close to 100 during the most intense period of the storm although the largest flow rates were largely based on eyewitness accounts. They surmised that the rise in discharge late in the event is explained by the high infiltration capacities of these soils. Once this capacity was met, the watershed response immediately changed, and the system behaved as a ‘‘tip bucket’’.

Results from Gaume et al. (2003) and Hawkins (1993) illustrate the potential importance of hydrologic thresholds in hydrologic modeling. Additionally, modeling processes for soils near saturation is very complicated because rapid changes in water potentials occur with very small changes in moisture content (Cloke et al., 2003). A number of preferential flow mechanisms can violate assumptions about water flux within bulk soil (Torres, 2002). Soils do not need to be fully saturated in order to conduct a significant amount of water. The movement of water through more conductive zones can actually flush water from the soil matrix. Torres and Alexander (2002) observed decreased soil water content and lower storage as pore water overcame capillary tension following a high-intensity irrigation spike to soil blocks in laboratory conditions. There may be a relationship between late, high-intensity spikes and extreme flood flows on soils with high infiltration rates (FATT, 2002; Gaume et al., 2003). When soils cease to absorb and store water, the source area can become suddenly large. Water movement under these conditions becomes a routing issue (i.e. roads, saturated overland flow, and preferential flow paths), rather than strictly a land-use issue. Although Hortonian or saturated overland flow is generally not thought to be a factor in forested watersheds, it may become one during extreme events (Sidle, 2006).

Given the lack of data collected specifically for the purpose, using metric models, such as the curve number method, to determine the specific causes of violent watershed responses to large rainfalls is a key challenge. Bayesian models and artificial neural networks (ANNs) are statistical paradigms that show significant potential for use in flood modeling. These methods differ from parametric approaches in that they utilize several probabilities (prior, conditional, marginal, and posterior) to address hypotheses concerning collected data. Probabilities are weighted and interpreted as ‘‘degrees of belief’’, rather than strict proportions or frequencies. This approach has been applied to hydrologic problems like flood frequency analysis (Tokar and Johnson, 1999; ASCE, 2000b,a; Thirumalaiah and Deo, 2000; Rajurkar et al., 2004). The primary advantages of this approach are: (1) the ability to account for the non-linear nature of flood prediction, and specifically systems that behave differently above or below certain thresholds, and (2) the ability to recursively ‘‘learn’’ or be ‘‘trained’’ both through iterative steps as new information is acquired. ANNs have been shown to be particularly well suited for use with difficult systems with data limitations (Tokar and Johnson, 1999).

6.4. Stream channel morphology

Most studies that evaluate land-use and flooding discharge express the effects as absolute or percentage increases in flow volumes, peak discharge rates, or specific yield. These representations are not particularly informative in terms of evaluating flood risk, particularly from an economic basis (Boyle et al., 1998). Discharge may be more precise, but the depth above bankfull stage dictates most of the impacts that a significant flood event might incur. The maximum stage achieved will be highly dependent on channel morphology in mountainous terrain. Flooding at the mouths of streams (where monitoring equipment may be located) can be considerably less severe than those realized in localized upper portions of a reach (FATT, 2002; Hicks et al., 2005), depending on storm morphology.

There are a number of stream classification systems that describe basic channel morphology or stream condition (e.g. Phankuch, 1975; Simon, 1989; Whiting and Bradley, 1993; Rosgen, 1994; Montgomery and Buffington, 1997). For example, the Rosgen system for stream classification identified nine stream and river types based on gradient, cross section, and plan view (Harrelson et al., 1994). Most high-order and intermittent streams in the Appalachians would be classified as Rosgen Aa+, A, B, or C stream types; channels are confined by steep side-slopes, with and without narrow floodplains. In an extreme event, the entire floodplain may transport water. The volume of discharge will be proportional to the cross sectional area which increases exponentially with stage. Discharge within a floodplain will increase asymptotically with stage height for approximated Rosgen streams profiles (Fig. 3). On Rosgen B (40 m wide, 2.5° slope floodplain, 45° side slopes), and C (100 m wide, 2.5° slope floodplain, 45° side slopes) stream types, changes in discharge at already large volumes should exhibit relatively small changes in stage height (10:1, volume:height, for smaller floods, and up to 30:1 for large floods). Stage height is more responsive to higher discharges for the Rosgen Aa+ (no floodplain, 45° slopes) and A (no

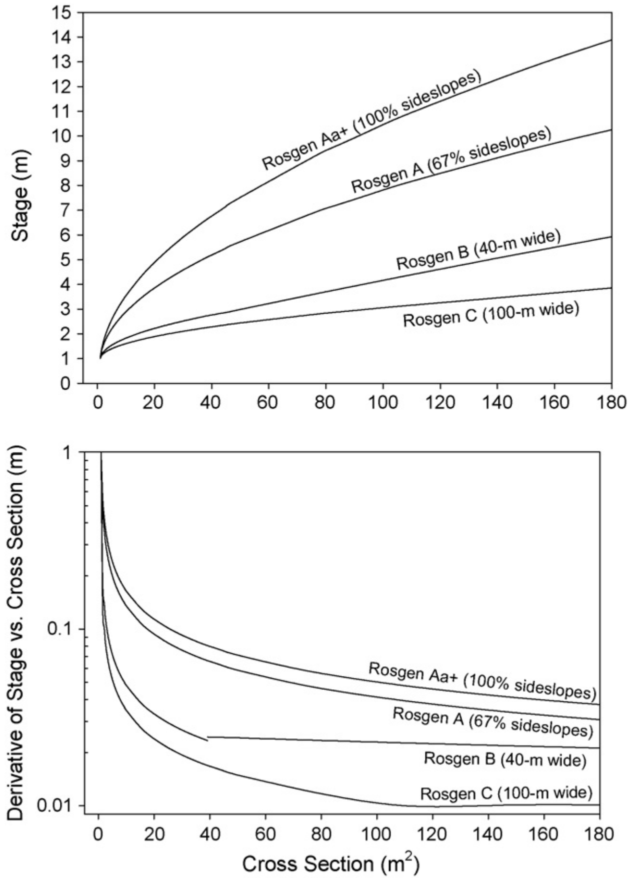

Fig. 3. (Upper) stage height versus active cross sectional areas, and (lower) change in stage versus active cross sectional areas, for four approximations of Rosgen stream types.

floodplain, 34° slopes); however, these streams still exhibit a ratio of approximately 3.5:1 for lesser floods, and approximately 6:1 for higher discharges. The ratio between discharge and stage would in practice be higher if they were measured in the field. As stage increases so would the hydraulic gradient; thus, a greater velocity (and therefore volume) of water would occur per unit increase in cross section. While the stage on narrow streams channels may be more sensitive to discharge, larger streams may be more at risk for erosion and changes in stream morphology due to energy associated with large flow volumes (Miller, 1990).

6.5. *Flood frequency*

6.5.1. *Observing extreme floods*

The Appalachian region has generated some of the largest area-discharge floods in the US, especially for smaller watersheds (Crippen and Bue, 1977). Additionally, the storms producing these floods have some of the largest 6 h rainfall amounts in the world (Crippen and Bue, 1977; Hicks et al., 2005). Notable floods are usually the result of large tropical systems, unusual thunderstorms, or rains coupled with spring snowmelt in the north (Neary et al., 1986; Neary and Swift, 1987; Sturdevant-Rees et al., 2001; Graybeal and Leathers, 2006), and therefore there is not necessarily a dominant flood season in some parts of the Appalachians (Lecce, 2000). Floods become a concern for rainfalls >50 mm day^{-1}, but risk is not necessarily directly related to rainfall amount (Miller, 1990). Numerous observational studies are available of extreme or "geomorphically effective" flood events in the region and throughout the world (e.g. Wolman and Gerson, 1978; Miller, 1990; Hawkins, 1993; Phillips, 2002; Yeo, 2002; Gaume et al., 2003; Holman et al., 2003; Gaume et al., 2004; Hicks et al., 2005). Geomorphically effective floods are those that are able to alter landscapes due to fluvial impacts that include soil and debris jams, sediment deposition, and downcutting.

The technical definition of floods and the general public's perception of floods are incongruous. Hydrologists generally define flood events as any streamflow outside of the flood stage, which is typically defined by the normal channel banks (Jarvis, 1936; Leopold and Maddock, 1954). The general populace typically consider floods to be events that damage lives and property (Barrows, 1948). Hydrologists typically use a systematic method for categorizing floods and rain events based on their annual return periods or intervals (Reis and Crouse, 1992). Return periods are the likelihood that a given flood flow will be attained or exceeded in a given year; it is inferred as the average interval in years between events of similar magnitude over the very long term. Due to climatic variability, events of a given return period can occur in shorter intervals than their timing implies, or that major floods would not occur in the same year. Return periods are calculated based on the observed record of maximum yearly flows fitted to an assumed probability distribution such as the log-Pearsons type-3 (Stedinger et al., 1993). Historical and paleoflood information can also be used to refine estimates of rare events (Stedinger and Baker, 1987). An alternative approach that may outperform annual series, especially for extreme hydrologic events, is partial series analysis (Madsen et al., 1997a,b; Madsen and Rosbjerg, 1997a,b). Despite its reported advantages, partial series is not yet widely utilized.

The length of the observed record and the maximum flood within that record are the dominant factors that define the flood frequency curve for a gauged system using annual series (IACWD, 1982). However, although there is a 25% probability that a 100-year flood will occur within a 30-year record (Fig. 4), the probability that it will happen twice in the same period is the square of this value (0.252 = 6.3%). Even where gauge data are available for relatively long time periods (20–50 years) and the flood peak distribution is fairly well known, there is little confidence in estimates of flood peaks greater than the 50- or 100-year interval (Kite, 1975; IACWD, 1982; Stedinger and Baker, 1987).

Although extrapolation may be used to estimate 500-year flood magnitudes, it can result in errors ≥15% (Thomas and Kirby, 2002) depending on floodplain geometry. This error could lead to classification errors for the return period of extreme events >100 years. Benson (1960) calculated the length of record required to predict extreme returns within 10% assuming a perfectly distributed 1000-year record. He found that to capture the 100-year event with a 95% and 80% confidence interval would require a length of record 115 and

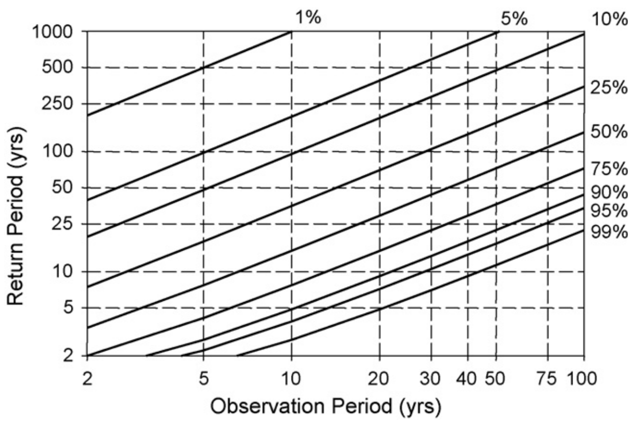

Fig. 4. Theoretical risks of floods of various return periods for given observation periods.

100 years, respectively (Table 1) (Newson, 1975). Therefore, data from Coweeta Hydrologic Laboratory (North Carolina) and Fernow Experimental Forest (West Virginia), which represent continuous records among the longest in the Appalachians relating to forestry practices, are currently suited for evaluating floods on the order of a 25-year return period.

Usable relationships between extreme rainfall and extreme floods are difficult to detect (Reich, 1970). Those studies that do exist were not monitored for extended periods after harvesting. Hornbeck (1973b) observed convergent slopes between a treatment and a control watershed at Hubbard Brook (New Hampshire) for one June containing 10 cm of rainfall for the month. However, the result does not account for the types and distributions of rainfalls within the 3 months reported over 4 post-treatment years, and there is no indication whether the paired-regression approach can be extrapolated beyond the 10 cm maximum rainfall. Lu (1994) found that forest harvesting effects converged for return flows over 25 years (Brooks et al., 2003). Other studies have similarly indicated that land-use has a limited role in controlling large rare floods (non-exceeding probability >90%), and that there is a convergence between harvest and non-harvest responses (Harr et al., 1975; Caissie et al., 2002; Robinson et al., 2003; Lavigne et al., 2004; FAO and CIFOR, 2005).

Spatial errors associated with observing extreme floods can occur when homogeneity assumptions are made about the spatial distribution of rainfall. Phillips (2004) reported that the maximum difference in precipitation between two watersheds within 10 km of one another (~300 km^2) might be as much as 14% to 33% in an extreme summer storm event. Hicks et al. (2005) indicated a six-fold difference within 10 km^2. A 100–1000-year flooding event at one scale may only be a 25-year event at a larger scale or even a common occurrence at a regional scale (Perry and Combs, 1998). For example, the odds that the center of a tropical system will go over any given location on the eastern seaboard of the US, including the Appalachians, is low. However, at the scale of the entire region several hurricanes make landfall every year. Likewise, cloudbursts from large convective systems that trigger locally significant flooding are likely to occur in any given summer, but the heterogeneity of the distribution of rainfall does not assure that any specific watershed will experience an extreme event. Additionally, complex climate cycles may affect localized weather patterns over periods of years, decades, centuries, and longer (Swift et al., 1988; Perry et al., 2001b).

Table 1
Length of record require to predict floods of given recurrence intervals (from Newson, 1975)

Recurrence interval, years	Length of record	
	95% confidence	80% confidence
2	40	25
10	90	38
25	105	75
50	110	90
100	115	100

The probability is low that a single, sufficiently instrumented watershed will capture the required information (e.g. precipitation, soil, and watershed dynamics) of both extreme and non-extreme conditions. Typically, instrumentation such as hourly recorders and water samplers are reserved for third-order or higher streams, further limiting our ability to look at forest impacts (Thornton et al., 2000). Additionally, in the Aude flood event in France (Gaume et al., 2004), one third of the equipment used to gauge the watershed was destroyed by the event, and data from many of those that survived were deemed unreliable. Within a given watershed, specific errors associated with (1) instrument error, (2) sensitivity of the stage–discharge relationship, (3) changes in cross section as the flood scours the channel, and (4) the conveyance of water on the floodplain further hinder the ability of a single watershed to capture sufficient information to accurately model flooding in a general sense.

Furthermore, the most extreme flooding often occurs on ungauged watersheds. Major forest hydrology research facilities were not originally designed to address extreme flooding. The infrequency of extreme events and the lack of adequate stream and soil hydrologic data is a real obstacle to understanding the hydrology of extreme floods (DeWalle, 2003).

6.5.2. *Defining extreme floods*

Defining ‘‘extreme floods’’ or ‘‘extreme rainfalls’’, their magnitudes, or their return periods, is a difficult problem. A 100-year rain event does not necessarily generate a 100-year flood (Hjelmfelt, 1980; Hawkins, 1993), and the changes in discharge associated with incremental increases in return period are not linear (IACWD, 1982). For instance, the change in stormflow between a 25-year event and a 50-year event may be considerably larger than the change between a 100-year event and a 500-year event. Secondly, return periods refer to the maximum event for an entire year, events of similar magnitude may happen more regularly, but are simply not the most significant within a given year. Therefore using return periods to define ‘‘extreme’’ conditions is not useful, and even less useful in light of the spatial issues mentioned in the previous section.

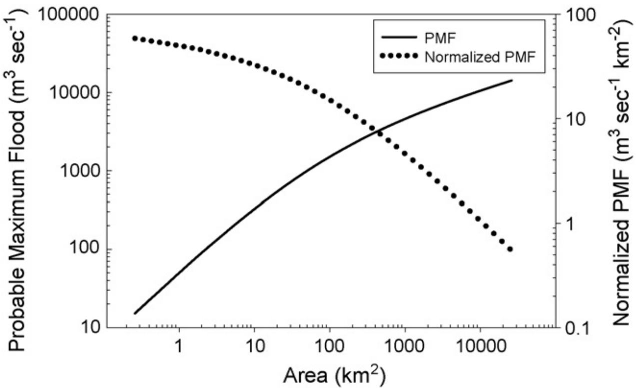

Fig. 5. Example of the positive asymptotic relationship between probable maximum flood (PMF) and area, and the negative exponential relationship between normalized PMF and area corresponding to the southern Appalachian flood region after Crippen and Bue (1977) and Crippen (1982).

Crippen and Bue (1977) and Crippen (1982) developed envelope curves to estimate the most severe flood magnitudes for watershed areas within various flood regions of the conterminous US. Other studies have also produced similar envelope curves e.g. (Miller, 1990; Hicks et al., 2005). As watershed area increases, the probable maximum flood (PMF) discharge increases asymptotically with area; normalized PMF decreases exponentially (Fig. 5). The proportion of PMF is potentially a more useful definition of extreme conditions because it would incorporate watershed area and regional differences and would allow for multiple extreme events to occur within the same year. Alternatively, another envelope curves, or curves based on the 50-, 100-, or 500-year return period, may be used.

The PMF may provide insight with regards to evaluating land-use effects. Assuming that the stormflow responses of smaller watersheds are more sensitive to conventional patterns of land-use change at relevant scales (Caissie et al., 2002; Hicks et al., 2005), it might also be assumed that the deviation from the initial slope of the PMF curve (~25 ha minimum) is related to the watershed size where land-use become less influential on extreme flood regimes. The slope of the PMF curve decreases by more than 50% as watersheds increase over 6.6–24 km^2 in the case of the Southern Appalachians, Northern Appalachians, and the Northeastern US (Table 2). It could be hypothesized based on this information that the influence of typical land-use dynamics on flooding is restricted to watersheds on the order of a few thousand hectares and smaller. This result is congruent with the differences between first and second order watersheds presented by Caissie et al. (2002). Watersheds of this size range would include all first-order streams, most second-order, and some third and fourth-order streams and predominantly perennial streams (Hansen, 2001). The median watershed size for intermittent stream inception is 6 ha, and ranges between 1 and 18 ha (Nutter et al., 1984; Paybins, 2003). The median perennial flow is associated with watershed sizes of 16 ha, and ranging between 4 and 60 ha. Although this may only apply to land-use effects on discharge, it should be noted that flood damage occurs downstream where flood stage is complicated by factors associated with floodplain development (e.g. human encroachment, impervious surfaces, channelization).

7. Summary and conclusions

The relationship between forests and floods, or more specifically the capacity of forests to prevent or mediate floods, has been debated in the US for almost a century. Many generalizations may be drawn from the current literature, but individual watersheds exhibit unique characteristics. Floods are

Table 2
Minimum and maximum slope, and watershed areas corresponding to 50% and 67% changes in slope, of the probable maximum flood (PMF) curves presented by Crippen and Bue (1977) and Crippen (1982) for the physiographic regions of the United States

Flood region[a]	Physiographic region[b]	Maximum slope (m^3 s^{-1} km^{-2})	Minimum slope (m^3 s^{-1} km^{-2})	Watershed area corresponding to 50% decrease in slope (km^2)	Watershed area corresponding to 67% decrease in slope (km^2)
1	Northeast and New England	46	0.24	6.6	21
2	Eastern Coastal Plain	86	0.19	2.5	7.4
3	Gulf Coastal Plain	63	0.13	6.6	19
4	Northern Appalachians	53	0.17	8.1	24
5	Southern Appalachians	160	0.11	3.1	8.6
6	Central Lowlands	92	0.33	8.2	25
7	Interior Low Plateaus	62	0.081	4.5	13
8	Interior Highlands	50	0.17	9.3	28
9	Osage Plains	95	0.069	3.3	9.3
10	Texas Coastal Plain	57	1.6	95	250
11	North Central Plains	49	0.11	6.4	19
12	Great Plains	120	0.39	8.0	24
13	Rocky Mountains	84	0.10	4.1	12
14	Colorado Plateau	36	0.046	1.8	4.8
15	Columbia Plateau	81	0.45	26	67
16	Great Basin and Desert	110	1.1	26	74
17	Pacific Mountain System	71	0.40	20	54

[a] Flood Regions based on Crippen and Bue (1977).
[b] Physiographic Names from USGS (2003).

the integrated product of multiple flowpaths delivering water to the stream. The amount of stormflow is most directly linked to the area in the watershed and volume of precipitation or snowmelt deposited on the site, stored, or transported to the stream. Flooding is also influenced by channel morphology and conveyance. Management practices potentially influence by altering the timing and volume routed via these component flowpaths, and by changing these hydrologic flowpaths. Many types of rainfall are capable of generating floods. Extreme rainfalls may overwhelm the capacity of water to infiltrate into the soil, or may rapidly exceed or bypass storage components within the soil and within the watershed. ''Extreme'' when used in the context of floods or precipitation only loosely refers to unusually large, geomorphically effective events; however, ''extreme'' is not well defined and does not provide a sense of proportion. The extremity of any given event is dependent on spatial and temporal scales.

There is no question that forests and watershed hydrology are linked. The current body of research indicates that forests have a significant effect on annual water yield, particularly baseflow. However, results are mixed with regards to the effect of forest operations on stormflow and peak discharge. Measured effects within the Appalachian region have been positive, negative, and neutral and there is little unequivocal evidence that forest operations greatly exacerbate extreme peak discharges. Harvesting can result in increased stormflows and peak discharges for small to medium events, but these effects diminish as events become larger (Thomas and Megahan, 1998).

The greatest potential forest operations have for affecting ''extreme'' peak flows is through routing via road systems or stream channel modification. Forest road systems directly influence both flow routing and sediment yields disproportionably to their area, and therefore they have the greatest potential for affecting ''extreme'' peak flows in the long-term. Roads systems (temporary and permanent) are a major component of forest harvesting and management, and have potentially longer effects if not properly located, constructed, maintained, and closed, and location is probably the more significant of these factors (Swift et al., 1988; Swift and Burns, 1999; Megahan et al., 2001; Ziegler et al., 2004).

It is difficult to partition flood flows based on the numerous factors and complex interactions that affect hydrologic responses. Forest harvesting is often mislabeled as deforestation. Deforestation is the permanent conversion of forests to alternative land-uses. Sites should recover from the effects of forest harvesting within 5–10 years after harvest. Although vegetation removals may increase pre-event saturation until vegetative cover can reestablish itself, elevated soil temperatures and baseflow rates decrease soil water content. The influence of forests and soils on flood hydrology is not constant, is seasonably variable, and diminishes with increased rainfall. Watersheds that have narrow stream channels, narrow floodplains, and shallow soils are most likely to have large increases in stream stage with corresponding rises in flood volumes. We estimate that the watershed sizes that are most likely to be sensitive to land-use are those between 20 and several thousand hectare. There is presently little research regarding the effect forests or forestry practices on the outcome of extreme floods.

The particular challenge in the Appalachian region is that forestry land-use generally occurs in areas that are already likely to have the most intense flood peaks (Hicks et al., 2005). The question remains, will these be increased by management? It may be possible for certain forest management practices to moderate certain floods by affecting storage and routing; whether they can moderate extreme floods remains an unanswered question. It has been suggested that forestry best management practices (BMPs) be modified to include management for increases in peak water yield to reduce flood risk (FATT, 2002; Mortimer and Visser, 2004). A primary function of forestry best management practices is to minimize erosion by lowering water velocities, preventing channelization, and redirecting water back onto the forest floor. Of the BMPs that are currently used, these practices should also prevent rapid routing of water that may increase stormflow and peak discharge. Proper road design, maintenance, and closure are key forest management activities. Considering that litigation may hold landowners responsible for land-use decisions, potential BMPs that may control flooding and the potentially conflicting goals they may entail needs to be carefully accessed.

If society chooses to alter vegetation management for the goal of avoiding extreme flows, we must also consider any repercussions with regards to other management issue (i.e. forest regeneration, disease, and pests). If society chooses to engineer stream reaches to specifically accommodate extreme flows, the repercussions with respect to stream habitat or low-flow must be considered as well. Finally, issues associated with developing areas beyond those prone only to frequent flooding, but exposed to infrequent, extreme events, cannot be ignored.

Our best datasets, with the longest periods on record, may not be entirely sufficient for evaluating the mechanisms controlling extreme events or even accurately designating their return periods. Due to the rarity of extreme events, and with the improvements in computing power and remote sensing, models have become increasingly employed to evaluate extreme conditions. We have found no models that have been specifically designed to address extreme flooding. While models can be useful tools for making reasonable first approximation of flood dynamics, the factors that control extreme floods appear to differ from those controlling lesser events (Sidle et al., 2006b). Thus, 'right' answers may be obtained for 'wrong' reasons.

The curve-number method is the most commonly used approach for evaluation of extreme floods because the models are readily available and easily applied. Mechanistic models, on the other hand, may be the most appropriate for this purpose because they at least rely on actual physical data although they are subject to overparameterization errors and often invoke incorrect flow pathways. Ultimately, all models are driven by their assumptions, and models are usually developed based on less-extreme watershed behavior; there are many reasons why extreme events may deviate from those assumptions. Models are useful for describing trends, or predict average behavior for systems that are well understood. Using models that are not designed for extreme events to partition sources of water in different land-use scenarios on specific watersheds, for a specific event, probably violates the rules of extrapolation and should be done with caution. Differences in model output may only reflect differences in assumptions being made by the user or a systematic bias, rather than actual differences in the field.

In order to model extreme floods they must be better defined and the factors that control them must be better understood. There is little information regarding the specific conditions that define the threshold between the ''standard'' and so-called ''violent'' watershed responses to rainfall. The violent response may be due to infiltration excess, saturation excess, hysteresis, or preferential flowpaths or a combination of these processes. There a possible relationship between high-intensity rainfall spikes after a period of soil wetting and extreme floods that may be due to sudden increase in the source area, the contribution of zero-order basins, or the lowering of soil water holding capacities once a precipitation spike has ended. Mechanistic models that specifically incorporate preferential flow and forest roads are probably the best equipped to gain understanding of these floods and formulate hypotheses for field experiments.

Acknowledgement

The authors would like to thank the USDA Forest Service, Northern Research Station for their support and financial assistance.

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