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# Channel network extent in the context of historical land use, flow generation processes, and landscape evolution in the North Carolina Piedmont

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ABSTRACT: Intensive agricultural land use in the 18th to early 20th centuries on the southeastern Piedmont resulted in substantial soil erosion and gully development. Today, many historically farmed areas have been abandoned and afforested, and such landscapes are an opportunity to study channel network recovery from disturbance by gullying. Channel initiation mapping, watershed area–slope relationships, and field monitoring of flow generation processes are used to identify channel network extent and place it in hydrologic, historical and landscape evolution context. In six study areas in the North Carolina Piedmont, 100 channel heads were mapped in fully-forested watersheds, revealing a channel initiation relationship of $380 = AS^{1.27}$, where $A$ is contributing area (m$^2$) and $S$ is local slope (m/m). Flow in these channels is generated by subsurface and overland flow. The measured relative slope exponent is lower than expected based on literature values of ~2 for forested watersheds with subsurface and overland flow, suggesting that the channel network extent may reflect a former hydrological regime. However, geomorphic evidence of recovery in channel heads within fully forested watersheds is greater than those with present day pasture. Present day channel heads lie within hollows or downslope of unchanneled valleys, which may be remnants of historical gullies, and area–slope relationships provide evidence of colluvial aggradation within the valleys. Channel network extent appears to be sensitive to land use change, with recovery beginning within decades of afforestation. Channel initiation mapping and area–slope relationships are shown to be useful tools for interpreting geomorphic effects of land use change. Copyright © 2012 John Wiley & Sons, Ltd.

KEYWORDS: channel initiation; land use change; gullies; streamflow generation; ephemeral channels

## Introduction

The extent of the channel network controls how rapidly sediment and water move through a watershed, at time scales from a single precipitation event to evolution of the landscape over thousands of years. First-order streams, as a stream network's outermost links, are tightly hydrologically, geomorphically, and biogeochemically linked to hillslope processes (Horton, 1945; Hack and Goodlett, 1960; Hewlett and Hibbert, 1967; Likens *et al*., 1977; Dietrich and Dunne, 1993; Gomi *et al*., 2002). Headwater streams are also major contributors of water, sediment, nutrients, and carbon to larger river systems (Gomi *et al*., 2002; MacDonald and Coe, 2007), so understanding the distribution of headwater streams is crucial for correctly modeling and interpreting hydrological, geomorphological, and biogeochemical processes at the watershed scale. Accurately delineating the extent of first-order channels remains as a significant challenge because of data limitations and uncertainty on the thresholds for channel initiation (Jaeger *et al*., 2007; James and Hunt, 2010; Brooks and Colburn, 2011; Orlandini *et al*., 2011). Channel network extent can also be profoundly disturbed by past and present human land use (Elmore and Kaushal, 2008), resulting in substantial changes in networks over time and with respect to topographic attributes.

Understanding the processes regulating channel initiation and channel network extent has been of fundamental interest to geomorphologists for decades (Horton, 1945; Hack and Goodlett, 1960; Carlston, 1965; Montgomery and Dietrich, 1988, 1992). Where channel networks do not initiate by landsliding or bedrock groundwater discharge, channel head position has been found to be a function of contributing area and local slope in landscapes around the world (Montgomery and Dietrich, 1992, 1994; Prosser and Abernethy, 1996; Hattanji and Matsushi, 2006; Hattanji *et al*., 2006; Imaizumi *et al*., 2010; James and Hunt, 2010). A number of relationships have been proposed to define a threshold combination of area and slope that results in channel initiation (Willgoose *et al*., 1991; Montgomery and Dietrich, 1992; Istanbulluoglu *et al*., 2002; Dalla Fontana and Marchi, 2003; James and Hunt, 2010), but most can be expressed by

$$c = A^x S^y \quad (1)$$

where $c$ is a threshold value that depends on climate, geology, soils, vegetation, and land use of a region, A is the upslope contributing area (typically m$^2$ or ha), and $S$ is local slope (typically m/m). Often, either area or slope is expressed as having an exponent of 1, which changes the other term's exponent to $A^{x/y}$ if $S^1$ or $S^{y/x}$ if $A^1$, essentially normalizing one exponent by the other. Such normalization is hereafter referred to as a relative exponent, following the convention of Desmet *et al.* (1999). Thus, the relative slope exponent is defined as $y/x$, and relative slope exponents will be reported throughout this paper for consistency.

Multiple studies of channel initation have found the relative slope exponent ($y/x$) to be approximately 2 (Table I), and several studies have *a priori* adopted a relative slope exponent in the range 1.7 to 2 and found a good fit to the drainage network (Dalla Fontana and Marchi, 2003; Giannoni *et al.*, 2005). Montgomery and Dietrich (1994) show that the exponent value of 2 derives from laminar infiltration-excess (Hortonian) overland flow and low gradient areas with saturation overland flow. These streamflow generation mechanisms are typical of many natural environments, from humid, forested landscapes to semi-arid and arid regions (Bonell, 1993; Puigdefabregas *et al.*, 1998).

Minimally disturbed landscapes, where channel initiation studies have generally been conducted, are assumed to represent channel networks that are well adjusted to the local climate, soils, and vegetation (Montgomery and Foufoula-Georgiou, 1993). In such landscapes, the channel network begins within the zones of convergent topography, in swales or valleys (Dietrich and Dunne, 1993). Unchanneled valleys are regions of non-erosive overland or subsurface flow, and their headward extent is controlled by variance in position of channel initiation locations on timescales of $10^4$ to $10^6$ years (Montgomery and Dietrich, 1992, 1994). At any given time, channel initiation can be described as a probabilistic function of slope, contributing area, median grain size, surface roughness, and precipitation excess, causing local variability within $c$ in Equation (1) within a region (Istanbulluoglu *et al.*, 2002) .

Landscapes undergoing disturbance from forest clearing, intensive agriculture, road building, wildfire, or severe storms can have channel networks that are not fully adjusted to other landscape properties. Such instability can manifest as rapid gullying and extension of the channel network (Ireland *et al.*, 1939; Prosser and Soufi, 1998; Istanbulluoglu *et al.*, 2002; Nyssen *et al.*, 2002; Parkner *et al.*, 2007). Gullies are typically characterized by a steep head cut resulting from concentrated overland flow or soil pipe collapse (Ireland *et al.*, 1939; Valentin *et al.*, 2005), and they may be initiated within the valley network or on planar or divergent hillslopes (Prosser and Soufi, 1998). In many gullied landscapes, infiltration excess overland flow occurs, when precipitation rates exceed the infiltration capacity of the land surface (Prosser and Abernethy, 1996; Poesen *et al.*, 1997).

A number of studies have characterized topographic thresholds for gullying, and widely varying exponents of area and slope have been reported (Table I). Some studies report no slope dependency on gully initiation, with gullying possible in all locations in a watershed (Prosser and Soufi, 1998; Vanwalleghem *et al.*, 2003), while others report relative slope exponents from $-1.2$ to 9.6 for gully initiation (see Table I for citations). The literature contains several different methods for calculating slope and contributing area and for identifying a relative slope exponent from the data distribution, but it is clear that there is far greater variability of relative slope exponents for gully initiation than for channel initiation, both globally and within a given region (Table I). That variability may reflect the range of processes responsible for gully initiation, and it limits the possibilities for applying literature values to predict gully extent in other areas.

Within the range of gully relative slope exponents, several studies found a good fit between gully locations and a relative slope exponent of 1 (Moore *et al.*, 1988; Desmet *et al.*, 1999; James and Hunt, 2010), giving equations of the form

$$c = AS \quad (2)$$

If flow depth increases with contributing area, as commonly assumed when flow confinement does not change, then (2) is equivalent to boundary shear stress ($\tau_b$):

$$\tau_b = \rho g D S \quad (3)$$

where $D$ is the depth of overland flow, $\rho$ is fluid density, and $g$ is gravitiational acceleration. Gully initiation then occurs where boundary shear stress exceeds a critical shear stress ($\tau_c$) for surface resistance to erosion. Gully erosion is also dependent on available sediment and sufficient capacity to transport that sediment (Valentin *et al.*, 2005; Hattanji *et al.*, 2006).

Landscapes gullied by past agricultural practices create conditions that are an opportunity to understand the recovery of channel networks following the cessation of a disturbance and explore the persistent influences of flow generation mechanisms on channel initiation, network extent and landscape evolution. In this study, channel initiation sites in multiple watersheds in the North Carolina Piedmont were investigated to understand the role of historical, intensive agricultural land use and gullying in setting the present day channel network extent. We combine channel initiation mapping with investigation of watershed area–slope relationships to place the modern and historical geomorphology in a landscape evolution context and to examine the utility of such an approach for interpreting geomorphic effects of land use.

## Setting

The Piedmont physiographic province of North Carolina is a ~240 km wide area that lies between the mountainous Blue Ridge and the nearly flat Atlantic Coastal Plain. The region has modest relief from fluvial dissection and regional tilting, with elevations ranging from 450–100 m. Metamorphic and igneous rocks in the Piedmont range from 300 to 750 million years old (Hoover, 1949; Rogers and Coleman, 2010). The mid-Atlantic and southern Piedmont was not glaciated during the Pleistocene, and the rate of soil production by bedrock and saprolite weathering corresponds to erosion rates over million year time scales (Pavich, 1989). In the North Carolina Piedmont, average annual precipitation is 1140–1180 mm, delivered in 40 to 50 storms, with no distinct wet or dry season. The annual mean temperature is 15°C, with mean temperatures in January and July 4°C and 26°C, respectively (State Climate Office of North Carolina, 2012).

Beginning in about the 1750s, land clearing, intensive agriculture and road building affected the North Carolina Piedmont. The predominance of agricultural land use peaked after 1865, and it declined in the early to middle 1900s. During this period, agriculture was predominantly cotton and row crops, and a lack of erosion control practices resulted in extensive sheet wash and gully erosion. Trimble (1974) estimated that most areas of North Carolina experienced 11 to 18 cm of soil erosion and Ireland and colleagues (1939) estimated that 25–75% of soil A horizons were lost to erosion. Where the clay-rich,

**Table I.** Relative slope exponents determined from field datasets, where channel or gully initiation was attributed to overland flow and/or shallow subsurface stormflow. Classification as either channel or gully was based on the terminology used in the source paper. The cited papers presented a variety of forms of Equation (1), and all equations have been rearranged to compute the relative slope exponent. The originally published form of the equation is also reported here, and includes equations with both upslope contributing area (A) and upslope contributing area per unit contour length ($A_c$). In calculating the relative slope exponents, $A$ and $A_c$ are equivalent

| Location | MAP (mm) | Vegetation | Relative slope exponent | Original equation form | r$^2$ | N | Citation |
|---|---|---|---|---|---|---|---|
| | | | | **Channels** | | | |
| southern Sierra Nevada, California, USA | 260 | open forest and grass | 1.7 | $S=0.35A^{-0.6}$ | lower limit* | 33 | (Montgomery and Dietrich, 1988) |
| Samoeng, Thailand | 1200–2000 | cultivated | 1.9 | $A=3097S^{-1.9}$** | 0.48** | 7 | (McNamara *et al.*, 2006) |
| Kanozan, Japan | 1466 | forest | 2.0 | $A=170S^{-2.0}$ | 0.43 | 26 | (Hattanji and Matsushi, 2006) |
| Ashio Mtns, Japan | 1581 | forest | 2.1 | $A=580S^{-2.1}$ | 0.37 | 29 | (Hattanji and Matsushi, 2006) |
| Ashio Mtns, Japan | 1478 | forest | 2.47 | $A=747S^{-2.47}$ | 0.56 | 34 | (Hattanji *et al.*, 2006) |
| Akaishi Mtns, Japan | 2800 | forest | 2.3 | $A=4658S^{-2.3}$ | 0.18 | 50$^{\dagger}$ | (Imaizumi *et al.*, 2010) |
| Marin County, California, USA | 760 | grassland | 2.5 | $S=0.27A^{-0.4}$ | lower limit* | 80 | (Montgomery and Dietrich, 1988) |
| Coast Range, Oregon, USA | 1500 | logged forest | 2.5 | $S=0.25A^{-0.4}$ | lower limit* | 71 | (Montgomery and Dietrich, 1988) |
| | | | | **Gullies** | | | |
| Idaho, USA | 1000 | intensely burnt forest | −1.2 | $A=5425S^{1.2}$ | 0.6** | 27 | (Istanbulluoglu *et al.*, 2002) |
| New South Wales, Australia | 800 | pine plantation | | "poor topographic separation of gullied and ungullied elements" | | | (Prosser and Soufi, 1998) |
| New South Wales, Australia | 500 | sparse grass | 1.59 | $A=1421S^{-1.59}$ | 0.52 | | (Prosser and Abernethy, 1996) |
| Loess Plateau, Shaanxi, China | 300-500 | cultivated | 1.6 | $A=306S^{-1.6}$** | 0.37** | 24 | (Wu and Cheng, 2005) |
| | | | 4.2 | $S=0.1839A^{-0.2385}$ | lower limit | | |
| Butare, Rwanda | 1166 | cultivated | 1.7 | $S=0.6A^{-0.6}$ | lower limit | 6 | (Moeyersons, 2003) |
| Alentejo, southern Portugal | 490 | cultivated or abandoned fields | 2.4 | $S=0.077A^{-0.414}$ | 0.52 | 16 | (Vandekerckhove *et al.*, 2000) |
| Alentejo, southern Portugal | 500-600 | cultivated | 2.9 | $S=0.02A^{-0.35}$ | lower limit* | | (Vandaele *et al.*, 1997) |
| Alentejo, southern Portugal | 490 | cultivated or abandoned fields | 3.3 | $S=0.083A^{-0.303}$ | 0.26 | 24 | (Vandekerckhove *et al.*, 2000) |
| Alentejo, southern Portugal | 550 | cultivated or abandoned fields | 3.4 | $S=0.09A^{-0.29}$ | 0.38 | 40 | (Nachtergaele *et al.*, 2001a) |
| Northwestern France | 700-1000 | cultivated | 2.5 | $S=0.06A^{-0.4}$ | lower limit* | | (Institut Geographique National, 1984) |
| Central Belgium | 700-800 | cultivated | 2.5 | $S=cA^{-0.4}$ | lower limit$^{\dagger\dagger}$ | 58 | (Nachtergaele *et al.*, 2001b) |
| Central Belgium | 700-800 | cultivated | 2.5-2.9 | $S=0.025A^{-0.4\ to\ -0.35}$ | lower limit* | | (Vandaele *et al.*, 1997) |
| Central Belgium | 700-800 | cultivated | 2.5-3.3 | $S=0.08A^{-0.4\ to\ -0.3}$ | lower limit* | | (Vandaele *et al.*, 1996) citing Poesen *et al.*, in prep |
| Loess Belt, Belgium | 700-800 | cultivated | 5 | $S=1.2A_c^{-0.2}$ | lower limit | 50 | (Desmet *et al.*, 1999) |
| Central Belgium | 800 | forest | | "statistically demonstrated that no correlation exists" | | 43 | (Vanwalleghem *et al.*, 2003) |
| Arhem, Northern Territory, Australia | 1389 | open forest and grass | 3.6 for areas >1100 m$^2$ | $c=A^{0.28}S$ | not reported | 44 | (Hancock and Evans, 2006) |
| | | | | not reported for areas <1100 m$^2$ | | 96 | |
| Almeria, southeast Spain | 182 | rangeland | 3.7 | $S=0.101A^{-0.267}$ | 0.44 | 52 | (Vandekerckhove *et al.*, 2000) |
| Colorado, USA | 300-500 | sagebrush and trees | 3.8 | $S=0.16A^{-0.26}$ | lower limit* | 17 | (Patton and Schumm, 1975) |
| Middle Veld, Swaziland | 650-1000 | cultivated or rangeland | 3.8 | $S=0.3044A^{-0.261}$ | 0.50 | 27 | (Morgan and Mngomezulu, 2003) |
| Middle Veld, Swaziland | 650-1000 | cultivated or rangeland | 5.6 | $S=0.2489A^{-0.1795}$ | 0.10 | 26 | (Morgan and Mngomezulu, 2003) |
| Middle Veld, Swaziland | 650-1000 | cultivated or rangeland | 8.2 | $S=0.1302A^{-0.1222}$ | 0.07 | 30 | (Morgan and Mngomezulu, 2003) |
| Middle Veld, Swaziland | 650-1000 | cultivated or rangeland | 9.1 | $S=0.2142A^{-0.1103}$ | 0.35 | 30 | (Morgan and Mngomezulu, 2003) |
| South Downs, UK | 750-1000 | cultivated | 4 | $S=0.09A^{-0.25}$ | lower limit* | | (Boardman, 1992) |

(*Continues*)

**Table 1.** (Continued)

| Location | MAP (mm) | Vegetation | Relative slope exponent | Original equation form | $r^2$ | N | Citation |
|---|---|---|---|---|---|---|---|
| northeast Portugal | 740 | cultivated | 4.4 | $S = 0.102A^{-0.226}$ | 0.21 | 50 | (Vandekerckhove *et al.*, 1998) |
| East coast, North Island, New Zealand | 2400 | cyclone-disturbed forest | 4.7 | $S = 0.787A^{-0.213}$ | not reported | 13 | (Parkner *et al.*, 2007) |
| Guadalentin, southeast Spain | 300-330 | cultivated | 7.0 | $S = 0.146A^{-0.142}$ | 0.15 | 37 | (Vandekerckhove *et al.*, 2000) |
| Guadalentin, southeast Spain | 300-330 | cultivated | 7.5 | $S = 0.157A^{-0.133}$ | .06‡ | 50 | (Vandekerckhove *et al.*, 1998) |
| Guadalentin, southeast Spain | 225-483 | cultivated or abandoned fields | 7.7 | $S = 0.15A^{-0.13}$ | 0.15 | 46 | (Nachtergaele *et al.*, 2001a) |
| Guadalentin, southeast Spain | 280 | cultivated | 9.6 | $S = 0.227A^{-0.104}$ | .06‡ | 40 | (Vandekerckhove *et al.*, 2000) |
| Lesvos Island, Greece | 405 | rangeland | 7.2 | $S = 0.285A^{-0.139}$ | 0.16 | 50 | (Vandekerckhove *et al.*, 2000) |
| | | | | **This Paper** | | | |
| Piedmont, North Carolina, USA | 1140-1180 | forest | 1.27 | | 0.65 | 100 | |
| | | | 1.18-1.33 | | lower limit | | |

*Threshold visually identified from lower envelope curve of data by Vandaele *et al.* (1996) following precedent of Patton and Schumm (1975)
**Relative slope exponent and $r^2$ calculated here.
† <16 identified by field surveys, rest identified by DEM only
†† Chosen to match Vandaele *et al.* (1997), appropriate threshold for 55 of 58 points in dataset
‡ Reported as not statistically significant at $P < 0.1$

less permeable B horizon was exposed, runoff increased and gullying often resulted. In some areas up to 90% of croplands were dissected by gullies up to 3 to 12 m deep (Ireland *et al.*, 1939). Where land became too gullied for continued agriculture use, it was abandoned and forested vegetation returned (Ireland *et al.*, 1939). Often the same areas were later cleared and used for agriculture again, and where land was too steep for farming, road building and timber harvest often produced similar soil erosion and gullying (James *et al.*, 2007).

Substantial changes in soil texture and reduction in A horizon thickness were associated with plowing, and these changes were associated with a significant reduction in capillary pore space and increase in runoff (Hoover, 1949). Soils that had the most desirable physical properties and hydrologic characteristics prior to European settlement were farmed and eroded the most intensively, resulting in an overall reduction in variability of soil properties across the Piedmont (Hoover, 1949).

Gullies persist in the forested landscapes of the Southern Piedmont (Galang *et al.*, 2007; James *et al.*, 2007). Historical photographs and field surveys reveal that areas reforested before 1939 have more frequent, deeper, and longer gullies than areas where agriculture was abandoned after 1939 (Galang *et al.*, 2007). These gullies continue to deliver water and sediment to larger streams.

Today, agriculture and second growth forests are the dominant land uses in the North Carolina Piedmont, but there are also three rapidly growing metropolitan areas. At present, about 40% of the North Carolina Piedmont is forested, but urban land use in the Piedmont is expected to increase by 10 to >25% over the next 50 years (Wear and Greis, 2011), largely at the expense of forested areas.

Data were collected from six forested study areas in the North Carolina Piedmont (Figure 1, Table II). Each study area is predominantly covered with second growth mixed hardwood-conifer forests, and has documentation or evidence of historical agricultural activity. Three study areas are owned by private land conservancies: Allison Woods (AW); Redlair – North (RN); and Redlair – South (RS). Allison Woods was owned by the same family for over 250 years, and the property was farmed until at least the 1930s (Allison Woods Foundation, 2007). The Redlair study areas comprise multiple small farms that were cultivated for cotton and general crops from the 1830s or earlier to almost 1950. By 1960, all presently forested areas had been planted with pines or allowed to regenerate hardwoods (Rankin, 2012). The Lake Norman (LN) study area encompasses a forested portion of Lake Norman State Park, which was donated to the State of North Carolina in 1962 by Duke Energy (North Carolina Division of Parks and Recreation, 2012). The final two study areas are located within the Birkhead Mountains Wilderness in Uwharrie National Forest: Uwharrie – East (UE) and Uwharrie – West (UW). Beginning in the 1760s, the land was cultivated by many small farmers, but the tract was consolidated into a single plantation in the 1850s. In 1931, the federal government purchased the land and, in 1984, established it as a wilderness area (USDA Forest Service, 2012).

Soils in the study areas have a variety of loamy textures (Table II) overlying thick clay and silty clay layers (Soil Survey Staff, 2012). Saturated hydraulic conductivity of the clay and silty clay layers ranges from 1.4 to 5 cm/h. Below the clay, there may be more loamy horizons, which are often clay-rich, and weathered bedrock. Outside of floodplains and riparian areas, water tables are >2 m below the land surface, except during rain events. At AW, the dominant soil series has 25 cm of clay loam on top of 76 cm of silty clay. At LN, the dominant soil series classifications have 15–23 cm of gravelly-sandy loam to sandy loam over >50 cm of clay. At RN, the A horizon consists

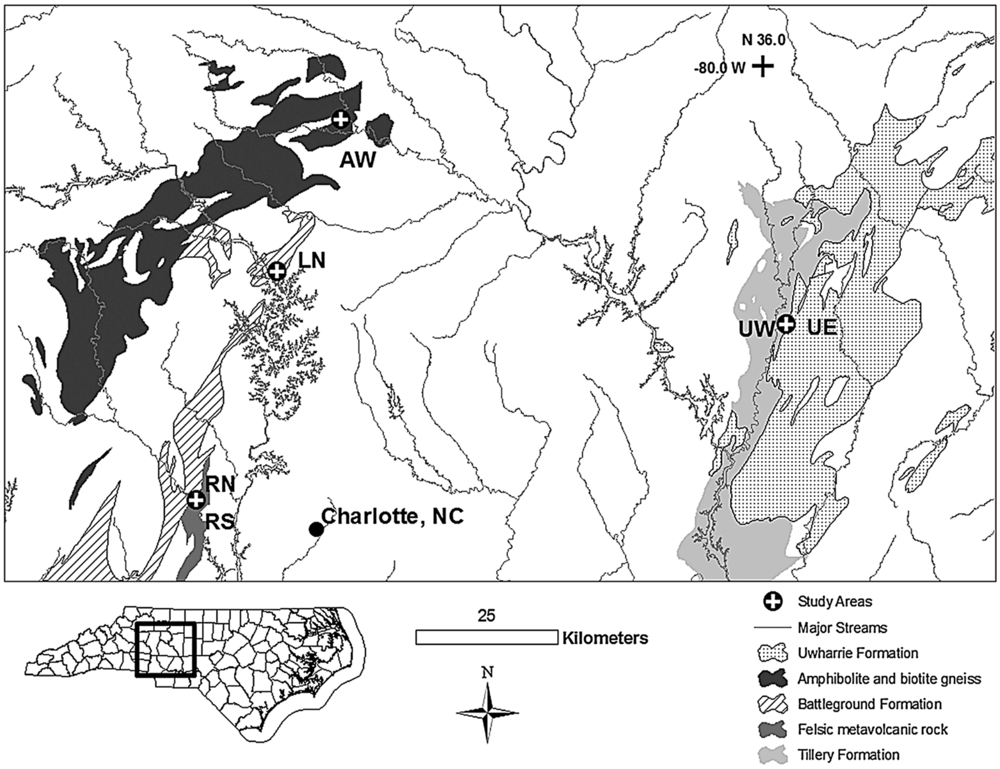


**Figure 1.** Map showing study areas relative to selected geologic formations of the North Carolina Piedmont. Geologic units are from the Geologic Map of North Carolina (Geological Survey of North Carolina, 1985).

**Table II.** Characteristics of study areas. Geologic units are from the Geologic Map of North Carolina (Geological Survey of North Carolina, 1985)

| Site | Location | Lithology | Dominant A horizon soil texture | Area ($km^2$) | Channel heads |
|---|---|---|---|---|---|
| Allison Woods (AW) | N 35.91 W 80.82 | amphibolites and biotite gneiss | clay loam | 1.18 | 14 |
| Lake Norman State Park (LN) | N 35.67 W 80.94 | quartz sericite schist (Battleground Fm.) | sandy loam to gravelly sandy loam | 1.19 | 20 |
| Redlair - North (RN) | N 35.31 W 81.09 | quartz sericite schist (Battleground Fm.) | gravelly silt loam | 1.43 | 25 |
| Redlair - South (RS) | N 35.30 W 81.09 | felsic metavolcanic rock | sandy loam | 0.39 | 14 |
| Uwharrie - East (UE) | N 35.59 W 79.94 | felsic metavolcanic rock (Uwharrie Fm.) | silt loam | 0.86 | 24 |
| Uwharrie - West (UW) | N 35.59 W 79.95 | metamudstone and meta-argillite (Tillery Fm.) | cobbly loam | 0.11 | 5 |

of 20 cm gravelly silty loam and sandy loam over 80 cm of clay. At RS, 18–20 cm of sandy loam to sandy clay loam overlie 53 cm of clay, according to soil series classifications for the area. At UE and UW, soils have 15 to 50 cm of very cobbly loam to silty clay loam over 23 to 100 cm of clay and silty clay (Soil Survey Staff, 2012). Relief at the sites ranges from 37 to 92 m. The gentlest average slopes (0.1 m/m) are found at AW, which is underlain by amphibolites and biotite gneiss. The steepest average slopes are found at the two sites underlain by felsic metavolcanic rocks: RS (0.21 m/m) and UE (0.19 m/m).

## Methods

Channel heads to 100 ephemeral streams were identified in the field at six study areas (Figure 1). Channel heads were identified by the upstream limit of erosion and flow within definable banks (Montgomery and Dietrich, 1989). Channel heads were commonly bowl-shaped depressions on the order of 1 m wide and 0.1 m deep, within larger hollows or locations of convergent topography, on the order of 10 m wide and 1–10 m deep. Downstream of the channel head, there were incised stream banks and evidence of recently flowing water (i.e., debris jams of leaves and sticks, sticks oriented downstream, or bare soils). Contributing areas upslope of the mapped channel heads were fully forested with mature hardwood forests, as verified in the field.

The mapped channel heads contribute flow to 41 small watersheds (<0.9 $km^2$), which are primarily forested and underlain by one rock type (Figure 2). Field observations were performed in the watersheds of each mapped channel head to ensure that there were no impervious surfaces or agriculture within the upslope contributing area. Not every channel head in each watershed was mapped, because of upslope areas with roads or residential or agricultural land use or because of lack of access. In the RS study area, 14 additional channel heads were mapped in watersheds with present day pasture land use, in order to assess the effects of current agricultural practices on channel head morphology and position.

 

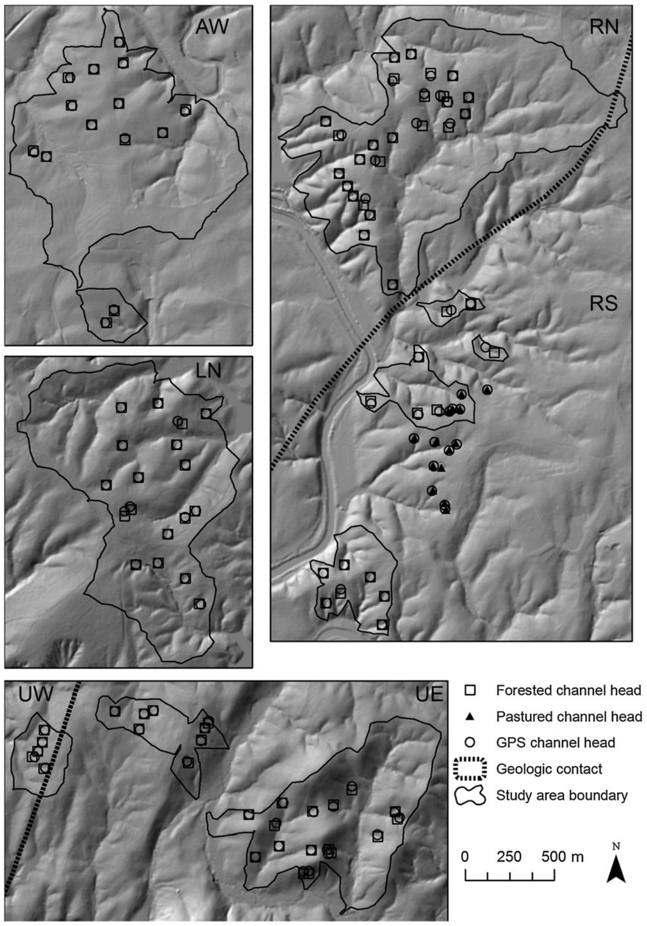


**Figure 2.** Maps showing GPS and corrected channel head positions relative to topography in the six study areas. Two letter designations name each study area as located in Figure 1. Because of their proximity, RN and RS are depicted in a single map, as are UW and UE. Outlines indicate the boundaries of the aggregated watershed areas were used to generate Figure 5. Geologic contacts are from the Geologic Map of North Carolina (Geological Survey of North Carolina, 1985).

Channel head positions were recorded with map-grade global positioning system (GPS) units, including a Garmin GPSMAP 76CSx handheld, Garmin GPS MAP 76CSx with backpack configuration, and Trimble Geo XT hand held. The median distance between coordinates with different GPS units at the same channel head is 2.63 m. GPS positions were plotted on a 6.096 m (20 ft) digital elevation model (DEM), based on Light Detection and Ranging (LiDAR) data (North Carolina Department of Transportation, 2007).

GPS coordinates taken at the field observed channel head sometimes plotted in areas of divergent topography on the DEM, even though field descriptions indicated that the channel head was positioned in a convergent topographic position or along a valley axis. In these cases, field notes and flow accumulation and slope maps were used to select a DEM pixel inferred to contain the channel head (Figure 2). This pixel selection process caused GPS points to be moved an average of 8.4 m. Twenty-two points were not moved, 54 points were moved by one to two DEM pixels, and only 8 points were moved by more than 5 DEM pixels. Dense forest canopy cover in the study areas impeding GPS performance is a likely source of the discrepancies between field descriptions and GPS points.

After correcting channel head positions, contributing area was calculated for the selected pixel, while slope was averaged for the selected pixel, and the upslope and downslope pixels with the greatest contributing areas (i.e. over a length of 18.3 m along the valley axis). Both contributing area and slope were defined using the D-infinite algorithm (Tarboton, 1997). Regression equations for area and slope of channel heads in each study area and for the entire dataset were fitted using JMP, Version 8 (SAS Institute Inc., Cary, NC). The effects of pixel selection on the relationship between area and slope at channel heads was evaluated by comparing the regression equation for all 100 channel heads with the regression fitted only to the 22 points not moved from their GPS positions.

In order to understand how channel head positions are situated in a landscape context, contributing area and slope for all pixels within each study area were also computed using the D-infinite algorithm. Since each study area had >3000 pixels, results were binned into 100-pixel groups sorted by contributing area, for visualization purposes.

In order to understand flow generation mechanisms and flow frequency at channel heads, observations were made during and following rain events between 27 July 2010 and 1 April 2011 at six ephemeral streams in the RN and RS study areas. The streams were selected on the basis of stream lengths, cross-sectional areas, and slope of the adjacent terrain to be representative of the ephemeral networks in the region. Each channel head was instrumented with a crest stage gage, built to United States Geological Survey specifications with PVC pipe (Sauer and Turnipseed, 2010). Crest stage gages were also installed in the unchanneled valleys upslope of the channel heads for the three streams at the RN study area. In order to assess whether streamflow generation occurred during saturated or unsaturated conditions, two channel heads were also instrumented with Campbell Scientific 616 soil moisture reflectometers in the upper 30 cm of soil, used to record changes to soil moisture every 5 min for the duration of the study. Four sensors were positioned in the uphill hollows 1 m above the channel heads in a semi-circle mirroring the shape and position of the channel head, and one sensor was inserted within the channel at the channel head. The accuracy of the sensors was field checked two times per month and after significant precipitation events using a Campbell Scientific CS620 hand-held reflectometer, calibrated to soils at the sites. Precipitation was recorded with a standard and a tipping bucket rain gage in a clearing between RN and RS.

## Results

### Field observations

In fully forested watersheds, all mapped channel heads were located within or downslope of colluvial hollows, rather than incised onto the divergent or planar surfaces of hillslopes. Channel heads had an average width of 1.1 m and average depth of 0.2 m. Channel heads did not coincide with gully initiation points, which were observed but not mapped. Colluvial hollows upslope of channel heads may represent former gully locations, and active gullying occurred downstream of channel initation, if at all.

At nearly all mapped channel heads, across all soil and rock types, undercutting was observed beneath exposed roots, and many channel heads had visible macropores. All channel heads were incised into soil, and no channel heads in forested watersheds were found with bedrock exposures at or near the point of channel initiation. All of the mapped channel heads initiated ephemeral streams. In the study area, ephemeral streams, active only during and for a few hours following precipitation events, form the headwaters of the channel network. Typically, ephemeral channels are 80 to 300 m long upslope of confluences with larger streams or small seeps that initiate intermittent or perennial flow in first-order streams.

Between 27 July 2010 and 1 April 2011, only two of 41 precipitation events ≥1 mm generated flow or standing water in

 

**Table III.** Flow generating precipitation events at Redlair North and Redlair South. Activated channel heads had standing or flowing water recorded by crest stage gages

| Event date | Activated Redlair North channel heads | Activated Redlair South channel heads | Standard rain gage total precipitation (mm) | Tipping bucket total precipitation (mm) | Maximum 60 minute precipitation (mm/h) |
|---|---|---|---|---|---|
| 8/19/2010 | 3 | 3 | 108 | NR* | 50.9* |
| 2/4/2011-2/5/2011 | 1 | 1 | 66 | 40.1 | 12.3 |
| 3/9/2011 | 0 | 1 | 25.4 | 18.6 | 7.1 |

*Event not recorded by tipping bucket rain gage. Precipitation intensity estimated based on the total precipitation recorded in the standard rain gage and the duration of precipitation at the Gastonia airport, 10.5 km southwest.

the channel heads at RN, while three events generated water in the RS channel heads (Table III). Only one event, an intense thunderstorm early in the morning of 19 August 2010, produced flow at all six of the monitored channel heads. This event had a precipitation intensity of approximately 50 mm per hour. The two other precipitation events that produced flow at the channel heads were lower intensity spring storms, but they had the wettest antecedent soil moisture of any events during the study period.

Measurements in the hollows above the channel heads combined with observations made during rain events allow us to identify flow generation mechanisms operating in the forested Piedmont. Soil moisture in the hollows was spatially variable and saturation was reached only briefly, if at all, in any area outside of the channel. Inflow from macropores was observed in several channel heads during rain events, and crest stage gages installed along the valley axis 3 m upstream of three RN channel heads never recorded overland flow. Based on these observations, subsurface stormflow appears to be the most frequent mechanism for flow generation in the channel heads and ephemeral channels. Saturation overland flow may occur during winter and spring storms with wet antecedent conditions. Infiltration excess overland flow is likely the most rare flow generation mechanism, and the NRCS reports infiltration capacities of 14.5–36.1 mm/h for the soils at RN and RS (Mockus *et al.*, 2007). Precipitation rate reached or exceeded the infiltration capacity only during one to two events during the study period. Such intense summer storms and the occasional tropical cyclone may generate significant infiltration excess overland flow, even in forested areas.

## Channel head positions

The average channel head contributing area was 7800 m$^2$ and the average slope was 0.13. Channel head contributing areas ranged from 1000 to 33,000 m$^2$, and there was a wide range of contributing areas at each site. Four sites (RN, RS, UE, UW) had average contributing areas of 5100 to 6500 m$^2$, while AW and LN had much larger average contributing areas (11,000–12,000 m$^2$). Channel head slopes ranged from 0.04 to 0.35, with average channel head slope at RN, RS, UE, and UW in the range 0.14–0.16, and less steep average channel heads at AW and LN (0.08–0.09).

For the 100 Piedmont channel heads, there is an inverse relationship between contributing area (m$^2$) and slope (m/m) (Figure 3). Rearranging the regression to the form of Equation (1), the relationship is given by:

$$380 = AS^{1.27} \quad (4)$$

This relationship is relatively strong ($r^2 = 0.65$) and statistically significant ($P < 0.001$). If the best fit equation is forced to have a relative slope exponent of 1, the resulting equation is $680 = AS$ ($r^2 = 0.62$). Similarly, the best fit equation with a relative slope exponent of 2 is $77 = AS^2$ ($r^2 = 0.45$). Both of these equations are statistically significant, but the equations with relative slope exponent of 1.27 and 1 better fit the data, as shown by the higher r$^2$ values.

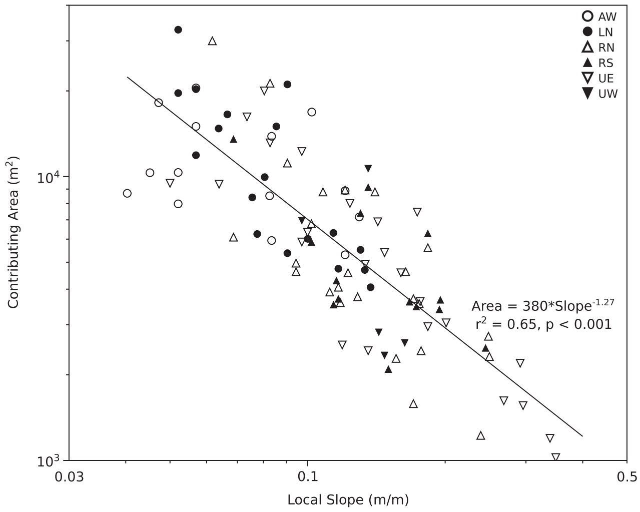


**Figure 3.** Plot of local slope and contributing drainage area of mapped channel heads. Regression line is fit to all data.

Following the precedent of earlier work (Patton and Schumm, 1975; Vandaele *et al.*, 1996), a line was drawn through the lower limit of the data to represent a minimum threshold area–slope relation for channel initiation. The resulting equation is $195 = AS^{1.18}$ if all data are included and $151 = AS^{1.33}$ if one point is excluded as an outlier. These equations have similar relative slope exponents to the best fit equation for the dataset, which makes use of all of the data, so the lower limit threshold equations are not considered further.

The effect of channel head pixel selection (i.e. moving GPS points that fell outside of areas with convergent topography) was examined by comparing the regression results for the full dataset (Equation (4)) to a subset of 22 channel heads where the GPS location was the same as the selected pixel. For those 22 points, the best fit equation is $562 = AS^{1.06}$ ($r^2 = 0.74$). Thus, the pixel selection process decreased the constant threshold by 32% and increased the relative slope exponent by 20%. However, the advantages of the pixel selection method, in improving the correspondence between field notes and DEM location and increasing the number of points for analysis, may outweigh increased uncertainty in the constant and relative slope exponent.

Relative slope exponents for sites with statistically significant best fit lines are 1.62 for LN, 1.47 for RN, 0.98 for RS, and 1.34 for UE. The strongest relationship between channel head area and slope ($r^2 = 0.75$) is observed at UE, which also has the greatest range of channel head slopes. At AW and UW the area–slope relationships are not statistically significant. For

UW, this is likely due to the small number of channel heads mapped (n = 5). The AW site has the lowest average slopes for both the mapped channel heads and the study area as a whole, and it also has the smallest coefficient of variation (standard deviation divided by the mean) of channel head contributing areas. The coefficient of variation in channel head contributing areas at AW is 42%, versus 60% to 100% at other areas. The combination of gentle slopes and relatively low variability in contributing area results in a lack of fit for an area–slope initiation function. Nonetheless, the similarity of slope and area for channel heads and study areas between AW, UW and the sites with statistically significant relative slope exponents warrants the inclusion of AW and UW channel heads in the overall dataset.

## Effects of present day land use on morphology and position

In the RS study area, 14 channel heads were mapped in fully forested watersheds and an additional 14 channel heads were measured and mapped in watersheds with current livestock pasture (Figure 2). The channel heads with upslope pasture had an average width of 0.9 m and an average incision depth of 0.8 m. Channel heads ranged from 0.2 to 2.1 m wide and 0.1 to 3 m deep. The two deepest channel heads were incised to saprolite or bedrock, and three channel initation points occurred at gully heads >1 m deep. Five channel head positions had been artificially stabilized by placed wood, concrete chunks, or rubbish. Artificially stabilized channel heads had similar slopes and areas as unstabilized channel heads in pastured watersheds (Figure 4). Nine of 14 channel heads were within 10 m of the pasture boundary, with two channels extending into the pasture. Informal observation in other disturbed areas adjacent to our study sites suggests that the channel initiation morphologies and positions described here are typical. While no measurements of soil moisture or infiltration capacity were made in currently pastured areas, we frequently observed overland flow upslope of channel initiation sites. We attribute this to compaction and low infiltration capacity in grazed areas (Daniel *et al.*, 2002; Pietola *et al.*, 2005).

In addition to morphological differences, channel initiation positions were different between fully forested and pastured watersheds (Figure 4). Slopes were similar, and average contributing area was slightly greater for the pastured watersheds, but the difference was not statistically significant. There was a substantially larger range of contributing areas (900 to 30,000 m$^2$) for pastured watersheds than for forested ones (2000 to 13,000 m$^2$). An F-test for equality of the two standard deviations showed a statistically significant difference ($P = 0.0006$) between the pastured and fully forested watersheds. The area–slope relationship for channel heads in pastured watersheds has a greater range and is much less well-defined than for fully-forested watersheds at RS, with the relative slope exponent (1.43) not statistically significant ($r^2 = 0.18$, $P = 0.14$).

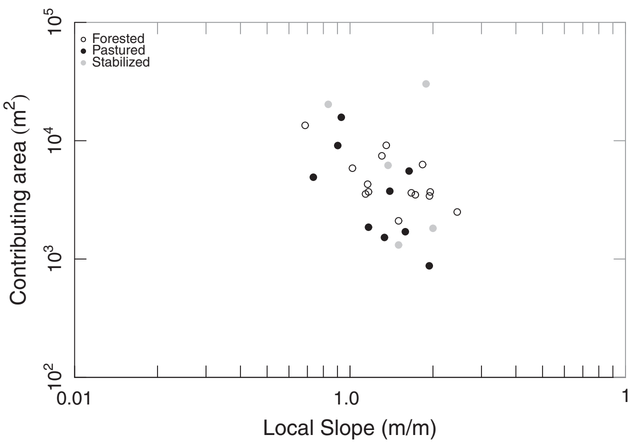


**Figure 4.** Local slope and contributing drainage area of mapped channel heads in fully forested versus partially pastured watersheds in the Redlair South study area. Filled gray circles indicate artificially stabilized channel head positions in partially pastured watersheds.

The similar range of slopes at channel initiation sites regardless of modern land use is unsurprising, since the slope is a relatively fixed property of the landscape at short time scales. The greater variation in contributing areas for channel initiation may be explained by at least three factors. Soil compaction and vegetative hydraulic roughness within the pasture may produce overland flow that has limited capacity to incise a channel until it enters a forested location with less compaction and lower roughness (Prosser *et al.*, 1995). Livestock trampling may also obscure morphological indicators of channels within pastures. A third explanation is that continuing human influences on channel head position through stabilization by placing materials to prevent upslope retreat of the channel head creates a wide range of contributing areas and decreases the topographic predictability of channel initiation. The occurrence of present day stabilization at pastured channel heads, along with abundant historical documentation of stabilization, suggest that direct human influences on channel initiation and position cannot be neglected when assessing channel network extent in current or former agricultural areas.

## Process zones

In order to understand the relationship between the current extent of the channel network and long-term landscape evolution, contributing area and local slope was examined for 41 watersheds in the six study areas. The watersheds have a maximum drainage area of 330,000 m$^2$ and include all of the mapped channel heads. A single area–slope plot for each study area was created, by aggregating 1 to 12 watersheds. Thus, the plots do not represent the profile of a single stream, but rather are a depiction of the overall landscape of each study area (Figure 5). The scaling relationships between area and slope in plots such as those in Figure 5, have been used to infer geomorphic process zones in a variety of landscapes (e.g. Montgomery and Foufoula-Georgiou, 1993). These inferred process zones are illustrated on the area–slope plot for RS (Figure 6).

DEM pixels with small contributing areas are located on hillslopes and in unchanneled valleys. The smallest contributing areas have a convex form (Figure 6, region A), which converts to a concave form between 100 and 300 m$^2$. This inflection is typically coincident with a transition from divergent to convergent topography associated with the hillslope to valley transition, as defined by Montgomery and Foufoula-Georgiou (1993). Dividing the contributing area of the convex–concave transition by the DEM pixel width, hillslope lengths of 15 to 45 m are estimated. These hillslope lengths are consistent with field observations. Pixels with contributing areas between ~300 and 1000 m$^2$ represent unchanneled valleys (Figure 6, region B), since the smallest channel head contributing area is 1000 m$^2$. For contributing areas <1000 m$^2$ (i.e. hillslopes and unchanneled valleys), the average slopes range from 0.11 at AW to 0.21 at RS. Pixels with contributing areas >1000 m$^2$ are a mixture of unchanneled valleys, channel initiation sites, and channels.

At large contributing areas, most sites exhibit rapidly declining average slopes with increasing area (Figure 6, region D). A decrease in slope with increasing contributing area that follows

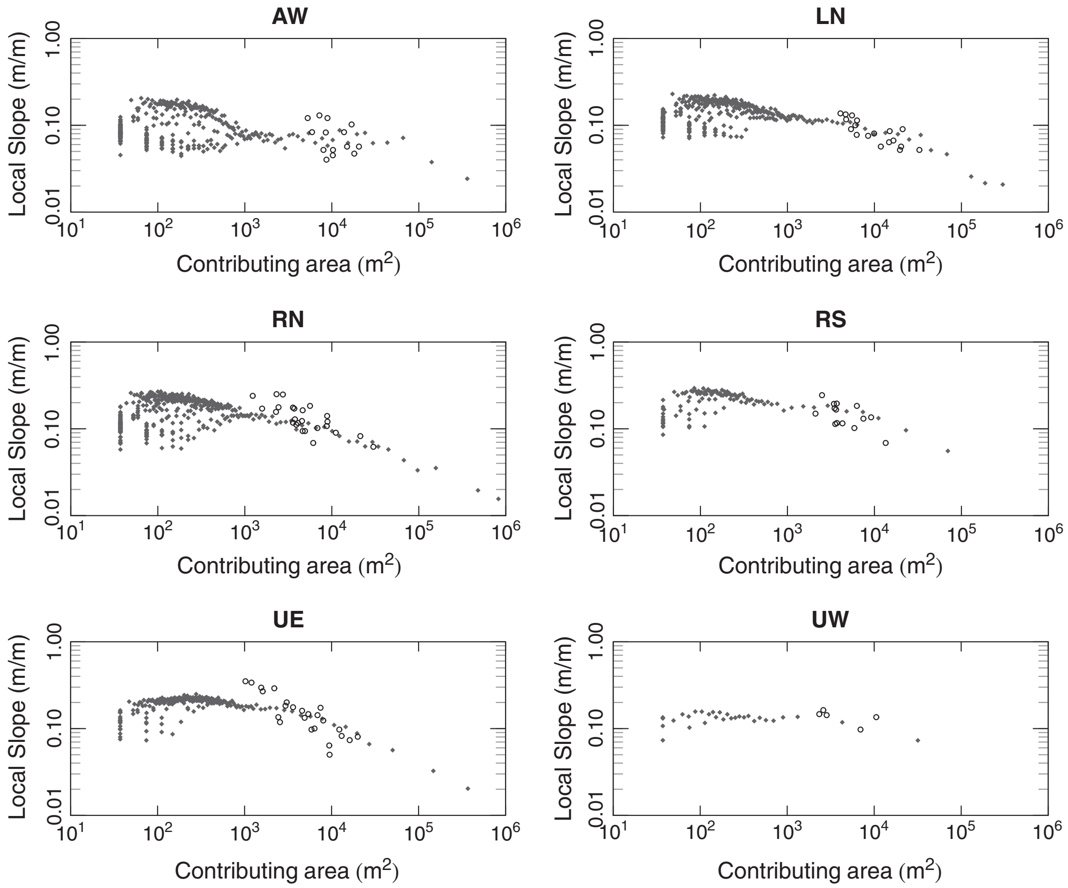


**Figure 5.** Area–slope relationships for study areas. Filled diamonds represent watershed DEM pixels in 100 pixel bins. Open circles represent channel head locations.

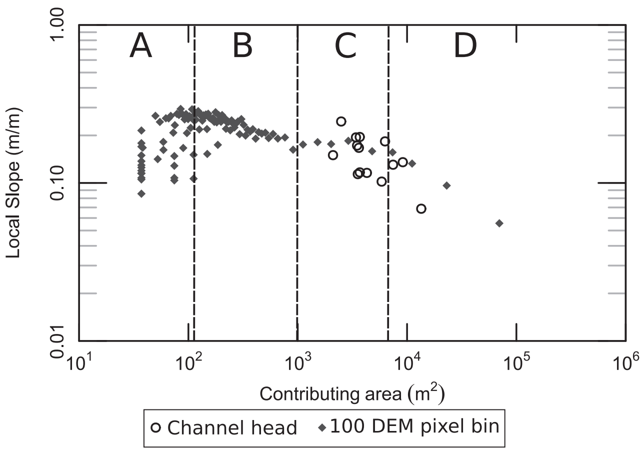


**Figure 6.** Area–slope relationship for Redlair South study area, illustrating regions with different scaling properties.

a power law relationship is representative of a concave longitudinal profile, and such a pattern is generally interpreted to result from fluvial erosion (Montgomery and Foufoula-Georgiou, 1993). At contributing areas greater than about 2000 m$^2$, RN and UE follow this pattern. Study areas LN, RS, and AW exhibit this power law relationship at contributing areas greater than 5000, 7300, and 66,000 m$^2$, respectively. The UW site is too small for meaningful analysis of slope trends for areas >1000 m$^2$.

The LN, RS, and AW study areas each have a region that has little trend in slope in the region between unchanneled valleys and fluvial domains on the area–slope plots (Figure 6, region C). For LN, slope is nearly constant between 1000 and 5000 m$^2$, and for RS slope does not vary much between 900 and 7300 m$^2$. AW exhibits a much larger range of areas with nearly constant slopes (800 to 66,000 m$^2$). Between sites, the average slope in this region is variable, from 0.17 at RS to 0.07 at AW. At greater contributing areas, all three sites exhibit a power law decrease in slope with increasing contributing area. Similar inflections in the area–slope relationship have been interpreted as occurring at the transition from debris flow channels to fluvial channels in other regions (Montgomery and Foufoula-Georgiou, 1993; Stock and Dietrich, 2003), but there is no recorded history or observable geomorphic evidence of debris flows at any of the field areas. Debris flows are thought to be rare in low relief landscapes like the Piedmont (Eaton *et al.*, 2003). At LN, the mapped channel head with the smallest contributing area occurs at 4000 m$^2$, just before the inflection between the constant slope and fluvial domains. At RS and AW, channel heads occur primarily in the constant slope region.

Landscape concavity is expressed as a power equation of the form $S = A^{-\theta}$, where $\theta$ is the concavity as in Whipple and Tucker (1999). Fitting the power equation to all bins in the unchanneled valley, constant slope, and fluvial domains results in very low $\theta$ and over-predicts slopes in the fluvial domain. Therefore, concavity was determined by fitting a power equation to the area–slope plot for bins in the fluvial domain. For LN, RN, RS, and UE, $\theta$ is between 0.35 and 0.46, while for AW it is 0.63, but there it is based only on a few bins. Most of these values are slightly lower than average for streams in the eastern and midwestern USA ($\theta = 0.5$–$0.6$), but are within the reported range of values ($\theta = 0.11$–$2.3$) (Flint, 1974).

## Discussion

Given the very long time scales of geologic and climatic stability in the Piedmont, any channel network instability seems likely to result from human activities, principally agriculture, over the last several centuries. Agricultural impacts on channel form and extent are well-documented in the historical literature of the region (Ireland *et al.*, 1939; Trimble, 1974). Comparisons between morphology and position of channel heads in watersheds with continuing agricultural impacts and those that have fully afforested demonstrate both the effects of agriculture and on-going recovery of the channel network. This recovery is evident within decades of afforestation, in the decrease in channel head dimensions and the lack of channel initiation at gully heads. The larger minimum area required to initiate channels in forested areas than pastures can also be interpreted as recovery through a decrease in channel network extent. A decrease in network extent, or downslope retreat of channel head positions, occurs if the rate of incision by stormflow is less than the rate of colluvial infilling of the channel (Montgomery and Dietrich, 1994). Afforestation and a subsequent increase in infiltration capacity could result in less frequent or less powerful overland flows, allowing infilling and a downslope retreat of channel heads. We suggest that historical gullying influences present day channel network extent, but that in fully forested watersheds, channel initiation is not solely regulated by the location of former gully heads.

The ability to use channel network extent to infer the hydrologic and geomorphic history of the landscape is significantly limited by several factors. First, we have no way of measuring or estimating how much the threshold for channel initiation ($c$ in Equation (1)) may have changed during and following the agricultural period. This threshold is partially determined by soil properties, and extensive loss of the A horizon has been documented throughout the region (Ireland *et al.*, 1939; Trimble, 1974). It is therefore likely that agricultural practices resulted in the mutual adjustment of both the threshold and relative slope exponent for channel initiation. Second, there are no maps of channel networks in the study areas either before the onset of intensive agriculture or during its peak. Consequently, we cannot quantify the actual extent of network expansion during the agricultural era or the potential contraction it has experienced in the afforestation period. Third, the lack of maps forces us to infer what the relative slope exponent of the pre-agricultural channel network was likely to have been, based on studies in other areas.

Flow generation mechanisms in the study area are typical of forested watersheds throughout the southeastern USA and other regions (Hewlett and Hibbert, 1967; Dunne and Black, 1970). However, the relationship between slope and area at channel initiation sites in the North Carolina Piedmont is substantially different from that previously reported for forested landscapes in other regions (Table I). In particular, the channel initiation function has a lower relative slope exponent than expected based on the literature for forested landscapes with similar flow generation mechanisms. If a relative slope exponent of ~2 is taken to represent well-adjusted conditions for channel networks resulting from saturation overland flow or laminar infiltration excess overland flow (Montgomery and Dietrich, 1994), as our limited observations suggest are the dominant present-day flow generation mechanisms in the study areas, then significant deviation from this exponent is a sign of a channel network poorly adjusted to present flow generation conditions.

We interpret the present relative slope exponent as a relict of past land use, in which flow generation mechanisms were likely different from today. During the period of intensive agriculture, when infiltration capacity was lowered by loss of topsoil and exposure of B horizon clays (Hoover, 1949), it is likely that infiltration excess overland flow was more common than it is now. Such interpretation is consistent with descriptions of sheetwash and erosion from that period (Ireland *et al.*, 1939; Hoover, 1949). Measurements in a range of settings suggest that vegetated hillslopes typically have a laminar to transitional overland flow regime (Emmett, 1970), but during fallow periods, it is possible that surface roughness in the study areas was low enough to permit the development of turbulent overland flow. In the case of turbulent infiltration excess overland flow, Montgomery and Dietrich (1994) combined Equation (3) and the Manning equation to show that channel initiation occurs at

$$c_t = AS^{\frac{7}{6}} \tag{5}$$

where $c_t$ is a threshold value that depends on $\tau_c$, $\rho$, $g$, infiltration excess rainfall rate, Manning's roughness coefficient, and DEM resolution. The similarity between the relative slope exponent identified in this study (1.27) and that suggested by Montgomery and Dietrich (1994) (1.17) raises the possibility that turbulent overland flow may have contributed to the extent of the present channel network. If a relative slope exponent of 2 is appropriate for channel networks with saturation overland or laminar infiltration excess overland flow generation mechanisms (Table I), then recovery from historical agriculture disturbance and gullying should be characterized by a channel network with a relative slope exponent that approaches 2 over time. More work is needed to understand how, and over what time scales, such a change in the relative slope exponent might occur.

Examining how changes in channel network extent occur on decadal time scales following agricultural abandonment and afforestation requires a study area with consistent soils and land use histories, but different and known dates of abandonment (e.g. constrained by aerial photographs). The study areas in this project each had different soils and likely had somewhat different land use histories and time since agricultural abandonment, so the differences in thresholds or relative slope exponents cannot be inferred to represent different time points on a single trajectory. It may be difficult to identify a chronosequence of forested watersheds with ephemeral channels in the Piedmont, given the close ties between soil type and the intensity and timing of land use (Trimble, 1974). In that case, future work should also focus on how time since afforestation affects the thickness, development, and hydrologic properties of the soil A horizon, because of its important role in affecting runoff generation mechanisms.

Placing the channel network in a landscape evolution context, the observational evidence and area–slope plots suggest that the geomorphic legacy of intensive agriculture is most manifest in the upper portions of small watersheds in the study area. As described below, the headward extent of the valley network may reflect the extent of gullying during the agricultural era, while slightly farther downslope, aggradation has had a persistent geomorphic signature. The concavity of the fluvial domain is similar to that of other river systems, suggesting that the geomorphic legacy of agricultural land use is less dramatic there than farther upslope.

Valley extent is thought to reflect a balance between diffusive slope processes and the maximum extent of the channel network on time scales of $10^4$ to $10^6$ years (Montgomery and Dietrich, 1992, 1994). Today, channel initiation in forested watersheds occurs within unchanneled valleys and is not characterized by gully heads at the channel initiation point. The existence of unchanneled valleys upslope of the channel initiation sites suggests that the present day channel network is less extensive than the maximum channel network in the geologic record for this landscape. Historical gullying may have produced or enlarged these now unchanneled valleys, but the current study cannot rule out their existence prior to the agricultural era. If the observed colluvial hollows and unchanneled valleys are remnants of gullying, then the topographic legacy of the agricultural era will persist until diffusive processes smooth the landscape or a new incision pulse extends deeper or farther upslope than that of the agricultural era.

In the current study, area–slope plots show a region with constant slope between the unchanneled valley and fluvial domains (Figure 6, region C), and this region has not previously been described in the literature. We interpret this region as an aggradational feature, produced by overwhelming volumes of colluvium from intense erosion of agricultural hillslopes. Gullies may have cut through portions of these sediment bowls, and modern channel heads exist within this domain, but with relatively small contributing areas and low slopes (0.07–0.17), sediment delivery exceeded transport capacity, and hollows and valleys filled with colluvium. Happ *et al*. (1940) described intermittent stream valleys with small drainage areas as the first places to receive and store sediment in the South Carolina Piedmont, and Costa (1975) identified both sheetwash deposits at hillslope toes and colluvial fans at junctions of intermittent streams as important sites of sediment storage. Costa (1975) estimated that 52% of eroded material was stored in sheetwash and colluvial deposits in the Maryland Piedmont. These previous descriptions of colluvial sediment storage high in the landscape support our interpretation of the constant slope region of the area–slope plots as a colluvial or aggradational domain resulting from historical land use.

From an applied standpoint, this study demonstrates the strong slope dependency of channel initiation in Piedmont watersheds. Contributing areas for channel initiation varied by a factor of 33, emphasizing the inappropriateness of defining a simple contributing area threshold in this landscape, as is done in common GIS-based frameworks. Previous work has suggested that a threshold stream power, represented by an area–slope product with a relative slope exponent of 1, adequately characterizes channel network extent in the South Carolina Piedmont (James and Hunt, 2010). Our results suggest that a relative slope exponent of 1 is appropriate in some areas of the North Carolina Piedmont (e.g. the RS study area), but that other areas may be better represented by an exponent of ~1.5 (e.g. LN and RN study areas). It is clear that the best strategy for quantifying channel network extent is to map a sufficient number of channel heads to define a site-specific channel initiation equation. In the absence of such detailed mapping, an equation of the form

$$c = AS^y$$

where $y$ is 1.0 to 1.5 seems most appropriate for the Piedmont of North and South Carolina and may be effective in abandoned, afforested agricultural areas in other regions.

## Conclusions

In the North Carolina Piedmont, the contributing area and local slope at which channel initiation occurs vary widely within and between sites, but is best described by the equation: $380 = AS^{1.27}$. This equation may have predictive power for application in other southeastern Piedmont sites, but needs to be independently tested. Causes of variation in the channel initiation area–slope relationship between sites may include land use history or soil properties. While these causes need to be further examined, the variation emphasizes the importance of selecting channel initiation thresholds that are as site-specific as possible when delineating channel networks from digital elevation data. The relative slope exponent of 1.27 is substantially lower than previously documented for forested watersheds, and it may reflect the legacy of agricultural erosion. Present day flow generation occurs most commonly by subsurface storm flow, but also occurs by saturation overland flow and occasional infiltration excess overland flow. These flow generation mechanisms are typically associated with channel networks with relative slope exponent ~2, suggesting that the Piedmont's present channel network is not adjusted to present hydrological conditions and may reflect a former hydrological regime.

Channel networks in the Piedmont reflect the legacy of intensive agricultural erosion from the 18th to mid-20th century and recovery following abandonment and afforestation. This legacy includes an extensive ephemeral channel network and colluvial aggradation, but is not manifested by present day channel incision into hillslopes. Ephemeral channel heads differ morphologically and in position between fully forested watersheds and those with continuing agriculture impacts, suggesting that geomorphic recovery is occurring in those watersheds where agriculture has ceased. This recovery may eventually lead to a less extensive channel network than today, possibly with a different relative slope exponent. The humid climate, thick, easily erodible soils, and gentle slopes of the southeastern Piedmont may make this a landscape capable of returning to a form similar to that existing prior to disturbance.

Channel initiation mapping and area–slope relationships derived from DEMs provide hydrological and landscape evolution context for understanding the geomorphic effects of historical and contemporary land use. While the southeastern Piedmont has a rich history of geomorphic investigation that enhances the interpretations possible for our study areas, channel initiation mapping and area–slope relationships may be powerful tools for rapidly assessing historical land use effects in areas with little or no previous study, particularly if disturbed and relatively undisturbed areas can be compared in a single region. These techniques may also be useful for understanding the geomorphic context of present day land use change, such as in the North Carolina Piedmont where afforested areas are being rapidly urbanized.

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