Evolution of transient topography on passive margins: A study of landscape disequilibrium in the southern Appalachian Mountains

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ABSTRACT

The mechanism through which the Appalachian Mountains have maintained moderate relief some ~300 Myr after the cessation of mountain building has long puzzled geomorphologists. As recent studies have shown that Appalachian exhumation has occurred at slow rates consistent with isostatic rebound of thickened crust, the driving forces behind localized episodes of accelerated incision and the associated rugged topography have been difficult to explain given the absence of tectonic uplift. This study uses previously undocumented relict fluvial gravels and knickpoint location to confirm the role of drainage rearrangement in producing local base level drop and subsequent basin-scale transient incision in the southern Appalachians. This process is fundamentally driven by the high potential energy of streams flowing across the elevated, slowly eroding Blue Ridge Plateau relative to the present Atlantic and landward interior base levels. Gravel deposits confirm that repeated capture of landward-draining Plateau streams by Atlantic basin streams, whose immediate base level is 250-300 m lower, forces episodic rapid incision and overall erosional retreat of the Blue Ridge Escarpment along the Plateau margin. The distribution of knickpoints, bedrock gorges, and relict surfaces in the interior of the Plateau indicate that the New River, which drains to the continental interior, is actively incising the low-relief Plateau surface due to episodic drops in landward base level. The origin of landward base level perturbation is unclear, but it may be the result of glacially-driven shortening and steepening of the lower New River during the Pleistocene. Collectively, these data indicate that rapid base level drop through drainage reorganization can energize streams in otherwise stable landscapes and accelerate fluvial incision and relief production without uplift of the land surface. This process is likely quite significant in post-orogenic settings, where inherited drainage patterns may not reflect the most direct, and thus energetically appropriate, path to

present base level. Passive margins may therefore never achieve a topographic steady-state, despite uniformly slow and constant uplift due to isostatic rebound.

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ATTRIBUTIONS

This dissertation benefited greatly from the guidance and support of my doctoral committee. The committee consists of Dr. James A. Spotila (chair), William S. Henika, Dr. Robert J. Tracy, Dr. W. Lee Daniels, and Dr. William A. Ranson. Spotila, Henika, and Tracy are faculty members in the Virginia Tech Department of Geosciences, Daniels is faculty member of the Virginia Tech Crop and Soil Environmental Sciences Department, and Ranson is a faculty member of the Furman University Department of Earth and Environmental Sciences. All committee members provided valuable suggestions and encouragement that contributed to the success of this research. Specific contributions of committee members listed as co-authors of Chapters 2 and 3 are described below.

Chapter 2 was submitted to *Geomorphology* as “New physical evidence of the role of stream capture in active retreat of the Blue Ridge Escarpment, southern Appalachian mountains” by Philip S. Prince, James A. Spotila, and William S. Henika. Philip Prince delineated study areas, prepared drainage basin morphology data, conducted field work with the assistance of William S. Henika, and produced clast roundness data for fluvial transport estimates. Prince also wrote the manuscript, with valuable contributions to scientific interpretation and context provided by Dr. James A. Spotila. Artwork in Figure 2.10 was drawn by Philip Prince in 2010.

Chapter 3 was submitted to *Geology* as “Stream capture as driver of transient incision in a tectonically quiescent setting” by Philip S. Prince, James A. Spotila, and William S. Henika. Philip Prince produced stream profiles and organized and conducted field work with the assistance of William S. Henika, an expert in the geology of the study area. Unrelated geologic mapping in the area by Henika also provided additional field sites following the initial identification of two gravel deposits in Christiansburg, Virginia by Prince. Prince wrote the manuscript, with stylistic and contextual support from Dr. James A. Spotila. Artwork in Figure 3.4 (as well as Figure 4.1) was drawn by Philip Prince in 2011.

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Chapter 1

Introduction

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For more than a century, geomorphologists have sought to understand the development of the complex topography of the Appalachian Mountains. As the persistent erosive action of rivers and streams has long been obvious, the processes by which elevation and relief are maintained despite constant downwearing have been of particular interest. While the structural and tectonic history of the greater Appalachian orogen has become increasingly understood, the range of forcings controlling the origin and evolution of the present-day landscape are yet to be fully constrained. Many regions within the Appalachians appear too rugged for the tectonically quiescent passive margin setting. This ruggedness often takes the form of juxtaposed high- and low-relief domains within a single drainage basin, implying complex and locally variable erosional dynamics despite regionally consistent climate and tectonics. These topographic contrasts are apparent at many scales within the Appalachian landscape, and their uncertain origin and significance to landscape evolution at large has long attracted scientific study.

W.M. Davis developed the earliest published theory of Appalachian topographic evolution from observations in dissected Appalachian Plateau of western Pennsylvania (Davis, 1889, 1899a). Drawing on the "base level" concept of Powell (1875), Davis recognized that erosion is fundamentally driven by the gravitational energy of streams relative to their base level, where potential energy becomes zero. By associating gravitational potential energy with active incision and downwearing, Davis theorized that topographic relief constantly decays towards a low-energy, nearly flat surface (or "peneplane") which is slopes very gently towards base level and lacks the energy to evolve further. He thus believed that topography and active incision were the result of uplift of the land surface, which energizes streams relative to base level and initiates a new phase of erosion towards the ultimate "peneplane" morphology. As the various stages of this "Geographical Cycle" are, in theory, reflected in the appearance of a landscape, early geomorphic analysis of the Appalachians focused on correlating landforms with the progress of denudation following uplift. The Davisian theory was widely accepted following its publication, and into the mid 20th century the contrasts in ruggedness within the Appalachian landscape were attributed to uplifted "peneplanes" in varying stages of dissection.

While Davis' "geographical cycle" and its consideration of potential energy relative to base level were revolutionary for the 19th century, the advent of plate tectonic theory (e.g. Wilson, 1965a) revealed the shortcomings of the model, in particular its emphasis on repeated localized "uplift" events. Following the recognition that orogenesis in the Appalachians ceased in the late Paleozoic, the present appearance of the Appalachian landscape was increasingly attributed to the effects of prolonged downwearing following the cessation of orogenesis. Hack (1960, 1973) regarded Appalachian topography as an erosional "relict" produced by steady post-orogenic denudation. The result of this prolonged erosion was a landscape in "dynamic equilibrium," where topography has become adjusted to independent variables (e.g., bedrock erodibility) to produce a landscape in which erosion is slow and spatially uniform. The dynamic equilibrium concept was initially based on the topographic contrasts between outcrops of differing lithology; hard rocks regularly do support ridges, mountains, and steep streams to erosionally keep pace with lower relief zones supported by weak rock. The identification of a thickened crustal "root" beneath the Appalachians and the resultant isostatic rebound further supports a landscape in dynamic equilibrium, as constant uplift slow uplift could be expected to maintain topography that exists in steady-state with downwearing long after the end of tectonic uplift (Davis, 1903; Schumm, 1963; Pinet and Souriau, 1988; Prowell and Christopher, 2000). Recent ^{10}Be erosion rate studies have produced results consistent with Appalachian dynamic equilibrium, where basins of varied lithology and relief all appear to be eroding at constant rates (e.g., Matmon et al., 2003b; Sullivan et al., 2007).

Despite the general acceptance of Appalachian dynamic equilibrium, the ruggedness of many portions of the Appalachian landscape appears inconsistent with slow, spatially-uniform erosion (Spotila et al., 2004). While some degree of bedrock control on topography and stream gradient is apparent within much of the Appalachians (Hack, 1957, 1960), many Appalachian drainage basins exhibit rugged topography and steepened stream reaches, or knickpoints, within weak bedrock units. The frequent drainage basin-specific distribution of these "non-Hack" landforms suggests perturbation to some local base levels has disrupted topographic equilibrium, but the origin of perturbation is uncertain. The localized distribution of topographic disequilibrium is inconsistent with tectonically-driven uplift, which would be expected to generate more widespread incision which would consistently affect basins within the uplifted area. Many authors have suggested Cenozoic climate change has led to increased incision rates (e.g. Mills, 2000; Hancock and Kirwan, 2007), but this explanation would also be expected to produce more widespread incision than is observed. In order to fully understand the long-term evolution of Appalachian topography, the origin and significance of localized topographic disequilibria to orogen-scale denudation must be better constrained.

Understanding the processes which drive locally accelerated erosion in the Appalachians is also relevant to the topographic evolution of passive margins worldwide. The rifting events from which passive margins result likely produce surface uplift, and the opening of a new ocean basin along the rift axis represents the establishment of a new base level in close proximity to elevated topography. Rift flanks are typically characterized by topographic escarpments, which are known to experience rapid erosion during and immediately after rifting (e.g. Steckler and Omar, 1994). The potential long-term effects of the rifting process on passive margin landscapes are, however, unclear. Rapid erosion may decay background exhumation rates soon after rifting as topography re-adjusts to lithology following the cessation of uplift. Conversely, the topgraphic interface between the landward high topography and the new seaward base level may remain dynamic in the long term. The potential energy of elevated, landward-draining streams relative to the new base level may be tapped through stream capture, driving ongoing incision and topographic evolution long after rift-related uplift ceases (Matmon et al., 2002). As evidence for both of these models exists in the Appalachians and on other passive margins, identifying robust evidence of active Appalachian incision, as well as its origins, will provide a useful constraint on the evolution of post-rift topography around the world.

To investigate the origin and evolution of disequilibrated topography in the Appalachian Mountains, this study has focused on zones of localized incision along the margin of the Blue Ridge Plateau and within the upper New River basin of the southern Appalachian Mountains. These study areas were chosen for obvious decoupling between lithology and topography as well as their potential significance to global-scale geodynamic questions. The Blue Ridge Plateau is bounded to the southeast by the steep Blue Ridge Escarpment, whose crest generally coincides with the regional drainage divide which separates elevated streams of the Plateau surface from the Piedmont surface ~300 m below. Previous work by Spotila et al. (2004) has shown that the Blue Ridge Escarpment, like other passive margin escarpments, has been shaped by focused erosion seaward of its present location. However, the mechanism of that erosion, as well as whether or not the Escarpment is still evolving, remain unclear. The New River is a major landward-draining river system of the Appalachians which appears to exhibit the effects of accelerated stream incision into an otherwise stable low-relief surface underlain by mechanically weak lithologies. Both study areas exhibit rugged topography and knickpoints that are not lithologically controlled, suggesting erosional dynamics are dictated by other forcing factors.

Understanding the controls on incision within these study areas will enhance the understanding of topographic process in the Appalachians and around the world.

The following data sets were developed using traditional methods of topographic analysis and field reconnaissance to reconstruct pre-incision drainage networks in order to determine the origin and evolution of localized landscape disequilibrium in the absence of major surface uplift. Chapters 2 and 3 describe the use of stranded fluvial gravels to confirm the role of episodic stream capture in producing localized, rapid incision and extremely rugged topography along the margin of the Blue Ridge Escarpment. The long term effect of these repeated capture events is ongoing erosional retreat of the escarpment and drainage divide ~200 Ma after rifting, which has not previously been confirmed by field or analytical studies. Chapter 4 applies the same methods of topographic interpretation and field work to document the region-scale spread of transient incision, and thus relief production, in the New River basin following episodic landward base level drop. Although the origin of this base level drop is uncertain, it appears to be late Cenozoic in age and may result from downstream glacial re-routing of the paleo-New River system. Collectively, these new data suggest that basin-scale rapid incision occurs frequently on passive margins when the potential energy of slowly eroding, elevated drainage basins is tapped by a drop in immediate base level, typically through drainage rearrangement. Localized rapid erosion is thus "energetically" possible without uplift as long as local base level drop can occur sufficiently frequently or rapidly to outpace lowering of the relict highland surface through "background" denudational processes. This result provides the first field-based validation of physical (Hasbargen and Paola, 2000) and numerical (Pelletier, 2004) modeling studies which suggest that passive margin landscapes may never reach a "dynamic equilibrium" condition. These data also reveal the potential for traditional methods of topographic analysis to produce useful constraints on landscape evolution when modern dating techniques fail to do so.

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Chapter 2

New physical evidence of the role of stream capture in active retreat of the Blue Ridge Escarpment, southern Appalachians

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ABSTRACT

The Blue Ridge escarpment of the southern Appalachian Mountains is a striking and rugged topographic feature of the ancient passive margin of eastern North America. The crest of the escarpment generally coincides with an asymmetric regional drainage divide, separating steep streams of the escarpment face from low gradient streams of the Blue Ridge upland. Recent exhumation and erosion rate studies suggest that the escarpment has evolved by inland erosional retreat, but the mechanism, timing, and magnitude of retreat remain poorly understood. Longitudinal stream profiles and slope-drainage area relationships of several upland basins draining the divide have led to the identification of 14 previously unknown fluvial terrace deposits preserved at the escarpment crest. These relict terraces and the associated beheaded drainages indicate the role of large stream capture events in producing ongoing escarpment retreat through landward divide migration and subsequent topographic adjustment. Terrace location and preservation suggest rectilinear drainage patterns and divide asymmetry generate discrete high order captures and episodes of rapid localized retreat that collectively produce slower evolution of the escarpment at large. While overall retreat magnitude and rate remain unknown, roundness of terrace alluvium suggests that the most recent captures have locally produced tens of kilometers of retreat within the limited preservation lifetime of the deposits. In contrast with recent numerical modeling and cosmogenic studies, these data show the potential for stream capture and divide migration to sustain passive margin escarpment evolution long after the cessation of rifting. The fluvial record of divide retreat preserved atop the Blue Ridge escarpment suggests the potential for using field methods to better constrain the histories of younger, taller, and potentially more dynamic passive margin escarpments.

INTRODUCTION

Major seaward-facing topographic escarpments are common features of rifted passive margins. Such “great” escarpments are believed to initiate during rift-flank uplift and subsequently evolve through erosional processes (Ollier, 1984; Kooi and Beaumont, 1994; Gallagher and Brown, 1997). Rapid erosion focused on the steep flank of a rift may produce inland retreat (Ollier, 1984; ten Brink and Stern, 1992; Young and McDougall, 1993; Tucker and Slingerland, 1994; Seidl et al., 1996), but the controls on the rate, timing, and mechanism of escarpment migration following rifting are poorly understood. Some great escarpments, such as southwestern Africa and southeastern Australia, appear to have migrated little since the early stages of development (Moore et al., 1986; Gilchrist et al., 1994; Bishop and Goldrick, 2000; Cockburn et al., 2000; Matmon et al., 2002; Persano et al., 2002), while other escarpments, such as southeastern Brazil and the Blue Ridge escarpment of eastern North America, may have continued to retreat into the late stages of escarpment development (Bohannon et al., 1989; Gallagher et al., 1994; Steckler and Omar, 1994; Brown et al., 2000; Spotila et al., 2004). The existence of a universal paradigm applicable to great escarpment evolution is thus questionable.

Further constraining the chronology and mechanisms of great escarpment retreat, particularly along mature passive margins, is significant to improving our understanding of the evolution of passive margin landscapes. After rifting, escarpments may be rapidly excavated by downwearing seaward of a fixed drainage divide (van der Beek and Braun, 1999; Matmon et al., 2002) or experience slow but steady retreat in parallel with the divide (King, 1962; Fig. 2.1). However, no clear global relationship exists between escarpment age (determined by age of rifting) and distance from the coast; the Sri Lanka escarpment (180 Ma, 65 km from the present coastline;Vanacker et al., 2007), the Drakensberg escarpment (130 Ma, 150 km from the present coastline; Moore and Blenkinsop, 2006), and the southeastern Australia escarpment (85-100 Ma, 60 km from the coastline; Weissel and Seidl, 1998) imply different overall retreat distances and long-term average rates. The Blue Ridge escarpment (BRE), a passive margin escarpment often overlooked in great escarpment studies, occurs along the oldest passive margin in the world (initial rifting at ~ 200 Ma; Pique and Laville, 1995; McHone, 1996), yet it is located within 70 km of inland rift basins and maintains steep, youthful topography (Spotila et al., 2004). The timing of exhumation of the area seaward of the BRE is comparable with that observed seaward of the ~100 Myr younger southeastern Australia escarpment, each with low temperature (U-Th)/He apatite cooling ages of ~ 85-110 Ma (Persano et al., 2002; Spotila et al., 2004). The persistent ruggedness of the BRE, despite its age, makes it unique among great escarpments, and further constraining the mechanism and potentially the timing of retreat is necessary to understanding long-term BRE evolution. The results of this study may, in turn, find application to other great escarpments, where establishing additional controls over long-term retreat mechanisms may enhance the interpretation of existing cosmogenic and thermochronologic data sets.

In this paper, we examine fluvial terraces preserved at the crest of the BRE and physical characteristics of beheaded upland drainage basins as a possible record of ongoing parallel escarpment retreat. Terraces were systematically located through the analysis of upland drainage basin morphology. Rounded to well-rounded alluvium from the terraces was subsequently sampled for clast roundness measurements, which were used to estimate clast transport distance. The use of field-based data aids in testing recent models for the origin and evolution of the BRE generated by apatite thermochronometry (Spotila et al., 2004), cosmogenic radionuclides (Sullivan et al., 2007), and drainage basin geometry (Bank, 2001). Terraces provide new field evidence of significant parallel retreat of the escarpment and divide continuing throughout the Cenozoic. Analysis of beheaded drainages provides an additional indicator of escarpment retreat and drainage divide encroachment. The preservation and predictable location of surficial and topographic evidence of BRE retreat also suggests the possibility of using field techniques to better characterize the evolution of younger, and potentially better preserved, passive margin escarpments.

BACKGROUND

Long-term Evolution of Great Escarpments

Clearly identifiable great escarpments occur along approximately one-third of the world's passive margins (Ollier, 1984). These escarpments exist as topographic steps separating low-elevation, low-relief coastal plains from elevated continental interiors. Great escarpments are typically ~ 0.3-1 km high and exist as narrow bands (~ 5-20 km) of rugged topography separating the lower relief coastal plain and upland surfaces. They are generally classified as either "shoulder-type" or "arch-type," depending on the position of the regional divide in relation to the crest of the escarpment (Matmon et al., 2002). The hypothesis of escarpment evolution by landward retreat after rifting is based on observations of active rift margins, where dynamic continental uplift and base level drop seaward of the divide produce focused erosion and steep, seaward-facing landforms.

Recent studies of thermochronometry, cosmogenic radionuclides, and offshore sedimentary data have suggested some escarpments evolve by initially rapid erosion that slows considerably as seaward-flowing streams begin to adjust to rift-related base level (Drakensberg escarpment (Fleming et al., 1999), Namibia (Rust and Summerfield, 1990; Cockburn et al., 2000;

Bierman and Caffee, 2001), Sri Lanka (von Blanckeburg et al., 2004; Vanacker et al., 2007), and southeastern Australia (van der Beek and Braun, 1999; Persano et al., 2002)). Continued knickpoint retreat on these seaward-flowing rivers may, however, increase escarpment sinuosity over time with little retreat of the topographic front (Weissel and Seidl, 1998; Heimsath et al., 2000; Matmon et al., 2002). In contrast, other data suggest significant escarpment retreat continues long after rifting. Inferred exhumation histories of escarpments surrounding the Atlantic basin, such as eastern Brazil, southwest Africa, and eastern North America are consistent with continued late-stage retreat (Gallagher et al., 1994; Brown et al., 2000; Spotila et al., 2004). Drainage patterns along the Western Ghats escarpment suggest continuing capture of upland streams, leading to overall parallel retreat of the drainage divide and escarpment (Harbor and Gunnell, 2007).

Numerical modeling has shown the influence of pre-rift topography, particularly the location of the regional drainage divide, on the rate and mechanism of escarpment evolution (Fig. 2.1). Modeling by van der Beek and Braun (1999) and van der Beek et al. (2002) showed that initial rapid evolution and subsequent stability are consistent with downwearing from a preexisitng inland divide (arch-type of Matmon et al., 2002). Large streams flowing seaward from the divide are energized by rift-induced base level drop and rapidly excavate an escarpment at the divide as they equilibrate to the new base level. Following this equilibration, the lower-energy seaward stream headwaters accomplish minimal additional retreat (Fig. 2.1, A). The more traditional model of prolonged, steady escarpment retreat involves the formation of a new, asymmetric regional divide, which retreats in parallel with the escarpment as a result of constant, focused erosion on the steep flank and upland stability (King, 1962) (Fig. 2.1, B).

While cosmogenic data indicating very slow retreat of many mature escarpments are consistent with the downwearing model (e.g., Cockburn et al., 2000; Brown et al., 2002), slow escarpment erosion could result from reduced relief and thus be consistent with the parallel retreat model as well. Climatic effects have also been suggested as a means of accelerating or slowing parallel retreat (Partridge, 1998). Thermochronological data do not clearly distinguish between downwearing or parallel retreat because of low geothermal gradients and modest escarpment relief of ~ 1 km or less (Braun and van der Beek, 2004). The contrast between these models makes constraining retreat mechanism an important step toward describing the origin and evolution of escarpments. Downwearing and parallel retreat scenarios have been modeled to ultimately produce escarpments of similar morphology despite different rates (Fig. 2.1), suggesting that the present morphology of escarpments may not provide clues to the mechanism of their evolution (Braun, 2006). The application of field methods aimed at determining the role of the drainage divide may add a useful additional constraint to the modeling of escarpment evolution.

Understanding escarpment retreat mechanisms has equally profound implications for the maintenance of rugged landforms over time. The relief-reducing processes of Davis (1902) and Penck (1953) explain low-relief upland and lowland surfaces, but fail to account for passive margin escarpment steepness. Parallel slope retreat as explained by King (1957) allows for the maintenance of relief over time, as the high potential energy of steep hillslopes focuses erosion while allowing upland areas to remain comparatively stable. This concept is now viewed as largely dependent upon lithology and structure, such as resistant caprocks and escarpment-parallel jointing that stabilize uplands and ease removal of material from the escarpment face (Munro-Perry, 1990; Moore and Blenkinsop, 2006; Gunnell and Harbor, 2008). In contrast, Penck (1953) hypothesized that uplands should erode along with the steeper escarpment face, reducing relief over time. This long-standing debate of landform evolution may find some resolution in the study of old passive margin escarpments.

The Blue Ridge Escarpment

The BRE stretches over 500 km along the eastern edge of the southern Appalachian highlands (Fig. 2.2). Rising in 300-600 m of relief above the 200-300 m Piedmont, the BRE represents an abrupt topographic boundary between the lower-relief Piedmont and Blue Ridge upland (or simply, the “Upland”) surfaces. Long wavelength slopes average 24^{o} in the BRE zone in Virginia, with steeper slopes developed on granitoid rocks in North Carolina (Spotila et al., 2004). In contrast to most high-relief zones in the Appalachians, the steep topography of the BRE does not coincide with the outcrop of a resistant lithology. Bedrock units on either side of the escarpment offer essentially equal resistance to erosion, and contacts between units cut across the BRE at varying angles. The BRE also shows no clear relationship to the nearby Brevard/Bowens Creek fault zone (Fig. 2.3A, 2.3B), which crosses the BRE in western North Carolina (Fig. 2.3) (Dietrich, 1959; Hack, 1973, Virginia Division of Mineral Resources, 2003). Landward of the BRE, the Blue Ridge Upland forms a broad plateau, termed the Floyd Surface by Dietrich (1957), which extends southwest from Roanoke, Virginia and is characterized by rolling hills, thick soils and saprolites, and aggraded valleys (Fig. 2.2). Average slope atop the Upland in Virginia is ~ 10^{o}, but the surface becomes increasingly rugged to the southwest (Spotila et al., 2004). The Upland extends ~ 40 km from the crest of the BRE northwest to the Valley and Ridge fold-thrust belt (Figs. 2.2, 2.3A, 2.3B). Northeast of Roanoke, the Blue Ridge highlands rise above the Piedmont as a narrow ridge isolated from the highlands farther west. This ridge morphology appears to have developed as a result of Atlantic basin rivers (e.g., Roanoke, James, and Potomac) breaching the western margin of the Upland and rapidly eroding the weak sedimentary rocks of the eastern Valley and Ridge (Harbor, 1996). In contrast, the Blue Ridge Upland forms a plateau that remains physically and fluvially connected to the westward-draining areas of the Valley and Ridge and Appalachian Plateau to the north and west. Several large westward-flowing streams with headwaters on the Upland surface ultimately drain portions of all three provinces (Fig. 2.2).

As a shoulder-type escarpment, the majority of the crest of the BRE coincides with the Eastern Continental Divide (ECD), an asymmetric regional drainage divide between Gulf of Mexico (west-flowing) and Atlantic (east-flowing) streams (Hayes and Campbell, 1894; Davis, 1903; Wright, 1927; Dietrich, 1957; Hack, 1973) (Figs 2.2 , 2.3). The low gradient of the westward-flowing streams draining the Upland contrasts strongly with the steep headwaters of streams draining the narrow BRE zone (Fig. 2.4). The steep escarpment slopes provide energy to Atlantic basin streams plunging toward the Piedmont whose erosive power is evident in cascading bedrock reaches and oversteepened fluvial profiles. Hack (1973) viewed the atypical steepness of the escarpment slope streams compared to Upland and nearby Piedmont streams draining the same lithology as evidence of fluvial disequilibrium along the BRE. These streams flatten rapidly upon reaching the low-relief Piedmont where generally symmetrical drainage divides and lower stream gradients suggest a more equilibrated landscape. Some Atlantic basin streams draining the escarpment appear to have captured Upland streams, such as the Dan River of Virginia (stream 19, Fig. 2.3A), whose headwaters meander across the Upland before turning 90^{o} and dropping over 600 m to the Piedmont (Dietrich, 1957; Hack, 1973). Apparent redistribution of fish species of the New River basin into streams comprising the headwaters of the Atlantic-draining Roanoke River basin (streams 19, i, and ii of Fig. 2..3A) provides further qualitative evidence of the capture of westward-flowing drainages (for a review, see Jenkins et al., 1971). These topographic and biological data provide anecdotal evidence of stream capture and episodic divide and BRE retreat.

The BRE has long been viewed as a feature shaped by erosional retreat related to the asymmetry of the divide (Davis, 1903; Wright, 1927; White, 1950; Dietrich, 1957, 1959; Harbor, 1996) but has seldom been regarded as a great escarpment produced by rift-flank uplift (e.g., Ollier, 1984; Tucker and Slingerland, 1994). Pazzaglia and Gardner (2000) suggested that the BRE is a great escarpment that was excavated from a fixed inland divide after Mesozoic rifting and base level drop. Other hypotheses for its origin have favored active tectonic or geodynamic origins. White (1950) proposed that normal-sense reactivation of the nearby Brevard/Bowens Creek fault zone produced the escarpment. Numerous workers have cited isostatic rebound related to thickened Appalachian crust or flexural response to offshore sediment loading as drivers of BRE formation and evolution (Wright, 1927; Pratt et al., 1988; Battiau-Queney, 1989; Hubbard et al., 1991; Pazzaglia and Brandon, 1996; Pazzaglia and Gardner, 2000). Spotila et al. (2004) presented thermochronological constraints on BRE evolution. Cooling histories of the Upland and Piedmont are distinct, suggesting that they do not represent a single offset "surface." These data also showed no difference in cooling ages across the Brevard/Bowens Creek fault zone, suggesting that reactivation of the structure did not form the BRE. The greater exhumation that has occurred on the Piedmont during the past 100 Ma is consistent with the erosional removal of a topographic bulge seaward of the present escarpment and, hence, long-lived retreat. While these data argue against local faulting or uplift as the origin of the BRE, they do not distinguish between escarpment formation through differential erosion by parallel retreat or downwearing from a preexisting or fixed inland divide. Thermochronology also fails to indicate distance and timing of retreat or association with an initiating structure in the present Piedmont. Verifying the mechanism of erosional retreat and constraining the kinematics of retreat are essential to testing the varied hypotheses of BRE origin.

Sparse field evidence of parallel retreat, in the form of well-rounded alluvium preserved near the BRE and divide and the character of beheaded streams, has previously been reported (Wright, 1927; Dietrich, 1957, 1959; Bank, 2001). Rounded cobbles and boulders scattered in dry valleys or on hilltops near the divide are believed to have been stranded by a reduction in drainage basin area and competence of westward-flowing Upland streams through landward escarpment and divide migration (Dietrich, 1957, 1959). This proposed origin of the alluvium is consistent with qualitative observations suggesting drainage basin reduction in westward-flowing Upland streams that now originate at the ECD. Many Upland streams begin to meander very close to their headwaters at the escarpment lip, while Piedmont streams of similar size show no meanders (Spotila et al., 2004). Wide, flat valleys and swamps are common at headwaters in close proximity to the divide, atypical of the headwaters of other streams atop the Upland surface. Quantitative study of fluvial profiles and slope-drainage area relationships of these streams may aid in verifying basin reduction as a product of divide and escarpment retreat. Establishing the provenance of escarpment-crest alluvium should further enhance understanding of retreat kinematics. The use of field-based proxies for escarpment evolution has proved somewhat useful in other locations, such as localized deposits preserved seaward of the Western Ghats escarpment (Widdowson and Gunnell, 1999; Harbor and Gunnell, 2007), weathering surfaces and duricrusts of the South African upland (Partridge and Maud, 1987), and the offshore stratigraphic record of erosion along the BRE (Poag and Sevon, 1989; Naeser et al., 2006) and southwestern Africa (Rust and Summerfield, 1990). Alluvium preserved at the escarpment crest and evidence of drainage basin reduction may thus directly reflect the history of divide retreat and provide concrete physical constraints on the evolution of the BRE.

METHODS

To obtain empirical constraints on the kinematics of BRE retreat, we have investigated the morphology of beheaded streams on the Upland and characterized associated relict alluvium preserved atop the divide. Both of these sources of data should reflect any loss of basin area from parallel retreat of the divide and escarpment. Progressive encroachment of the divide into established Upland basins may have removed steep headwater reaches of fluvial profiles, leaving behind the lower gradient of downstream morphologies. By comparing these potentially beheaded drainage profiles to unaltered streams atop the Upland, estimating the magnitude of basin area or stream length lost may be possible. Alluvium of obvious fluvial origin preserved at the headwaters of beheaded streams may exceed the competence of the host stream given present relief and discharge and be excessively rounded for present channel length. Qualitative comparison of clast roundness to an established relationship to transport distance may provide an estimate of the magnitude of divide and escarpment migration. Together, these data may confirm the mechanism of BRE retreat and permit a generalized restoration of the BRE through paleobasin reconstruction.

Graphical and quantitative methods can be applied to the description of stream channel and basin geometry. A smooth, concave-up longitudinal profile is regarded as an indicator of steady-state equilibrium in a fluvial system (Hack, 1957, 1973; Snyder et al., 2000; Roe et al., 2002; Whipple, 2004; Bowman et al., 2007; Goldrick and Bishop, 2007; LaRue, 2008), with disruptions in smooth concave-up shape indicating disequilibrium, such as that produced by faulting or basin capture. The loss of headwaters to divide migration could be manifest as linear profiles whose steepened headwaters, and thus concave-up shape, have been lost to headward erosion of Atlantic drainages. Longitudinal profiles provide a qualitative basis of comparison, whereas the slope-drainage area relationships of streams can be quantitatively compared. Local channel slope and drainage area are related by the power law

$$S = kA^{-\theta} \tag{1}$$

where S is local slope, k is the steepness index, A is drainage area upstream of the point of the slope measurement, and θ is the concavity index (Flint, 1974). Values of θ should be generally consistent within tectonically and lithologically similar regions (Kirby and Whipple, 2001; Whipple, 2004). Beheaded streams lacking a concave-up profile shape should present less negative θ values and lower slope at a given drainage area (or stream length) than unaltered Upland streams of comparable size.

We constructed longitudinal profiles and slope-area relationships based on 1:24,000-scale USGS topographic maps for 20 westward-flowing streams with headwaters at the ECD (the ECD-draining streams) and 18 westward-flowing control streams from the Upland interior (the Upland controlstreams). ECD-draining profiles were first compared to Upland control profiles to identify characteristics consistent with divide migration and basin loss, such as low-gradient headwaters and overall linear (not concave-up) shape. Average Upland control and ECD-draining profiles were produced by averaging slopes from all streams in each population in 1000-m increments. The qualitative data gathered from profile analysis was then used to guide our selection of streams for slope-area analysis. When the effects of divide and escarpment retreat on the drainage pattern of the study area are considered, it is apparent that not every stream rising at the divide and flowing west would have been beheaded. During landward migration, the major regional divide would occasionally “overtake” pre-existing subordinate divides once located in the Upland interior. Headwaters of streams rising at these subordinate divides would then rise at the ECD and escarpment crest, but would not have been affected by beheading. A random selection of ECD-draining streams would mix these equilibrium streams with beheaded streams, potentially obscuring evidence of divide and escarpment retreat. Accordingly, we applied an intentional bias to focus attention on ECD-draining streams that have most likely been beheaded. We chose the five most linear and apparently anomalous streams draining the ECD westward to estimate stream length lost through comparison to streams draining the same surface but rising landward of the ECD. Ten Upland control streams whose profiles showed smooth, well-developed concave-up shape were selected to quantify an Upland control stream slope-area relationship describing streams unaffected by divide retreat. This slope-area relationship was combined with the average length-area power law relationship from the same ten streams to obtain an expression for length (*L*) as a function of slope (*S*):

$$L = (S/0.0335)^{-1.68} \qquad (2)$$

This expression describes the relationship between increasing channel length and decreasing local slope of Upland control streams unaffected by divide retreat. Slopes at or near the headwaters of many ECD-draining streams are anomalously low and more consistent with slopes observed well downstream of typical Upland stream headwaters. We substituted the unusually low initial slope values (defined by the length of the first contour interval in the stream) of the five most linear streams draining the asymmetric divide for *S* in Eq. (2) to estimate channel length lost to divide and BRE retreat.

To constrain retreat using relict alluvium preserved at the escarpment crest, we first completed a systematic search for terrace deposits at the headwaters of ECD-draining streams that appear truncated by the divide. We focused on streams originating in broad gaps or low relief areas at the escarpment crest. The gentle gradient of these headwaters areas, characteristic of downstream reaches of Upland control streams landward of the escarpment and divide, presented qualitative evidence of basin loss and offered the greatest potential to preserve relict alluvium at the surface. Field reconnaissance was conducted at the headwaters of all ECD-draining streams whose morphologies indicated beheading to locate physical evidence of basin loss from divide retreat, such as mature alluvium preserved in wind gaps very near or atop the ECD. Four of the headwaters terraces that hosted the most abundant and best-preserved rounded clasts were selected for transport distance estimates based on clast roundness.

Clast roundness is known to increase with progressive fluvial transport such that the shape of clasts in alluvium preserved at the divide may be inverted to obtain the magnitude of channel length lost to BRE and divide retreat (Mills, 1979; Lindsey et al., 2007). Sadler and Reeder (1983) offered an empirical and field-expedient method for relating the roundness of quartzite clasts to transport distance. Their regression was produced in the San Bernardino Mountains, California, using clasts of clearly discernible provenance, permitting a direct "outcrop-to-basin" determination of transport distance that could be related to roundness. We chose the method and regression of Sadler and Reeder (1983) because it was developed for a similar lithology and was produced empirically using clasts of known origin and transport distance. Their results showed good agreement with a quartzite clast flume experiment of Kuenen (1956) as well as other empirical studies (Tricart and Schaeffer, 1950; Hovermann and Poser, 1951; Hollerman, 1971; Goede, 1975). Clast shape is expressed quantitatively by the Cailleux Roundness Index, or CRI (Cailleux, 1947), defined as

$$\mathrm{CRI} = \{(2^*R_c)/L\}^*1000 \quad (3)$$

where R_c is the radius of curvature of the sharpest corner and L is the length of the long axis. To avoid low estimates for thinly bedded quartzites or tabular veins, R_c is measured in the orientation of maximum projection (parallel to short axis, orthogonal to long and intermediate axes). After Sadler and Reeder (1983), we only measured clasts with long axes ranging from 4 to 10 cm to avoid transport over-estimates related to the greater ease of rounding of very large clasts. We measured CRI for ~ 30-60 of the most rounded unbroken clasts present in each terrace to obtain a maximum estimate of transport distance (i.e., using an intentional bias).

RESULTS

Longitudinal profiles indicate morphological contrasts between ECD-draining and Upland control streams. A number of ECD-draining profiles are nearly linear and lack the steep headwater reaches that produce the concave-up shape of Upland control profiles (Figs. 2.5A, 2.5B). The more pronounced concave-up shape of the average Upland control profile implies fluvial equilibrium with lithology and regional base level, whereas the comparatively linear shape of the average ECD-draining profile suggests the loss of headwaters to fluvial beheading. This trend in profile shape is apparent in both short (~ 6 km) and long (~ 25 km) streams (Figs. 2.5A, 2.5B).

Knickpoints, or local convexities, occur along profiles of both populations (e.g., streams 10 and 31 of the control group; 5 and 34 of ECD-draining streams), but do not alter the overall shape of profiles (Figs. 2.5A, 2.5B). These may represent active rejuvenation of a relict westward-flowing drainage network by intermittent lowering of the landward base level. Slight contrasts in lithologic resistance may also produce the isolated convexities, but the effects are localized and are not mapped (Virginia Division of Mineral Resources, 2003). The unusually steep profile of stream 11 likely results from the resistant Precambrian-Cambrian quartzites underlying its basin; it was excluded from average Upland profile calculation (Fig. 2.5A). Several ECD-draining streams show concave-up profiles or exhibit slightly steepened headwater reaches (3, 6, 23, 33) (Fig. 2.5A). These profiles may represent preexisting equilibrium streams of the Upland interior that were "overtaken" in place by landward divide migration and thus have not lost drainage area. Alternatively, this morphology may be the result of an earlier pulse of New River incision which migrated headwardly through the drainage network and rejuvenated the basins to their headwaters at the ECD. Aside from these variations, however, a clear distinction between the two stream populations is evident in the individual and averaged profile shapes (Fig. 2.5A).

A comparison of the steepness and concavity indices of streams draining the asymmetric divide to the Upland control streams further delineates morphological distinctions between the two groups and provides an estimate of channel length lost to divide retreat. We focused our analysis on five streams rising at the divide that presented particularly linear profiles and low headwater slopes: 4, 22, 35, 36, and 38 (Figs. 2.3A, 2.3B, 2.5A, 2.5B). ECD-draining streams show a more gradual downstream decay of channel gradient, resulting in less negative concavity index (θ) values. Average concavity index (θ) for the ECD-draining streams is -0.219, compared to -0.401 for the ten Upland control streams analyzed. Morphological distinctions between the two stream populations are also apparent in steepness index (k; Eq. 2) values; average steepness index (k) for ECD-draining streams is 0.157, while the steeper headwater reaches of the Upland control streams yield an average (k) value of 0.401. These slope-area data reflect the trend observed in the longitudinal profiles and are consistent with basin loss resulting from divide and escarpment retreat. Substituting the low headwater slopes of the five ECD-draining streams into Eq. (2) provides a loose estimate of stream length lost to divide retreat. Estimated channel loss results are presented in Fig. 2.6. Average estimated channel loss from BRE and divide retreat is

~ 8 km (Fig. 2.6C). Chestnut Creek (35) and Flat Creek (38) suggest the largest channel losses (18.0 and 11.1 km, respectively). While loose estimates, these analyses of channel concavity appear to reflect significant channel loss and are thus consistent with divide retreat and sufficient stability of the Upland to preserve the relict basin morphology.

Combined analysis of longitudinal profiles, slope-area relationships, and 1:24,000-scale topography led to the identification of 14 individual fluvial terrace deposits in wind gaps and low relief areas atop or near the ECD (Fig. 2.3A, 2.3B). These terraces are physical evidence of both parallel retreat of the assymetric divide and stability of the Upland surface. The terraces are characterized by accumulations of rounded to well-rounded vein quartz or quartzite clasts up to ~ 25 cm long (Fig. 2.7). No polymineralic (lithic) clasts were observed in any of the terrace deposits. Fine matrix material typically is sandy clay; but some deposits, particularly site N (Figs. 2.3B, 2.7), contain a matrix dominated by quartz sand and refractory minerals. Clay content and redness appear to be proportional to the extent of clast weathering. The effects of floroturbation (i.e., tree throw) and agricultural disturbance are apparent (Fig. 2.7). Terrace I (Fig. 2.3A) may contain primary depositional features indicated by roughly parallel lines of cobbles ~ 1 m below the surface in a roadcut (Fig. 2.7), but the extensive weathering of the deposit complicates this interpretation. Clasts are very abundant in the deposits, a number of which approach a clast-supported structure. All clasts show signs of weathering, but the extent of weathering is variable between deposits located in close proximity to one another. For example, site E contains numerous clasts showing little evidence of chemical weathering; but site I, located ~ 15 km away (Fig. 2.3A), contains only extensively weathered and pitted cobbles, most of which are broken. The absence of a trend between clast weathering or soil development and terrace location suggests ongoing, but episodic, asymmetric divide retreat.

The locations of the terrace deposits are related to the local topography. Deposits concentrate in wind gaps and low-relief areas within broad, region-scale sags along the divide (Fig. 2.8). In the Virginia study area (Fig. 2.3A), these sags cluster at 825 m (~ 2700 ft) and 760 m (~ 2500 ft) elevation. Many terraces occur just at or above the headwaters of streams whose profiles show little or no concave-up shape, low headwater gradient, and a lack of rejuvenation from landward base level drop (Fig. 2.5A, 2.5B). Preserved unconsolidated alluvium is not, however, found at the headwaters of all such streams or in all sags along the asymmetric divide. Several preserved terraces are located seaward of the divide (e.g., D, J, L, M; Fig. 2.3A). These terraces share the same elevations and topographic features as other sites on the Upland margin and seem to be perched on topographic remnants of a once-continuous Upland surface. For example, terrace D appears to align with stream 24 and is preserved at the same elevation as terrace E, possibly reflecting divide retreat through two distinct episodes of stream capture (the headwaters of stream 30 were captured to form stream 19, followed by additional capture of stream 30 headwaters by stream ii; Fig. 2.3A). The preservation of unconsolidated alluvium atop these small Upland remnants separated by deep, narrow gorges suggests that sudden, rapid base level drop resulting from stream capture can produce strongly differential erosion in mature landscapes. The apparent stability of these Upland remnants adjacent to active gorge development invites comparison to the contrast in Upland and stream valley erosion rates at Dolly Sods, West Virginia, described by Hancock and Kirwan (2007).

Roundness of the clasts preserved in remnant terraces at the crest of the BRE imply significant transport and hence considerable loss of stream length by divide and BRE retreat. We measured clast roundness for ~ 30-60 clasts from four terraces (N, E, B, A) (Fig. 2.9). Roundness was converted to an estimate of transport distance using the relationship of Sadler and Reeder (1983). Terraces N and E show the highest roundness and estimated transport distances for individual clasts, with a maximum value of ~ 270 km (clast shown in Fig. 7F). Average transport distance and filtered average (excluding highest and lowest 10% of values) are lower, but all suggest stream length loss of > 10 km (Fig. 2.9). Given that not all clasts within a terrace would have been transported the same distance, the average roundness values of the top 10%, ranging between 47-185 km, may be more useful metrics. Some of the variation between terraces may relate to "paleoorder" or paleoflow direction of the streams that deposited the terraces. The two terraces with lower average transport values (A, B) occur in tributaries of the same master stream, the Little River (Figs. 2.3A, 2.9). In contrast, the more rounded clasts of terraces E and N occur in valleys that are directly contiguous with escarpment-orthogonal trunk streams.

DISCUSSION

Anomalous drainage basin morphologies and fluvial terraces provide strong qualitative evidence for ongoing divide and escarpment retreat resulting from the repeated capture and dissection of Upland drainage basins. These data suggest a model of escarpment evolution that differs from the recent preexisting, or fixed, inland divide plateau degradation model derived from cosmogenic, thermochronologic, and numerical modeling studies of other escarpments (Moore et al., 1986; Gilchrist et al., 1994; Bishop and Goldrick, 2000; Cockburn et al., 2000; Matmon et al., 2002; Persano et al., 2002; van der Beek et al., 2002; Braun and van der Beek, 2004). The morphology of Upland drainages rising at the divide is consistent with a state of transient adjustment between the Upland and BRE (Dietrich, 1957; Hack, 1973; Gasparini et al., 2007). Examples of topographic disequilibrium observed cannot be rectified with a near-stationary BRE that stopped retreating significantly long ago (Sullivan et al., 2007). We instead propose a BRE history similar to the structurally and fluvially controlled retreat model suggested for the Western Ghats escarpment by Harbor and Gunnell (2007) and Gunnell and Harbor (2008). We view BRE evolution as an episodic, yet ongoing, process where significant long-term parallel retreat is accomplished through the repeated capture and rapid dissection of Upland drainage basins.

While the overall distance of escarpment migration remains uncertain, altered basin morphologies and clast roundness allow for loose estimates of recent (10^6 y) retreat. The vein quartz clasts measured here certainly equal, and likely exceed, the resistance to rounding of the quartzite clasts of Sadler and Reeder (1983), suggesting that our transport estimates (i.e. ~ 50-100 km based on maximum roundness) are probably conservative. Structural control of drainage immediately southeast of the BRE by northwest-southeast trending fractures or joints suggests that tens of kilometers of transport could have been orthogonal to the escarpment, suggesting transport distance could provide a reasonable basis for a loose retreat estimate. Headwater gradient of channels truncated by the divide suggests ~ 10-20 km of channel loss, reasonably consistent with averaged clast transport estimates. Flow direction and meander interval are, however, significant unconstrained variables; and the escarpment-parallel flow and high degree of sinuosity present in streams of the study areas would contribute to an overestimate of retreat. Clast transport distance will thus remain a rough estimate of retreat magnitude until sediment provenance is clearly established.

The systematic preservation of terrace deposits at low gradient headwaters supports the conclusion that apparent channel loss is the result of divide and escarpment retreat. The persistence of surficial alluvial deposits and anomalously low headwater slopes at the escarpment crest suggests Upland denudation is outpaced by landward retreat of the divide and escarpment. Although the terraces are not dated, clast weathering is similar to Upland-sourced vein quartz clasts from > 1 Ma New River terraces preserved under the same climate conditions in the nearby Valley and Ridge (Ward et al., 2005). While a potential age of more than 1 Ma for our deposits suggests great stability of the surface on which they are preserved, Upland denudation rates of ~ 10 m/Myr based on thermochronometry (Spotila et al., 2004) and cosmogenic dating (e.g., Sullivan et al., 2007) suggest a limited lifetime of surficial deposits atop the Upland. If the terraces are several million years old and were stranded by the capture of escarpment-orthogonal drainages (terraces D and E; Fig. 2.3A) tens of kilometers in length, local capture-driven retreat rates of 1-10 km/Ma are plausible. This implies that local divide and BRE retreat rates may, in the case of major capture events, outpace Upland lowering by as much as three orders of magnitude. These speculative retreat rates would, however, only result from the occasional capture of large Upland drainage basins with discharges capable of producing very rapid erosion upon capture. Indeed, smaller but more frequent captures producing comparatively modest local retreat rates probably account for considerable retreat over the long term. In either case, BRE and divide retreat rates exceeding Upland lowering by an order of magnitude or more would be limited to captured basins undergoing dissection and not be applicable to the entire BRE at any given time.

When considered in the context of the present Upland drainage network, the respective location, roundness, and freshness of terrace alluvium may offer some insight into the paleodrainage network and its role in escarpment retreat. Terraces A and B have lower average roundness values and are drained by the BRE-parallel local trunk stream, the Little River (stream 1; Fig. 2.3A), and may thus be the remnants of short, lower order tributary streams only truncated by the last few kilometers of asymmetric divide retreat. Terrace E is drained by a stream directionally and topographically continuous with the local trunk stream, Big Reed Island Creek (stream 30; Fig. 2.3A), which flows roughly orthogonal to the BRE. The greater roundness of terrace E clasts may reflect longer transport in the larger basin of a paleo-trunk stream. Clasts from terrace N, which is also drained by a stream showing general continuity with the French Broad River (Fig. 2.3B), provided the highest overall transport estimates from all terraces studied. Terraces N and E, which host the most rounded alluvium, also showed the least amount of weathering, suggesting relatively recent stranding and thus rapid erosional destruction of their depositing basin upon its capture. This comparatively fresh alluvium may be viewed as additional qualitative evidence of the rapid dissection and local divide and escarpment retreat associated with large capture events.

We propose a conceptual model of BRE evolution that highlights the role of stream capture in producing escarpment retreat through prolonged, but punctuated, retreat of the ECD (Fig. 2.10). Headwardly eroding Atlantic basin streams of the steep divide flank (Fig. 2.10, panel 1) can tap the potential energy of westward-flowing Upland drainages through stream capture (panel 2). The capture process may initiate with diversion of Upland groundwater before the westward-flowing surface channel is physically diverted to the Atlantic basin. Connection to the seaward base level, which is hundreds of meters lower, greatly steepens and energizes the captured stream, and rapid incision propagates headwardly through the basin as it equilibrates to the new base level (panel 2). Structural weaknesses (faults and fractures/joints) controlling flow directions will facilitate rapid dissection of the captured basin and allow headward erosion to encroach upon neighboring Upland basins, eventually producing additional captures along strike (panel 3). Adjustment of captured streams to the seaward base level rapidly dissects the detached Upland remnants and moves the locus of topography landward in response to the new location of the divide. Continued headward erosion along structural weaknesses sets up future landward captures and re-starts the retreat process (panel 4). In order for this retreat model to continue in the long term, capture and retreat must occur more rapidly than lowering of the Upland (landward) base level to maintain divide asymmetry and the energetic potential of Upland streams relative to the seaward base level. Minor (tens of meters) episodic incision of Upland streams due to landward base level drop could potentially propagate headwardly through the entire drainage network and ultimately steepen the headwaters of westward-flowing Upland streams, increasing symmetry of the divide. A topographic ridge may form along the ECD, acting as a physical impediment to future capture events. Over the long term, the combined effects of repeated Upland base level drops will reduce contrast in landward and seaward base levels and thus reduce the potential energy of Upland streams available for capture (bottom of panel 4). Divide asymmetry and retreat thus form a positive feedback loop; asymmetry facilitates rapid retreat, which in turn maintains the divide morphology and the energetic potential for continued capture and retreat. A sufficiently stable landward Upland base level is thus equally vital to preserving retreat potential in the long term and allowing the process to continue after lengthy intervals between localized captures.

While we view BRE evolution as a long-lived and ongoing process, the local, basin-scale retreat rates of ~ 1-10 km/Ma proposed are certainly not applicable to the whole feature for the entirety of its inferred ~ 200 Ma lifetime. Cosmogenic dating indicates the BRE is no longer experiencing significant retreat along much of its length (e.g., Sullivan et al., 2007). This suggests rate-controlling events must be spatially isolated but occur in enough locations with sufficient regularity to produce an overall parallel retreat over the lifetime of the BRE. The long-term BRE retreat rate may therefore best be viewed as a "weighted average" of local and short-lived rapid retreat events interrupting long periods of quiescence where escarpment evolution is negligible. When these rare, localized but extremely rapid retreat rates are considered over the length of the feature, possibly for ~ 200 Ma, a more reasonable "lifetime" parallel retreat rate is suggested.

Our proposed retreat model is heavily dependent on a preexisting rectilinear drainage network on the Upland surface. Lineament analysis of areas of the escarpment in which alluvium is fresh and divide asymmetry most pronounced indicate the role of orthogonal drainage-controlling weaknesses in sustaining the retreat process (Figs. 2.10, 2.11). Streams in the Appalachian Blue Ridge and Piedmont generally trend along two orientations: orogenic (and BRE) strike-parallel, controlled by weak lithologies or structures (e.g., Brevard fault zone), and orogenic strike-orthogonal, controlled by fractures and joints (Fig. 2.11). The enhanced erodibility of these drainage-controlling features is fundamental to BRE evolution. Orthogonal flow directions allow for more frequent captures at high order points along Upland streams that facilitate a large increase in stream power and rapid dissection of the captured basin as illustrated by our conceptual model (Fig. 2.10). Where flow directions are essentially opposite across the divide, headward erosion captures insignificant Upland area and does not energize the capturing drainage, producing minimal slow retreat. Terrace I (Fig. 2.3A), preserved in such an area, contains clasts showing considerable pitting, breakage, and deep staining, all indicative of more extensive weathering than that experienced by the other terraces. In contrast, orthogonal flow directions increase the potential for capture of high-order channels and rapid adjustment of topography to the new divide location. Orthogonal drainages also facilitate capture across low-relief subordinate divides of the Upland (Fig. 2.10, panel 3), allowing basin-by-basin capture to propagate along BRE strike. The impact of orthogonal flow directions and weak drainage-controlling structures on retreat rate is clearly seen where fluvial evidence of retreat is best preserved, such as in terraces E and N (Figs. 2.3A, 2.3B). Easily eroded fractures and lithologies permit rapid basin dissection and speed the landward progression of headward erosion, the source of subsequent captures further inland.

Considering the potential for repeated episodes of local but rapid retreat combined with extremely slow Upland lowering, interpretation of the BRE as the final remnant of a Mesozoic rift-flank feature may be reasonable. The initial BRE topography almost certainly formed well seaward of its present location, and quartz mylonite clasts in terraces B and E (Fig. 2.3A) indicate the Brevard/Bowens Creek fault zone was a drainage controlling feature atop the Upland and did not reactivate to initiate the BRE as suggested by White (1950). This idea is further supported by fluvial evidence of retreat atop the ECD and BRE where it is located well seaward of the Brevard Zone in western North Carolina (Fig. 2.3B). While the clasts we examined represent only the most recent captures and retreat, they still suggest sufficient transport to restore parts of the escarpment near Triassic basin border faults 70 km southeast. Whether the BRE initiated at these structures or further southeast remains unknown. The present Piedmont drainage network suggests that the mechanism indicated by our study could have supported continued BRE retreat from an initial location over 100 km to the southeast. Persistence of divide asymmetry related to Upland stability (Sullivan et al., 2007) combines with active incision along the escarpment front (e.g., streams 19 and ii; Fig. 2.3A) to suggest that the landform will continue to evolve and maintain its topographic youth despite an ~ 200 Ma erosional history.

Field evidence of the role of structurally controlled stream capture and punctuated retreat of the BRE complements studies of other passive margin escarpments. Our results illustrate the value of applying more traditional methods of geologic study to constrain passive margin escarpment history. Alluvium has not yet been used as a proxy for retreat of the Western Ghats, but the combination of appropriate structure, variation in lithology, and apparent capture frequency (Harbor and Gunnell, 2007) suggests the likelihood of terraces preserved at or near its crest. The low relief, low erosion rates, preservation of soils, and rectilinear drainage network of the Sri Lankan Upland (von Blanckenburg et al., 2004; Vanacker et al., 2007), qualitatively similar to features of the Blue Ridge Upland, may have preserved evidence of parallel escarpment retreat driven by stream capture. The episodic retreat model highlights the importance of divide location and preexisting drainage network as controls over escarpment evolution (van der Beek et al., 2002). Our data also indicate the importance of careful selection of cosmogenic erosion rate sample locations. As the high order capture events that ultimately drive retreat are spatially isolated and transient because of the rapid dissection of captured basins, escarpments may appear (and likely are) stable along most of their length at any given time (Fleming et al., 1999; Cockburn et al., 2000; Sullivan et al., 2007; Vanacker et al., 2007). The active process may be difficult to observe unless cosmogenic samples are taken in the immediate vicinity of a large knickpoint formed by a recent capture event. Cosmogenic dating of captured basins undergoing dissection, such as the Dan River Gorge (stream 19 of Fig. 2.3A), will be an important way of testing the episodic retreat model and assessing its potential applicability to sinuous escarpments, such as southeast Australia, where gorge incision is known to outpace upland lowering (Nott et al., 1996).

CONCLUSIONS

Fluvial terraces and beheaded stream valleys preserved atop the ECD at the BRE crest provide clear evidence of significant escarpment retreat associated with landward divide migration during the Cenozoic. Stream morphology and terrace location suggest this ongoing retreat process is driven by the episodic capture of large Upland drainages by steep streams flowing east to the Atlantic Ocean. The structurally controlled rectilinear drainage network and low relief of the Upland facilitate repeated captures, which produce rapid dissection of the captured basin and localized landward migration of the zone of maximum relief. When considered along the entire strike length of the BRE, the combined effect of these local, but rapid, retreat events may lead to parallel divide and escarpment retreat over the long term. While the roundness of terrace clasts suggests many tens of kilometers of transport, uncertain provenance and paleodrainage networks preclude the use of this data in establishing a total retreat distance or point of origin for the BRE. Nevertheless, the size and roundness of alluvium preserved at the BRE crest suggest extensive Upland drainage basins have been destroyed by escarpment and divide retreat.

In contrast to recent numerical modeling and cosmogenic studies, our results indicate that landward migration of the drainage divide produced by stream capture events can produce continued significant retreat of mature passive margin escarpments. The persistence of the retreat process long after the demise of initial rift topography may be related to a combination of extremely slow Upland denudation and comparatively rapid escarpment retreat. The freshness of terrace alluvium and intact preservation of relict stream gradients suggest that local retreat rates resulting from large captures may outpace Upland lowering by three orders of magnitude. This differential erosion maintains maximum divide asymmetry and the potential energy of Upland streams relative to the Atlantic base level, preserving the energetic driver of the retreat process. This model of episodic retreat is consistent with areas of slow retreat along the BRE and other escarpments indicated by cosmogenic dating, but it also predicts that very localized retreat rates in other areas could be at least three orders of magnitude higher. Additional cosmogenic dating along the BRE will be necessary to validate our model, and its relevance to other passive margin escarpments is presently unclear. A single paradigm of passive margin escarpment evolution may not exist, and these features may evolve through a range of mechanisms governed by drainage networks, divide location and asymmetry, structure, and lithology. In either case, the physical evidence of the role of divide migration in BRE retreat provides a useful additional constraint on the evolution of this feature and may suggest a new source of data regarding the retreat mechanism of other passive margin escarpments.

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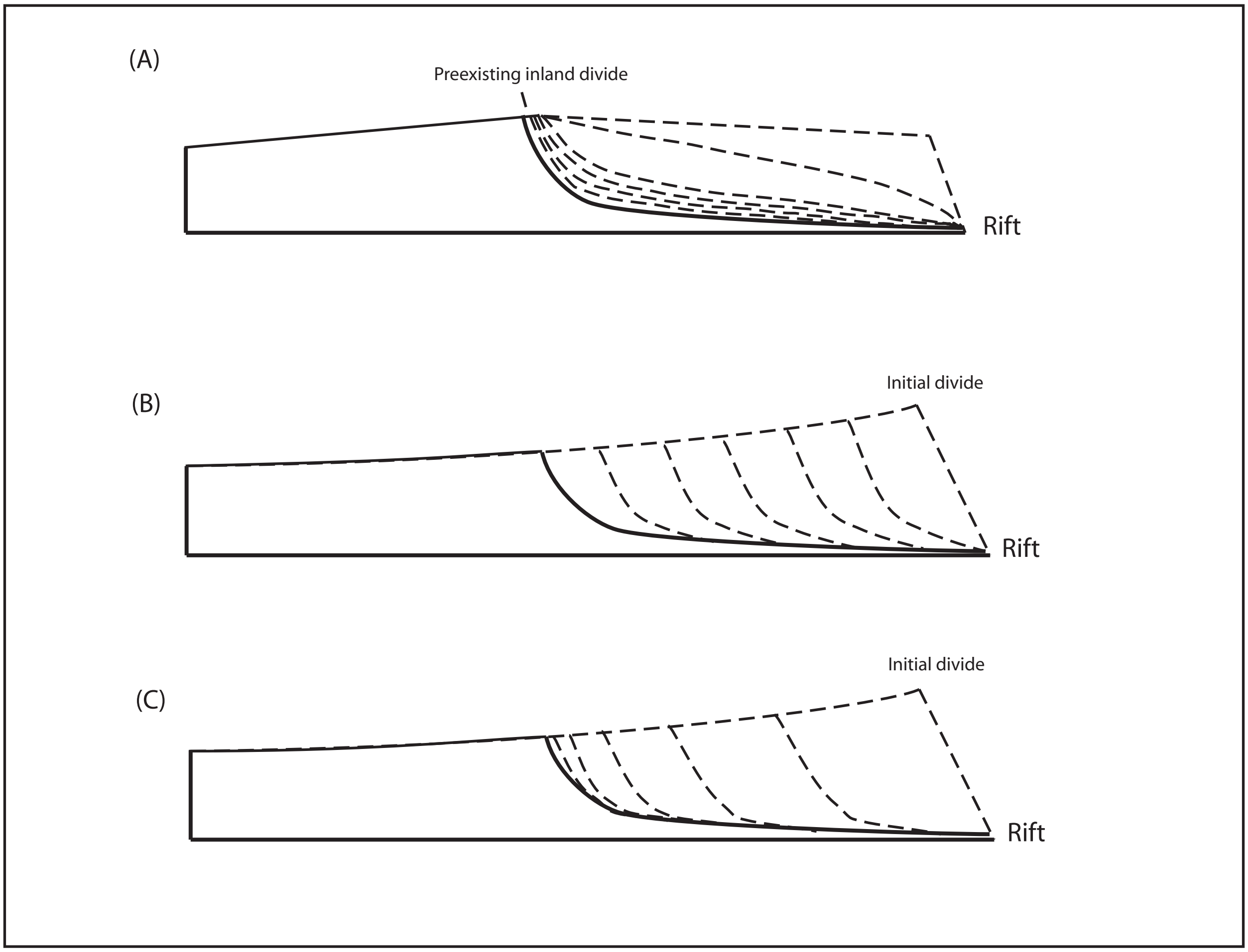
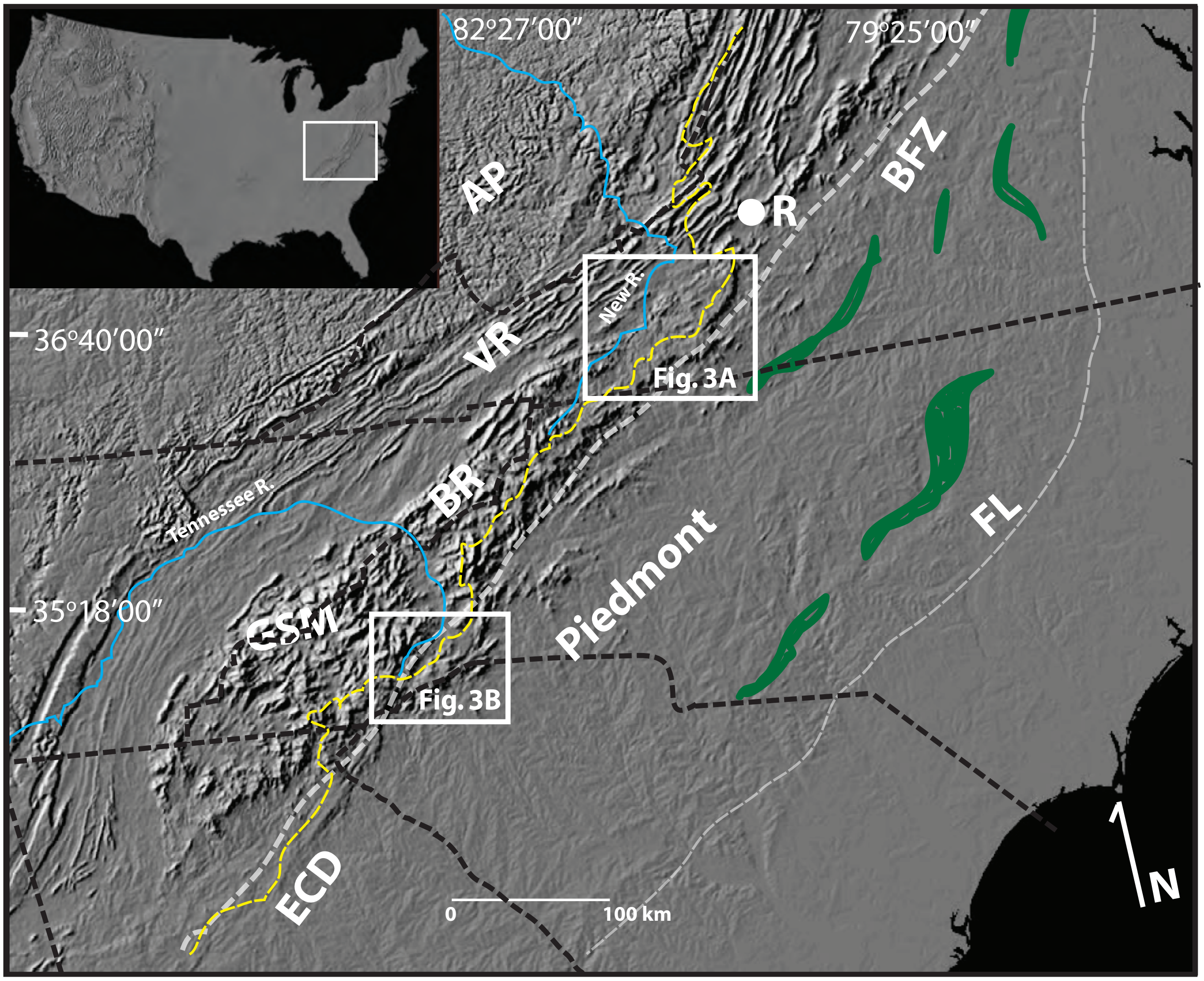
Figure 2.1: Theoretical models of passive margin escarpment retreat.



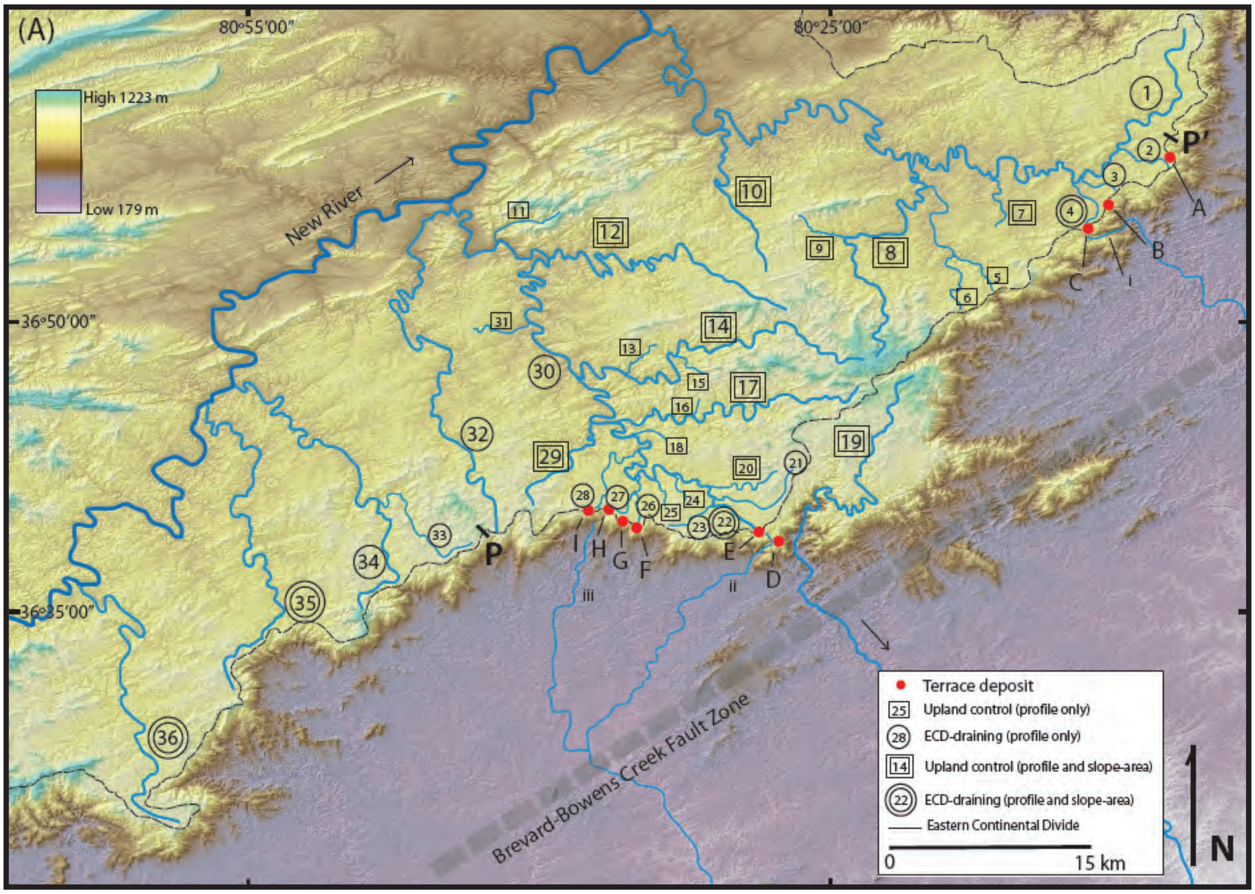
Comparison of three popular models of passive margin escarpment evolution. Dotted lines trace the development of escarpment topography across equal time intervals. All models ultimately produce escarpments of similar morphology despite differing mechanisms and rates of retreat. (A) Rift-related base level drop seaward of a preexisting drainage divide steepens large streams and rapidly excavates an escarpment at the divide. Subsequent retreat is minimal. (B) Rift flank uplift produces an asymmetric divide atop the rift shoulder. Focused erosion on the steep flank of the divide produces steady parallel retreat of divide and escarpment. (C) The parallel retreat mechanism of (B) decelerates because of the decrease in escarpment relief. After van der Beek et al. (2002).

Figure 2.2: Location of the Blue Ridge Escarpment.



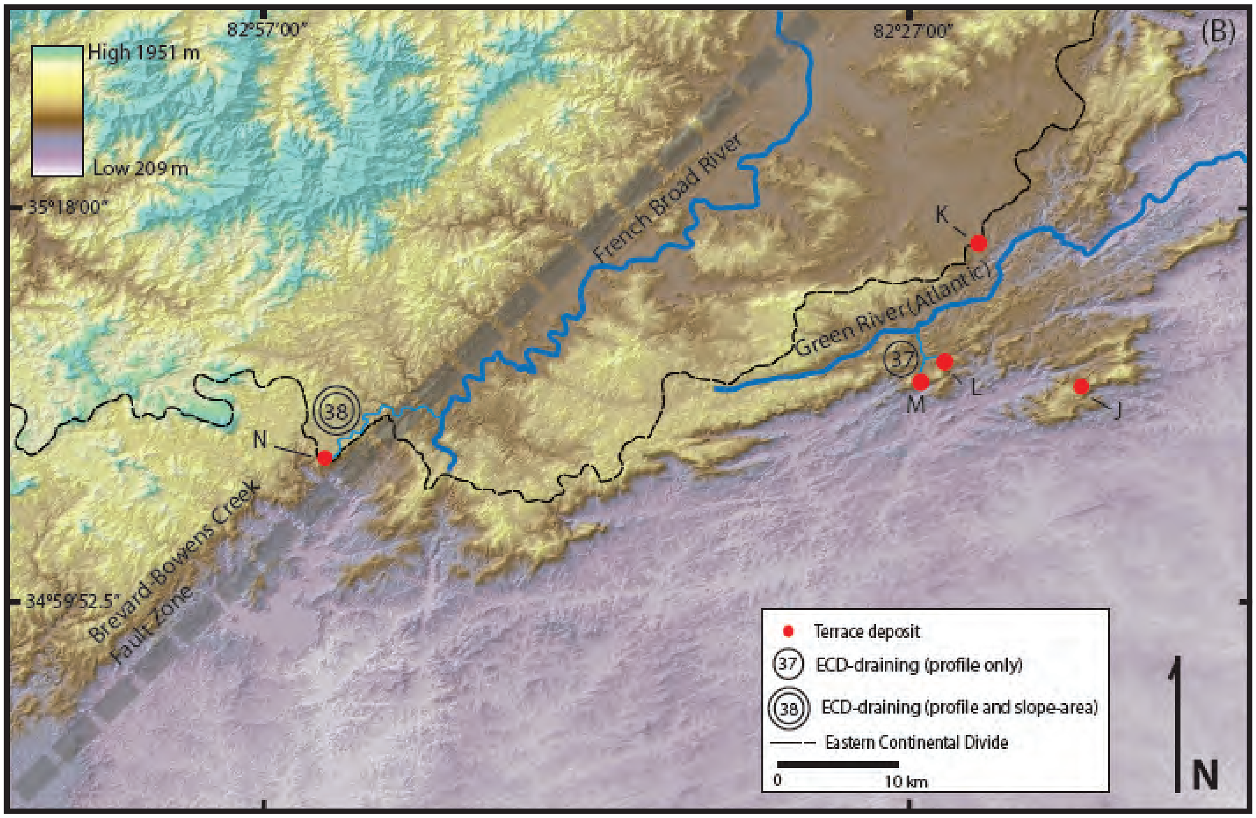
Location of the Blue Ridge Escarpment and Eastern Continental Divide (ECD) in relation to relevant topographic and geologic features of the southern Appalachians. The Escarpment is the area of steep topography separating the Blue Ridge (BR) and Piedmont physiographic provinces. South of Roanoke, Virginia (R), the Escarpment coincides with the ECD (yellow line). Mesozoic basins in the Piedmont are shown in green. Brevard/Bowens Creek fault zone (BFZ) is identified by red line. Fall Line (FL) is denoted by gray line. VR=Valley and Ridge, AP=Appalachian Plateau, GSM=Great Smoky Mountains. Boxes denote locations of Figs. 2.3A and 2.3B.

Figure 2.3A: Map of the Blue Ridge Plateau, southwest Virginia



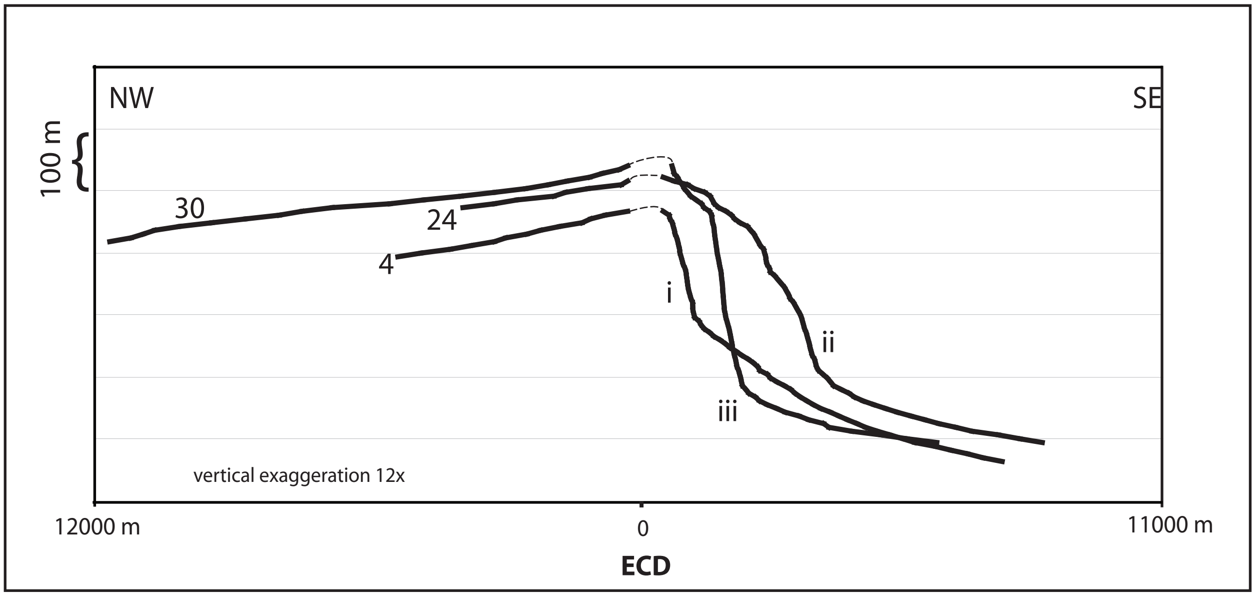
Shaded relief map of Virginia study area (Fig. 2.2) showing streams (numbers) and terrace deposits (letters) used. Larger numbers indicate large streams whose profiles are presented in Fig. 2.5B. P and P' are endpoints along a topographic profile of the ECD (Fig. 2.8). ECD and Brevard fault zone (BFZ) are indicated. Map is based on 10-m resolution DEM. Streams are labeled as follows: 1=Little River, 2=Payne Creek, 3=Silverleaf Branch, 4=Meadow Creek, 5=Thomas Grove Church tributary to Dodd Creek, 6=W. Fork Dodd Creek, 7=Oldfield Creek, 8=W. Fork Little River, 9=Spurlock Creek, 10=Big Indian Creek, 11=Rock Creek, 12=Greasy Creek, 13=Burks Fork tributary, 14=Burks Fork, 15=Adams Branch, 16=Chisholm Creek, 17=Laurel Fork, 18=Bear Creek, 19=Dan River (Atlantic Basin), 20=Stone Mountain Branch, 21=Big Reed Island Creek headwaters, 22=Pine Creek, 23=Puckett Church Branch, 24=Pipestem Branch, 25=Sulphur Springs Branch, 26=Grassy Creek, 27=Grassy Creek tributary, 28=Pine Creek Ward's Gap, 29=Little Snake Creek, 30=Big Reed Island Creek, 31=S. Prong Buckhorn Creek, 32=Little Reed Island Creek, 33=E. Fork Crooked Creek, 34=Crooked Creek, 35=Chestnut Creek, 38=Brush Creek, i=Rennett Bag Creek, ii=Ararat River headwaters, iii=Waterfall Branch.

Figure 2.3B: Map of the Blue Ridge Plateau, western North Carolina.



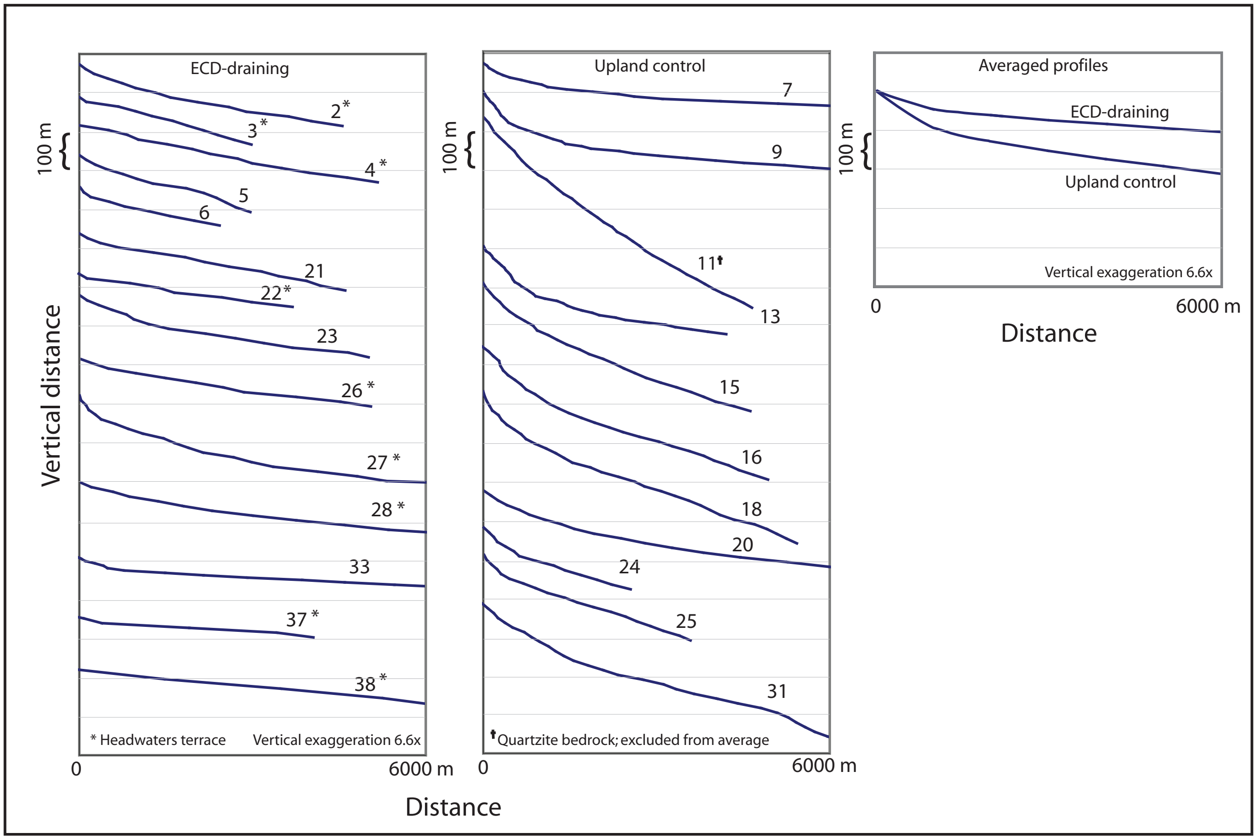
(B) Shaded relief map of North Carolina study area showing streams and terrace deposits used. Map is based on 10-m resolution DEM. 37=Terry Creek, 38=Flat Creek. Terraces J, L, and M and Terry Creek are drained to the Atlantic Ocean by the Green River, a stream analogous to the Dan River (19) of the Virginia study area.

Figure 2.4: Profiles of streams rising on opposite sides of the eastern continental divide.



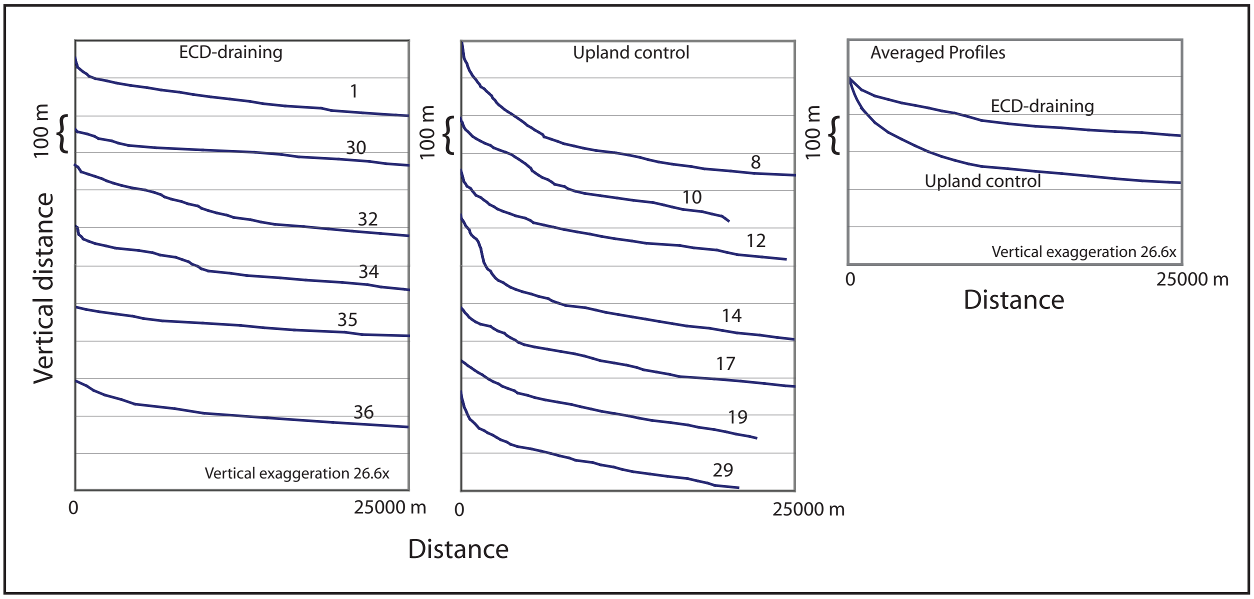
Long profiles of three pairs of Upland and Atlantic-basin streams (Fig. 2.3A) illustrating the asymmetry of the Eastern Continental Divide. The overall convex profile of the Ararat River (ii) indicates a large zone of active incision that may result from a recent capture event that stranded terrace E (Fig. 2.3A).

Figure 2.5A: Longitudinal stream profiles.



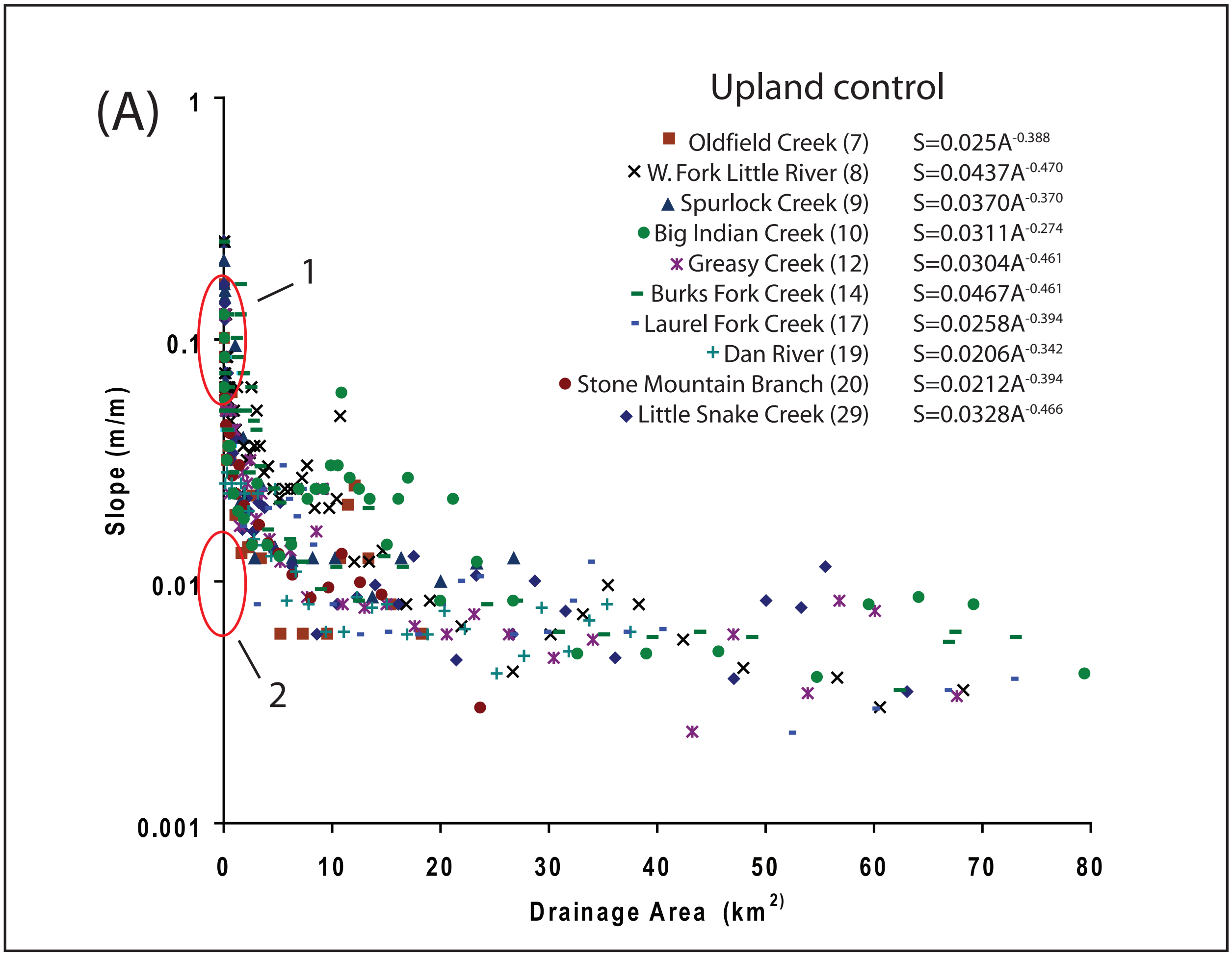
Long profiles of short (< 6 km) ECD-draining streams (left) and Upland control streams (right) (Figs. 2.3A, 2.3B). Average profiles were developed by averaging slope values over 1000 m intervals for every stream within each of the four groups. Locations shown in Figs. 2.3A and 2.3B. Asterisks denote streams with terrace deposits at their headwaters.

Figure 2.5B: Longitudinal stream profiles of major drainages.



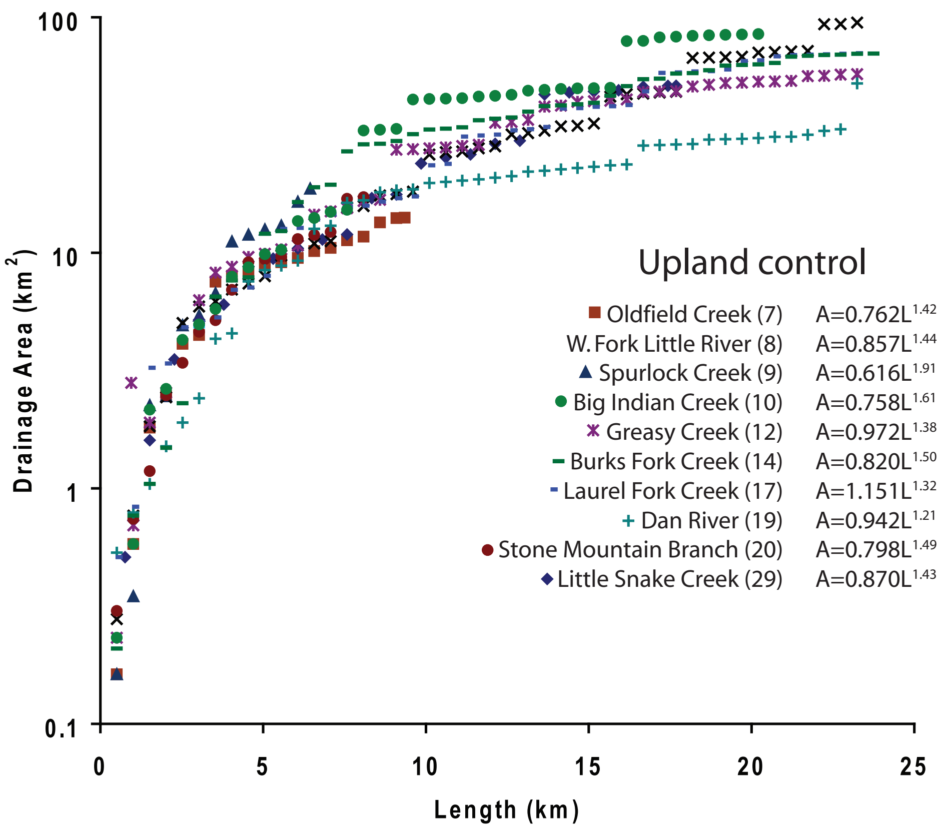
Long profiles of ~ 25-km-long ECD-draining streams (left) and Upland control streams (right) (Figs. 2.3A and 2.3B). Little Snake Creek (29; right) drains a high elevation, high relief area that was likely a local divide within the Upland, and is therefore treated as an Upland control stream despite its present proximity to the ECD. Average profiles constructed in the same manner as Fig. 2.5A.

Figure 2.6A: Slope vs. drainage area relationships of Blue Ridge Upland streams



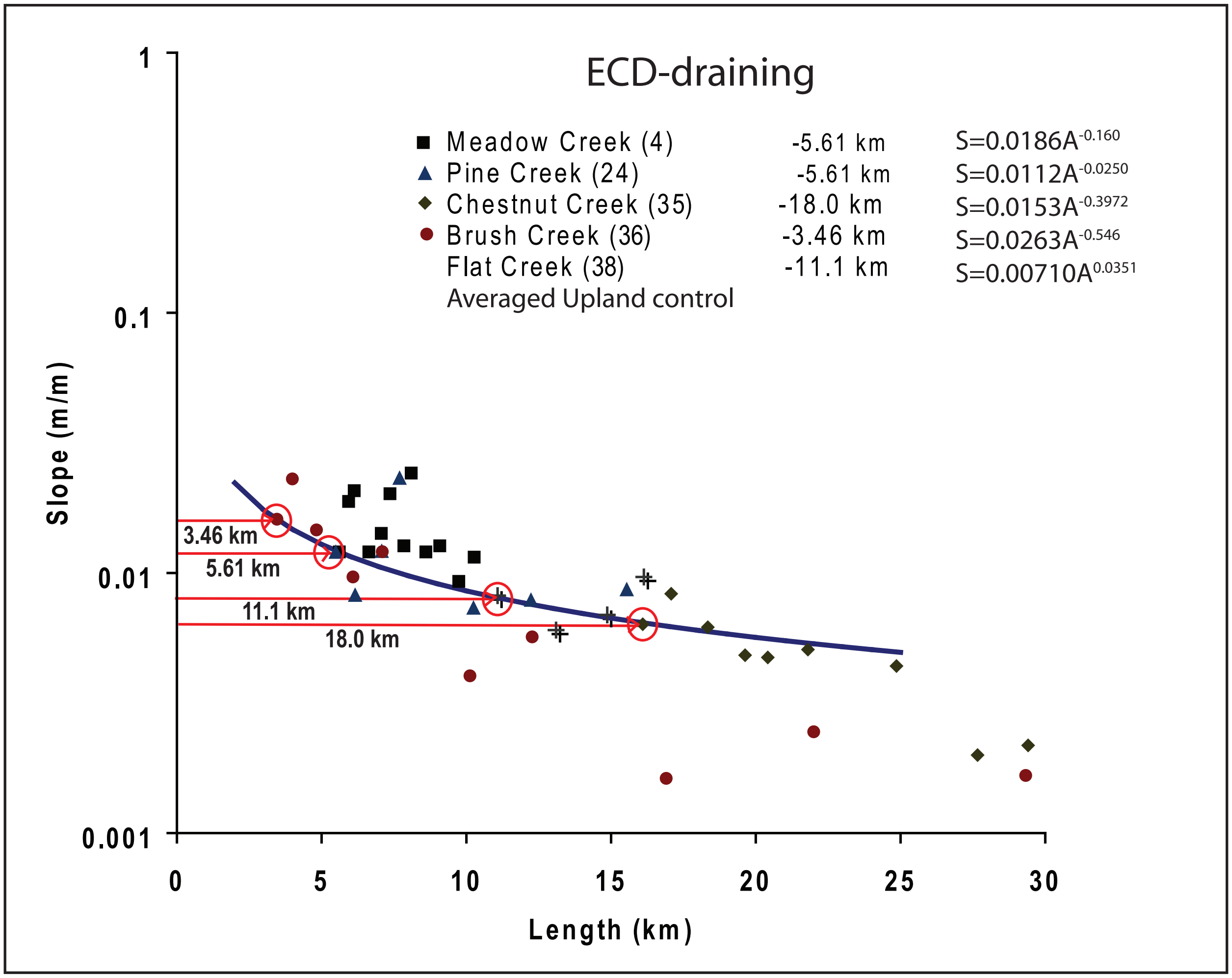
(A) Slope-drainage area relationships of ten Upland control streams showing plotted raw data and best-fit power law expressions. Average Upland control slope-area relationship is described by the power law S = 0.0314A-0.403. Ellipse 1 indicates the range of Upland control stream headwater slopes; ellipse 2 indicates range of headwater slopes of beheaded ECD-draining streams (Fig. 2.6C). (C) Headwater slope values of five beheaded ECD-draining streams shifted to fit the average Upland control slope-length power law curve. ECD-draining headwater slopes are anomalously low and consistent with slopes encountered kilometers downstream from Upland control stream headwaters. Red arrows indicate the magnitude of the required shift and thus the Upland control channel length necessary to obtain the low headwater slopes of the ECD-draining streams.

Figure 2.6B: Drainage area vs. length relationships.



Drainage area-channel length relationships of the ten Upland control streams from Fig. 2.6A. Average relationship is described by the power law A = 0.855L1.47. Average slope-area (Fig. 2.6A) and area-length (Fig. 2.6B) power laws were combined to obtain the average Upland control slope-length power law (L = (S/0.0335)-1.68) plotted in Fig. 2.6C.

Figure 2.6C: Slope vs. length relationships.

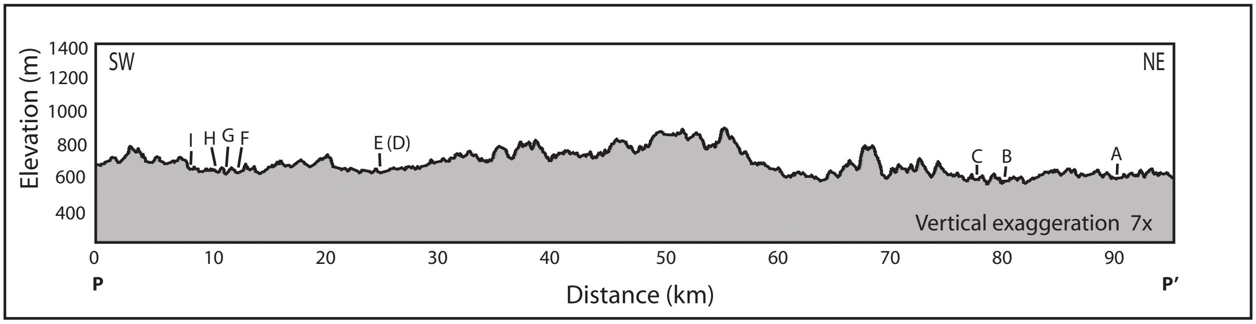


Headwater slope values of five beheaded ECD-draining streams shifted to fit the average Upland control slope-length power law curve. ECD-draining headwater slopes are anomalously low and consistent with slopes encountered kilometers downstream from Upland control stream headwaters. Red arrows indicate the magnitude of the required shift and thus the Upland control channel length necessary to obtain the low headwater slopes of the ECD-draining streams.

Figure 2.7: Selected images of terrace deposits

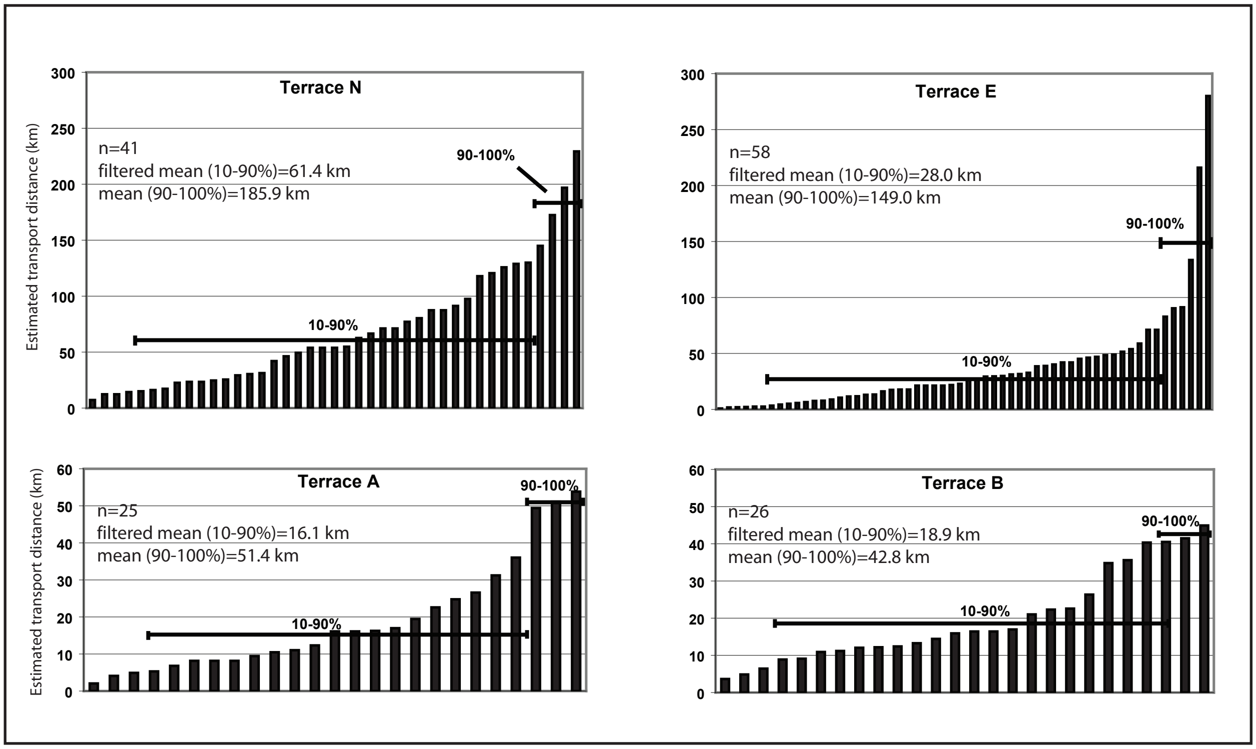
Selected images of terraces showing typical field characteristics. All photos are taken within ~ 100 m of the ECD. Locations of terraces are shown in Figs. 2.3A and 2.3B. Scale card arrow is 10 cm long. (A) Large, well-rounded clasts in sandy clay matrix from terrace E. (B) Unweathered vein quartz clasts in quartz sand matrix from terrace N. (C) Rounded to well-rounded clasts exposed by floroturbation (tree-throw) from terrace B. (D) Small boulder from terrace E. (E) Cobble lines exposed by roadcut from terrace I. (F) Selected clasts from terrace E. Note rounding at all clast sizes.

Figure 2.8: Elevation profile along the eastern continental divide.



Profile along the ECD from P to P' (Fig. 2.3A) showing terrace locations. Terraces cluster in low elevation, low relief areas along the ECD.

Figure 2.9: Rounding and transport estimates of clasts preserved in terrace gravels



Distribution of estimated clast transport distances from terraces N, E, B, and A (Figs. 2.3A and 2.3B). Transport estimates are based on measured values of Cailleux Roundness Index from each clast (Cailleux, 1947) after Sadler and Reeder (1983).

Figure 2.10: Conceptual model of stream capture-driven escarpment retreat.

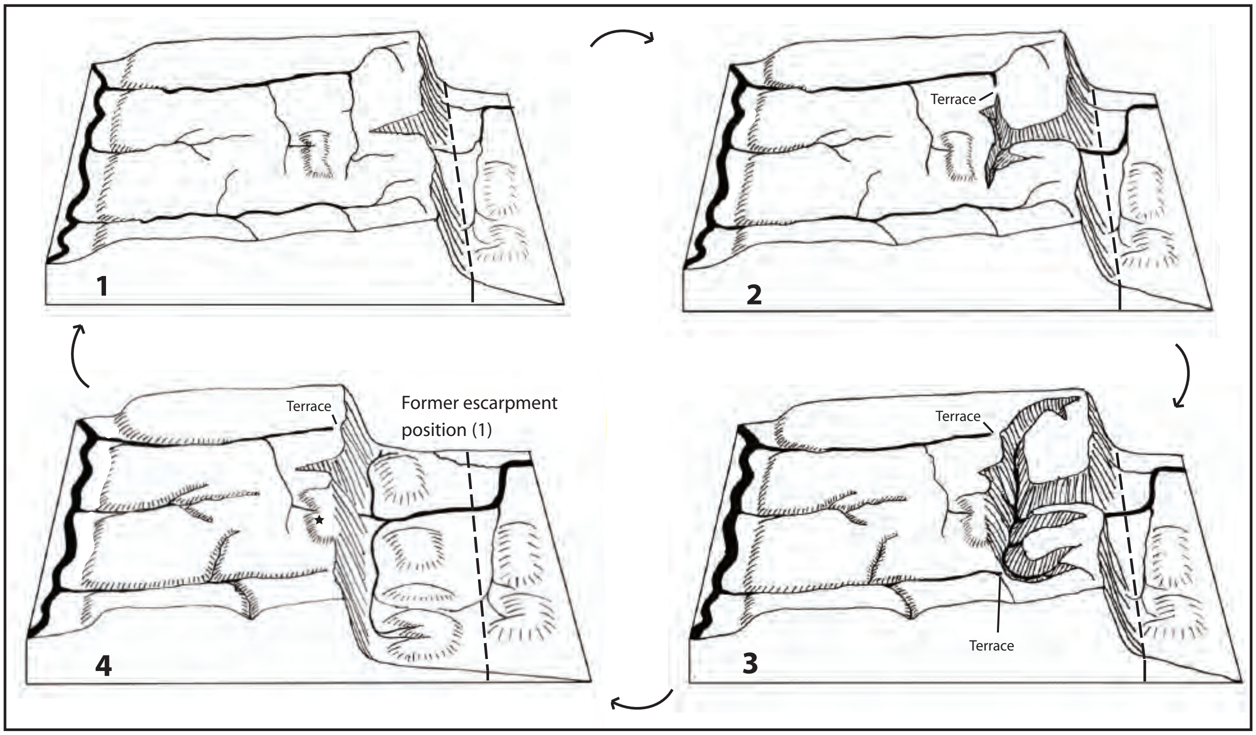
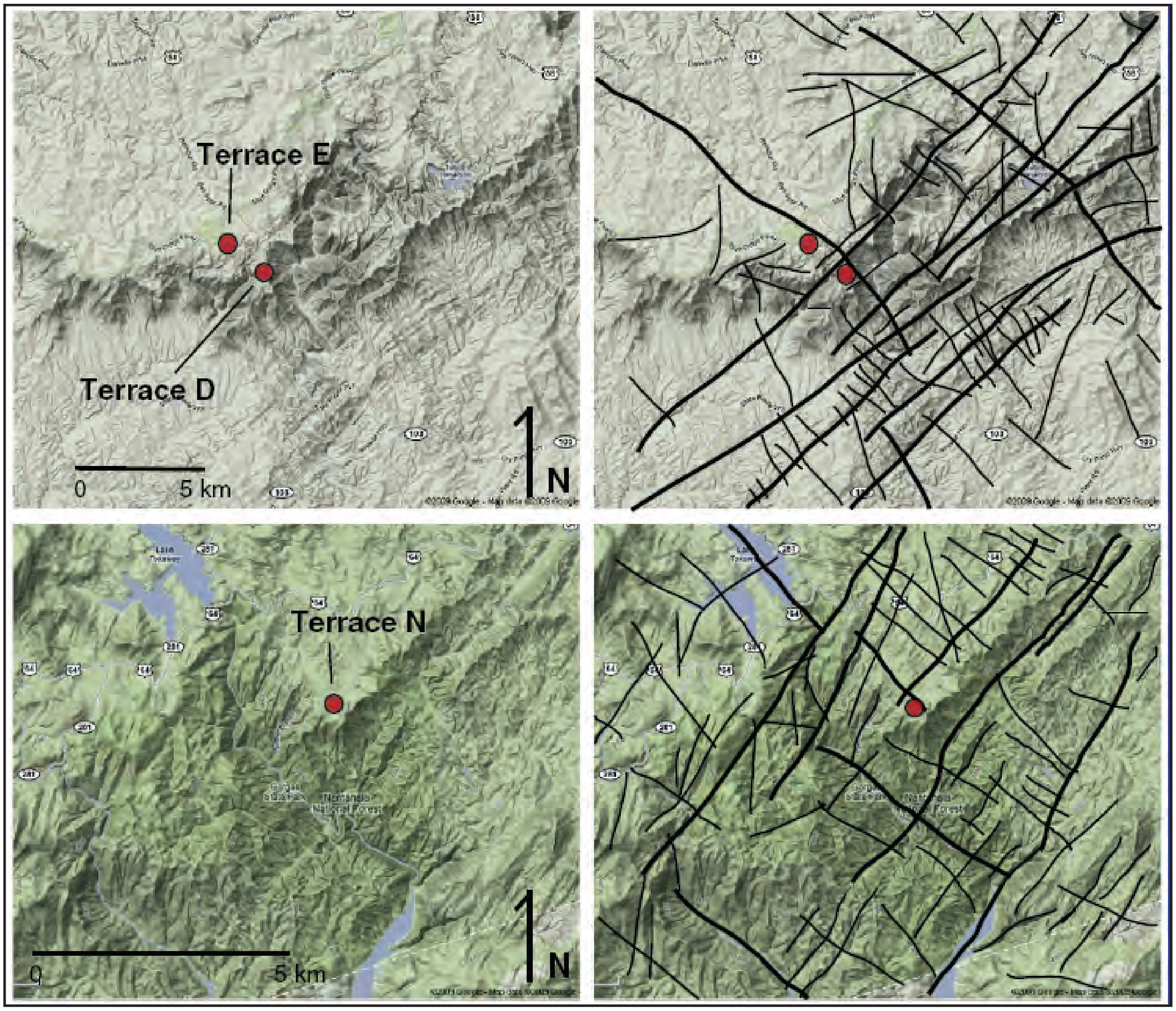


Illustration of the conceptual model of parallel divide and escarpment retreat driven by high order stream capture. The dashed line, indicating position of the escarpment in Frame 1, remains fixed throughout as a reference point. (Panel 1) Headward erosion progresses rapidly along an escarpment-orthogonal weakness (e.g., fracture) and approaches an initial capture point. (Panel 2) Capture and reversal of escarpment-parallel drainage leads to rapid dissection and the progression of headward erosion along drainage-controlling weaknesses. Because of the rectilinear drainage network and low Upland relief, two additional captures are now imminent. A terrace may have been preserved by the beheading of the preexisting drainage. (Panel 3) Capture of two more escarpment-orthogonal streams leads to rapid dissection of the entire basin affected by capture thus far. Rapid incision by the reenergized drainages isolates portions of the Upland and strands two new remnant terraces. (Panel 4) Complete dissection of the captured basin leads to parallel escarpment and divide retreat. A terrace deposit remains at the now underfed headwaters of the far drainage. The near drainage, though beheaded, has been rejuvenated by an Upland base level drop that has locally steepened the terrain and scoured away any residual alluvium. The small ridge now located at the escarpment crest was once a local divide within the Upland interior, and the stream draining it (starred) has not been affected by divide migration although it now drains the main regional divide. Stream 29 may be respresentative of the ECD "overtaking" a subordinate, landward divide (Figs. 2.3A, 2.5). Capture is not favored across this ridge or the more symmetric divide at the bottom of Frame 4, and this portion of the escarpment will enter a phase of stability. Headward erosion could proceed along another escarpment-orthogonal lineament to initiate a new cycle of capture and punctuated retreat of the far portion of the escarpment.

another escarpment-orthogonal lineament to initiate a new cycle of capture and punctuated retreat of the far portion of the escarpment.

Figure 2.11: Lineament trends relative to selected terrace deposits.



Shaded relief maps of two areas, each shown with and without interpreted lineaments, showing the roughly orthogonal lineament trends prevalent in areas of the BRE that host well-rounded and fresh terrace alluvium (see Figs. 2.3A and 2.3B for location). The thickness of the interpreted lineament trends is a proxy for influence over the drainage network. Northeast-trending lineaments are associated with thrust faults and weak, sheared lithologies of the Brevard/Bowens Creek fault zone. Northwest-trending lineaments are joints and fractures that pass uninterrupted across the Brevard fault zone, suggesting post-Paleozoic origin. The rectilinear drainage network resulting from the interaction of these weak features facilitates repeated large capture events as well as rapid dissection of the captured basin (Fig. 10). Shaded relief images are captured from Google Maps, based on 10-m resolution topography. ©2009 Google-Map data ©2009 Google.

APPENDIX A: U-Th-Pb DATING OF DETRITAL ZIRCONS FROM BLUE RIDGE ESCARPMENT TERRACE DEPOSITS

Introduction and description of technique

The relative stability and resultant longevity of the Blue Ridge Upland has allowed widespread preservation of fluvial terraces of unknown age or association with the modern drainage network (Prince et al., 2010). Such terraces are regularly encountered in wind gaps located along the crest of the Blue Ridge Escarpment (BRE) and eastern continental divide (ECD) in western North Carolina and southwest Virginia. Present consensus suggests that the rivers responsible for wind gap terraces slowly lost drainage basin through stream capture, producing northwestward migration of the eastern continental divide and Blue Ridge escarpment and leaving underfed streams or dry valleys inundated with sediment terminating in the gaps. Provenance of the sediments preserved in these dry or underfed drainages should be useful in determining the former extent and drainage networking of the Blue Ridge Upland. Although most of the terrace deposits contain only quartz (or possibly metasedimentary quartzite) cobbles and gravel, some refractory mineralogies, particularly zircon, have been preserved in finer terrace sediments. All rocks presently in outcrop on the Blue Ridge Upland should, as metasediments of Laurentian affinity, contain only Grenville-age zircons. Any zircons of Paleozoic age found in upland sediments must have originated in rocks to the east that currently outcrop at least 400m lower in the Piedmont physiographic province. Of particular interest would be felsic pluton-derived zircons from units such as the Mt. Airy granite and Henderson Gneiss. The limited size of these plutons would allow tight constraints on paleo-drainages as well as provided minimum distances of escarpment retreat.

Sampling Methods

Appropriate underfed stream valleys and wind gaps were selected after analysis of basin parameters such as longitudinal profile and slope-drainage area relationship. Field reconnaissance confirmed the presence of terrace material along the eastern continental divide at the headwaters of the selected drainages. Approximately five gallons of terrace material, including both cobbles and fine-grained matrix, were collected from each location. Two locations, Pine Creek, Virginia (PC-1) (terrace E, Fig. 2.3A) and Cannaday Gap, Virginia (CG-1) (terrace C, Fig. 2.3A) were sampled from the headwaters of small streams originating in the terrace deposits. Ward's Gap, Virginia (WG-1) (terrace I, Fig. 2.3A) and Flat Creek, North

Carolina (FC-1) (terrace N, Fig.2.3B) were sampled from subaerial terrace deposits within 400 m of the ECD (Fig. 2.12). Sampling locations were chosen to avoid possible contamination from road gravel or other lithic material introduced by non-natural means to the terrace areas. Terrace material was first washed to remove clay and plant debris. The resulting sediment was sieved to collect the <250 um fraction. WG-1 was processed using a Wilfly table to concentrate dense phases and remove quartz and mica. Approximately 250 g of the <250 um sediment were processed using LST (lithium heteropolytungstate liquid) to separate quartz from denser phases, such as zircon.

Magnetic separation by Frantz machine was conducted at 0.5 A to remove opaque and mafic minerals from the <250 um high density split. The resulting sediment was inspected by microscope to confirm the presence of sufficient zircon for 100 grain analysis. PC-1, WG-1, and FC-1 were selected for U-Th-Pb dating due to their proximity to rocks known to contain zircon of Paleozoic age and the high concentration of zircon in the final split. CG-1 was not selected for analysis due to the possibility of road gravel contamination and the low abundance of zircon in the final split.

PC-1 and WG-1 contained subhedral to rounded zircons as well as small euhedral zircons. No difference in magnetic susceptibility, a possible indicator of slight chemical variation and thus different origin, was observed between the two morphologies. The euhedral zircons were generally smaller than the subhedral/rounded zircons. Inclusions were common, and may have contributed to some of the low magnetic susceptibilities (approximately 0.75A) of some of the zircons.

FC-1 contained mostly high aspect ratio, euhedral zircons of a morphology very distinct from those observed in PC-1 and WG-1. Rounded zircons were not present. Zircons were relatively inclusion-free and did not exhibit the unusually low magnetic susceptibilities encountered in PC-1 and WG-1.

Samples were submitted to the Laserchron facility at the University of Arizona for final refinement and mounting to eliminate potential sample bias based on grain morphology.

Analytical Methods

U-Th-Pb dating was conducted at the University of Arizona Laserchron Center. (particulars of instrument and facility). 100 zircons were analyzed in each sample using a 35 um laser spot and 148 pulses to produce sufficient ablation. Six additional core-and-rim analyses were conducted on WG-1 with a 25 um spot to check for thin Paleozoic overgrowths. The first 50 zircons analyzed in each sample were chosen from randomly selected five-grain "units" throughout the mount to eliminate bias or emphasis on one area of the mount. The subsequent 50 analyses concentrated on euhedral grains in an attempt to locate Paleozoic magmatic zircon in the samples. Following the dating process, zircons with analyzed using cathodoluminescence to insure thin overgrowths present were in fact ablated during analysis.

Results and Discussion

All zircons analyzed provided only Grenville ages (1.2-1.4Ga) (Figs. 2.13-2.15). PC-1 and WG-1 contained two significant populations; one at approximately 1.2 Ga and a smaller population around 1.4 Ga. FC-1 ages were very tightly clustered just below 1.2 Ga. Concordia plots indicated no presence of younger rims or overgrowths surrounding the Grenvillian cores. U-Th concentrations indicated evidence of metamorphic overgrowth of Grenvillian age. Core-and-rim analysis of WG-1 indicated no Paleozoic overgrowths on the zircons.

The lack of Paleozoic zircon ages among the analyzed grains suggests no presence of Piedmont-derived sediment in fluvial terraces along the ECD in North Carolina and Virginia. Approximately one-half of analyses were focused on euhedral crystals as the most likely to be of younger magmatic origin, but the same suite of ages were produced by euhedral zircons and the larger, subhedral to rounded zircons. When beam size was reduced to 25um and cores and tips/rims of euhedral zircons were analyzed, consistent ages were produced from cores and rims indicating no Paleozoic overgrowth. Cathodoluminescence indicated core-and-rim and structures on a number of the analyzed grains. Numerous laser pits crossed subgrain boundaries, but the results for all analyses sampling what appeared to be multiple zircon crystallization events all yielded Grenville ages. U-Th concentrations, when plotted against Pb 206/U 238 ages, indicated overgrowths were formed during metamorphic events occurring around the same time as crystallization. Lead loss due to weathering and radiation damage affected a number of grains.

The absence of Paleozoic zircons in the sampled material does not allow for any conclusions regarding the former extent or drainage networking of the Blue Ridge Upland. Zircons analyzed yielded ages consistent with zircons that would be found in rocks and sediments originating on the upland surface today. The bimodal Grenville populations encountered in the Virginia samples (PC-1 and WG-1) are consistent with age suites produced by bedrock analyses by Bream (2003). The tightly grouped 1.15-1.2 Ga ages produced by FC-1 are consistent with zircon ages from the Toxaway Dome, a large body of Grenville age gneiss underlying the FC-1 sample site and observable in outcrop nearby. While age results obtained appear to provide no insight into the evolution of the Blue Ridge Upland, they may offer some useful information about the nature of the sediments found in upland terraces and the mechanism of divide migration. Zircon material was sampled from terrace material that suggested little mixing with local colluvium. The majority (all, in the case of FC-1) of clasts encountered in the sampled deposits clearly displayed considerable rounding by fluvial process. The lack of local, untransported material suggests that the sediment sampled is dominantly, if not almost entirely, of fluvial origin. Subdued local relief surrounding the terrace deposits and their location along the continental divide further reduce the possibility of colluvially mixing local zircon-bearing soil derived form *in situ* weathering into the terraces. These features of the deposits combine to suggest that most of the material within the terrace deposits has, in fact, been fluvially transported to its current location.

The abundance of Grenville-age zircon in the deposits indicates deposition of sediment sourced from a basin not encompassing nearby non-Grenville plutonic material and may be useful in demonstrating that stream piracy and migration of the ECD has been, for the most part, a gradual process. This is especially well-demonstrated in the case of FC-1, where the Ordovician Henderson Gneiss is located only 3-4 km away from the location of the sampled terrace. If a paleo-river draining both the Henderson Gneiss and Toxaway Dome were catastrophically captured, a combination of zircons from both sources should be preserved in terrace deposits formed by the river. More gradual capture would remove the more distant Henderson source from the basin while still allowing the more proximal Toxaway Dome to supply zircon-bearing sediment. The reduction in basin would in turn reduce stream competence and reduce the possibility of removing a great deal of previously deposited sediment from the basin. An analysis of sediment preserved deeper in the terrace might validate this hypothesis.

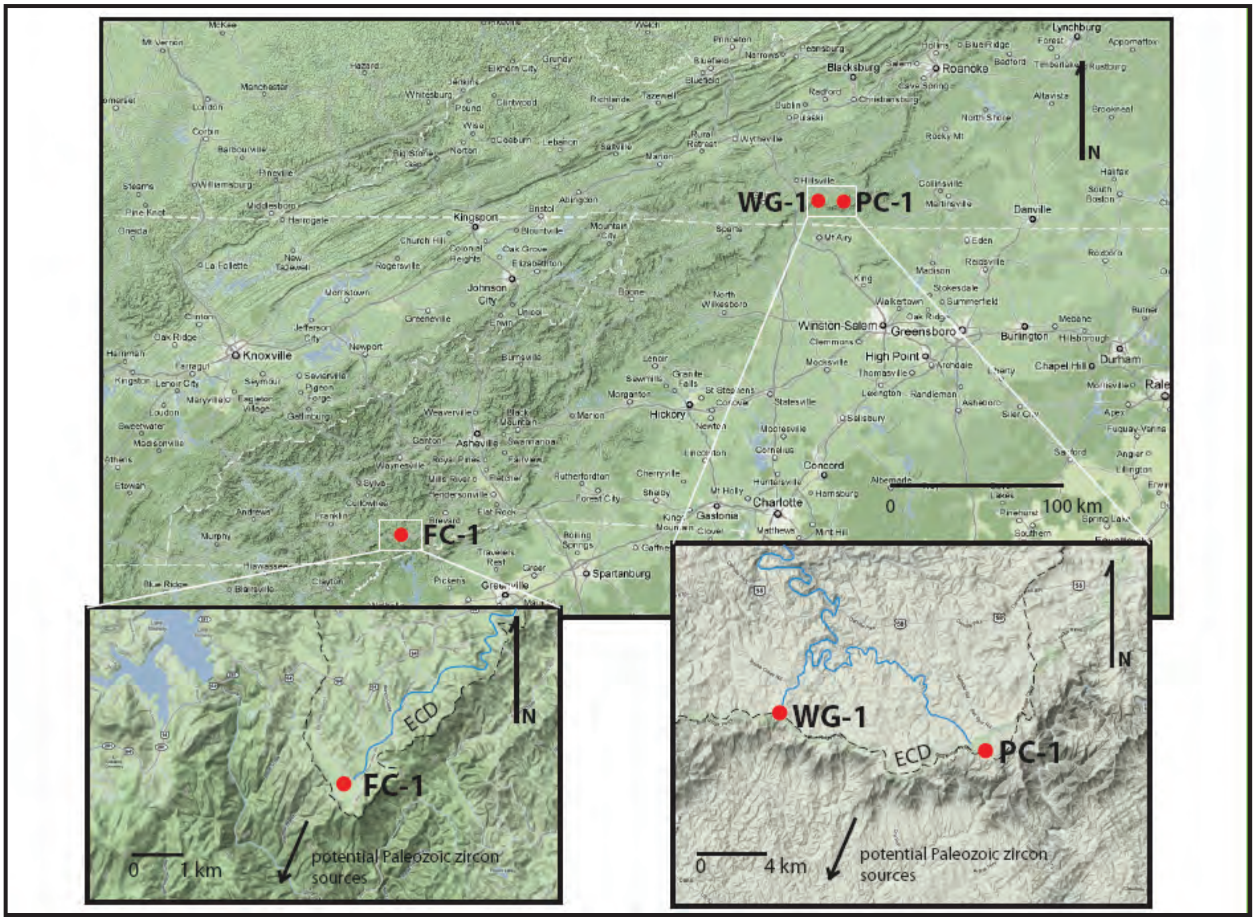
The volumetric contrast between Grenville zircon sources and Paleozoic zircon sources in the potential "paleo-basins" may also have contributed to the absence of Paleozoic zircons in the terraces sampled. The Henderson Gneiss and Mt. Airy granite are volumetrically minor when compared to the hosting metamsediments, and the outcrop extent of these units on a pre-existing land surface at least 400m higher than present cannot be known. As plutons of finite dimension, the Mt. Airy granite and Henderson Gneiss theoretically might not have been exposed at all on an older upland. Assuming these units were in fact exposed to provide zircons, considerably more metasedimentary rock would have been contributing zircons simultaneously and effectively “diluting” the zircon signature of the plutons. Only 100 zircons are analyzed per sample; over 1,000 are mounted for analysis in each sample. The samples, when prepared for analysis, consist of the zircon separated from five gallons of sediment collected from deposits covering several acres to thicknesses of at least one meter and often several meters. There exists the possibility that a small number of Paleozoic zircons were in fact present on the sample mount and simply were not analyzed. Also plausible is the presence of Paleozoic zircon in particular areas or depths of the terraces that were not sampled. Additional in-depth characterization of terrace deposits would be useful prior to subsequent studies. A more practical and economical approach to zircon provenance studies in such old deposits might be to concentrate on several hundred analyses from a number of depths and locations in one deposit.

References

Bream, B.R., 2003, Tectonic implications of geochronology and geochemistry of para- and ortho-gneisses from the southern Appalachian crystalline core [Ph.D. dissertation]: Knoxville, University of Tennessee, 296 p.

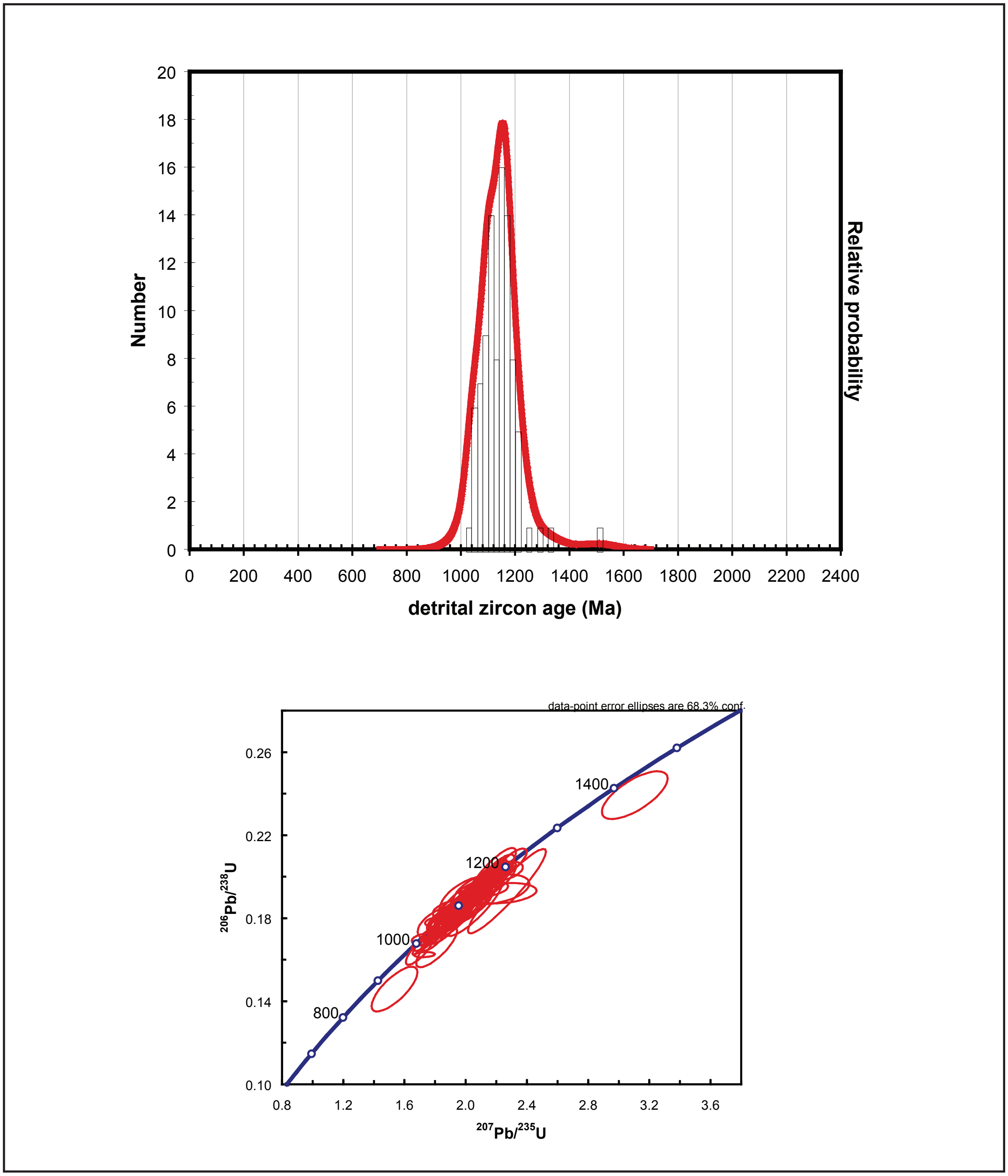
Prince, P.S., Spotila, J.A., and Henika, W.S., 2010, New physical evidence of the role of stream capture in active retreat of the Blue Ridge Escarpment, southern Appalachians: Geomorphology, v. 123, p. 305–319, doi:10.1016/j.geomorph.2010.07.023.

Figure 2.12: Detrital zircon sampling locations.



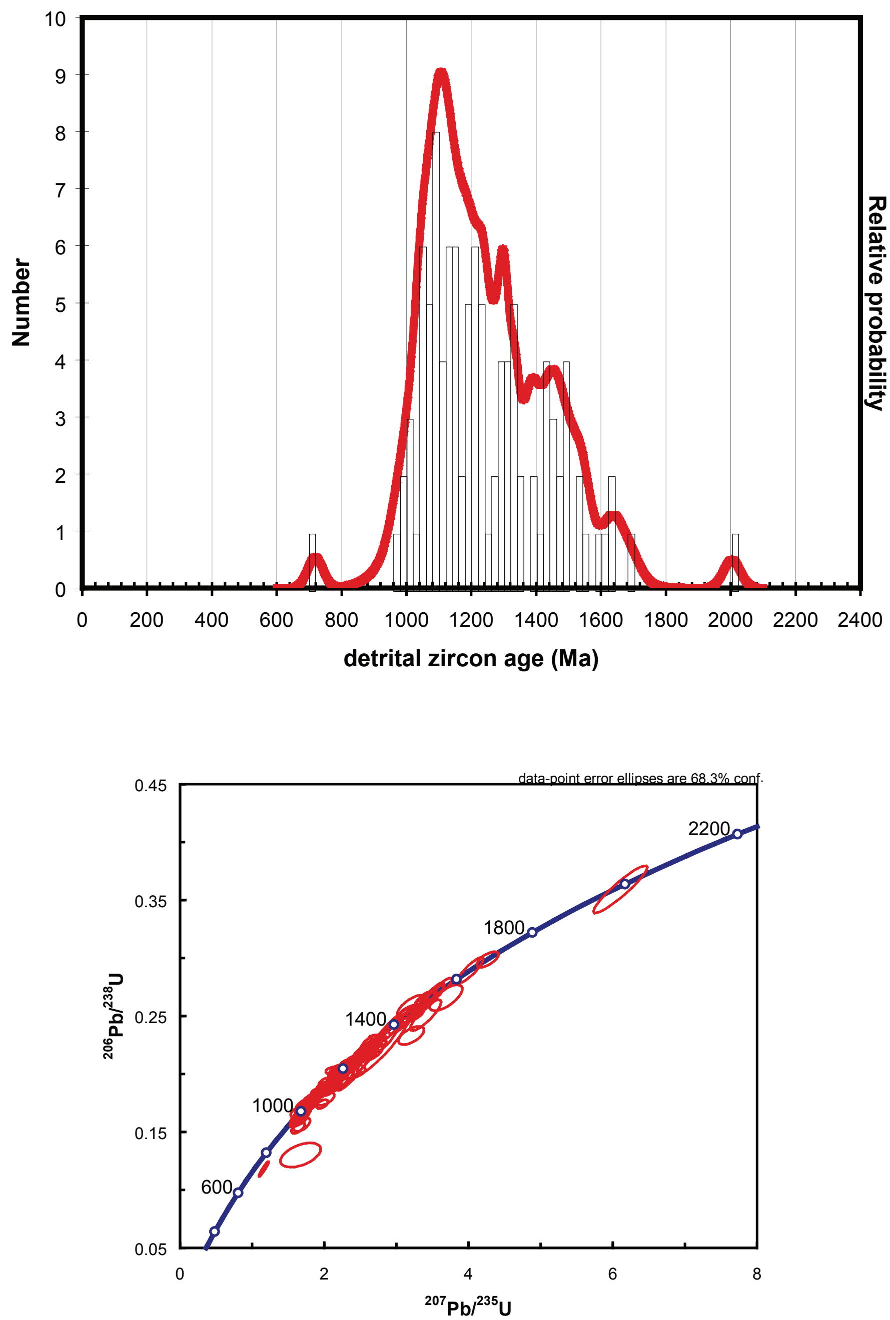
Location of detrital zircon sampling sites along the Blue Ridge Escarpment. Samples PC-1, WG-1, and FC-1were derived from terraces E, I, and N (respectively) of Figs. 2.3A and 2.3B. ©2009 Google-Map data ©2009 Google.

Figure 2.13: Ages of FC-1 zircons.



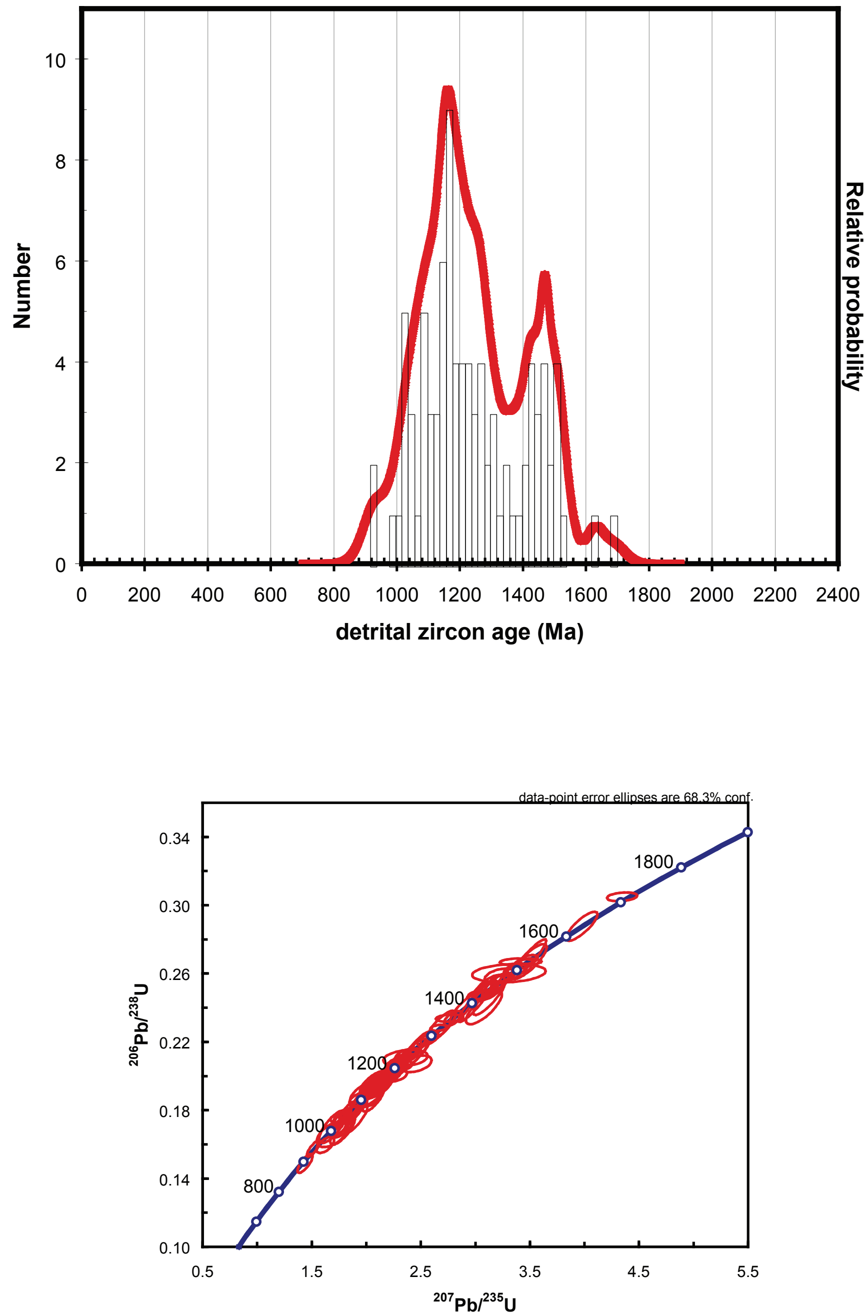
Age probability density and Pb/U concordia plot, FC-1 zircons. No discordance indicated.

Figure 2.14: Ages of WG-1 zircons.



Age probability density and Pb/U concordia plot, WG-1 zircons. No discordance indicated.

Figure 2.15: Ages of PC-1 zircons.



Age probability density and Pb/U concordia plot, PC-1 zircons. No discordance is indicated.

Chapter 3

Stream capture as driver of transient incision in a tectonically quiescent setting

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ABSTRACT

We use unique fluvial gravel deposits preserved atop a regional drainage divide to confirm the role of stream capture in driving ~250 m of incision in the transient Roanoke River basin of the Appalachian Mountains. Gravel provenance constrains the pre-capture position of the divide, indicating ~225 km^2 of basin area were abruptly connected to the base level of the capturing stream. The resulting wave of incision is presently manifest as major knickzones separating adjusting reaches from relict headwaters resembling streams of the New River basin, from which the Roanoke River was captured. The unusual preservation of the unconsolidated gravels on small relict surfaces adjacent to bedrock gorges indicates extreme spatial variability in erosion rates within the Roanoke basin, which is the first documented example of a transient passive margin basin connected to a capture event by stranded fluvial debris. Our results show the potential for stream capture across an asymmetric drainage divide to drive major transient incision independent of external forcings, such as climate change or tectonic uplift. A continuation of this process will lead to eventual capture of ~7,000 km^2 of the New River basin in the relatively near geologic future.

INTRODUCTION

Observations of transient landscape response to perturbations in boundary conditions are data upon which models of landscape evolution are built (e.g., Davis, 1899a; Leopold and Bull, 1979; Howard, 1994; Whipple and Tucker, 2002). Principles of topographic disequilibrium are primarily established from cases in active tectonic settings, where rapid base level drop produces a unique signature of bedrock gorges, elevation-accordant remnant surfaces, and migrating knickpoints separating adjusting and relict portions of the drainage network (e.g., Schoenbohm et al., 2004; Clark et al., 2006; Oskin and Burbank, 2007). This understanding is furthered by examples of topographic adjustment to extreme climate change, such as transition from fluvial to glacial conditions (Brocklehurst and Whipple, 2002; Brook et al., 2006). The nature of topographic response to changes in boundary conditions where active uplift is minimal, however, is poorly constrained, as landscape transience is less recognized in areas of slow erosion. The magnitude of perturbation necessary to induce transient response and the timescale of adjustment are also unknown. Transient landscapes in slowly eroding settings may thus enhance interpretation of non-tectonic controls on denudation and sediment delivery, but they have yet to be exploited.

The lack of landscape transience in subdued post-orogenic landscapes is largely due to the lack of abrupt base-level changes associated with tectonic uplift combined with topographic adjustment to independent variables, such as bedrock erodibility (Hack, 1973). In such settings, subtle changes in boundary conditions may result in only muted, gradual topographic adjustment and the appearance of a dynamic topographic steady-state (Hack, 1960). There are hints of disequilibrium in such settings, however, such as local discordance of erosion rate between advancing gorge heads and dissected relict plateau surfaces (Nott et al., 1996; Springer et al., 1997; Granger et al., 2001; Hancock and Kirwan, 2007). These examples suggest that local perturbations of topographic steady-state that are not externally forced may be intrinsic to slowly-eroding landscapes, but their origin remains unclear. A potential internal driver of transient topographic adjustment is stream capture (Hasbargen and Paola, 2000; Bonnet, 2009). Given the dependence of fluvial incision rates on slope and drainage area (Whipple and Tucker, 2002; Crosby and Whipple, 2006), capture could result in significant, visible adjustment of fluvial profiles and associated topography. Although stream capture is a well known phenomenon, to what degree does it represent a natural source of local disequilibrium in stable landscapes?

A hindrance to relating stream captures to landscape characteristics is the topographically ephemeral nature of such events (Bonnet, 2009). However, we have discovered unique fluvial gravels stranded by a major stream capture in the central Appalachians which can be directly linked to the transient response of a fluvial network. Along the rugged topographic margin of the Blue Ridge upland plateau, a competition of drainage basins has resulted in punctuated shifts of the eastern continental divide at the expense of the Mississippi basin (Hack, 1973; Gunnell and Harbor, 2010). Prince et al. (2010) identified gravels along the divide that connected divide migration to capture, but these deposits could not be clearly linked to basins in a state of transient adjustment (Fig. 3.1). The Roanoke River basin represents a transient Appalachian basin (Fig. 3.1), where headwater streams follow low-gradient alluvial channels on the plateau surface before cascading ~250 m over steep knickzones. This incision has dissected the plateau margin into relict surfaces separated by bedrock gorges, with the butte-like relicts grading into the contiguous plateau upstream of gorgeheads (Gunnell and Harbor, 2010; Prince et al., 2010).

Given that elevation and channel morphology of the Roanoke headwaters imply a former connection to the New River, the Roanoke basin provides a rare opportunity to relate basin-scale transience to stream capture in a non-tectonic setting. We have applied the approach of Prince et al. (2010) to look for gravels along the New-Roanoke divide, testing for a former connection between the two basins to constrain the origin of Roanoke basin disequilibrium. A review of these methods can be found in the Data Repository.

OBSERVATIONS

Two fluvial gravel deposits stranded by Roanoke capture were discovered in low-elevation, low-relief zones along the New-Roanoke divide. Extensive mapping indicates that the deposits are localized remnants of discrete streams and not part of a regionally extensive alluvial cover. Both deposits are comparable in field setting and expression to the gravels identified by Prince et al. (2010). The Fishers View deposit (Fig. 3.1) is preserved atop clay-rich regolith and weathered gneisses of the Blue Ridge province and contains rounded to well-rounded clasts of metaquartzite, quartz mylonite, clear and blue vein-derived quartzite, and pyroxene granulite. The slightly weathered pyroxene granulite clasts represent the only polymineralic clasts yet found on the divide. Although the deposit is unconsolidated and unstratified, clast concentration varies within the deposit with some areas approaching clast-supported structure. Sandy clay fines are clearly distinguishable from the red clays derived from the weathering of local bedrock. The Crab Creek deposit is preserved ~4 km to the northwest atop weathered Cambrian-Ordovician dolomites of the eastern Valley and Ridge province (Fig. 3.1). The deposit grades from elevation-accordant hilltop cappings east of the divide to a continuous swath of alluvium in the wind gap headwaters of Crab Creek (stream 1, Fig. 3.1), a New River tributary. Excavation of one hilltop exposure has revealed a sharp contact between alluvium and local clays derived from carbonate weathering (Fig. 3.2). The Crab Creek deposit contains silicified oolite and chert clasts from the eastern Valley and Ridge as well as the same Blue Ridge-derived metaquartzite, quartz mylonite, and clear and blue vein quartzite lithologies observed in the Fishers View deposit. Clasts of Blue Ridge origin are smaller and more rounded than those of the Fishers View deposit, which is nearer the Blue Ridge source areas.

The range of clast lithologies allows basic paleo-basin reconstruction. Translucent and blue vein-derived quartzites indicate sourcing from a distinct suite of basement gneisses of the northern Blue Ridge upland (Virginia Division of Mineral Resources, 1993) (Fig. 3.1). The location of the pyroxene granulite source relative to the Fishers View deposit indicates generally southwest flow across the unit prior to capture-driven drainage rearrangement, although the relationship between this deposit and the Crab Creek deposit is unclear. Blue vein quartz pebbles from both deposits are strong provenance constraints, as field reconnaissance indicates outcrop of coarse blue quartz veins is limited to the headwaters area of stream 3, 18 km east of the Crab Creek deposit (Fig. 3.1). Blue quartz clasts also distinguish the deposits from nearby New River terrace gravels of Ward et al. (2005), which are dominated by "milk glass" vein quartz clasts sourced well southwest of the study area (Fig. 3.1). Translucent gray chert in the Crab Creek deposit was sourced from a dolomite unit at the eastern margin of the Valley and Ridge, suggesting capture occurred west/northwest of the present dolomite outcrop near the confluence of stream 11 and the Roanoke River (Fig. 3.1). Chert clasts are sub-rounded, while well-rounded translucent vein quartzites suggest up to 25 km transport based on the approach of Prince et al. (2010), consistent with the respective source locations relative to the Crab Creek deposit.

Knickpoint and relict surface distribution suggest incision has propagated as a wave throughout the captured basin. Sharp knickpoints separating unincised, sand- and gravel-bedded relict stream reaches from adjusting reaches persist in crystalline basement units ~15 km upstream from the confluence of stream 11 and the Roanoke River (Fig. 3.1, Fig. 3.3). Knickpoints in weaker sedimentary units of the Valley and Ridge are less pronounced, and incision appears to have "leaked" upstream of the steepest portion of the knickzone (Berlin and Anderson, 2009). Channels on sedimentary and basement rock closer to the inferred capture point lack sharp knickpoints but maintain oversteepened profiles (Norton et al., 2008) (Fig. 3.3). Headward erosion by these streams has produced additional captures along the basin margin. Capture propagation is apparent in the westernmost headwaters of stream 11 (Fig. 3.1), where flow is toward the New River before bending > 90° toward the Roanoke to produce a grapnel-shaped drainage pattern similar to that observed by Vogt (1991). Relict surfaces near the inferred capture point are small and discontinuous, while relict surfaces closer to the major basement knickpoints (streams 2, 3, 4, and 8) are larger and more intact, ultimately grading into the contiguous plateau (Fig. 3.1). Loose projection of relict reaches toward Crab Creek indicates the Roanoke headwaters have retained a gradient consistent with connection to the New River (Fig. 3.3). Basement knickpoints thus appear to represent a mobile yet impenetrable barrier to external forcings, beyond which the landscape continues to evolve under pre-capture boundary conditions.

DISCUSSION AND CONCLUSION

Fluvial gravels preserved at Crab Creek and atop the New-Roanoke divide confirm that transient Roanoke River incision is the result of capture from the New River basin. Gravels were sourced from identifiable rock units now in outcrop entirely within the Roanoke basin, confirming the former connection of the Roanoke headwaters to the New River. Since the capture, accelerated incision has been focused in large streams and has not yet propagated through the entire captured drainage network. Relict surfaces and knickpoints suggest the Roanoke headwaters remain in a state of transient adjustment. Unlike other transient Appalachian landscapes, the gravels we have identified clearly indicate a capture event as the origin of Roanoke headwaters disequilibrium. Our findings are corroborated by biological studies which have shown that the Roanoke River hosts several fish species that are of New River basin origin (Jenkins et al., 1971). Fluvial debris, knickpoint location and elevation accordance constrain the preservation and extent of the dissected paleolandscape, allowing a detailed characterization of the transient response of a ~225 km^2 basin to a rapid ~250 m drop in base level.

Extreme spatial variation in erosion rate that is inconsistent with Appalachian dynamic equilibrium (e.g., Hack, 1960; Matmon et al., 2003) is suggested by the preservation of the gravels. In order for the gravel deposits to persist, the upland surface must be stable, incision must be rapid, or both. Gravels are located atop both weathered crystalline rock of the Blue Ridge and dolomites of the Valley and Ridge, indicating lithology does not control relict surface preservation. Within adjusting portions of the Roanoke basin, crystalline and sedimentary rock units support similar bedrock stream gradients and comparably rugged terrain, consistent with spatially-limited decoupling of topography, stream gradient and lithology along the migrating Atlantic-Gulf of Mexico divide suggested by Hack (1973) (Fig. 3.1, Fig. 3.3). Knickpoints are not localized to structures or lithologic contacts, suggesting the features represent the spatially transient “leading edge” of adjustment to the new base level. We collectively interpret these morphological factors as evidence of the control exerted by the rate of onset and magnitude of the base level drop on adjustment of the fluvial network (Whipple and Tucker, 2002; Pazzaglia and Brandon, 2001). Knickpoint migration through actively adjusting areas must have been sufficiently recent and rapid that topography and stream gradient have not yet had time to reflect contrasts in erodibility.

While the age of the deposits is unknown, we can make crude inferences based on existing Appalachian data. Both deposits have maintained a sandy clay matrix distinct from host soils derived by in situ bedrock weathering, suggesting minimal mixing through bioturbation or colluviation that would be expected during long-term preservation. We interpret the limited degree of clast weathering as evidence of the youth of the capture event. Quartzite clasts show no friability and minimal pitting, and the survival of pyroxene granulite clasts (Fishers View) suggests the gravels are younger than the Blue Ridge Escarpment gravels of Prince et al. (2010) (B, Fig. 3.1). Quartzite clast preservation is comparable to similar cobbles from ~1–2 Ma New River terraces 25 km to the northwest (Ward et al., 2005). If the gravels are this young, the 17 km of knickpoint retreat and extensive dissection observed would require truly exceptional post-capture erosion rates. The very existence of the gravels does support the extreme rates we propose; denudation rates of ~10 m/Ma documented elsewhere on the low-relief Blue Ridge Plateau surface (Spotila et al., 2004; Sullivan et al., 2007) suggest a limited potential for preservation of unconsolidated surficial deposits.

Our results demonstrate that stream capture can produce major landscape disequilibrium in tectonically quiet settings. We observe a fluvial response analogous to that described in transient landscapes from tectonically- or climatically-forced settings (Harkins et al., 2007; Norton et al., 2008). Preservation of the gravels suggests relict surface stability may be exceptional, unlike active tectonic transient landscapes where background erosion rates may be too high for long-term preservation of relict surficial deposits (e.g., Harkins et al., 2007; Norton et al., 2008). Relict upland surface stability is reflected in the persistence of the elevated Blue Ridge Plateau ~200 Myr after rifting, which provides a source of potential energy to be converted to kinetic energy by capture. The plateau thus functions as an "energy reservoir" which, when tapped by Atlantic streams, drives localized rapid erosion atypical of the Appalachians at large. Assuming the landward base level elevation is maintained, the erosional mechanism we describe should continue to produce transient incision events until the Blue Ridge Plateau has become entirely dissected and adjusted to the Atlantic base level (Fig.3. 4). We expect additional captures by stream (11) to occur quickly through the weak, fractured shale RS

(Fig. 3.1), leading to the capture of the Little River and then the New River near present Graysontown, Virginia (GT, Fig. 3.1). If the capture which stranded the Crab Creek deposit occurred ~2 Ma and the “grapnel” headwaters of stream 11 reflect the progress of additional captures, New River capture should occur within 3 Myr. This will suddenly connect ~7,000 km^2 of Blue Ridge plateau to the Atlantic base level and cause a rapid wave of incision to propagate through the captured basin. Ultimate dissection of the New River basin will reduce the Blue Ridge plateau to a ridge morphology, indicating the potential of a single, large-scale capture event to alter landscapes at a regional scale and generate a major sediment flux to terrestrial or marine basins (Fig. 3.4). Drainage rearrangement along asymmetric passive margin drainage divides may therefore represent an internal driver of transience capable of operating long after the cessation of tectonic uplift and in the absence of climatic forcing.

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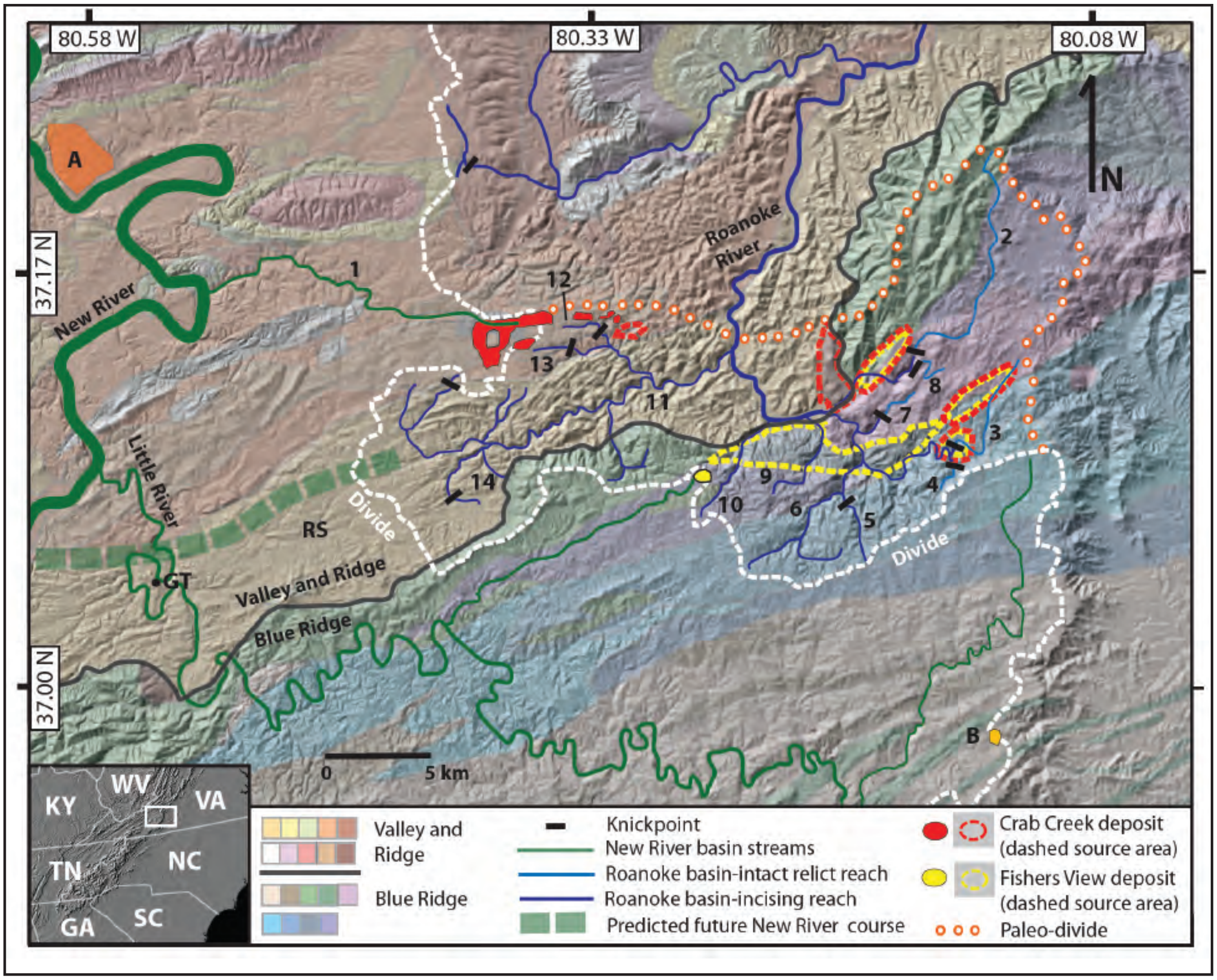
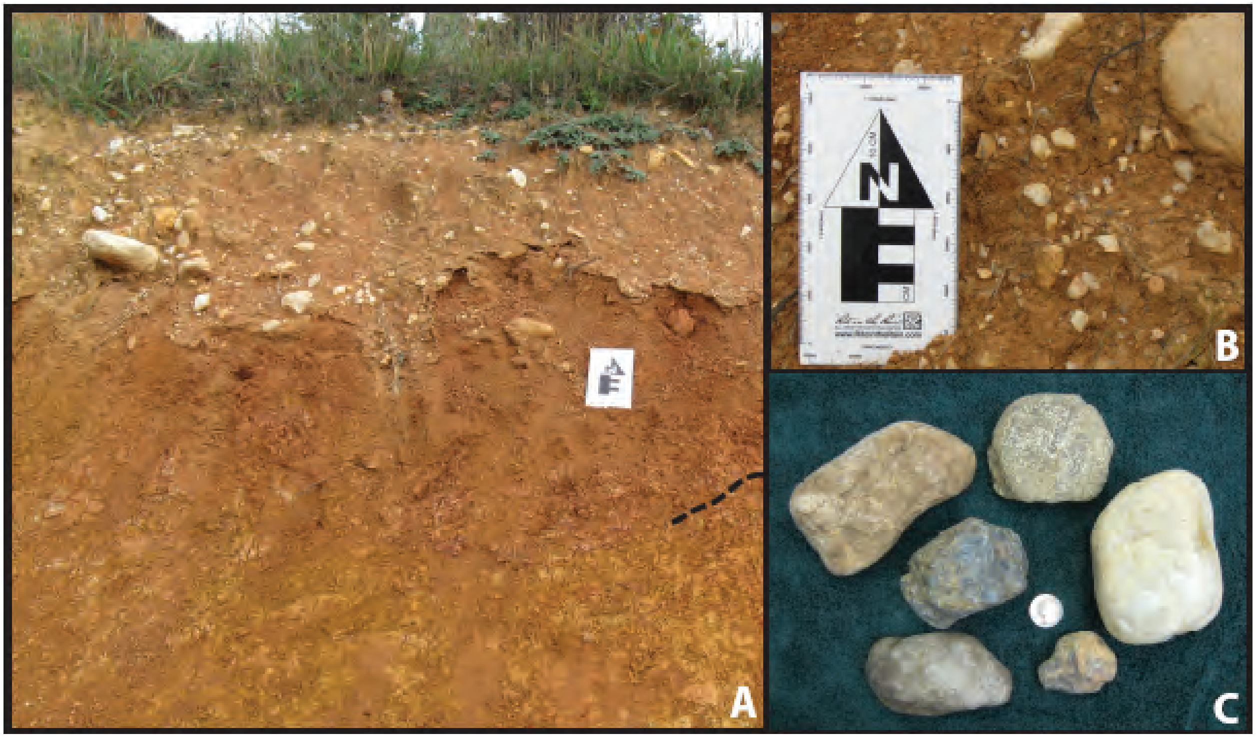
Figure 3.1: Map of the northern Blue Ridge Plateau



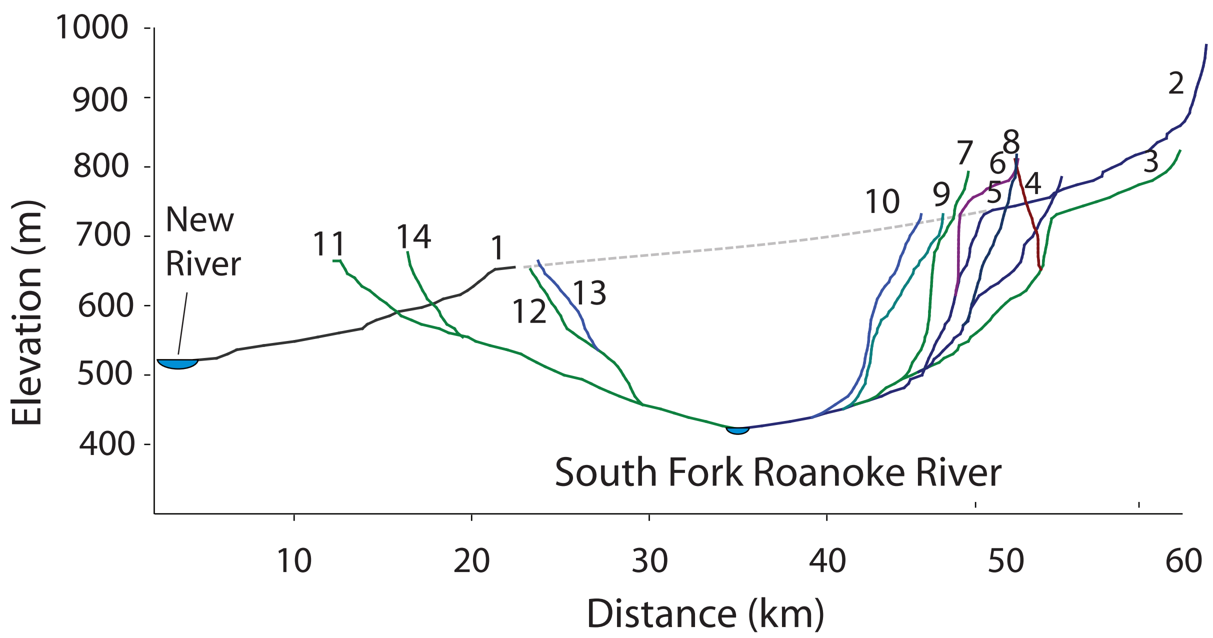
Map of study area combining bedrock geology (Virginia Division of Mineral Resources, 1993) with 10 m resolution DEM (www.seamless.usgs.gov). Inset shows general location of study area. Numbered drainages correspond to Fig. 3.3. Stream 1 is Crab Creek, beheaded by the capture. The "grapnel" shape of the westernmost headwaters of stream 11 is inferred to represent subsequent captures following the main capture which stranded the Crab Creek deposit. A-New River gravels of Ward et al.(2005), B-Gravel of Prince et al. (2010), GT-Graysontown, Virginia, RS-Rome shale.

Figure 3.2: Photographs of paleo-Roanoke River gravels.



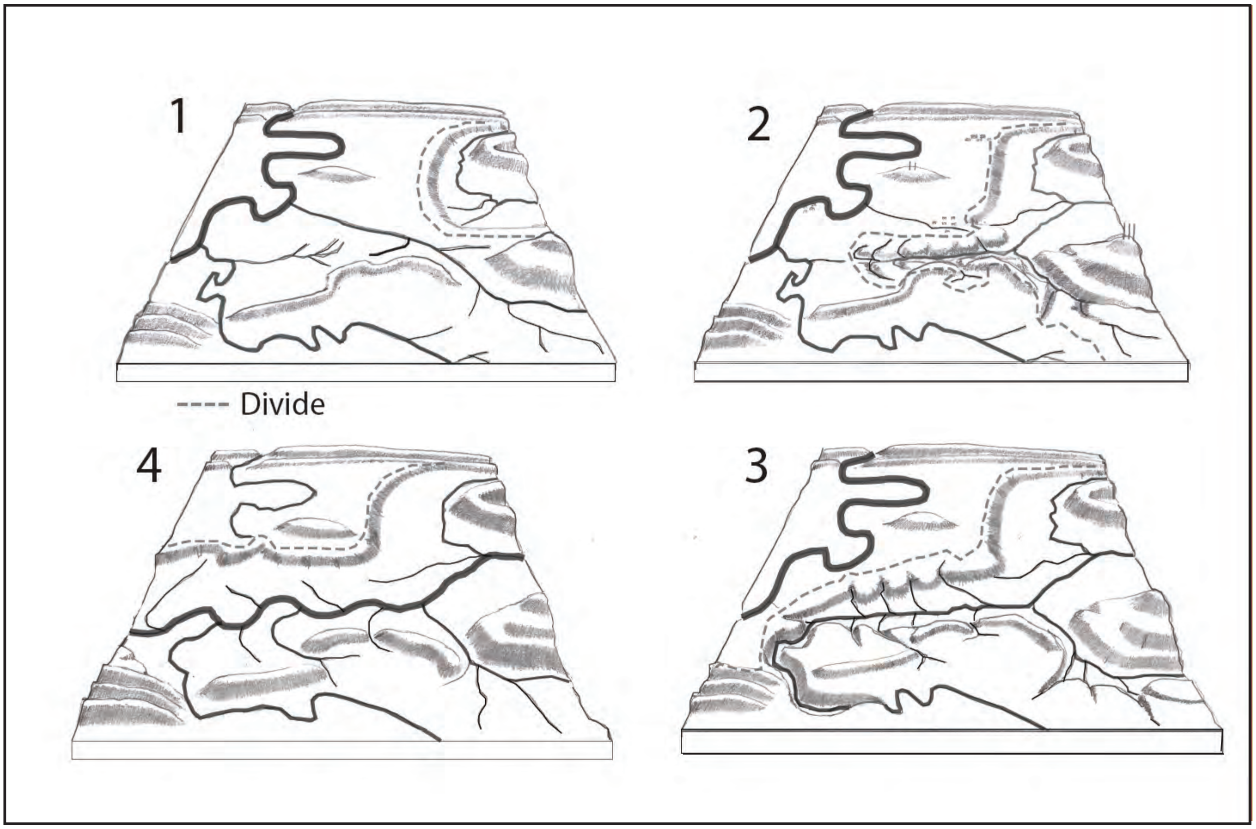
A) Outcrop of Crab Creek deposit, Christiansburg, Virginia showing alluvium (red)-residuum (tan) contact. B) Detail of Crab Creek alluvium showing vein quartz pea gravel and metaquartzite cobble. Arrow is 10 cm. C) Clasts from Crab Creek and Fishers View deposits. Clockwise from top: pyroxene granulite, glassy vein quartz, chert, metaquartzite, blue vein quartz, quartz mylonite. United States quarter for scale.

Figure 3.3: Longitudinal stream profiles.



Longitudinal profiles of streams within the study area (see Fig. 3.1). Knickpoints in streams 2, 3, 6, 7, and 8 separate relict alluvial headwaters from incising reaches. Hypothetical projection of relict reaches beyond knickpoints grades smoothly into Crab Creek headwaters. Lower Crab Creek is adjusting to New River incision discussed in Ward et al. (2005).

Figure 3.4: Conceptual model of Roanoke capture sequence.



Conceptual model illustrating the effects of stream capture events on the topography of the study area. Dashed line represents the drainage divide. Frame 1: Pre-capture drainage pattern suggested by gravel deposits. Frame 2: Present-day configuration showing effects of the capture event which stranded the Crab Creek deposit. Frame 3: Future capture of Little River near Graysontown, Virginia (see Fig. 3.1). Frame 4: Post-New River capture.

APPENDIX B: DESCRIPTION OF TOPOGRAPHIC ANALYSES USED TO IDENTIFY RELICT GRAVEL DEPOSITS

Fluvial gravels stranded by capture were located using the methods of topographic analysis and field reconnaissance described in Prince et al. (2010). These methods focus on identifying channels which have lost headwaters to capture events, leaving beheaded channels downstream of the capture point and driving rapid incision above the capture point. This rapid incision rapidly dissects the captured basin and obscures the appearance of the former drainage network, complicating identification of beheaded channel. Beheaded channels and their associated gravels must thus be identified by their topographic signature, consisting of atypically linear longitudinal profiles, excessive valley size with respect to drainage area, and little or no topographic expression of the drainage divide at the headwaters. As the drainage would have flowed through a valley prior to capture, these altered headwaters will occur at topographic lows along the asymmetric divide. These parameters are the physical manifestation of truncation of a concave-up longitudinal profile expected of a stream which has achieved some degree of energetic balance with its host substrate (e.g. Hack, 1960). While slope-area relationships can potentially quantify headwaters loss, topographic and longitudinal profile analysis can qualitatively identify its effects to guide field reconnaissance for stranded gravels. The tools and methods applied in this study are reviewed below.

Longitudinal profiles of streams rising at the divide and draining to the New River were constructed using 1:24,000 scale topographic maps. Streams with anomalously linear profiles are noted for field inspection, as this profile shape indicates truncation as well as preservation of the remnant channel. Analysis of hillshade maps produced from 10 m DEM data (www.seamless.usgs.gov) were also used to identify low-relief areas along the crest of the divide. Valleys ending in wind gaps are excellent candidates, but any low-relief zone along the divide is of interest. The use of the hillshade effect highlights topographic contrast and accentuates topographically muted low-relief headwaters occurring along the divide. Hillshade images also reduce the distraction produced by roads, towns, and land use notations on 1:24,000 scale topographic maps.

Once potential beheaded streams were identified, the locations were checked against a 1:24,000 scale topographic map to determine if they occurred at low elevation with respect to the crest of the divide at large. Lineament patterns in the study area were also considered, as joints and fracture traces exert considerable control over the Appalachian drainage network. Areas along the divide which exhibited low relief, low elevation, and alignment with a topographic lineament intersecting the divide were selected as starting points for field reconnaissance. This method increases the efficiency of field work by focusing reconnaissance on starting locations. The results of Prince et al. (2010) indicate stranded gravel deposits are localized to beheaded channels, and random field inspection along the trace of the divide could easily continue for days without locating gravel. Once deposits have been identified, however, field time can be allotted to mapping the extent of gravels and determining where no gravel is preserved.

As the unconsolidated deposits are preserved in low-relief areas, outcrop may not be readily apparent. Field inspection should focus on roadcuts, as the wind gap settings are frequently exploited by roads and railroads crossing the divide topography. Agricultural disturbance may also expose gravel, but will likely obliterate any primary depositonal features. Where no human disturbance has exposed soil, tree throw (floroturbation) typically exposes clasts. Once gravels are identified, field reconnaissance is conducted along the divide between deposits. Gravels are, however, expected to be localized based on the drainage density within the study area. Divide migration should occasionally "overtake" pre-existing subordinate divides, such that not every stream with headwaters on the main regional divide would be beheaded.

The well-preserved relict reach of stream 2 (Fig. 3.1) was loosely projected along probable structurally-controlled flow paths to intersect with the Crab Creek deposit. This rudimentary projection method reflects the elevation distance between the knickpoint lip and the Crab Creek deposit as well as the distance between the two features along the hypothesized flow path. Projection of the relict reach of stream 2 is only intended to indicate that the knickpoint lip remains at a higher elevation than the preserved downstream reach of its former channel. The elevation difference along the likely flow path is such that a gradient consistent with, or slightly less than, the gradient immediately above the knickpoint lip in stream 2 would be possible in a channel formerly connecting the relict reach to Crab Creek. The projection thus functions as a general proxy for preservation of the relict features within the transient portions of the Roanoke basin.

Reference

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Chapter 4

Topographic evidence of ongoing transient incision in the upper New River basin, southern Appalachian Mountains

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INTRODUCTION

Characterizing the transient response of drainage networks to perturbations in boundary conditions is essential to developing useful models of landscape evolution (e.g. Davis, 1899a; Leopold and Bull, 1979; Howard, 1994; Whipple and Tucker, 2002). Transient incision is typically associated with rapid base level drop in active tectonic settings, but locally accelerated incision is also known to occur in river systems draining passive margins (Nott et al., 1996; Springer et al., 1997; Granger et al., 2001; Hancock and Kirwan, 2007; Prince et al., 2010, in press). While these studies have indicated that landscapes in tectonically quiescent settings may not evolve through slow, spatially uniform steady-state erosion, their margin-scale relevance has been limited by their focus on short segments of trunk streams or small drainage basins. In order to fully understand the role of transient incision events in shaping passive margin topography, the basin-wide response to documented trunk stream incision must be examined (Crosby and Whipple, 2006).

The New River in southwestern Virginia, which drains the ancient passive margin of eastern North America, is an excellent example of a regional trunk stream known to have undergone episodic Cenozoic incision. Cosmogenic ^{10}Be dating of terraces up to 128 m above modern river level by Ward et al. (2005) confirmed numerous episodes of accelerated New River incision during the past ~2 Myr. Previous studies have also identified widespread fluvial gravel deposits on upland surfaces within the New River basin, providing additional qualitative evidence of trunk stream incision outpacing lowering of the surrounding landscape (Mills, 1986; Bartholomew and Mills, 1991). While these studies have provided useful constraints on the incision history of the New River itself, the response of the drainage network at large has not yet been explored. Dietrich (1959) suggested the New River was actively incising into two discrete low-relief surfaces, herein referred to as the NRV surface, developed on carbonates and shales of the Valley and Ridge, and the Blue Ridge Plateau surface, developed on metamorphic rocks of the western Blue Ridge. Dietrich's work was, however, limited to Floyd County, Virginia, and thus stopped short of characterizing the spread of incision at the basin scale. Documenting the spread of transient incision through the New River as well as its tributaries would offer a unique perspective on the basin-scale migration of transient incision within a passive margin setting.

Transient incision is known to produce a unique topographic signature of elevation-accordant remnant (relict) surfaces, bedrock gorges, and headwardly migrating knickpoints which separate adjusting and relict portions of the drainage network (e.g. Schoenbohm et al., 2004; Clark et al., 2004; Oskin and Burbank, 2007). The spatial distribution of these topographic features can thus serve as an indicator of the headward progress of transient incision, particularly when the potential role of bedrock geology can simultaneously be considered (Fig. 4.1). A 220 km reach of the New River, as well as its tributaries along that reach, were selected as a study area in which to evaluate the basin-wide consequences of documented episodic incision events within a regional trunk stream (Ward et al., 2005). The study area extends from the Brush Mountain Water Gap, immediately downstream of the Ward et al. (2005) terraces, to ~25 km upstream of Fries, Virginia (Fig. 4.2). This study area encompasses the "Great Valley" portion of the NRV surface, underlain by Cambrian shales and carbonates, as well as all of the Blue Ridge Plateau surface within Virginia. Methods of topographic characterization of the streams and topography intended to identify the effects of transient incision within the study area are described below.

METHODS

Knickpoint Identification

The term "knickpoint" classically refers to an abrupt steepening of stream gradient or a local convexity in channel profile (Whipple and Tucker, 1999; Crosby and Whipple, 2006), regardless of the origin of the convexity. As the purpose of this study is to characterize the spread of transient incision events through the New River basin, it is important to identify knickpoints that represent the mobile leading edge of incision events. The methods of channel analysis described below are intended to refine the simple longitudinal profile-based definition of "knickpoint" by also considering the lithology and topographic context associated with a profile convexity. As an increase in channel gradient, and thus a profile convexity, can arise where rivers pass from resistant to weak lithology (Fig. 4.1), analysis of longitudinal profiles alone may not always reflect the progress of transient incision.

Channel length above convexities must also be considered, as Crosby and Whipple (2006) concluded that knickpoints may become effectively immobile within the drainage network once they have migrated to a low threshold drainage area, for which channel length is considered a proxy due to Hack's Law (Hack, 1957). Once knickpoints have migrated to this threshold point in channels, they no longer reflect the presence of actively migrating incision within a basin. This study therefore only considers convexities occurring more than 3 km downstream from channel headwaters. While this is a somewhat arbitary value, it is larger than the ~1.3 km threshold channel length reported by Crosby and Whipple (2006) on very weak rock in New Zealand. While the strength of rock units within this study area relative to those in the Crosby and Whipple (2006) study is uncertain, focusing on knickpoints at higher channel length increases the likelihood of identifying transient features. Following this reasoning, knickpoints occurring at high drainage area (represented by channel length >20 km) are also denoted. These features represent an even greater perturbation to accepted "equilibrium" stream morphology (Hack, 1957; Flint, 1974), and their presence is a robust indicator of actively migrating incision. Accordinlgy, longitudinal profile construction primarily focuses on large tributaries of the New River (Fig. 4.2). Tributaries to these streams were subsequently analyzed to aid in interpretation of convexities in the large streams. Due to the irregular distribution of large New River tributaries on the NRV surface, a number of small New River tributaries were profiled to aid in tracking headward incision within the New River itself (Fig. 4.2).

Knickpoints were identified and plotted using a combination of 1:24,000-scale topographic maps, satellite images, and 10 m resolution DEM hillshade maps (Figs. 4.3, 4.4). Due to the resolution limitation of available maps, a steepened zone within a channel had to cover a total vertical displacement of at least 6.1 m (20 ft) to be identified. After longitudinal profile and 1:24,000-scale topographic map analysis identified convexities occurring at greater than 3 km channel length, convexity positions were plotted on a 10 m DEM hillshade map showing regional geology (Virginia Division of Mineral Resources, 1993) to determine the lithologic context of the convexity location (Fig. 4.5). Convexities occurring where channels pass from resistant lithologies (i.e. quartzite) to weak lithologies (i.e. carbonate, shale) without entering a gorge developed in the weak lithology were excluded, as they did not clearly represent the leading edge of a headwardly migrating incision event (Fig. 4.1). Convexities occurring at the head of a gorge or clearly incised channel, even if it coincided with resistant outcrop, were noted as knickpoints due to topographic evidence of headward migration. Convexities occurring within a uniform lithology were also interpreted as related to transient incision. As the work of Ward et al. (2005) suggests multiple incision events have affected the study area, more than one transient knickpoint could be expected to occur within an already incised reach of channel. Convexities occurring downstream of an identified transient knickpoint were evaluated in the context of lithology as well as the profiles of downstream tributaries to determine whether or not they are transient features (Fig. 4.1).

Satellite images were used to supplement topographic maps in the identification of incising reaches within low-gradient large streams flowing across carbonate strata underlying the NRV surface. The weak mechanical nature of the carbonate substrate produced very broadly convex reaches in larger streams (streams 25 and 26; Figs. 4.2, 4.5), where gradient gradually increase by a factor of only 2 to 2.5, compared with the ~4-fold increase typically observed in knickpoints developed in small carbonate streams draining the NRV surface and all streams of the Blue Ridge Plateau. If the large carbonate streams showed extensive bedrock exposure limited to the convex zone along with downstream gorge deepening and hanging tributaries, the broad convexity was classified as a transient knickpoint.

Relict Surface Identification

Following knickpoint identification, a shaded relief map color-coded by elevation (Fig. 4.6) was prepared for the purpose of delineating the topographic effects of incision into the NRV and Blue Ridge Plateau surfaces. As the New River is the regional master stream whose incision sets temporary base levels for its tributaries, elevation map color ramp was adjusted so that color changes in the New River channel corresponded to a major New River knickpoint (Fig. 4.6). The distribution of the downstream color at any New River color break generally represents the extent of headward migration of a temporary base level, represented as an elevation range to account for stream gradient. DEM color-coded elevation mapping is a useful reference to illustrate the progress of dissection of the low-relief NRV and Blue Ridge Plateau surfaces, represented by light blue and brown, respectively. It must be noted that the map functions on the basis of contrast between adjacent color domains within the respective NRV and Blue Ridge Plateau surfaces. For example, light blue gorge coloring on the Blue Ridge Plateau implies incision into the brown domain, not correlation with the NRV surface. This is a product of the limitation of the ArcGIS software, and is unavoidable if the NRV and Blue Ridge Plateau surfaces are to be illustrated together as part of the New River basin.

Relict topographic domains constituting the NRV and Blue Ridge Plateau surfaces were delineated as portions of the landscape occurring at elevations above, or upstream of, the clear break between steep, narrow gorges and aggraded stream valleys and gentle hillslopes (Fig. 4.6). As knickpoints at gorgeheads communicate a new base level to the landscape, relict topography represents stream reaches and land surface which continue to evolve under pre- incision boundary conditions. The term “relict” implies that these reaches and the associated portions of the drainage basin have not yet been affected by the incision events apparent within the study area, but does not imply a constrained age or static preservation. “Relict” is used in transient landscape literature to describe low(er)-relief topographic and fluvial domains beyond the uppermost knickpoints in a transient basin (*e.g.* Schoenbohm et al., 2004; Clark et al., 2005; Crosby and Whipple, 2006; Berlin and Anderson, 2009). Interpretation is generally qualtitative, and relies primarily upon obvious topographic contrasts, knickpoint position and the presence of alluvial channels within relict domains versus bedrock channels within, and potentially below, knickzones. While channels may flatten and become alluvial between knickzones, these reaches are not classified as relict as they flow through gorges with oversteepened walls and hanging tributaries which should continue to adjust to the new temporary base level. The use of “relict” in this study is intended to maintain consistency with the existing body of transient landscape literature, where it has become an accepted descriptor of pre-incision landforms which will not respond to the new base level until reached by knickpoints and thus new boundary conditions.

The color-coded hillshade elevation map of the study area was also used to identify perched meander cutoffs analogous to the San Juan River “Loop” of Leopold and Bull (1979). These features are apparent on DEM-derived maps, but are often difficult to resolve on topographic maps or with satellite images due to varied land use and, frequently, minor incision into the hosting relict surface. Perched cutoffs are excellent indicators of spatially mobile incision as well as former temporary base levels. Their preservation can also provide evidence of the spatial variability of incision within transient basins. As incision is rapidly within large active channels but much more slowly through very small basins, “Loop”-type features which no longer carry a large drainage can remain intact even when perched tens of meters above an incised active stream.

RESULTS

The unique signature of bedrock gorges, elevation-accordant remnant surfaces, and knickpoints above tributary junctions common to transient landscapes (e.g. Schoehnbohm et al., 2004; Clark et al., 2006; Oskin and Burbank, 2007) are readily apparent within the studied portion of the New River basin. Four knickpoints were identified in the New River between the

Brush Mountain water gap and the North Carolina-Virginia border. 118 knickpoints were identified in tributary basins throughout the study area (Figs. 4.3-4.10), with many knickpoints occurring in large streams at high drainage area (Fig. 4.3). Tributaries entering the New River in the farthest downstream portion of the study area contain more knickpoints than upstream tributaries, consistent with their outlets having been passed by headwardly migrating New River knickpoints which have not yet entered the upper reaches of the study area (Fig. 4.7). With increasing distance south and west from the Brush Mountain Water Gap, incision into the relict portions of the NRV and Blue Ridge Plateau surfaces decreases in depth and distance migrated away from the New River (Figs. 4.6, 4.10). This distribution of incised and relict domains is consistent with actively migrating incision within the study area.

A fifth New River knickpoint that is now obscured by the waters of Claytor Lake was identified by comparing knickpoint distribution and relative incision into the NRV surface in streams entering the New River above and below present-day Claytor Lake (Note: the lake is mentioned only as a reference location, and obviously has played no role in shaping the topography described) (Fig. 4.10). New River tributaries entering the river below Claytor Lake show the effects of the now-submerged knickpoint, while tributaries entering above the lake do not. This contrast is most strongly reflected in the major knickpoint of the Little River, which has no analog in upstream New River tributaries (Figs. 4.11-4.13). The existence of a New River knickpoint now obscured by Claytor Lake was confirmed in historical documents relating to navigation of the New River by barges carrying iron ore (Trout, 2003).

Six perched meander cutoffs were identified in the study area (Figs. 4.14-4.20). These features were apparent on the color-coded elevation map, and are also discernible in satellite images. Loop elevation above the modern stream ranged from 40 m in the northeasternmost (farthest downstream within the drainage network) loop (Little River, stream 10) to ~3 m in the southwesternmost (farthest upstream) feature (Chestnut Creek, stream 30) (Fig. 4.14). Field reconnaissance to the Big Reed Island Creek (stream 19; Fig. 4.17) loop was hindered by property access issues, but the presence of alluvium in the shallow subsurface was clear in roadcuts and shallow excavations. A well-defined terrace along Big Reed Island Creek appears to continue into the loop (Fig. 4.17). No other loop features have yet been visited due to difficulty of access and property owner issues.

Distribution of color fields on the DEM map indicates that the relict NRV surface slopes gently to the northeast, decreasing ~50 m in elevation along the trend of the New River (Figs. 4.6). The river itself drops ~150 m across the same distance, increasing its depth of incision in the NRV surface from ~30 m where it enters the Valley and Ridge province to ~130 m at the Brush Mountain water gap (Fig. 4.6). The Blue Ridge Plateau surface also slopes gently towards the New River, and large drainages on the plateau show the same downstream increase in incision depth. Color distribution is imperfect due to the slightly lower elevation of the Plateau within the Little River basin, but color field contrasts remains an effective means of characterizing the distribution of erosional regimes within the landscape. Along the courses of the large Plateau drainages, the onset of increasing gorge depth associated with major knickpoints is apparent in the close spacing of color fields within channels (Fig. 4.6). This trend is also discernible in NRV surface streams, although the tendency of carbonate bedrock knickpoints to "leak" incision above the knickpoint lip (Frankel and Pazzaglia, 2007; Berlin and Anderson, 2009) has allowed channels to become somewhat steepened throughout their entire length, partially obscuring the break between relict and incised reaches.

DISCUSSION

Knickpoint Distribution

The widespread occurrence of knickpoints above tributary mouths and evidence of progressing headward migration of incision into an extensive relict surface indicate that New River Valley topography is in a state of transient adjustment to an episodically lowering base level. A number of knickpoints occur within a single lithology (Fig. 4.5), demonstrating that stream gradient is not controlled by bedrock lithology alone (Hack, 1957, 1973). Some knickpoints have localized where channels pass from resistant to weak lithologies (i.e.streams 10 and 19), but these knickpoints are always coincident with gorgeheads, implying headward retreat of a mobile incision front has slowed atop resistant outcrop (Fig. 4.5). Tributaries downstream of lithologic contact knickpoints also contain knickpoints developed entirely within weak rock (i.e. streams 11 and 12, entering stream 10; Fig. 4.10), further suggesting the role of transient incision in knickpoint location. Relict surfaces are decreasingly dissected away from the New River and its largest tributaries, providing an additional qualitative indicator for the headward progress of incision throughout the drainage network (Fig. 4.6).

Knickpoint distribution in drainages of varying size suggests the New River basin is still actively adjusting to incision. All large New River tributaries (> 25 km channel length) cross knickpoints many kilometers downstream of their headwaters, with Chestnut and Crooked Creeks (streams 28 and 30, respectively) actually entering the New River through major knickpoints developed within a single, non-resistant lithology (Fig. 4.5, 4.9B). As accepted models of knickpoint retreat suggest drainage area exerts strong control over retreat rate (e.g. Howard and Kirby, 1983; Whipple and Tucker, 1999), the presence of these confluence knickpoints in large streams suggest they have recently begun to respond to New River incision. This interpretation is supported by the location of the mouths of Crooked and Chestnut Creeks immediately below the uppermost New River knickpoint (Fig. 4.9). The presence or absence of knickpoints in smaller streams in the study area appears to be controlled by the relative position of a knickpoint, or knickpoints, in their master stream (see stream 10 and tributaries, Fig. 4.10). The locations of small tributary (<10 km channel length) knickpoints (Fig. 4.7) suggests that the initiation of these features "follows" large stream knickpoints as they migrate headwardly. Observed distribution of knickpoints does not appear to be related to slowed retreat at low drainage areas, as described by Crosby and Whipple (2006). Instead, the location of a stream's mouth relative to the present location of major New River knickpoints seems to exert the strongest control on the presence, or absence, of knickpoints within its channel (Fig. 4.3; 4.5).

Analysis of the number and location of knickpoints in New River tributaries can be compared to the profile of the New River itself to indicate the progress of multiple incision events throughout the basin. This topographic evidence is consistent with the findings of Ward et al. (2005), which suggest late Cenozoic New River incision has been episodic. At present, the New River profile shows five knickpoints along the 220 km of channel studied. Streams entering the New River below Claytor Lake (Fig. 4.7) all contain multiple knickpoints, appearing to reflect the passage of many discrete incision events. These streams are incised up to ~100 m into the NRV surface (Fig. 4.6). Streams draining the NRV surface to the New River upstream of the present location of Claytor Lake show less complex incision histories, reflecting the ongoing headward migration of knickpoints through the New River system (Fig. 4.3). Streams draining the Blue Ridge Plateau show the same trend, with dissection of the Plateau decreasing in catchments entering the New River further upstream (Fig. 4.6).

This incision pattern is clearly illustrated by comparing the Big Reed Island Creek (stream 19) system with Crooked and Chestnut Creeks (streams 28 and 30), which enter the New River 42 km upstream (Fig. 4.9). The mouths of Crooked and Chestnut Creeks are also separated from the Big Reed mouth by two major New River knickpoints, whose passage has affected the Big Reed system but is yet to reach the Crooked and Chestnut basins. Big Reed has a major knickpoint ~60 km above its mouth at the head of an extensive gorge system, with another smaller knickpoint occurring ~17 km above its mouth. Chestnut and Crooked Creek contain major knickpoints ~5 km above their mouths, with very small knickpoints ~ 25 km above their mouths (Figs. 4.7-4.9). The gorges through which Crooked and Chestnut Creek flow are also not cut as deeply into the Plateau surface, whose elevation is consistent between the Big Reed and Chestnut/Crooked systems (Fig. 4.6). Although Big Reed Island Creek is larger than Chestnut or Crooked, it flows across much more resistant lithologies which would certainly slow the headward migration of incision. It also seems unlikely that incision in the Big Reed system could travel an order of magnitude faster than in Chestnut or Crooked if incision began simultaneously in all three streams. The contrasts between these basins therefore probably reflect, at least to some degree, the spatially transient character of New River incision. The morphology and incision history of streams thus depends on whether or not they have yet "seen" pulses of incision migrating through the drainage network.

Additional evidence for incision progressively migrating through the New River and into its tributaries is seen in differences between the profiles and NRV surface incision of the Little River (stream 10) and Big Reed Island Creek (stream 19), which flow across the same series of bedrock units (Fig. 4.5). Both drainages have a knickpoint located in shcist underlying the Blue Ridge Plateau >60 km above their mouth, but the Little River crosses a major knickpoint 33 km above its mouth as it passes from metamorphic rocks of the Blue Ridge into dolomites of the Valley and Ridge (Figs. 4.5, 4.7-4.10). Big Reed Island Creek crosses a comparatively modest knickzone as it enters the Valley and Ridge, even though it is developed on very strong bedrock (Figs. 4.8-4.10, B of Fig. 4.13) at a much smaller drainage area than the lower Little River knickzone. The Little River passes over additional minor knickpoints through a deep gorge developed entirely within dolomite before entering the New River (Figs. 4.5, 4.7-4.10). These features are also absent from the Big Reed channel. The additional knickpoints in the Little River are likely correlative to a series of rapids now under Claytor Lake, which are referenced in historical documents describing navigation of the New River by bateaux carrying iron ore. These rapids separate the temporary base level to which Big Reed Island Creek is connected from the lower base level to which the Little River presently connects (Fig. 4.10). Big Reed Island Creek is thus yet to experience an incision event that is now active in the Little River basin as well as in New River tributaries further downstream.

The significance of the major Little River knickpoint in understanding the spread of incision through the study area is further reflected in the lithology that supports the falls (Fig. 4.13). Field reconnaissance revealed that the steepest reach within the knickpoint is unexpectedly developed on weak phyllites of the Chilhowee and Erwin/Hampton formations, which vary locally between resistant meta-quartzarenite and more sheet silicate-rich compositions (Figs. 4.11, 4.13). The course of the river has likely migrated to its present location in order to cross from the Blue Ridge to the Valley and Ridge across the least resistant route possible, but the presence of an oversteepened reach atop the weak rock is an excellent indicator of actively migrating incision within the Little River channel. The phyllites exposed at Bear Falls (Fig. 4.11) are chlorite rich, and should only temporary support a steep gradient in stream the size of the Little River. Further upstream, the river passes from quartz-rich granitoid gneiss onto weak sheet silicate-rich rock across a ledge with a comparatively modest step in channel elevation (Fig. 4.12). The strength of the gneiss unit relative to the phyllite is apparent in the morphology of the individual rapids in the channel. Rapids take on a “chute” or “flume” morphology through the phyllite, whose weakness permits the channel to narrow in response to steepening. The strong and widely-jointed gneiss forces the channel to drop over a river-wide ledge where flow passes through openings produced by plucking of large blocks. This knickpoint morphology is also apparent in the lower knickpoint of Big Reed Island Creek (C of Fig. 4.13). When the incision pulse that is presently located just downstream reaches this resistant ledge, a large and impressive falls should form.

Meander Cutoffs

Perched meander cutoff loops provide further evidence of migrating episodic incision focused in large stream valleys (see Leopold and Bull, 1978; Crosby and Whipple, 2006) (Figs. 4.14-4.20). These features furnish a concrete constraint on a former temporary base level, and their preservation at varying elevation above modern channels offers an excellent physical indication of transient incision in the New River basin. While each of the six loops identified all appear to contain some amount of completely intact surface, progressive dissection can be seen spreading into the features from their outlets into the gorges of the modern channels. All loops are somewhat incised into the surrounding landscape, and small streams entering and passing through the loops cross knickpoints into the loops, follow flatter courses on the loop surface, and then drop again to the modern stream. This knickpoint distribution indicates incision was active prior to cutoff, and has continued. While the relative ages of abandonment of the loops is unclear, three loops located near one another in the Big Reed Island Creek (stream 19) system (Figs. 4.16-4.18) all occur at comparable distances (~15 m) above the associated active channel. This is anecdotal evidence that the loops carried active channels during the same period in the base level "history," and may have been cut off due to arrival of the same pulse of incision. While correlating the abandonment of loops throughout the basin is difficult, the intact portions of the features do have significance as "time capsule" landforms which should preserve an ancient channel sequence. Analysis of material preserved in these paleo-channels may be useful in constraining their age as well as climatic conditions operating at the time of abandonment.

Timing and Rate of Migration of Incision

Hypothetical correlation of knickpoints and meander cutoffs with the dated gravels and straths of Ward et al. (2005) may offer some crude constraint on the tempo of the transient incision events. If the elevation of the 50 m terrace of Ward et al. (2005) is considered in the context of knickpoints in the New River upstream of Kentland Farms, the Little River loop (40 m above modern river level) (Figs. 4.14, 4.15) may be a relict of the same temporary base level that formed the 50 m Kentland terrace. If this is the case, then the incision pulse that caused abandonment of the Kentland terrace has spread at least as far as the Little River loop and may presently exist as the rapids submerged under Claytor Lake. According to Ward et al. (2005), the 50 m Kentland terrace was abandoned at 955 ka. If this age is accurate and the Kentland terrace and Little River loop were stranded by the same headwardly-migrating pulse, the wave of incision has passed through 60-70 km of carbonate –floored New River channel in 1 Myr. This is an exceedingly rapid and highly speculative rate, and it does reflect uncertain paleo-surface correlations and an "end member" model of knickpoint migration where these features are robust boundaries between old and new base levels. The carbonate bedrock between Kentland and the Little River loop is weak, and incision would be expected to "leak" above knickzones and communicate some incision upstream of the steepest reach. It is clear, however, that the Little

River has experienced incision that tributaries above the reach of river covered by Claytor Lake have not, suggesting a spatially mobile incision front (Fig. 4.10). Even without robust dates, the preservation of low-relief, gravel-capped relict surfaces adjacent to deeply incised bedrock gorges suggests the migration of incision pulses is rapid, the relict surfaces are exceedingly stable, or both.

If the spread of incision observed does correlate with abandonment of any of the higher Ward et al. (2005) terraces, the material removed by active downcutting must represent a major sediment pulse to the Mississippi basin. Inspection of the New River and its tributaries below the study area confirms that incision has indeed spread headwardly throughout the basin from its present mouth at the Ohio River, and additional knickpoints are still migrating towards the study area. The effects of transient incision have been documented elsewhere in the Ohio basin both above (Springer et al., 1997) and below (Granger et al., 2001) the New River, suggesting the entire Ohio River system has been delivering sediment at an accelerated rate during the late Cenozoic. While it is uncertain whether episodic incision has characterized the entire Neogene (or older) history of Appalachian drainages, the widespread distribution of gravels on the NRV surface along with its low relief suggest it may have developed during a period of base level stability. As the highest terrace of Ward et al. (2005) appears to be associated with the greater NRV surface, its abandonment may indicate the onset of repeated relative landward base level drop and a resulting period of incision that is still active. Glacial rearrangement of the Ohio basin could provide a driver for regional base level fluctuation, but more robust constraint on the timing of incision is necessary. Inspection of the Gulf of Mexico sedimentary record could provide further insight into the timing of the incision events observed relative to the long term sediment flux from landward-draining Appalachian River systems.

Origin of Base Level Perturbation

While the effects of transient incision in the New River basin are apparent, the origins of the NRV and Blue Ridge Plateau surfaces and the episodic incision events dissecting them are unclear. The widespread distribution of gravel deposits on the NRV surface suggest the river meandered very broadly at previous, higher base levels, while its present entrenched course appears to have remained comparatively stationary following the onset of the incision events. Gravel deposits are yet to be documented on the Blue Ridge, but at least two gravel deposits that appear to reflect broad meandering of rivers across the surface are known to the author (Fig. 3).

As with the New River itself, Blue Ridge Plateau tributaries appear to have generally retained their courses since the onset of the recent incision events (excluding meander cutoffs) into the Plateau surface. Climate fluctuation related to Milankovich cycles was suggested by Ward et al. (2005) as the driver of New River incision, and, indeed, the periglacial climate of the last ice age could certainly have fueled accelerated incision. Rearrangement of the paleo-New River (Teays) basin by glacial damming in Ohio also appears to have shortened the system's path to ultimate base level, potentially initiating incision events at the point of rearrangement near Chillicothe, Ohio (Fig. 4.21). While constraining minimum age of the highest gravels on the NRV surface would provide the most robust constraint on the timing of the onset of incision, it is unlikely that a good result can be obtained. Ward et al. (2005) obtained a rough age of ~ 2 Ma for gravel 128 m above the modern New River at Kentland Farms, just upstream of the Brush Mountain water gap. The elevation and weathering of this deposit suggests probable correlation with the regionally extensive NRV surface, suggesting its dissection is late Neogene and could indeed reflect climatic fluctuation.

CONCLUSIONS

The distribution of knickpoints and relict surfaces within the studied portion of the New River basin suggest that numerous episodes of transient incision are presently migrating through the drainage network. Incised domains that appear to be "spreading" away from the New River through its tributaries reflect the basin-scale consequences of trunk stream incision events documented by Ward et al. (2005), particularly the episodic delivery of large amounts of sediment to the Mississippi basin. The presence of knickpoints developed in weak rocks in large streams suggests incision is actively migrating through the drainage network, and will continue to dissect relict portions of the New River basin. Relief should therefore be expected to increase in presently intact relict areas as incision continues to migrate headward. Inspection of the New River channel downstream of the study area indicates additional knickpoints migrating towards the study area which will inititate future episodes of transient incision.

While the actively incising stream reaches identified are probably quite dynamic, they are frequently juxtaposed against very stable relict landforms covered with alluvium which was deposited when the regional base level was higher than present. This rapid transition from stable to disequilibrated landforms and channels indicates that erosional dynamics are highly spatially variable within transient landscapes. The dominant erosional processes at a given location are thus stronlgy controlled by position within the New River drainage network rather than background uplift rates or lithology. Many unincised relict landforms, particularly perched meander cutoffs, are exceedingly well preserved. Such features have the potential to provide data regarding the timing of their abandonment, as well as prevaling climatic conditions at the time of their deposition. These features also provide an exceptional indicator of spatially and temporally transient incision, as some cutoffs are beginning to experience dissection at their outlets while the rest of the feature still "awaits" the spreading incision.

The transient condition of New River basin suggests that large portions of the Appalachians do not presently exist in dynamic equilibrium (Hack, 1960). While the study area does not show evidence of rapid surface uplift, episodic drops in landward base level have been sufficiently large and rapid to disequilibrate the New River system and initiate waves of incision. Although transient landscapes are most frequently associated with active tectonic settings, this result indicates that (presumably) non-tectonic passive margin base level perturbations can occur at rates and magnitudes sufficient to force fluvial and topographic adjustment. If passive margin river systems can be decoupled from base level by relatively minor perturbation, it is likely that they never achieve a true equilibrium with base level and lithology. However, the great stability of relict surfaces reported by Ward et al. (2005) may combine with the spatially mobile rapid erosion within transient reaches to produce long-term "average" rates of incision that are consistent with the dynamic equilibrium model (Hack, 1960). The time scale over which averaged disequlibria produce an apparent uniform lowering rate is unclear, but analysis of sediments within the Mississippi basin may aid in constraining long-term Appalachian incision history. Additional study of the New River system, as well as other transient landward-draining Appalachian basins, may enhance understanding of the controls on passive margin topographic evolution following the cessation of major tectonic activity.

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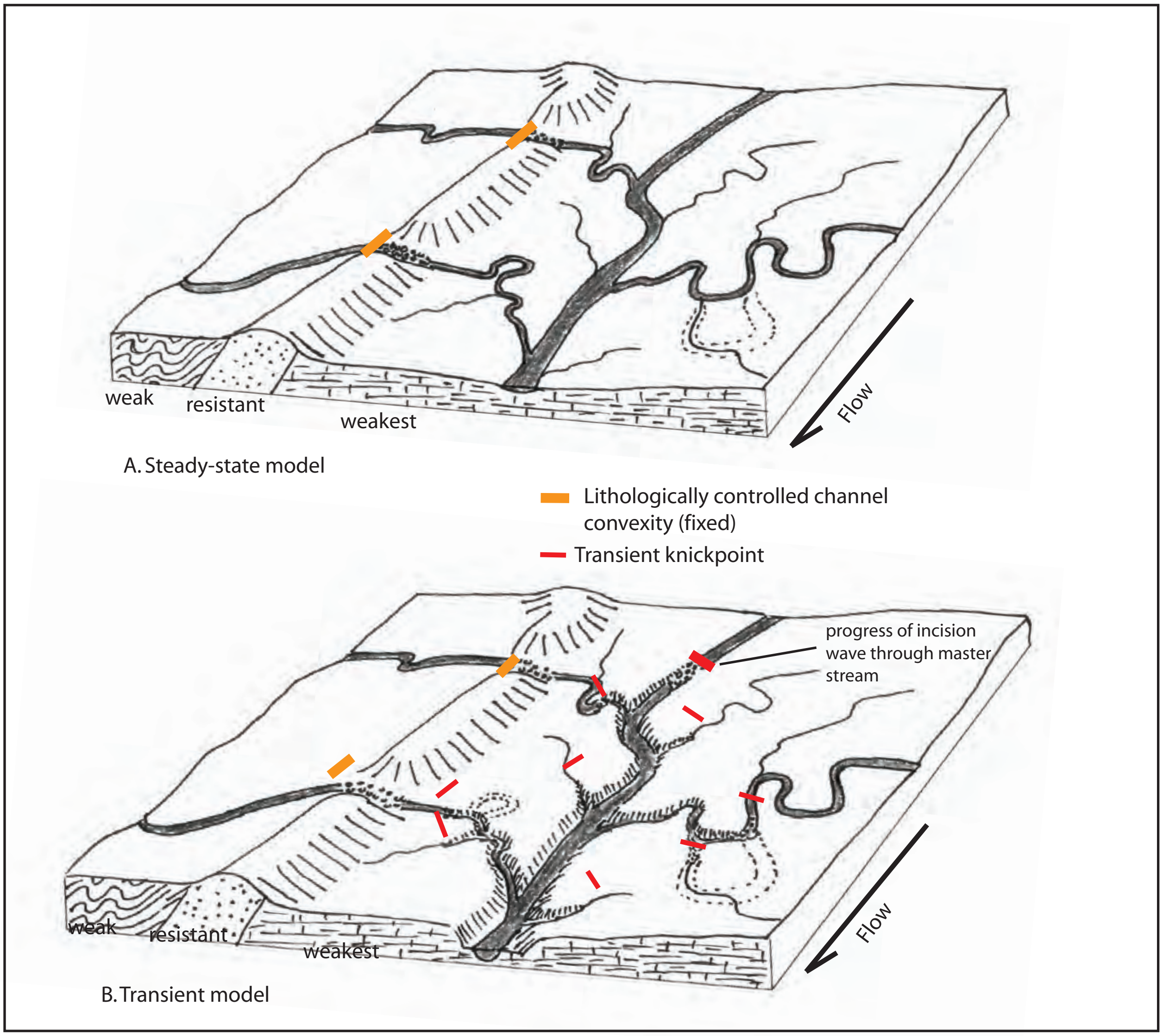
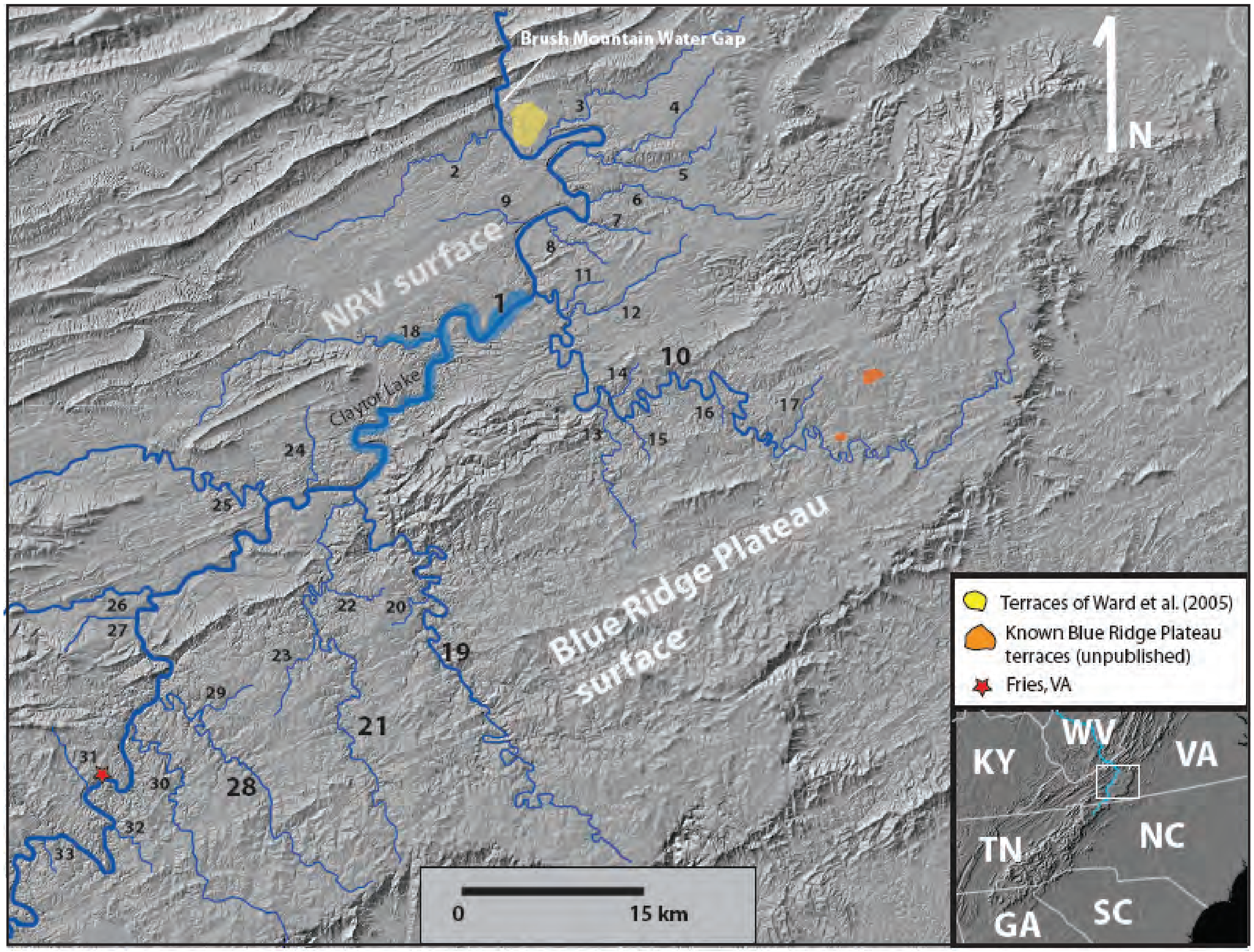
Figure 4.1: Conceptual illustrations of steady-state and transient landscapes.



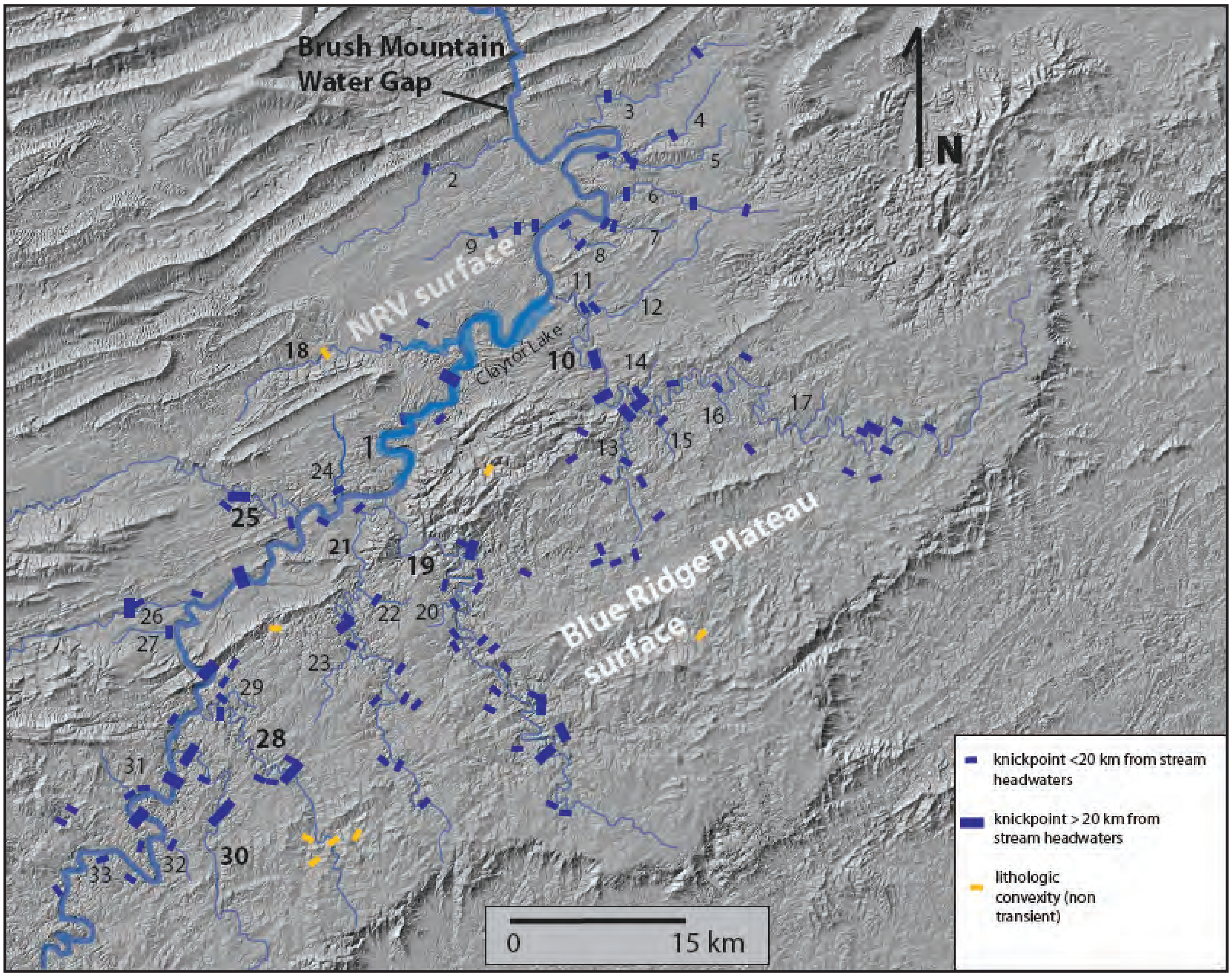
Cartoon illustration comparing knickpoint distribution and topography associated with steady-state, or dynamic equilibrium (Hack, 1960) (A), and transient (B) landscape evolution. In the steady state model, stream gradient and topography are entirely controlled by lithology. Resistant strata support steep stream gradients, and channel convexities arise when streams pass from resistant to weak bedrock. These channel convexities are spatially and temporally fixed, and would not be considered "knickpoints" within this study. In the transient model, knickpoints represent the spatially mobile leading edge of adjustment to a drop in local base level as dictated by the regional master stream. As the master stream incises rapidly, tributaries must also incise to adjust their channel elevation and gradient to the new local base leve Knickpoints represent the boundary between the upstream "relict" portions of the landscape, which continue to evolve under pre-incision boundary conditions, and the adjusted downstream domains. Because knickpoints spread through the drainage network, they can occur in weak or strong rock depending on where the front of incision is located at a given time. The pattern of knickpoint distribution is most strongly controlled by the progress of knickpoints through large streams, not by lithology.

Figure 4.2: Map of the New River basin in southwest Virginia.



10 m DEM with hillshade topography showing streams whose longitudinal profiles were constructed for the study. Inset shows regional setting. Following profile construction, tributaries of these streams were inspected for knickpoints using 1:24,000 scale topographic maps and satellite images. 1. New River 2. Back Creek 3. Toms Creek 4. Stroubles Creek 5. Slate Branch 6. Crab Creek 7. Plum Branch 8. Connollys Run 9. Falls Branch 10. Little River 11. "Loop" branch 12. Meadow Creek 13. Big Indian Creek14. Big Branch 15. Lost Bent Creek 16. Little Camp Creek 17. Camp Creek 18. Peak Creek (omitted due to Claytor Lake backwater 19. Big Reed Island Creek 20. Unnamed tributary 21. Little Reed Island Creek 22. Rock Creek 23. Mill Creek 24. Pine Run 25. Reed Creek (omitted due to overlap) 26. Cripple Creek 27. Ivanhoe Branch 28. Crooked Creek 29. Staunton Branch 30. Chestnut Creek 31. Stevens Creek 32. Oglesby Branch 33. Moore Creek. DEM source: www. seamless.usgs.gov.

Figure 4.3: Knickpoint distribution in the upper New River basin.

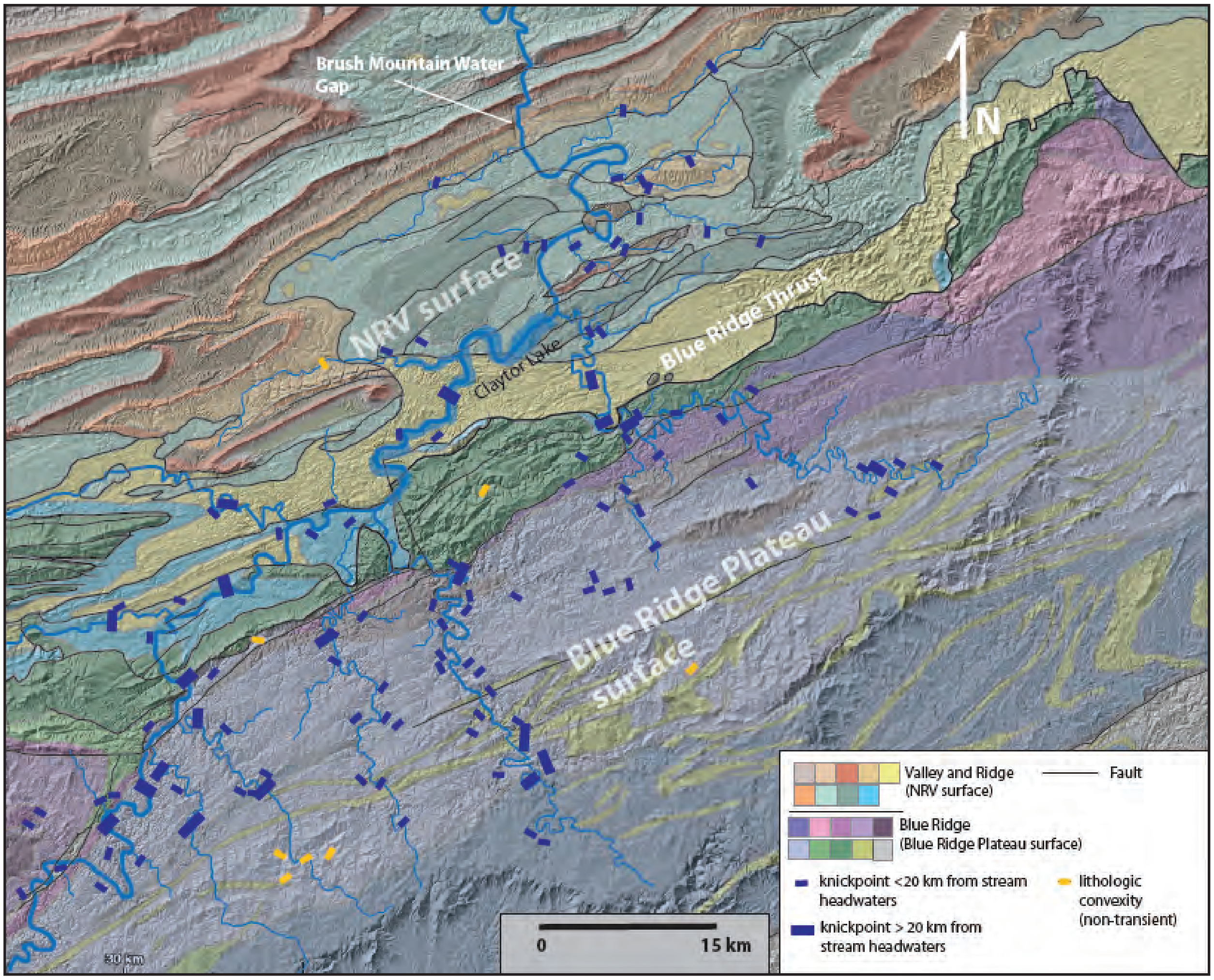


10 m DEM with hillshade topography showing knickpoint and lithologic convexity distribution within the study area. Knickpoints occurring at high drainage area, represented by heavy hash marks, are robust indicators of actively migrating incision (see Crosby and Whipple, 2006). DEM source: www.seamless.usgs.gov.

Figure 4.4: Satellite image indicating knickzone in weak carbonate rock.

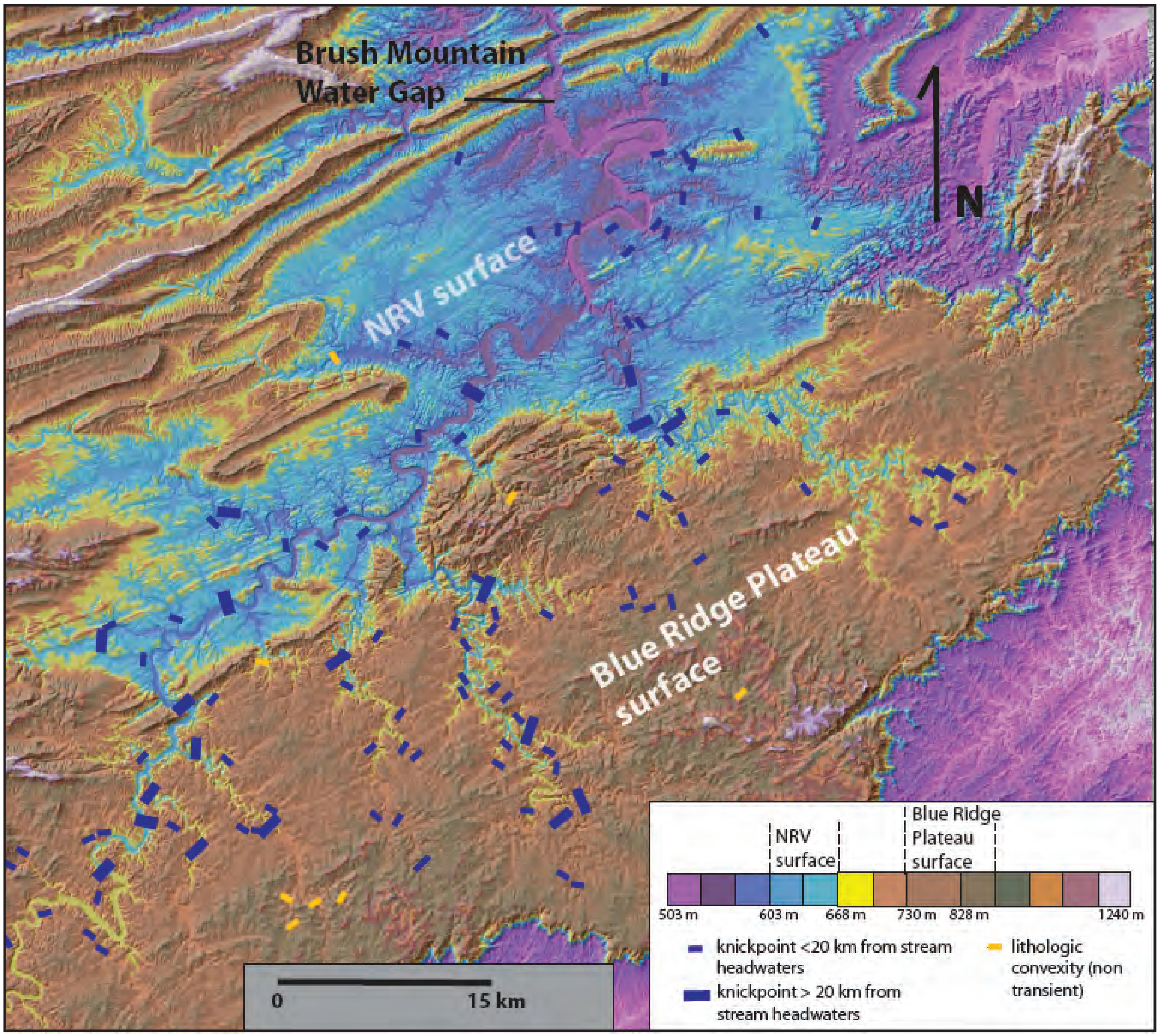
Google Earth satellite image of incising bedrock reach in Reed Creek (25; Fig. 4.2). This knickzone is developed entirely within Cambrian dolomites of the eastern Valley and Ridge. This incising reach is difficult to identify from the longitudinal profile due to the modest increase in gradient it represents; many other streams of similar size show comparable gradient increase as a product of apparent mechanical contrast in bedrock units. The rapids developed across bedrock ledges seen here are almost certainly not the product of mechanical properties, as no rapids occur upstream where the stream meanders across the same unit. This image indicates the usefulness of satellite imagery as a supplement to interpreting longitudinal profiles. Image source: Google Earth, image date Feb. 1, 2007. ©2010 Google-data ©2011 Commonwealth of Virginia.

Figure 4.5: Map of knickpoint locations relative to bedrock geology.



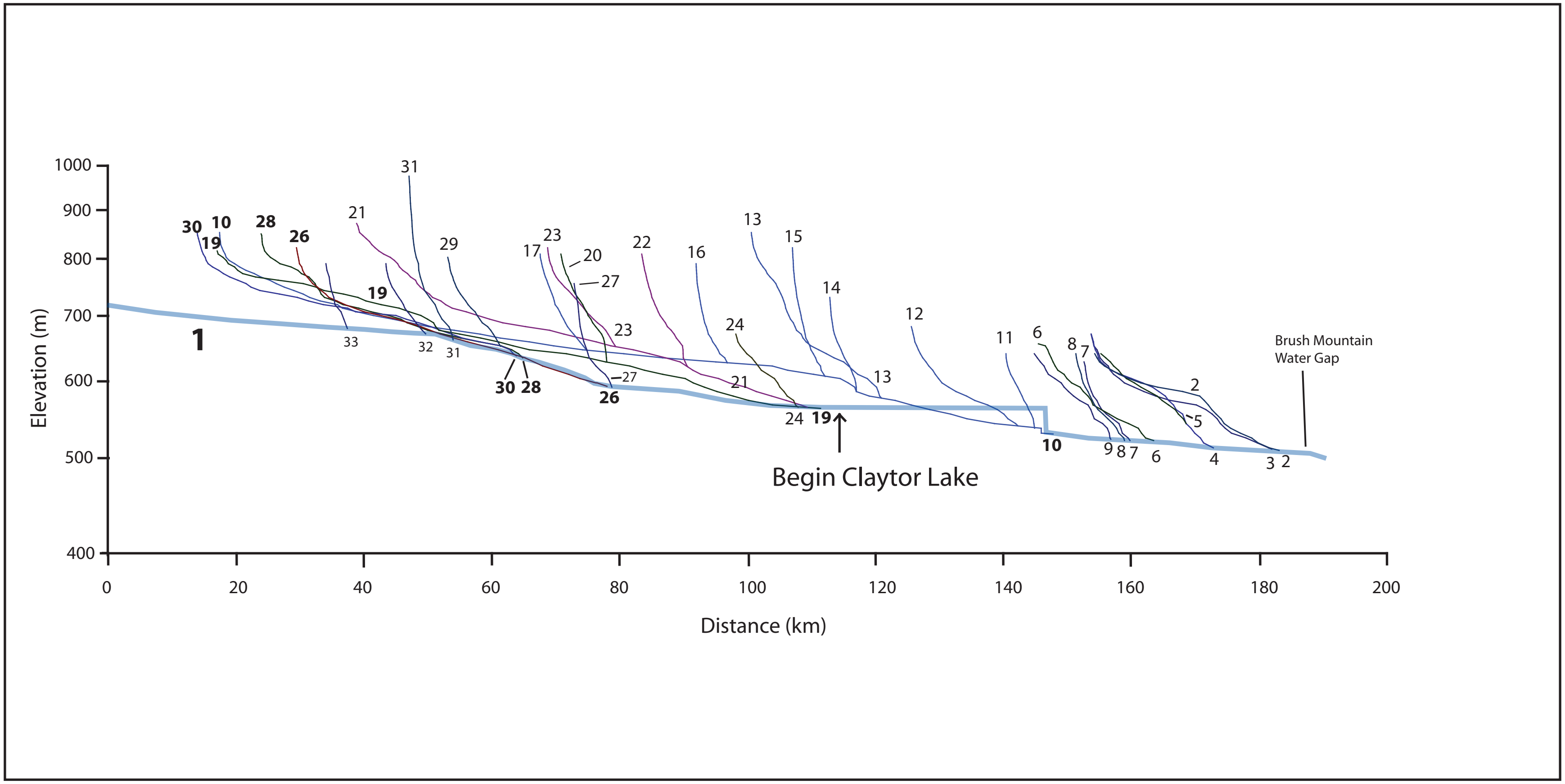
Bedrock geology (Virginia Division of Mineral Resources, 1993) superimposed onto 10 m DEM hillshade topography (www.seamless.usgs.gov). Note that two uppermost New River knickpoints occur within a uniform lithology upstream of outcrop of Blue Ridge quartzites (forest green; immediately southeast of Blue Ridge Thrust). Although some channel convexities appear to originate entirely from lithologic contrast, the position of >90% of knickpoints suggests headward migration of incision throughout the basin. DEM source: www.seamless.usgs.gov.

Figure 4.6: Colored elevation map showing knickpoint locations.



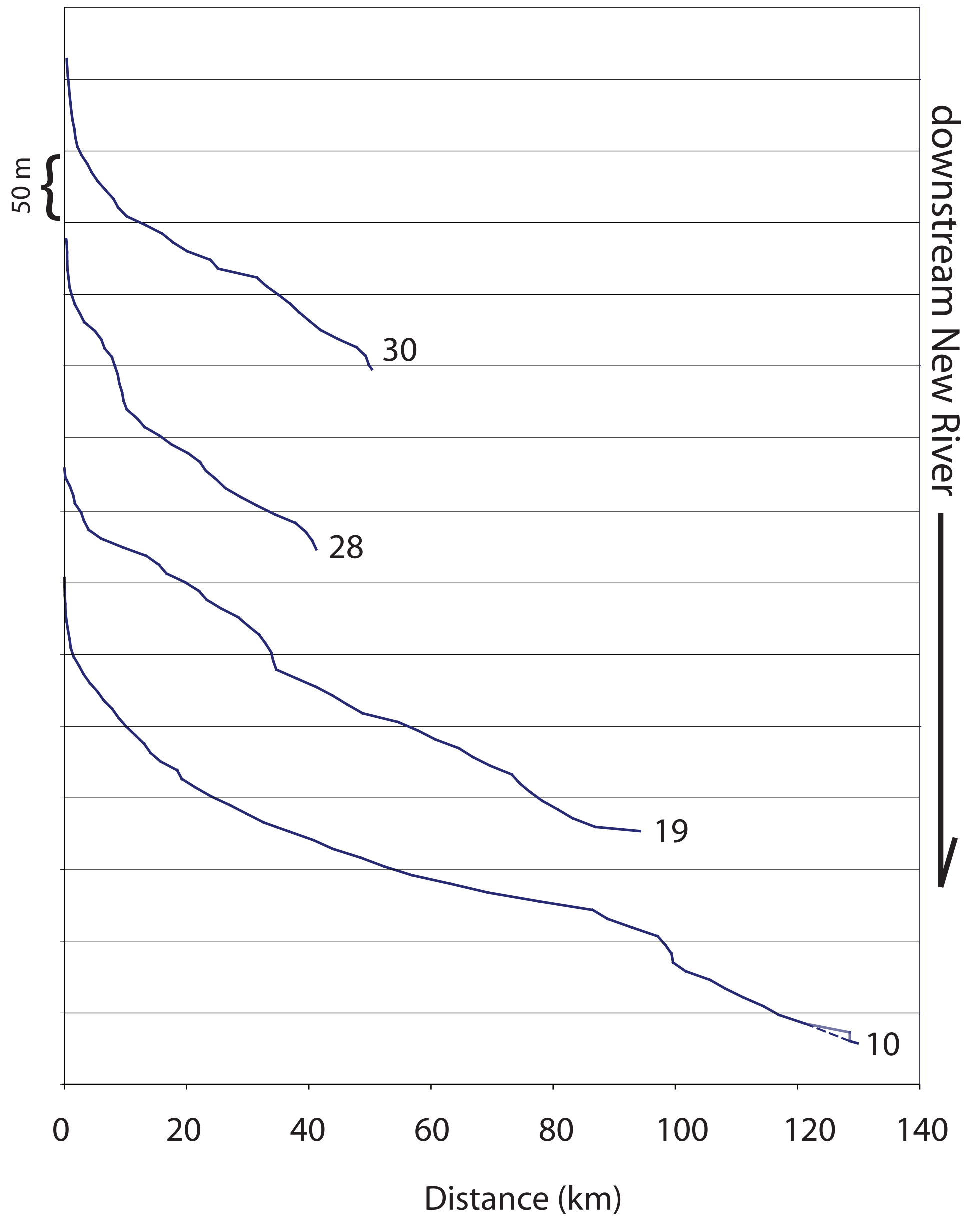
Knickpoint locations superimposed on color-coded elevation 10 m DEM with hillshade. 730-828 m color interval is adjusted to reflect the extent of the low-relief relict Blue Ridge Plateau surface, which hosts slightly greater relief between uppermost knickpoints and lithologically controlled topographic highs (lavender) than than the NRV surface. DEM source: www.seamless.usgs.gov.

Figure 4.7: Integrated stream profiles of the upper New River basin.



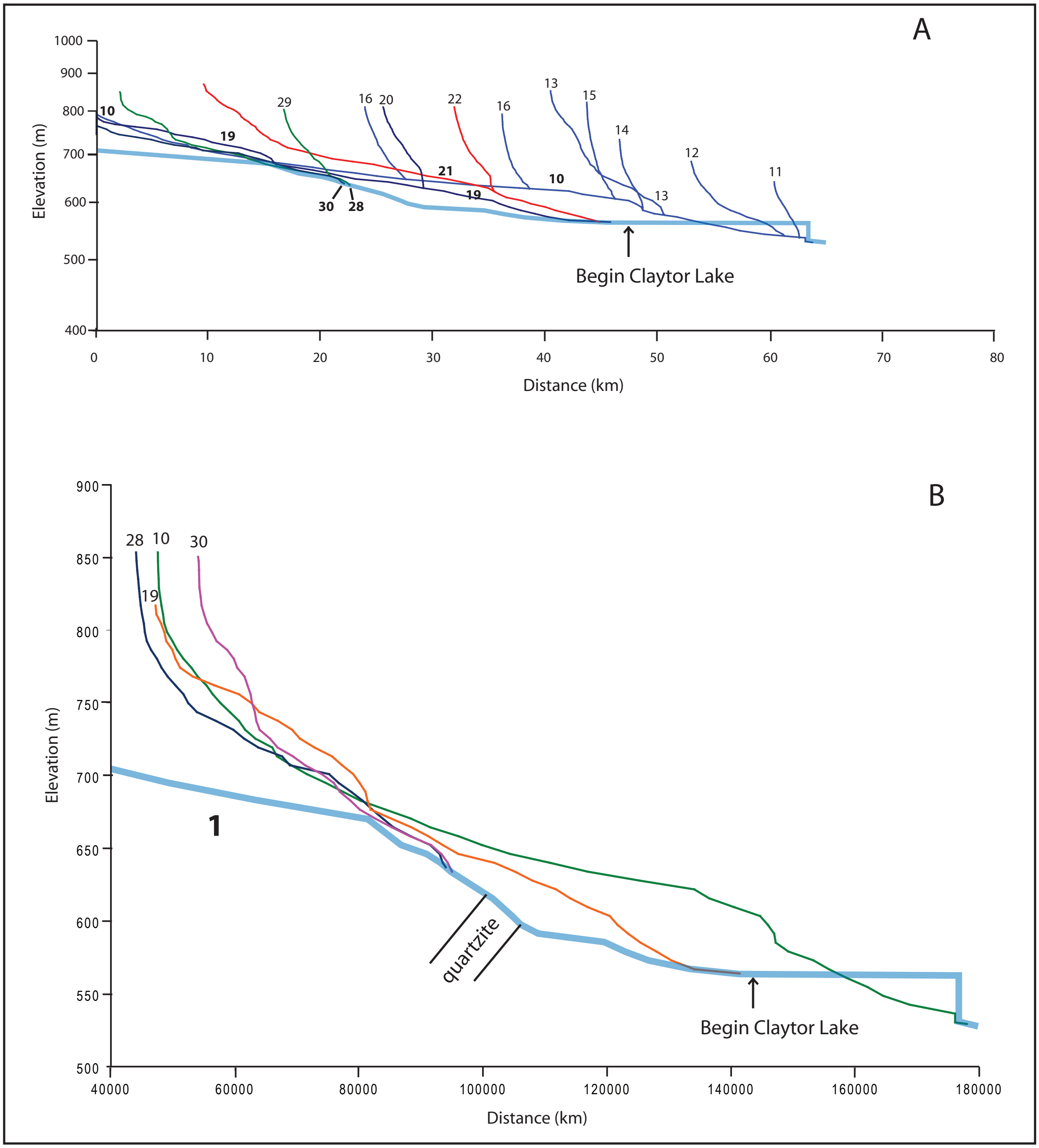
Stream profiles from the study area. Profiles are positioned based on the location of their confluence with the New River. See Fig. 4.2 for locations of streams. Elevation is represented on a logarithmic scale to account for the extreme gradient contrast between the New River and its smaller tributaries. Peak Creek (18) profile was omitted as much of it is obscured by the waters of Claytor Lake; Reed Creek (25) is not plotted as it is almost entirely overlapped by the New River profile and cannot be clearly seen. This overlap is, however, evidence of oversteepening within New River knickzone below the mouth of stream 26; the gradient of the New River should, by Flint's Law (Flint, 1974), be much lower than that of the much smaller Reed Creek.

Figure 4.8: Longitudinal profiles of major New River tributaries.



Profiles of four major Blue Ridge Plateau tributaries of the New River. Profiles are plotted in order of their confluence with the New River (Fig. 4.2). Note confluence knickpoints in Chestnut Creek (30) and Crooked Creek (28), which enter the New River at the upstream end of the study area where the uppermost wave of incision observed has only recently passed. The extent to which upstream drainage area controls rate of knickopint migration in the four streams is uncertain due to mechanical differences in underlying geology.

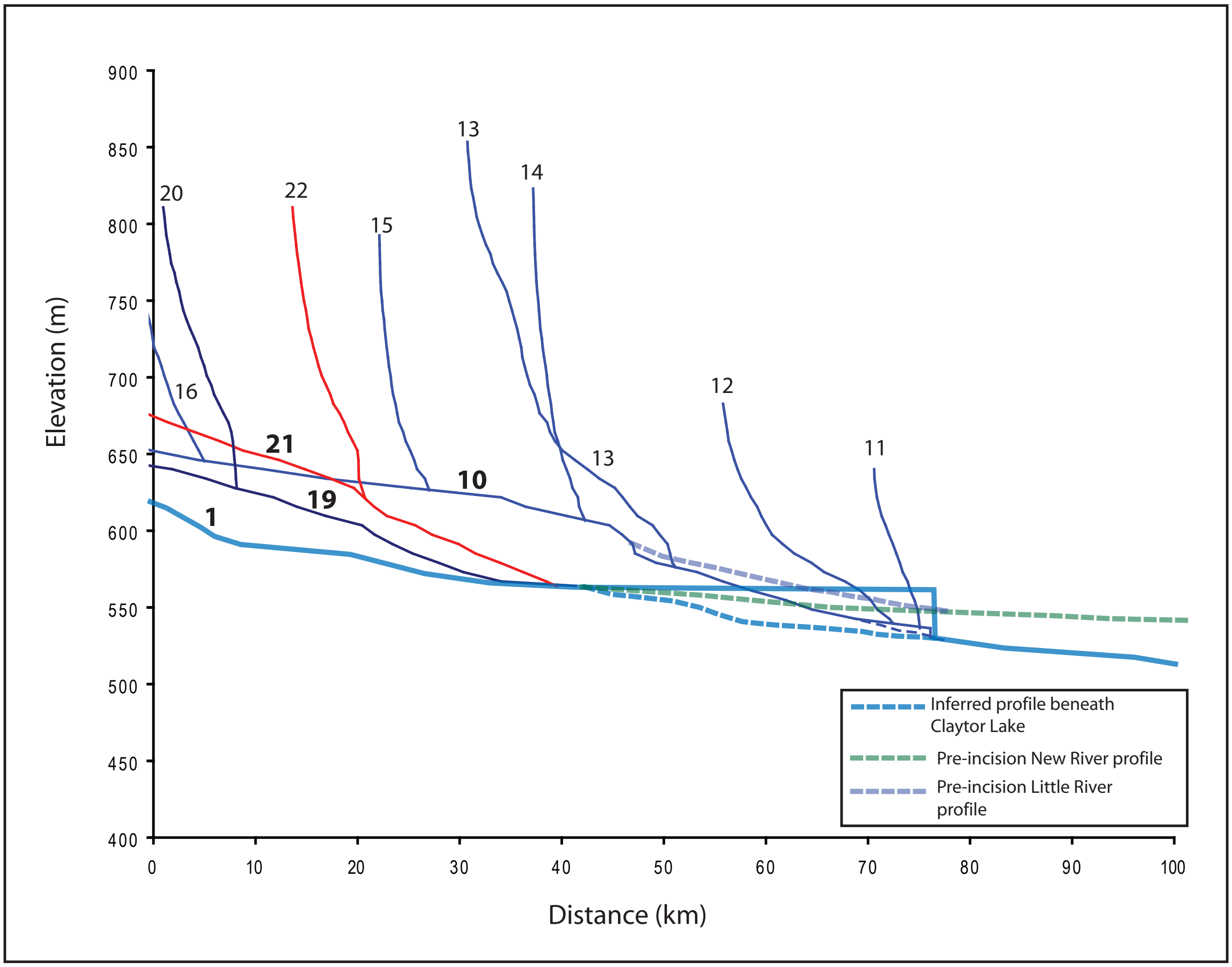
Figure 4.9: Longitudinal profiles showing knickpoint positions in New River tributaries.



Profiles of Blue Ridge Plateau streams plotted relative to the New River (Fig. 4.2). A). Selected profiles from the four major Blue Ridge Plateau basins in the study area. Note hanging tributaries. Logarithmic elevation scale. B). Profiles of the four major Blue Ridge Plateau tributaries of the New River with on arithmetic scale. Outcrop of quarzite in the New River is indicated (this outcrop cannot be extrapolated to other profiles). Headward migration of incision is apparent, as

Little River (10) and Big Reed Island Creek (19) have been affected by incision pulses that have not yet reached Crooked Creek (28) and Chestnut Creek (30).

Figure 4.10: Hypothesized pre-incision profiles of the New River and tributaries.



Hypothetical paleo-profiles of Little River (10) and New River (1) inferred from knickpoint distribution and incision into NRV and Blue Ridge Plateau surfaces (arithmetic scale) (Fig. 4.2). New River profile beneath Claytor Lake is based on historical documents (Trout, 2003). Paleo-profiles reflect gradients prior to the arrival of the wave of incision presently manifest as rapids now covered by Claytor Lake. The major knickzone in the Little River likely represents the accumulation of incision pulses atop slightly metamorphosed slates at the northwestern margin of the Blue Ridge Plateau. While weak themselves, the slates should be more resistant than the carbonates across which the river flows further downstream to the New River. The profile of the lower Little River probably resembled that of lower Big Reed Island Creek prior to steepening by the incision pulse that is presently located under Claytor Lake.

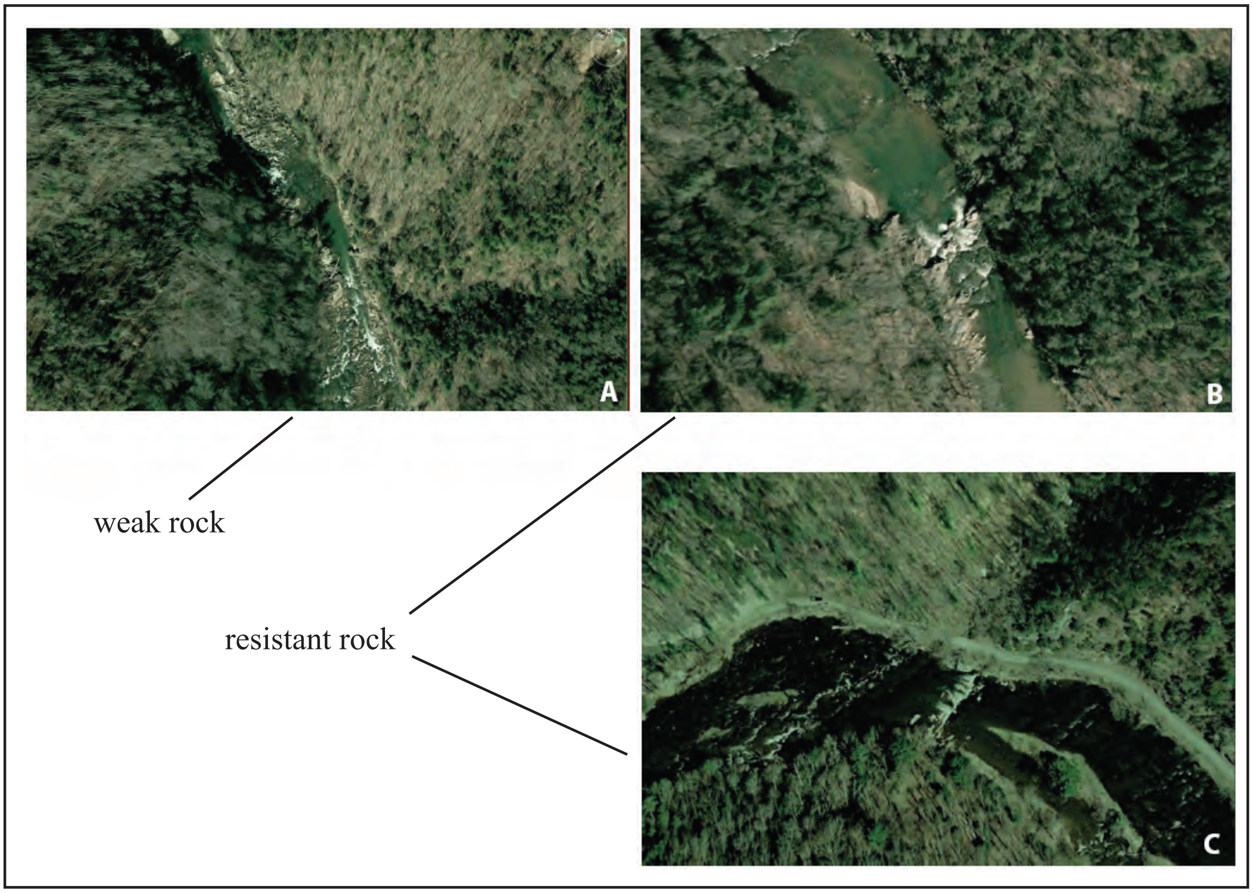
Figure 4.11: Upstream photo of Bear Falls of the major Little River knickpoint.

Bear Falls, the largest single drop in the lower knickpoint of the Little River (10; Fig. 4.2). Rapid is developed in weak slate/phyllite. Note potholes and flutes at right. Tops of protruding rock ribs appear to be the remnants of a gently sloping, beveled strath into which the river has incised ~2m. Retreat of the lip of the falls is apparent in the decreasing height of the rock ribs above the active channel. Potholes up to 1.5 m deep can be found within rock ribs. Another drop of similar height through the same lithology occurs just above Bear Falls. Photo courtesy of John Gannon.

Figure 4.12: Uppermost ledge in major Little River knickpoint.

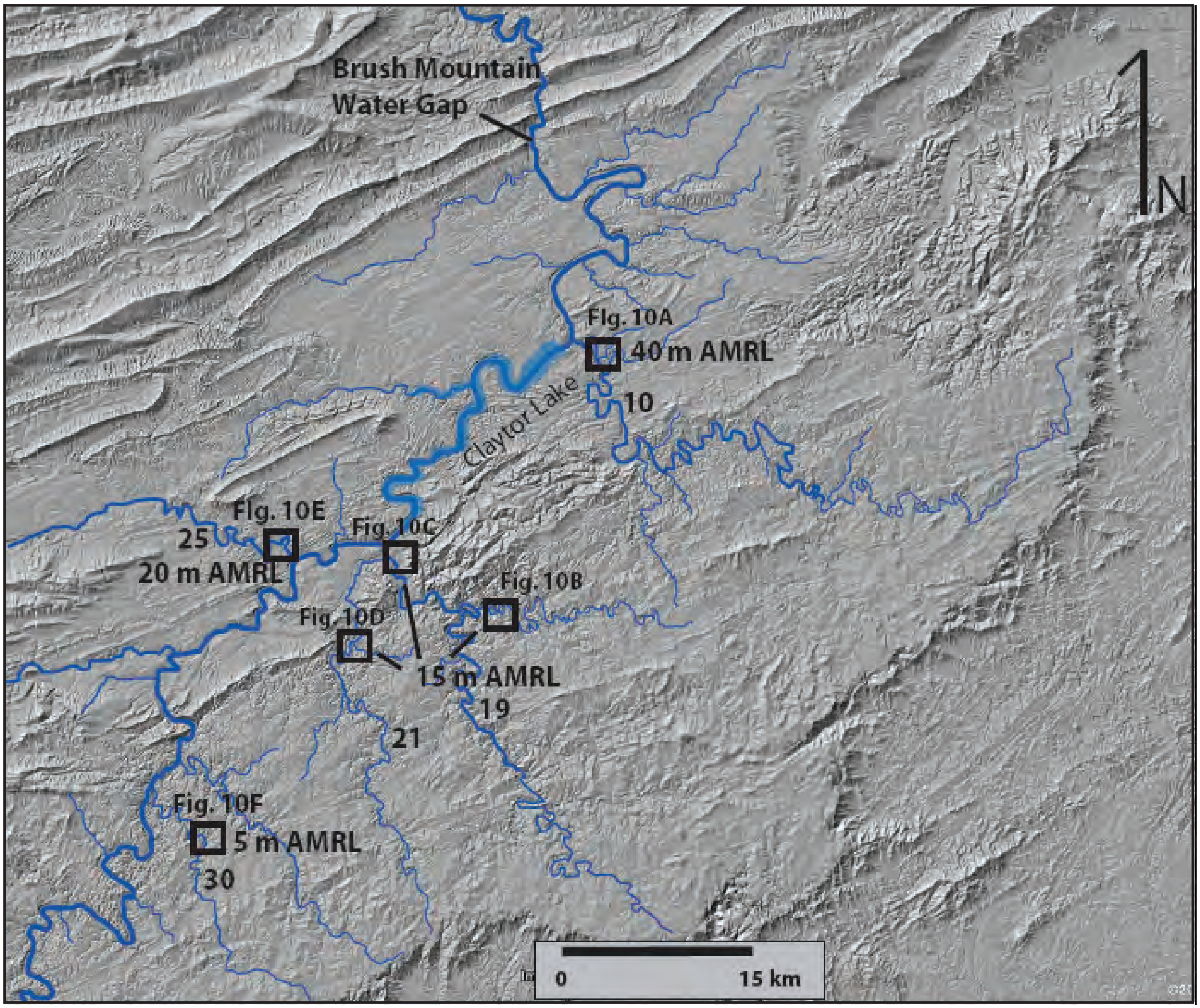
Uppermost ledge in the major Little River knickzone (see Fig. 4.13, panel B). Ledge is developed on lineated Mesoproterozoic biotite gneiss. Wide joint spacing, coarse texture, and high quartz content make the rock unit very resistant, particularly in comparison to the slates supporting the steeper reach ~1 km downstream. The morphology of this falls is distinct from Bear Falls; here, the strong ledge is intact across the streambed, forcing water to spill across its entire width. Bear Falls is comparatively channelized, as the weak, closely jointed and foliated rock is little obstacle to the channel narrowing expected with a sudden increase in gradient (see Fig. 4.13, panel C). When the wave of incision forcing steepness across Bear Falls reaches this unit, a major river-wide falls should form. Photo courtesy Amy Snyder. Paddler is John Gannon.

Figure 4.13: Knickpoint morphology on weak vs. strong rock



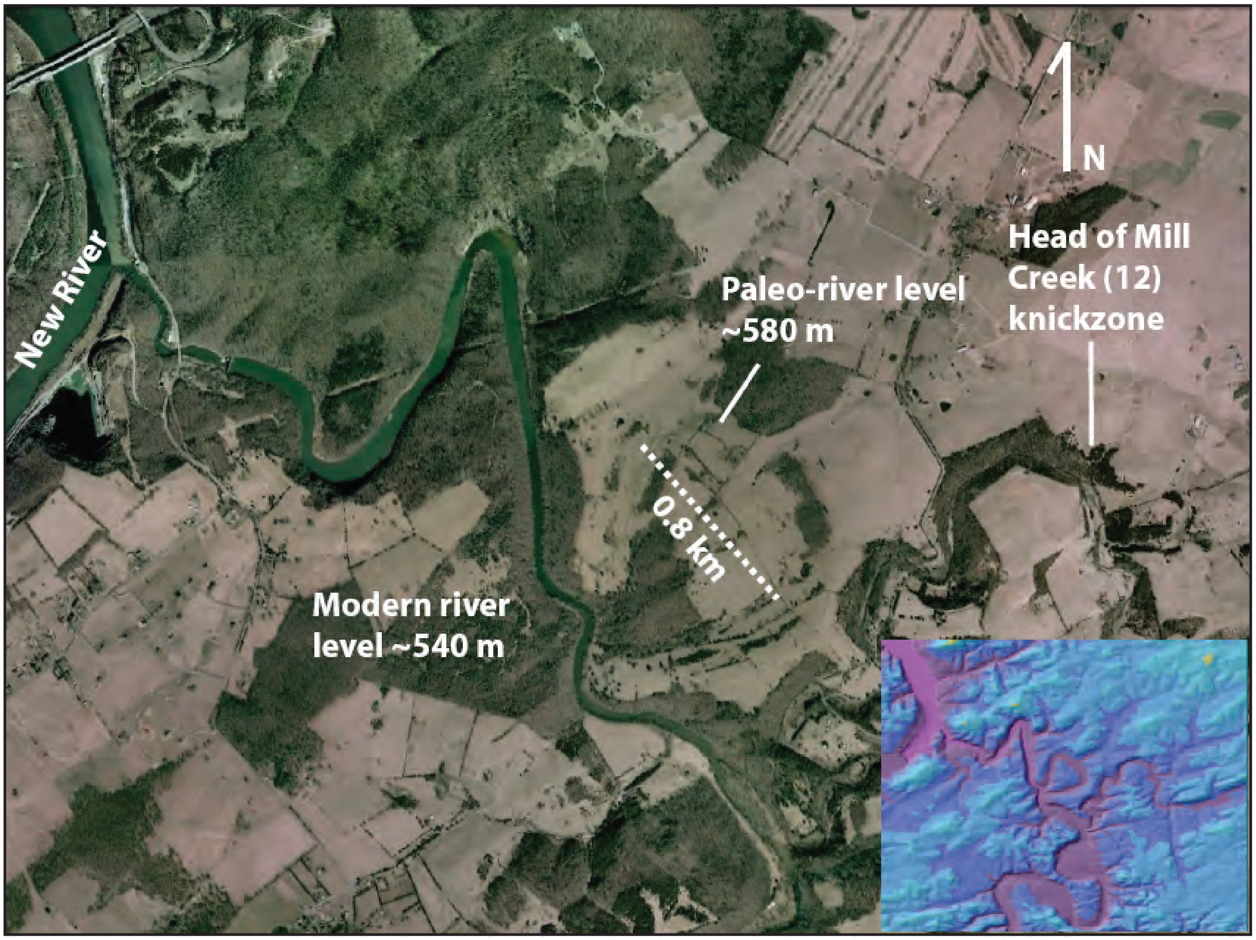
High drainage area falls and rapids within the study area. A. Bear Falls (upper left) and large drop immediately upstream, major Little River (10) knickpoint. Weak rock permits channel narrowing and produces flume- or chute-type rapids (see Fig. 4.11). B. Ledge at upstream end of major Little River knickpoint. Resistant ledge remains intact across the channel, producing a wide falls (see Fig. 4.12). C. Lower Big Reed Island Creek (19) knickpoint, developed on a very resistant bed of late Precambrian quartzite. As with the falls in B, the very resistant substrate produces an intact ledge and channel-wide falls. Image source: Google Earth, image date Feb. 1, 2007. ©2010 Google-data ©2011 Commonwealth of Virginia.

Figure 4.14: Locations of perched meander cutoffs on the Blue Ridge Plateau.



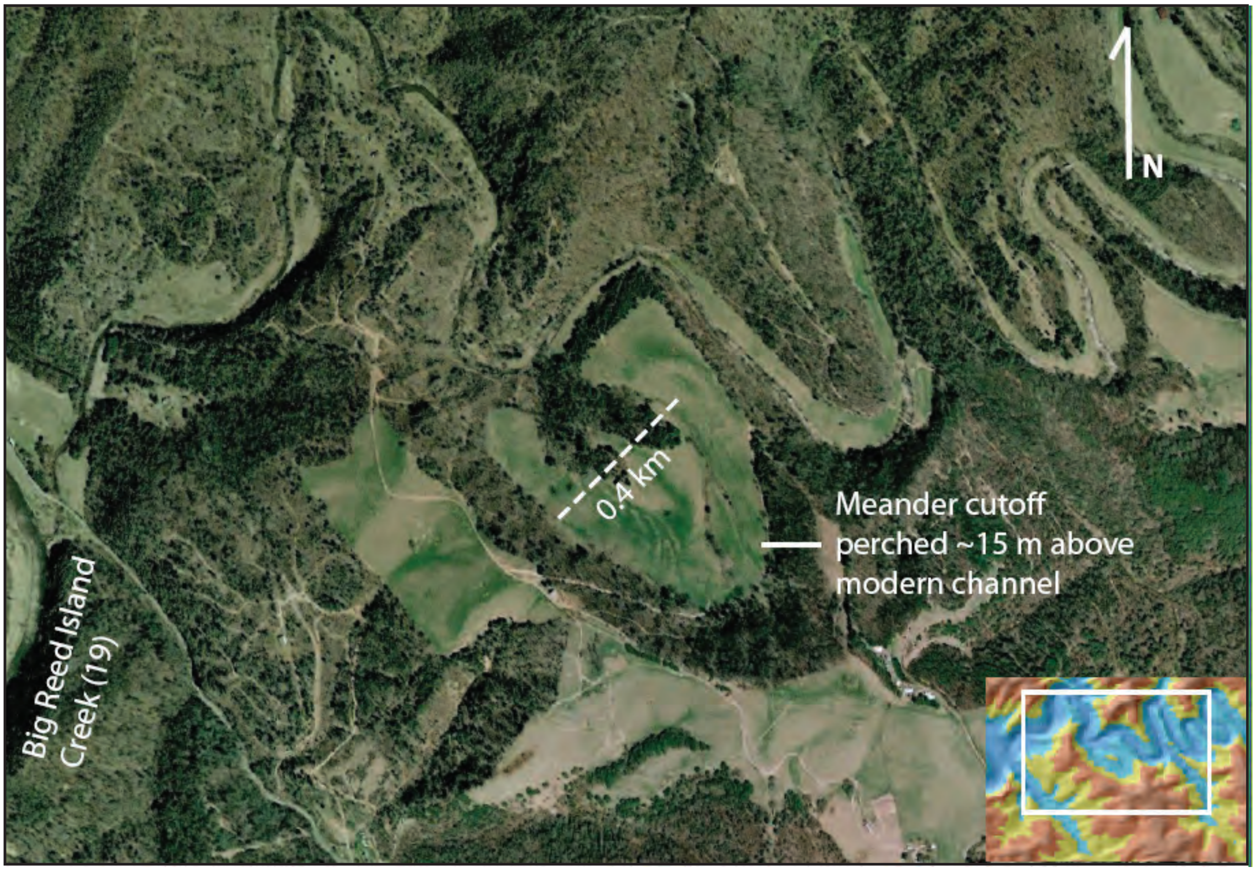
10 m DEM with hillshade topography showing locations of perched meander cutoffs (”loops”) within the study area. Elevation above modern channel is indicated. Details of the loops are found in the subsequent figures. DEM source:www.seamless.usgs.gov. 3-D topography ©2010 Google-data ©2011 Commonwealth of Virginia.

Figure 4.15: Perched meander cutoff above the Little River.



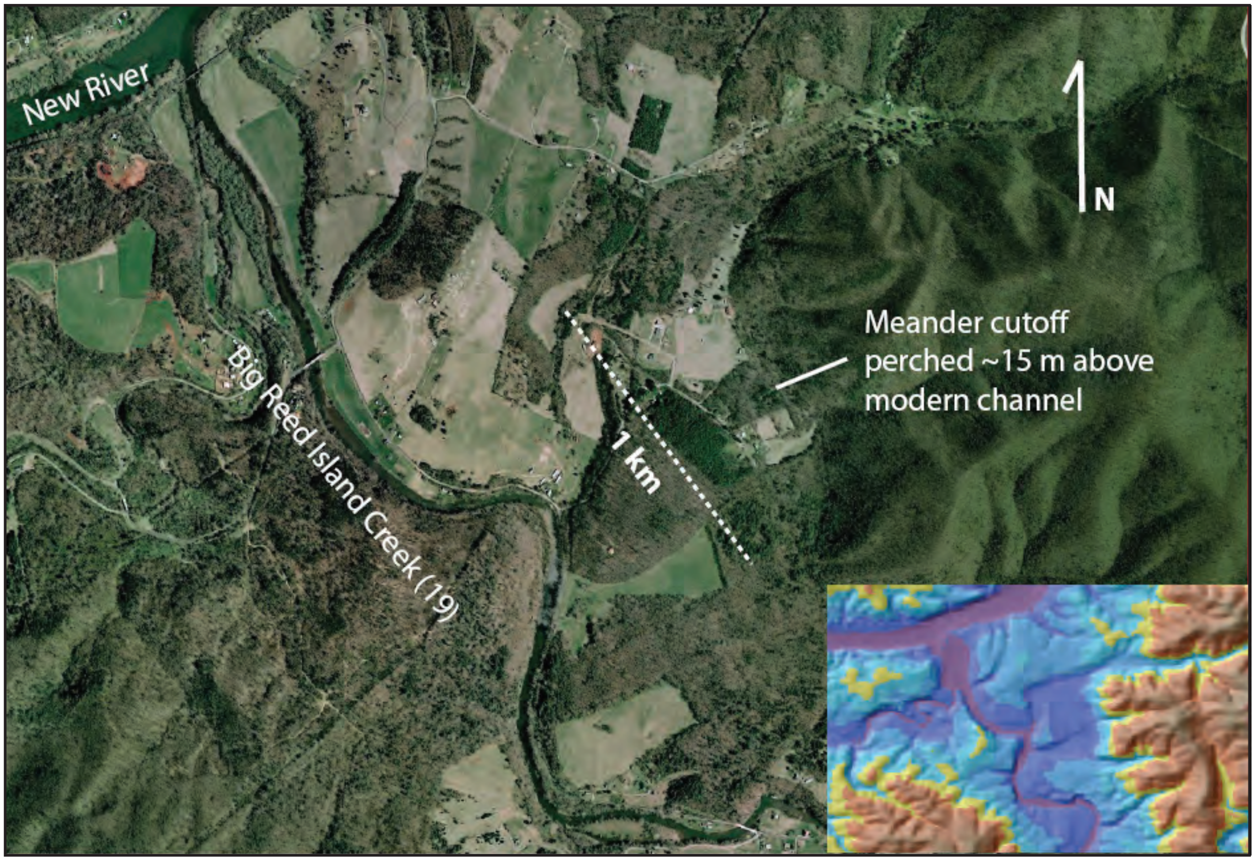
Satellite image of perched meander cutoff above the Little River (10; Fig. 4.2). Loop was moderately incised into NRV surface prior to abandonment. Inset shows color-coded DEM image of the feature. Image source: Google Earth, image date Feb. 1, 2007. ©2010 Google-data ©2011 Commonwealth of Virginia. DEM source: www.seamless.usgs.gov.

Figure 4.16: Perched meander cutoff above Greasy Creek.



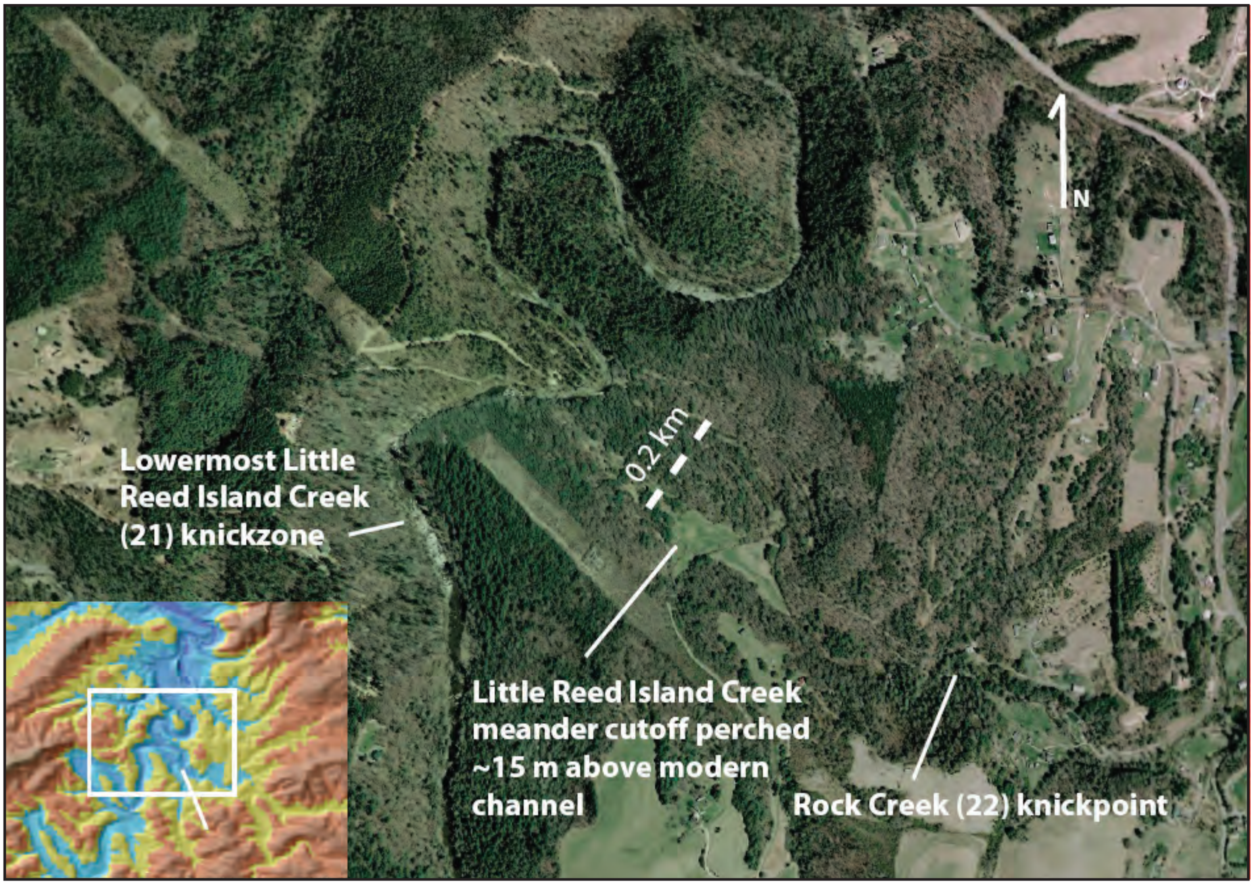
Perched meander cutoff above Greasy Creek, a major tributary of Big Reed Island Creek (19; Fig. 4.2). Loop is preserved at the same elevation above the modern channel as the loops in Figures 10C and 10D. Loop is deeply incised into the Blue Ridge Plateau surface. Image source: Google Earth, image date Feb. 1, 2007. ©2010 Google-data ©2011 Commonwealth of Virginia. DEM source: www.seamless.usgs.gov.

Figure 4.17: Perched meander cutoff above Big Reed Island Creek



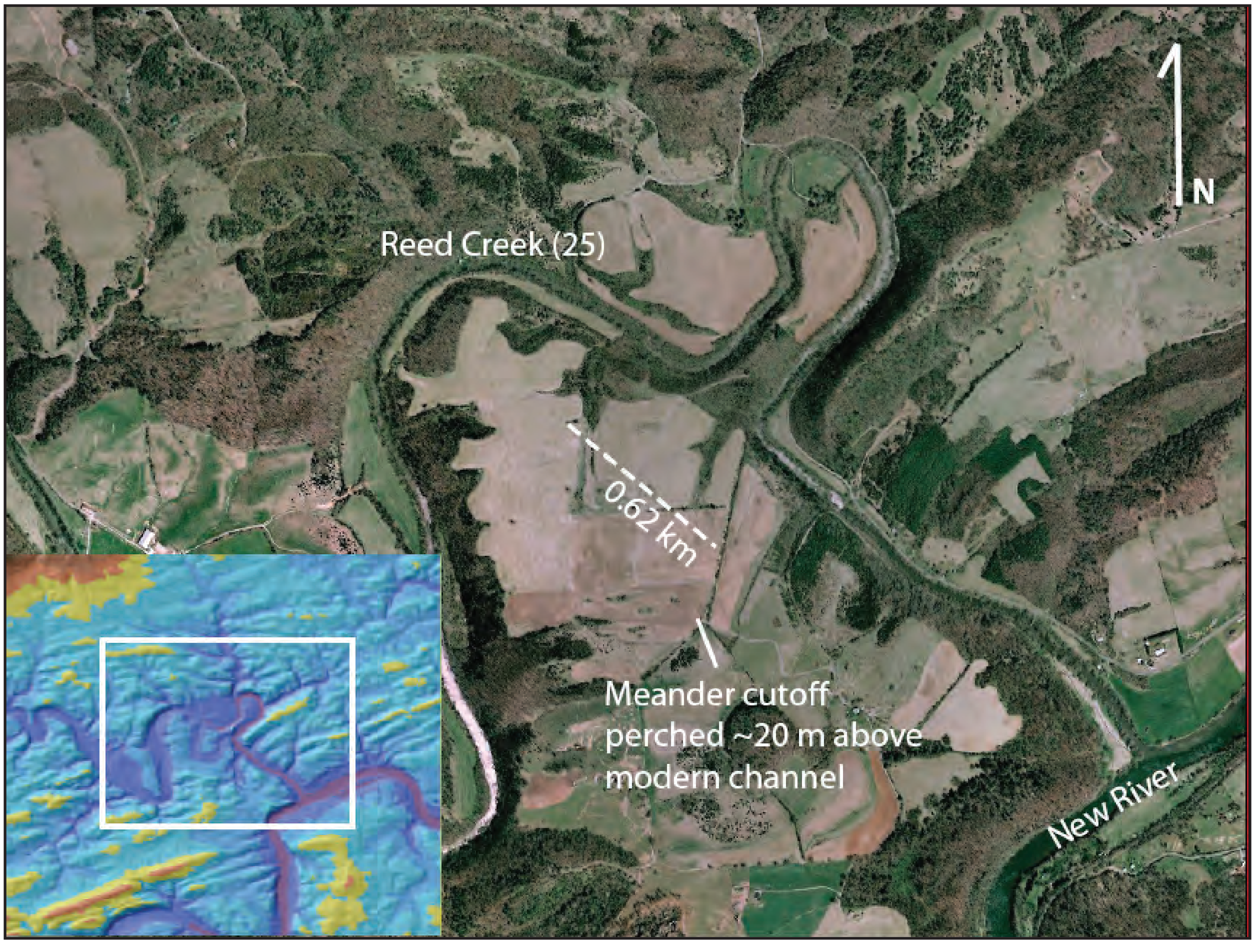
Perched meander cutoff above Big Reed Island Creek (19; Fig. 4.2) near its confluence with the New River. Loop is gently incised into NRV surface. Well-preserved terraces along Big Reed Island Creek pass continuously into the loop, although its outlets are being dissected by small surface streams. Image source: Google Earth, image date Feb. 1, 2007. ©2010 Google-data ©2011 Commonwealth of Virginia. DEM source: www.seamless.usgs.gov.

Figure 4.18: Perched meander cutoff above Little Reed Island Creek



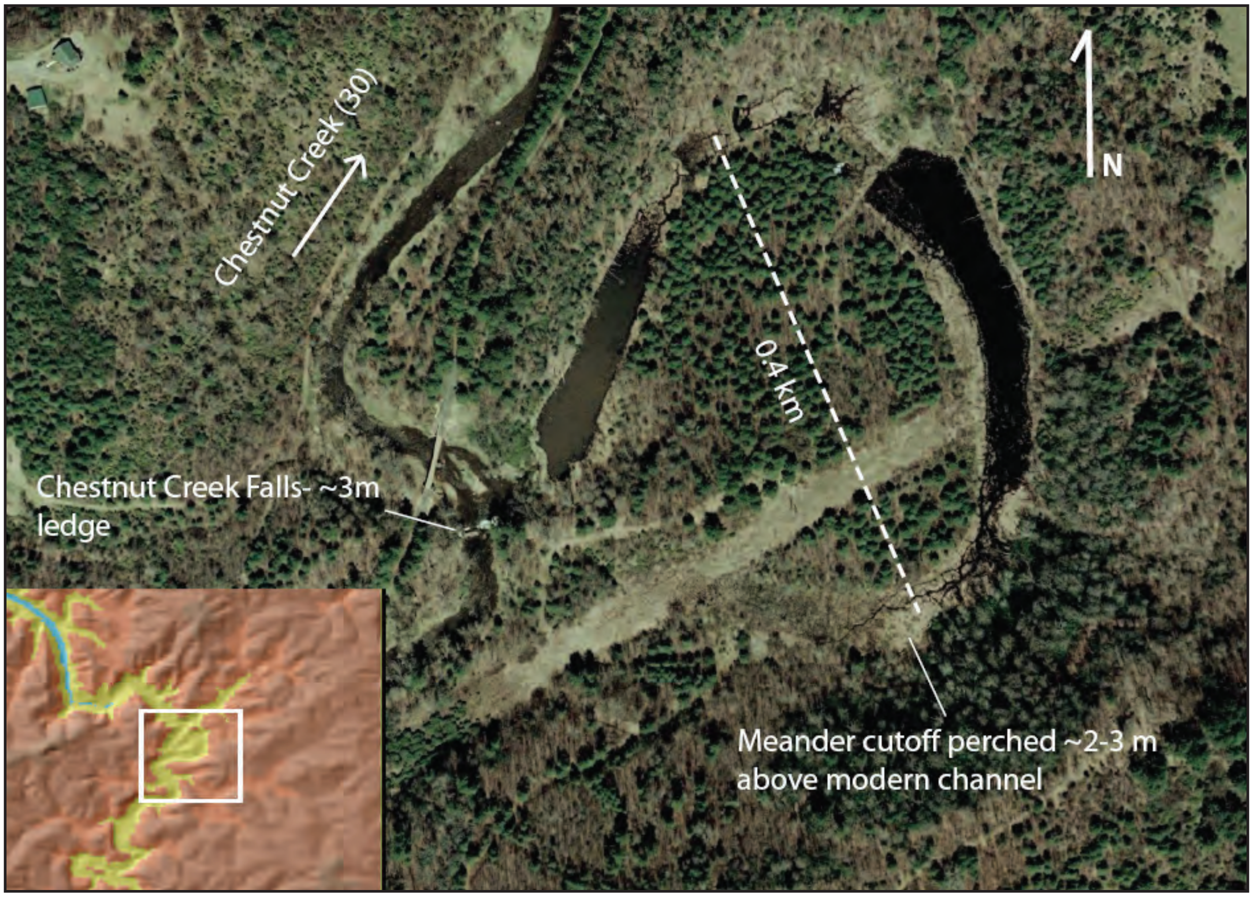
Perched meander cutoff above Little Reed Island Creek (21; Fig. 4.2). Loop is deeply incised into the Blue Ridge Plateau surface. Rock Creek (22; Fig. 4.2) enters the loop across a major knickpoint, flows along the loop surface, then drops again into the active Little Reed Island Creek channel. Image source: Google Earth, image date Feb. 1, 2007. ©2010 Google-data ©2011 Commonwealth of Virginia. DEM source: www.seamless.usgs.gov.

Figure 4.19: Perched meander cutoff above Reed Creek.



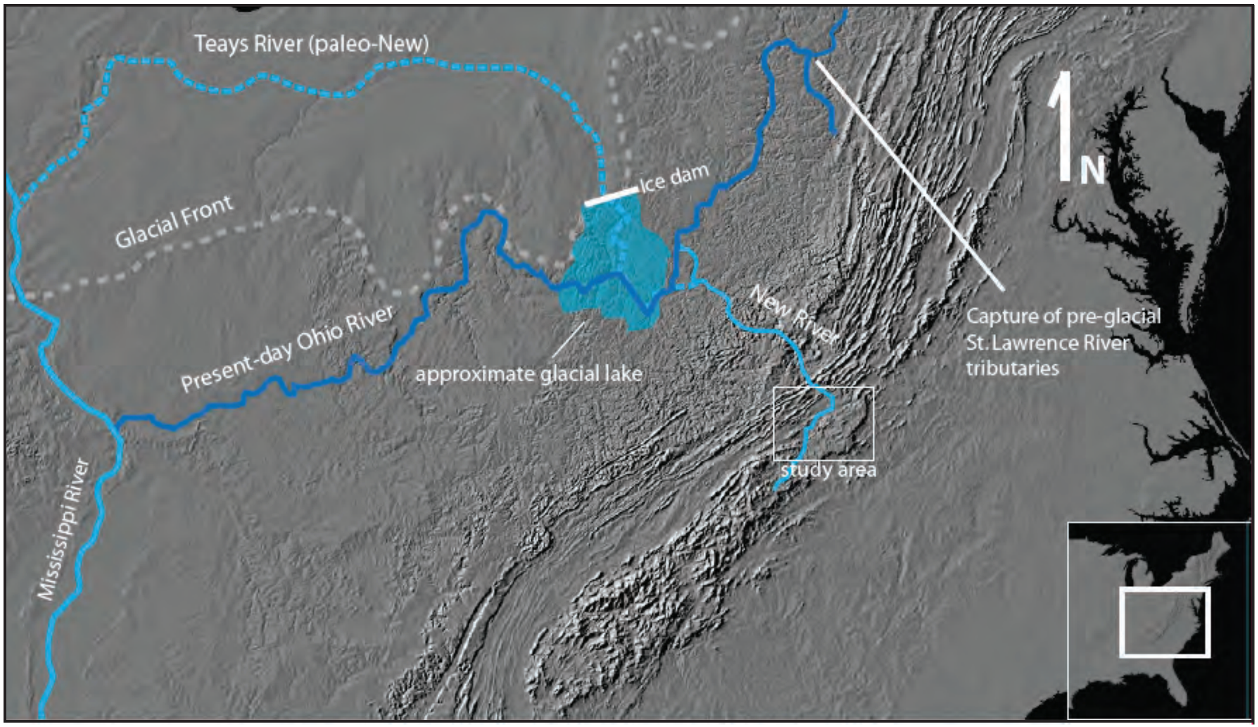
Perched meander cutoff above Reed Creek (25; Fig. 4.2). Loop is very gently incised into the NRV surface, and is barely discernible in satellite images or on 1:24,000 topography. Image source: Google Earth, image date Feb. 1, 2007. ©2010 Google-data ©2011 Commonwealth of Virginia. DEM source: www.seamless.usgs.gov.

Figure 4.20: Perched meander cutoff above Chestnut Creek.



Perched meander cutoff above Chestnut Creek (30; Fig. 4.2). Loop is deeply incised into the Blue Ridge Plateau surface. A large ledge, the leading edge of the Chestnut Creek confluence knick-point, is presently located between the loop arms such that the lower arm is perched higher above the active channel than the former entrance arm. Image source: Google Earth, image date Feb. 1, 2007 ©2010 Google-data © 2011 Commonwealth of Virginia. DEM source: www. seamless.usgs.gov.

Figure 4.21: Glacial rearrangement of the Teays River system.



Glacially-driven rearrangement of the Teays River system. The modern New River is the relict headwaters of the Teays system, which was glacially dammed near Chillicothe, Ohio. The glacial lake drained, or was captured and drained, to the southwest, producing the modern drainage pattern seen in the Ohio River system. The Teays system was considerably shortened by the rearrangement, which may have steepend gradient adequately to drive incision into the study area (white box) on the New River. Incision following glacial rearrangement of the Ohio system is believed to have caused capture of the present Ohio headwaters from the St. Lawrence River system (Farrand, 1988). DEM source: http/pubs.usgs.gov/imap/i2206/.

APPENDIX C: LITHOLOGY AND PRESERVATION OF NRV SURFACE GRAVELS

Overview

For a review of relict New River gravel deposits, see Houser (1981), Mills (1986), Bartholomew and Mills (1991), and Ward et al. (2005). New river gravels of the eastern Valley and Ridge province consist almost entirely of well-rounded vein quartzite and metaquartzite clasts from the Blue Ridge province to the southwest. Clast and terrace preservation show an inverse relationship to distance above the modern river level, such that the highest terraces contain the most weathered clasts and lack the very flat tread surfaces seen in lower elevation terraces. Clast weathering and the development of hummocky, but still low-relief terraces surfaces appears to function as a reasonable proxy for relative age of abandonment. While exact provenance is unknown, the clasts are clearly of metamorphic origin and have been transported to their present location by the New River when it flowed at a higher elevation. Their significance lies in their constraint of many elevation-accordant surfaces as remnants of a land surface connected to a formerly higher base level. The preservation of the gravels also indicates the great stability of the remnant surfaces and the spatially migrating character of adjustment to New River Incision.

Blacksburg-Christiansburg area deposits

Isolated pockets of well-rounded vein quartzite and metaquartzite pebbles, cobbles, and small boulders are located at approximately 630-650 m above sea level within developed areas of Blacksburg and Christiansburg. A particularly extensive deposit is located within, and northeast of, the Virginia Tech airport. Clasts are regularly uncovered by construction and gardening and are easily identified by their consistently advanced degree of rounding and metamorphic lithology. Exact provenance of the gravels is unclear, but mylonitized vein quartz clasts almost certainly originated within the Fries Fault Zone to the east/southeast of the deposits. Thin sections of a metaquartzite revealed the early stages of grain boundary migration consistent with lower greenschist facies metamorphism experienced by the late Precambrain metasediments that now form the Pilot Mountain ridge southeast of the deposit.

While the alluvium was clearly derived from the metamorphic Blue Ridge province southeast of its present location, its context within the paleo-drainage network is unclear. Some authors (Houser, 1981; Bartholomew and Mills, 1991) have suggested the Blacksburg deposits are the remnants of a “Blacksburg River” flowing to the New River from the northeast, but gravels identified in the headwaters of Crab Creek in Christiansburg constrain the paleo-course of a trunk stream draining the presumed "Blacksburg River" basin. The Crab Creek gravels are also lithologically distinct from the Blacksburg gravels due to the presence of blue quartz sand and pebbles as well as clear vein quartz from the northern Blue Ridge. The lithology of the Blacksburg gravels, along with their extreme roundness, thus suggest they were deposited by a paleo-course of the New River which flowed around the northeast end of Price Mountain, across Blacksburg, and then southwest to pass cross Brush and Gap Mountains through the same water gaps it uses today. This interpretation differs from Bartholomew and Mills (1991), but the Crab Creek gravels were not known at the time of their study.

The significance of the Blacksburg gravels to the Kentland Farms 128 m terrace of Ward et al. (2005) is unclear; both deposits are clearly of New River origin, but they may represent different episodes of base level quiescence or course abandonment. While the Blacksburg deposit is the most extensive, other NRV surface gravels can be found on hilltops of appropriate elevation between Kentland Farms and Blacksburg. The presence of fluvial gravels on elevation accordant hilltops supports the interpretation that the deposits are remnants of a formerly extensive low-relief surface that is being dissected. These gravels indicate great stability of remnant surface hilltops adjacent to actively incising bedrock stream valleys, reflecting the spatial variability of the erosional response to New River incision.

Gap Mountain deposit

An unusual deposit of relict alluvium persists in the US 460 roadcut across Gap Mountain northwest of Blacksburg. Originally identified by Bartholomew (unpublished data), the locally iron oxide-cemented gravel is preserved in a sag on the Gap Mountain ridge ~ 260 m AMRL. The deposits contains very well rounded vein quartz and metaquartzite clasts of Blue Ridge origin as well as sandstone, chert, and silicified oolite clasts derived from more proximal Valley and Ridge units. The presence of blue Ridge clasts indicates that a river with a drainage basin stretching at least 40 km to the southeast once passed through the gap, but whether that river was the New River itself or a large tributary is uncertain. The Gap Mountain deposit is most significant as physical evidence of the stability of the ridge crest in relation to the surrounding valley floors. The deposit is record of a drainage network with a local base level ~130 m higher than the level of the New River when it appears to have flowed across present-day Blacksburg. Assuming the remainder of the Gap Mountain ridge and other ridges have been as stable as the small surface in the gap itself, relief has increased by as much as ~270 m in the eastern Valley and Ridge during the late Cenozoic.

The age of the Gap Mountain deposit is uncertain, but Houser (1981) estimated abandonment at ~8 Ma assuming a regional long-term erosion rate of 30-40 m/My. This estimate assumes spatially consistent erosion, however, and does not account for the spatially and temporally variable erosion suggested by remnant surface preservation in the New River Valley. While constraining the age of the deposit is quite difficult, its elevation above the modern river level and above nearby deposits roughly dated to ~2 Ma (Ward et al., 2005) suggests the gap surface, and perhaps many of the ridge crests of the Valley and Ridge, may be some of the oldest land surface features in the world.

James Cave-Dublin-Fairlawn

Inspection of topography indicates the surface preserved in the Blacksburg and Christiansburg areas also persists on the west bank of the river. While gravels have not been extensively mapped in this area, preservation of relict fluvial debris inside James Cave northwest of Fairlawn indicates the area hosts the same type of relict surface as described in the Blacksburg area. Clasts within the cave are very well rounded quartzites of Blue Ridge origin. Their roundness and lithology confirm their transport and deposition via the New River. The clasts are currently preserved within the stream inside James Cave, and also persist in cave terraces which are perched above the present stream level. While the source of the clasts is not mapped, they have clearly been washed into the cave from the surrounding land surface.

The town of Dublin and the New River Valley airport occupy a very low relief surface at ~650 m elevation, the same elevation as the gravel deposits near the Blacksburg airport. The Dublin area is located at the headwaters of several New River streams in which knickpoints have not yet reached the headwaters. As previously mentioned, slow knickpoint migration in these streams probably reflects their small drainage basins and subsurface flow through karst systems. Indeed, the cave stream flowing through James Cave follows an alluvial course through the cave before it enters a knickzone which has deeply incised the carbonate bedrock at the eastern end of the cave.

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APPENDIX D: OBSERVATIONS OF THE PROGRESS OF TRANSIENT INCISION THROUGH NEW RIVER BASIN STREAMS

Streams of the NRV Surface

Upstream (southeast) of the Brush Mountain water gap, Back Creek (2) and Toms Creek (3) are extensively incised to their headwaters into the NRV surface. These streams are large, and flow along the Pulaski thrust, whose weak, brecciated carbonate outrcrop should provide little resistance to knickpoint migration. Tributaries of these streams remain oversteepened and continue to attack the relict NRV surface, which remains intact on larger interfluves. While the basin-wide incision of these streams reflects the spread of an old incision event (or events), knickpoints apparent in longitudinal profiles indicate the streams continue to adjust to a subsequent pulse of New River incision. The Back Creek knickzone has developed atop exposure of Missippian clastic units from the footwall of the Pulaski thrust. These resistant rocks have slowed knickpoint migration, and preserve a pre-recent incision reach at a slightly higher elevation than the analogous knickpoint (or knickzone) in Toms Creek, which is hosted by brecciated carbonates and clastics along the trace of the Pulaski thrust. Both profiles also show slight convexity immediately above their confluences with the New River, suggesting initiation of another knickpoint in response to a comparatively recent, and minor, pulse of New River incision. The spacing between the developing confluence knickpoints, the upstream knickzones, and the incised headwaters may indicate episodic incision events separated by periods of base level stability.

The Stroubles Creek (4)/Slate Branch (5) basin enters the New River from the southeast upstream of the Toms Creek confluence. These streams do not exhibit the advanced adjustment of Toms Creek or Back Creek because of the considerable distance their courses flow across moderately resistant clastic units exposed within the Price Mountain window. These resistant units have slowed the headward advance of knickpoints compared to the weak breccias of Toms and Back Creeks, preserving intact relict reaches within the headwaters of Stroubles Creek and Slate Branch in the Blacksburg area. Blacksburg is thus situated on a small plateau overlooking the Toms Creek basin; this relief is a direct result of more advanced Toms Creek adjustment juxtaposed against the unincised Blacksburg landscape through which Stroubles and Slate Creek knickpoints are yet to migrate. Longitudinal profiles of the creeks reveal multiple knickpoints occurring within the Price Mountain window outcrop upstream of the clastic-carbonate contact.

The presence of these knickpoints within a uniform lithology is evidence of the spatially transient character of the features, which are not fixed where streams pass from a resistant lithology to a weaker substrate.

Crab Creek (6), which drains Christiansburg, flows entirely across carbonate strata. Crab Creek is the beheaded remnant of Bottom Creek, which was captured into the South Fork Roanoke River basin (Prince et al., accepted). Crab Creek has thus lost ~225 km^2 of drainage basin due to capture. This loss of discharge has certainly slowed its response to New River incision, suggesting that its capture pre-dates the onset of major incision into the NRV surface. Crab Creek retains a short relict reach upstream from the point where it is crossed by US 460 northeast of downtown Christiansburg. Downstream of this point, Crab Creek follows an incised course containing three knickpoints. While these discrete knickpoints may reflect separate pulses of incision, they may also reflect subtle mechanical contrasts within the carbonate strata which slow headward migration and allow knickpoints to “stack” at the margins of more resistant beds. Small tributaries draining Christiansburg north to Crab Creek retain relict headwaters which pass over knickpoints into the present Crab Creek channel, producing the generally low relief which was undoubtedly a factor in the town’s location. Many cities and towns within the study area occupy areas of relict topography situated well above active stream channels.

Plum Creek (7) enters the New River ~1.5 km upstream of Crab Creek. Plum Creek contains a significant knickpoint ~2 km upstream of its mouth reflecting a recent pulse of New River incision. The knickpoint is developed atop carbonate strata and passes into Devonian shale substrate upstream. The Plum Creek headwaters are incised into the relict NRV surface west of Christiansburg, suggesting previous knickpoint migration through the stream to its headwaters. The headwaters remain oversteepened for the carbonate substrate and will continue to erode headwardly into the relict NRV upland.

The City of Radford is located on a beveled surface 60 m AMRL. Connolly’s Run (8) has incised the surface to grade with modern river level. This stream contains a knickpoint immediately above its confluence with the New River, but incision has occurred to the headwaters of the stream and likely reflects the migration of previous knickpoints. The Connolly’s Run knickpoint is developed within a single Cambro Ordovician carbonate lithology. As with the previously discussed drainages, the headwaters of Connolly’s Run, as well as its tributaries, remain oversteepened with respect to the present New River base level and will continue to erode headwardly into the relict surfaces to lengthen their channels and reduce overall gradient.

Falling Branch (9) enters the New River from the northwest slightly upstream of the mouth of Crab Creek. The Falling Branch basin shows adjustment to New River incision throughout, but headwaters incision is minor and likely related to incision "leaking" beyond downstream knickpoints developed on the weak carbonate substrate (Frankel et al., 2007; Berlin and Anderson, 2009). Inspection of the longitudinal profile reveals four separate knickpoints, two of which occur very close to the New River confluence. Response of surface channels to New River incision in this area is likely muted due to considerable subsurface flow through karst systems. The headwaters of Back Creek are encroaching on the Falling Branch basin from the North due to the more rapid adjustment of the Back Creek basin. As with Crab Creek, the significance of upstream knickpoints in the history of New River incision is unclear; however, the lowermost knickpoint clearly reflects response to the recent passage of a knickpoint in the New River channel.

Upstream of Radford, two more small tributaries on the west bank of the New River area incising the relict surface. As seen in tributaries downstream, these streams are incised throughout their basin but contain knickpoints well downstream of headwaters. The headwaters of these streams have begun to dissect the intact remnant occupied by the town of Dublin. Like Blacksburg and Christiansburg, Dublin is located atop a large portion of undissected NRV surface whose position within the drainage network has prevented extensive relief development.

The Little River (10) is the largest tributary of the New River in Virginia. Above its mouth on the east bank of the New River just below Claytor Dam, the Little River follows a meandering course incised into the NRV surface. A major knickzone developed on late Precambrian metaquartzites of the hanging wall of the Blue Ridge thrust separates the Valley and Ridge reach of the Little River from its upper reaches flowing over crystalline rocks of the Blue Ridge Plateau. Response of the Blue Ridge Plateau reach of the Little River will be discussed below.

From its confluence with the New River to ~9 km upstream, the Little River is generally adjusted to New River base level. The river follows an incised meandering course, with sharp gorge walls with bedrock exposure rising from the active channel. Many large, intact remnant surfaces can be observed owing to the lack of large surface streams in the area. Longitudinal profiles indicate a ~10 m knickzone representing the most recent New River incision events ~ 9 km upstream of the New River confluence. This knickzone is apparent in satellite images as numerous bedrock ledges exposed in the streambed. An additional knickzone likely exists closer to the New River confluence, but this feature is presently obscured by a dam and small impoundment. While the large basin of the Little River has facilitated rapid response to New River incision, tributaries remain disequilibrated to the present New River level. Mill Creek, the largest tributary of the Little River downstream of the Blue Ridge thrust, contains a knickzone just above its confluence with the Little River. Smaller tributaries maintain knickzones and over steepened reaches, but incision within these small drainages has not spread far beyond the trunk stream valleys. Bartholomew and Mills (1991) postulated paleo-Little River courses based on gravel deposits, but the relationship of these inferred paleo-courses to the NRV surface base level is uncertain.

The most significant relict feature within the Valley and Ridge reach of the Little River is a perched meander cutoff preserved ~40 m AMRL. The dry loop channel is partially adjusted to the present Little River level, with the southeastern arm more extensively adjusted due to the small tributary stream which flows through it. The uppermost portion of the loop is totally undissected. The loop is itself genlty incised ~40-45 m into the surrouding NRV surface. The wall of the loop channel are not steep, and lack the abruptness seen in the large risers separating terraces observed downstream in the New River. This constitutes anecdotal evidence of some degree of base level stability at the time the loop channel was active. Abandonment of the channel must have pre-dated the most recent 40 m of incision. The elevation of the loop above the present Little and New River channels is comparable to a broad strath terrace on the site of the Radford Army Ammunition plant as well as straths upstream of the loop along the Little River. These strath features, which are separated from modern river level or lower terraces by sharp risers, also appear to indicate a pause in incision which allowed lateral migration of the channel and beveling of the surface as well as formation of the meander which is presently preserved as the “loop.” These surfaces indicate two major incision events separated by a period of base level stability since initial abdandonment of the NRV surface.

Peak Creek (18) is a large west bank tributary draining Pulaski. Peak Creek has incised the NRV surface extensively owing to its large drainage area and carbonate substrate. A knickzone likely exists within the Peak Creek channel above its New River confluence, but the feature is today obscured by the waters of Claytor Lake. A minor knickzone presently exists just upstream of the town, and is developed atop weak Mississipian clastic units.The headward progress of this, or potentially earlier, knickzones up Peak Creek from the New River is preserved by knickpoints in Peak Creek tributaries, such as Thorne Springs Branch. This small stream retains a knickzone developed entirely atop carbonate substrate ~1 km upstream of its confluence with Peak Creek. The town of Pulaski is constructed in the Peak Creek floodplain/meander belt. Pulaski is situated 70 m lower than nearby Dublin, which is located atop an NRV surface remnant. Pulaski is surrounded by bluffs capped by remnants of the NRV surface, and small Peak Creek tributaries rising on relict surfaces maintain transient reaches as they adjust to New River incision as it is transmitted through Peak Creek.

Reed Creek (25) enters the New River near the present headwaters of Claytor Lake. Like Peak Creek, Reed Creek is a large drainage which shows extensive adjustment to New River incision in its dissection of the NRV surface. While the topography of the Reed Creek basin is qualitatively analogous to that previously described, the distance upstream from the "reference surfaces" is reflected in the increased elevation of the remnant surfaces and the overall magnitude of incision which has occurred. Inferred NRV surface at the mouth of Reed Creek is approximately 660 m, and surfaces increase in elevation with increasing distance from the river to approximately 700-710 m at Wytheville, which occupies a large intact remnant surface similar to Dublin, Christiansburg, and Blacksburg. Remnants of the NRV surface near the mouth of Reed Creek suggest ~70 m of incision compared to ~130 m of incision apparent near the Brush Mountain water gap (Ward et al., 2005). The discrepancy between slopes of the NRV surface and present slope of the river (based on incision magnitude) reflects the presence of knickpoints between Reed Creek and Brush Mountain. These knickpoints have not yet reached the mouth of Reed Creek to communicate continued base level lowering to the Reed Creek basin and New River tributaries further upstream. Inspection of satellite images reveals a series of small knickpoints approximately 21 km upstream of the mouth of Reed Creek. While these knickpoints are not large enough to be resolved by 1:24,000 scale topographic maps, they clearly represent a zone of bedrock incision within an otherwise alluvial channel.

A "loop" feature like the one preserved on the Little River is located ~25 m above the modern Reed Creek channel ~4 km above its confluence with the New River. The paleo-meander is not deeply incised into the surrounding relict surface, and is not apparent in satellite imagery. DEM analysis, however, shows the feature clearly.

Cripple Creek (26) enters the New River 17 km upstream, and topography near the confluence reflects the trends in relict surface elevation and incision described above. NRV surface remnants reflect ~65 m incision at the mouth of Cripple Creek. Interpretation of remnant surfaces upstream of the confluence is complicated by lithologically-controlled hills and ridges, but overall remnant elevation is ~670 m. Satellite imagery reveals a zone of minor knickpoints ~12 km above the mouth of Cripple Creek. This zone of minor bedrock incision is similar to the feature described in Reed Creek, but has not migrated as far upstream. As Reed and Cripple Creek have similar drainage areas on the same lithology, the discrepancy in knickpoint migration should reflect the time elapsed since the last pulse of New River migrated past their respective mouths.

Streams of the Blue Ridge Plateau Surface

The Blue Ridge portion of the Little River (10) is separated from the Valley and Ridge reach described above by a major knickzone developed on the Blue Ridge quartzite strata and adjacent mylonitized gneisses in the hanging wall of the Fries Fault. The Little River drops ~ 40 m across the knickzone, which appears to have "collected" incision pulses migrating headward from the NewRiver. While the position of the knickzone does coincide with mapped clastic units, field inspection reveals that the knickzone extends into comparatively weak mylonitized gneisses. Upstream of the knickzone, the river is incised ~ 80 m into the Plateau. Large tributaries are typically well-adjusted to the Little River in this reach. The ~80 m of inicision appears to have spread throughout the Little River basin, and inspection of the longitudinal profile upstream of the quartzite knickzone shows no additional major knickpoints. This observation is corroborated by inspection of satellite images which reveal an incised alluvial channel with very little bedrock exposure. The West Fork of the Little River, the largest tributary to the Little River above the quartzite knickzone, is well-adjusted to the modern Little River level. The resistant Blue Ridge quartzite units have effectively blocked the most recent ~40 m of incision from progressing headward into the Blue Ridge Plateau. Longitudinal profiles of minor tributaries indicate some minor incision has spread above the main knickzone, but the reach of the Little River upstream of the quartzite outcrop remains detached from the present local New River base level.

Big Reed Island Creek (19) enters the New River near Allisonia Virginia. The Big Reed Island Creek basin occupies the Blue Ridge Plateau southwest of the Little River basin. Dietrich (1959) recognized that remnant surfaces in the Big Reed basin tended to be somewhat higher in elevation than those of the Little River basin. He attributed this to the relative locations of the mouths of these streams; while the basins share a divide on the Plateau, Big Reed enters the New River 35 km upstream from the Little River. The Big Reed Island Creek basin is thus connected to a New River elevation that would be notably higher than that of the Little River mouth even during a period of stable base level. Transient adjustment within the Big Reed basin also reflects the considerable separation of the stream mouths; the upstream location of the Big Reed mouth has not yet been passed by knickpoints which have passed the Little River mouth. The lowermost reach of Big Reed passes through a dolomite unit comparable to the substrate of the lower Little River, but Big Reed is not as deeply incised into the dolomite. Incision into the dolomite decreases towards the Big Reed mouth, while it increases towards the Little River mouth, which has recently been passed by a New River knickpoint. Big Reed, and its large tributary Little Reed Island Creek, cross the same resistant units as the Little River before entering the Valley and Ridge, but they exhibit smaller knickpoints producing a smaller elevation change in the streambed.

The contrasts between the Big Reed and Little River systems collectively indicate that New River knickpoints that have passed the Little River confluence have not yet reached the Big Reed confluence. These knickpoints are not readily identifiable in the New River longitudinal profile, and are almost certainly obscured by Claytor Lake. Big Reed and Little Reed still contain knickpoints well upstream of the Blue Ridge-Valley and Ridge boundary, and the reaches upstream of these knickpoints appear entirely relict. Below these knickpoints, both streams flow through gorges developed into the Plateau surface, and their tributaries enter the gorges across knickpoints. Headwaters reaches of many smaller tributaries are also entirely relict, preserving large areas of remnant topography atop interfluves and between the Big Reed and Little Reed gorges. Remnant topography also exists between the gorges of Greasy Creek and Burks Fork, large tributaries of Big Reed which enter sufficiently far below the headwardmost Big Reed knickzone to have adjusted extensively to Big Reed base level. Laurel Fork enters Big Reed within its headwardmost knickzone, and still retains intact relict headwaters despite its size.

A perched meander cutoff loop is preserved ~15 m above the modern Big Reed Island Creek channel ~2.5 km upstream of confluence with the New River. Dissection of the loop is limited, although its northwest arm does carry a small surface channel which appears to sink into karst. The loop is occupied by numerous private residences, but the presence of alluvium is apparent in roadcuts. Terraces of apparently correlative elevation with the undissected portion of the loop are visible along Big Reed Island Creek above and below the loop outlets. Greasy Creek, a major tributary entering Big Reed Island Creek ~21 km above its confluence with the New River, also hosts a meander cutoff perched ~15 m above the active channel.

An interesting factor in the response of Big Reed Island Creek to New River incision is the loss of a ~25 km reach of headwaters to capture. The paleo-Big Reed headwaters is today known as the Dan River, but gravel deposits confirming its former connection to the New River basin were identified by Prince et al. (2010). The loss of considerable Big Reed discharge due to the capture would slow the headward advance of knickzones, but the timing of initial incision into the Blue Ridge Plateau and the capture are uncertain. Inspection of the beheaded Big Reed valley, today called Pine Creek, indicates an unincised morphology that does not reflect New River incision prior to the capture.

A paleo-fluvial feature similar to the Little River loop is preserved above Little Reed Island Creek (21) ~18 km upstream from its confluence with the New River. Inspection of 10 m DEM topography, 1:24000-scale topographic maps, and satellite images indicate a meander cutoff on the southeast bank of the stream which is now preserved ~15 m above the active channel. Rock Creek, a small tributary of Little Reed, follows the northeast arm of the loop and has incised in response the Little Reed incision. The southwest arm does not carry an active channel, and thus appears essentially intact. Rock Creek (22) enters the loop across a major knickpoint known locally as Eagle Falls; this feature would once have entered Little Reed immediately below its knickpoint. The knickpoint thus reflects incision of Little Reed prior to loop cutoff. Post-cutoff incision has now reached the bottom of the Rock Creek knickpoint, increasing its overall height.

The southernmost portion of Virginia's Blue Ridge Plateau is drained by two large streams, Crooked Creek (28) and Chestnut Creek (30), which enter the New River upstream of its crossing of the resistant Blue Ridge quartzite outcrop. The mouths of the two creeks are separated by less than 2 km, with Crooked Creek entering the New River further downstream.

Both streams enter the New River knickzone across their own knickpoints, which appear to begin exactly at their mouths and have experienced very little headward migration. These knickpoints are developed entirely within late Precambrian metasediments of the Blue Ridge Plateau, which are not particularly resistant and would be expected to facilitate rapid knickpoint migration. The small distance of headward migration is consistent with the location of the stream mouths relative to the headwardmost New River knickpoint in Virginia, which is located ~11 km upstream. These streams are thus responding to New River incision event(s) which passed the Little River and Big Reed systems long ago, and have only recently breached the resistant quartzite strata which the New River crosses just downstream of the Crooked Creek mouth. Comparatively deeper incision of the Big Reed system into the Plateau surface suggests Crooked and Chestnut Creeks will experience additional incision when knickpoints presently crossing the quartzite strata advance beyond their mouths.

Chestnut and Crooked Creeks have dissected the Blue Ridge Plateau upstream of their New River confluence knickpoints. Longitudinal profiles of both streams show additional knickpoints further upstream that presumably represent earlier episodes of New River incision. The Blue Ridge Plateau surface within the Chestnut and Crooked basins appears to be continuous with, but less dissected than, the Plateau surface in the Big Reed Island system. Zones of talc-rich ultramafic metamorphic rock in the Chestnut basin and unmapped but field-confirmed migmatitized schist in the Crooked basin produce local topographic anomalies, and these lithologies may have "collected" (Crooked) or blurred knickpoints. In any case, as with the Little River and the Big Reed Island Creek systems, Chestnut and Crooked Creeks have responded to past New River incision events yet retain entirely relict headwaters.

Chestnut Creek contains a "loop" cutoff feature ~10 km above its confluence with the New River. A perched meander loop is preserved above the active channel on the east bank of the stream. Beaver damming has produced two ponds within the low gradient feature. While no active streams now use the loop channel, the north arm is perched ~3 m higher above the channel than the south arm due to a minor knickpoint located in the Chestnut Creek channel (Chestnut Creek Falls) between the two arms. The knickpoint has migrated headward past the northern (downstream) arm, but has not yet reached the upstream arm, which is perched only 2-3 m above the active channel. Like the other two "loop" features, the Chestnut Creek loop provides evidence of active, headwardly migrating pulses of incision spreading throughout the drainage network. While it is not possible to correlate all three loop abandonment events, it may be reasonable to conjecture that abandonment of the loops, which exist on bedrock, may be related to the onset on incision following a period of base level quiescence. The decreasing elevation between loop and active channel moving headward within the New River system may thus reflect the upstream migration of waves of incision. Noteworthy is the presence of an “apparent” loop several kilometers upstream of the Chestnut Creek feature; this loop is the result of straightening the stream to facilitate railroad construction. A new channel was blasted across the neck of land in a tight meander bend, creating a knickpoint and apparent cutoff. Historical documents indicate that the loop closer to the New River is, in fact, a natural feature.

Two major knickpoints occur in the New River ~4 km and ~12 km upstream of the mouth of Chestnut Creek. The knickpoint nearest Chestnut Creek occurs in non-resistant Blue Ridge metasediment units, while the more distant knickpoint appears to coincide with outcrop of a sheared basement unit. Upstream of these knickzones, the New River maintains a much lower gradient profile, despite passing repeatedly passing between non-resistant Blue Ridge metasediment units and more resistant gneisses at the southern limit of the study area. Small tributaries above these knickpoints do steepen as they enter the river, but whether they are slowly responding to a pulse of incision well upstream or “leakage” of incision from the knickpoints immediately downstream. The New River is incised into the plateau above the knickpoints and beyond the study area, suggesting it has experienced episodes of transient incision pre-dating those described in this study.

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