

Field Trip to the Fall Zone, Cape Fear River, Raven Rock State Park

Sean Gallen

Department of Marine, Earth, and Atmospheric Sciences,
North Carolina State University
sgallen@ncsu.edu

Frank J. Pazzaglia

Department of Earth and Environmental Sciences, Lehigh University
fjp3@lehigh.edu

Introduction

The pre-meeting field trip for the 2013 EarthScope National Meeting focuses on the landscape, surficial deposits, and bedrock of the upper Coastal Plain and Fall Zone in central North Carolina, exposed along the Cape Fear River at Raven Rock State Park (Fig. 1). The purpose of this field trip is to provide an intriguing set of observations regarding the long-term landscape evolution of the eastern United States as a platform for framing questions that might be directly addressed by EarthScope and related research. These questions include, but are not limited to the assembly, structure, and evolution of the lithosphere of the eastern U.S., the dynamic flow of the underlying asthenosphere, Appalachian orogenesis, post-Appalachian rift initiation and evolution, and the crustal scale structures that may play a role in the observed seismicity. We cannot hope to see examples of all of these interesting research targets in a half-day trip, but we do intend on providing a platform for these and other questions based on the observables and recent work that will be discussed at Raven Rock State Park.

Our intent is to focus on landscape evolution and the surficial processes, namely river incision and watershed-scale erosion that both respond and dynamically interact with the lithospheric processes that uplift and deform rocks. Field trip participants are likely familiar with the venerable arguments of the cycle of erosion (Davis, 1899) and dynamic equilibrium (Hack, 1960) so this field trip will steer clear of trying to support or falsify either model. Rather, during the trip, we will briefly review what we know and identify how particular observations and data compel or constrain our thinking on this subject. For the eastern U.S. including the field trip location, the interested reader is directed to Doering (1960), Daniels et al. (1978), Owens, (1989), Pazzaglia and Brandon (1996), Spotila et al (2004), Gallen et al., (2013), Portenga et al. (2013) and McKeon et al., (submitted, GSA Bull) for more thorough reviews of the basic long-term landscape evolution arguments.

Landscape evolution is best quantified as changes in the mean elevation, mean relief, rate of erosion, and unsteadiness of that erosion, the latter being a good measure of the degree of disequilibrium and transient processes operating on the landscape. Here, data and observations for long-term (million to tens of millions of years) and intermediate (hundreds of thousands to

million year) scale erosion and landscape change are presented. Excellent examples of how paleorelief could be inferred from geomorphic data are found in McKeon et al. (GSA Bull., submitted) and Gallen et al. (2013).

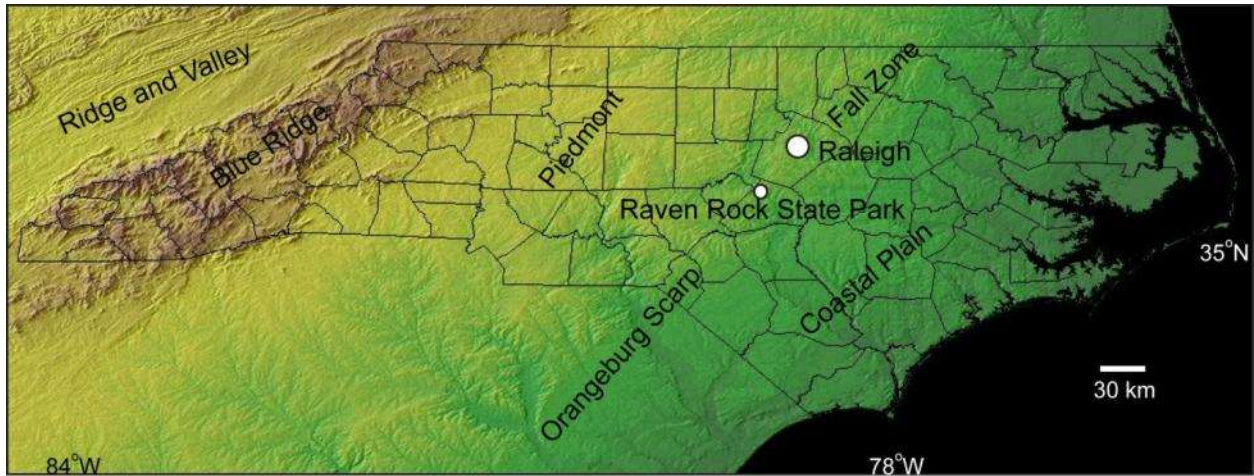


Fig. 1a. Color shaded relief map of North Carolina showing some of the major physiographic and topographic features and field trip location. Modified from <http://www.ngdc.noaa.gov/mgg/topo/img/nc.jpg>.

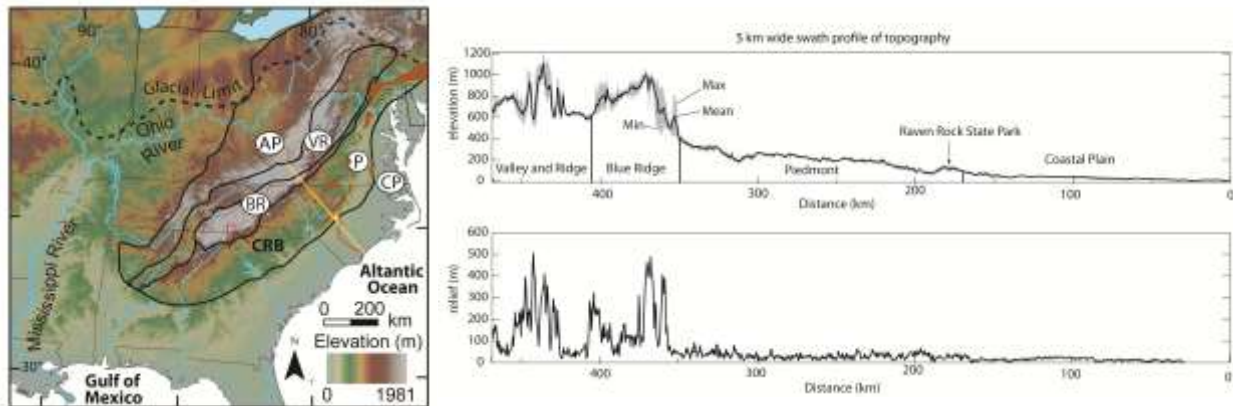


Fig. 1b. Topography and swath cross-section (yellow line) for the southern Appalachians, centered on the Cape Fear River and Raven Rock State Park field trip stop. AP = Appalachian plateau, VR = Valley and Ridge, BR = Blue Ridge, P = Piedmont, and CP = Coastal Plain. Mesozoic rift basins shown in orange. CRB denotes location of Cullasaja River basin (Gallen et al., 2013, see text below).

The geology of eastern North America including North Carolina (This section is updated and modified from Pazzaglia et al., 2011, EarthScope-GeoPRISMS Conference field trip, Lehigh University, with contributions from M. Withjack and R. Schlische).

Eastern North America encompasses the Appalachian Mountains and the archetype Atlantic passive margin and as a result is a source of formative thinking related to continental assembly, orogenic evolution, continental rifting, (reviewed in Sheridan and Grow, 1988; Faill, 1997a,b) and post-rift geodynamic evolution. Key paradigms such as the Wilson cycle (Wilson, 1966; Oliver et al., 1983) and Cenozoic eustasy (Haq et al., 1987; Miller et al., 2008) are based on data and research on this margin. Heterogeneity of the Atlantic passive margin lithosphere is both the result and consequence of the diverse tectonic events that it has experienced over the past billion years (Figs. 2, 3). These events include various arrangements of subduction polarity during Grenville and Appalachian compressive orogenesis (Faill, 1997a,b); continental margin segmentation, rifting and breakup, leading to the opening of the Atlantic Ocean (Withjack et al., 2011, Schlische et al., 2003); voluminous igneous activity associated with the Central Atlantic Magmatic Province (CAMP); the post-rift evolution (Hutchinson, 2005); and unsteady Cenozoic epeirogeny (Pazzaglia and Brandon, 1996). The passive margin geodynamic evolution is superposed on a lithosphere that marks the transition from fully continental Precambrian Grenville basement in the west, to fully oceanic Jurassic Atlantic sea floor in the east.

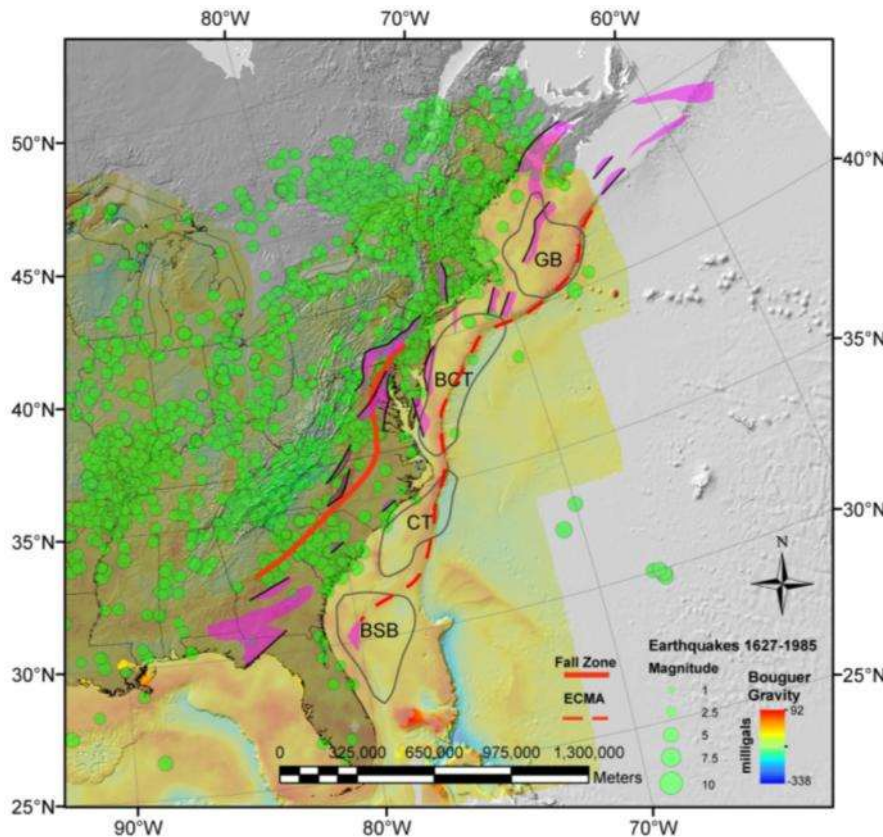


Fig. 2. Map of Eastern North America (ENAM) showing the Bouguer gravity anomaly draped on Appalachian topography and Atlantic Ocean bathymetry. Green circles are earthquake locations. Purple polygons are syn-rift basins with border faults in black. Black-outlined polygons are approximate boundaries of shelf-slope basins; GB = Georges Bank basin, BCT = Baltimore Canyon trough, CT = Carolina Trough, BSB = Blake Spur Basin. Topographic, bathymetric, and geophysical data are all from online USGS GIS data repositories

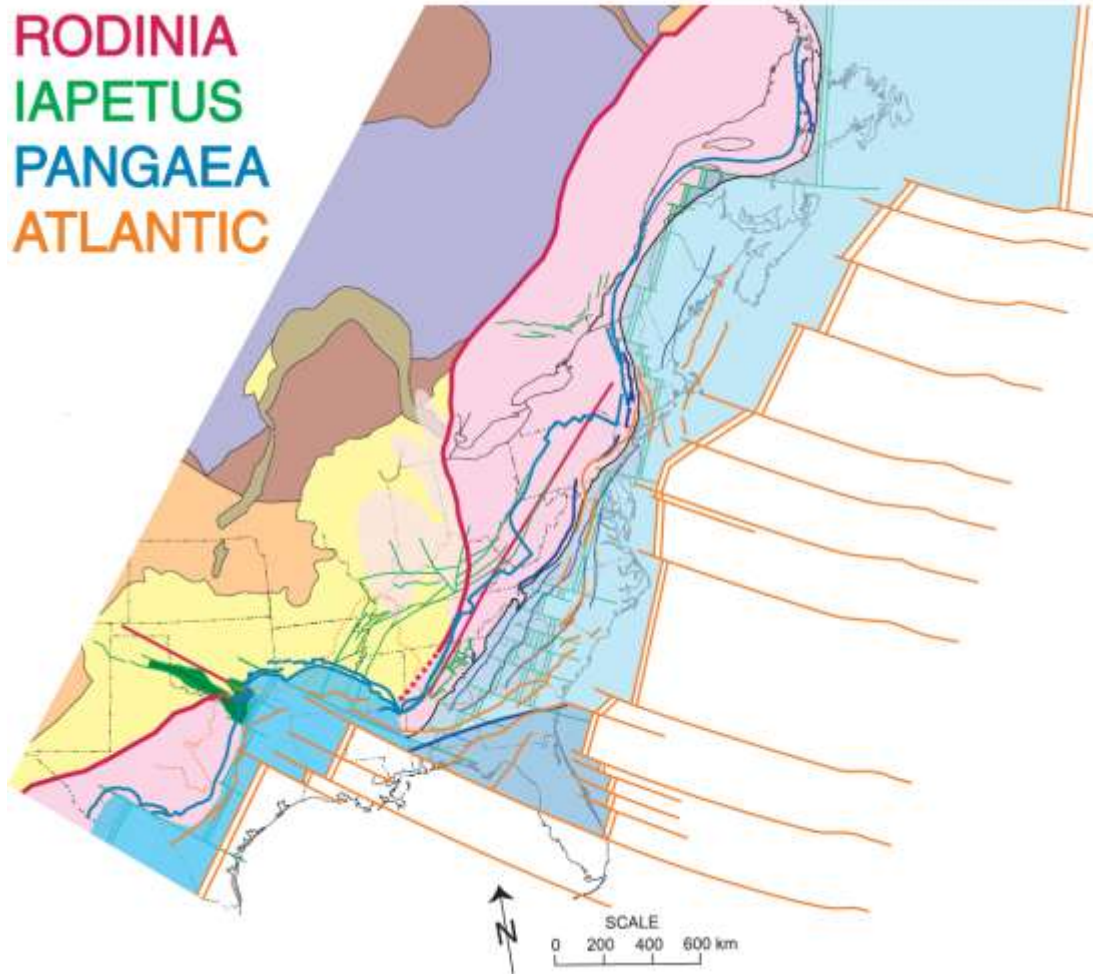


Fig. 3. ENAM margins from the late Proterozoic to the present. (Thomas, 2006).

Much of the ENAM lithosphere was formed during the late Proterozoic Grenville orogeny during assembly of the Rodinia supercontinent, and crust of this age constitutes the basement of the modern passive margin (Whitmeyer and Karlstrom, 2007). A latest Proterozoic-early Cambrian passive margin formed by rifting of Rodinia and the opening of the Iapetus Ocean (Thomas, 2006). Large segments of Grenville crust attenuated and separated from North America during this rifting would later be involved in Appalachian orogenesis. Opening of Iapetus also resulted in two continental rifts, the Catoctin rift and Rome trough that ultimately subsided and were covered by a thick wedge of passive margin siliciclastics and carbonates. Low-magnitude, but persistent seismicity remains concentrated in these rifts.

The Appalachians were constructed on top of this Rodinian rift system and passive margin following a protracted period of collisional tectonics during the Paleozoic and closing of the Iapetus Ocean, culminating in the Permian (Hibbard et al., 2006; Fig. 4 and posters). The traditional interpretation holds that the three great clastic wedges preserved in the Appalachian foreland are related to three pulses of orogenesis during the Paleozoic – the Late Ordovician

Taconic orogeny, the Late Devonian Acadian orogeny, and the Pennsylvanian-Permian Alleghenian orogeny.

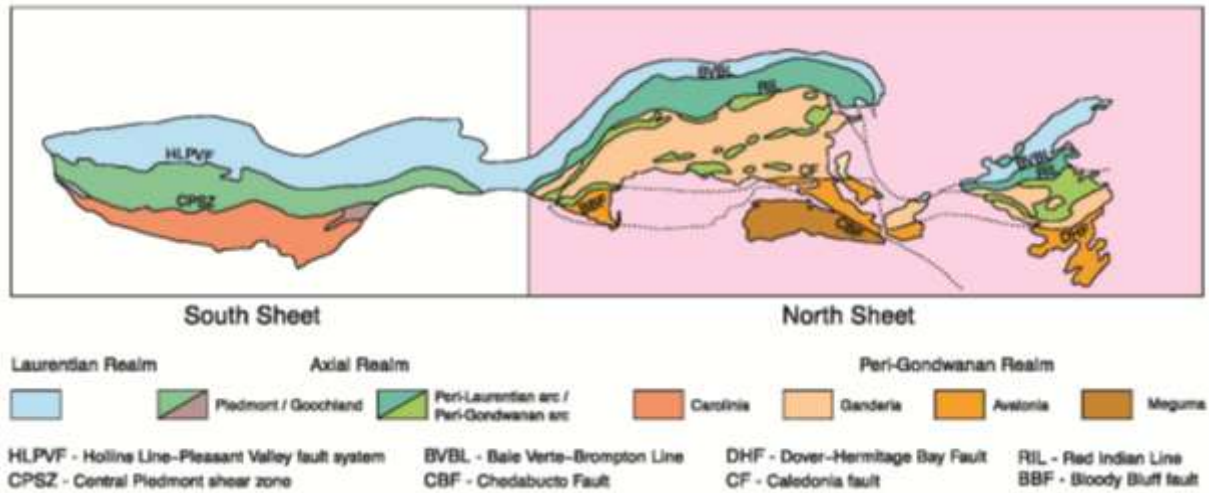


Fig. 4. Summary lithotectonic elements of the Appalachians (Hibbard, 1996).

By the Early Permian (~280 Ma) the Appalachians were a lofty mountain chain modeled to be similar in mean elevation, relief, and width to the modern central Andes (Slingerland and Furlong, 1989). Deformation during this orogeny propagated far westward into the foreland, imbricating much of the former foreland basin, and shedding a thick molassic wedge west across the craton. That wedge was responsible for up to 11 km of burial in the anthracite fields of eastern Pennsylvania (Levine, 1986), 4 km in central West Virginia (Reed et al., 2005), and up to 2 km of burial in the mid-continent (Hegarty, 2007). Deep erosional exhumation of the Appalachian core during and after the culmination of crustal thickening first overfilled, and then unroofed the Appalachian foreland with the detritus being transported westward to the mid-continent and beyond (Riggs et al., 1996; Rahl et al., 2003; Hegarty et al., 2007). The contact between Piedmont and late Triassic rocks shows that erosion reduced Appalachian topography to several hundred meters or less of local relief when rifting began.

Extension during the Mesozoic reactivated many of the pre-existing Grenville and Appalachian structures, producing a series of wide, deep fault-bounded rift basins from northern Florida to the Grand Banks of Canada (Withjack et al., 1998, 201; Withjack and Schlische, 2005; Faill, 2003; Fig. 5). The remnants of these basins, exposed on the margin today, provide a wealth of geologic information about syn-rift and post-rift depositional and deformational processes (Olsen et al., 1996; Withjack et al., 1998). Rifting began in an apparently distributed fashion, often reactivating sutures of Paleozoic accreted terranes, then localized to the present margin location, leaving behind a number of abandoned Mesozoic rift basins adjacent to the successfully rifted margin. Most aspects of this localization, including the role of sutures, however, remain unclear. Rift localization off the east coast of the US and its conjugate was roughly coincident with one of the most voluminous but short-lived volcanic events in Earth's history (i.e., the Central Atlantic Magmatic Province, CAMP), and much of the rifting along the margin was

correspondingly volcanic. In contrast, the northernmost portion of this margin (offshore Newfoundland) is distinctly magma-poor. Here, in many but not all places (like the northern Scotian margin) rifting has left behind wide tracks of highly thinned continental crust and exposed, serpentinized mantle along the margin. Not only does the style of breakup change substantially between these magmatic end-members, variations in magmatism and deformation are also seen on smaller scales between adjacent segments. Fault segmentation is apparent across the margin from abandoned rift basins onshore to oceanic crust offshore, but many questions remain about the development and evolution of segmentation through time. This margin and its conjugate are relatively well preserved and uncomplicated by subsequent tectonic events, making it an excellent setting in which to examine the deformation, magmatism and segmentation that led to continental breakup.

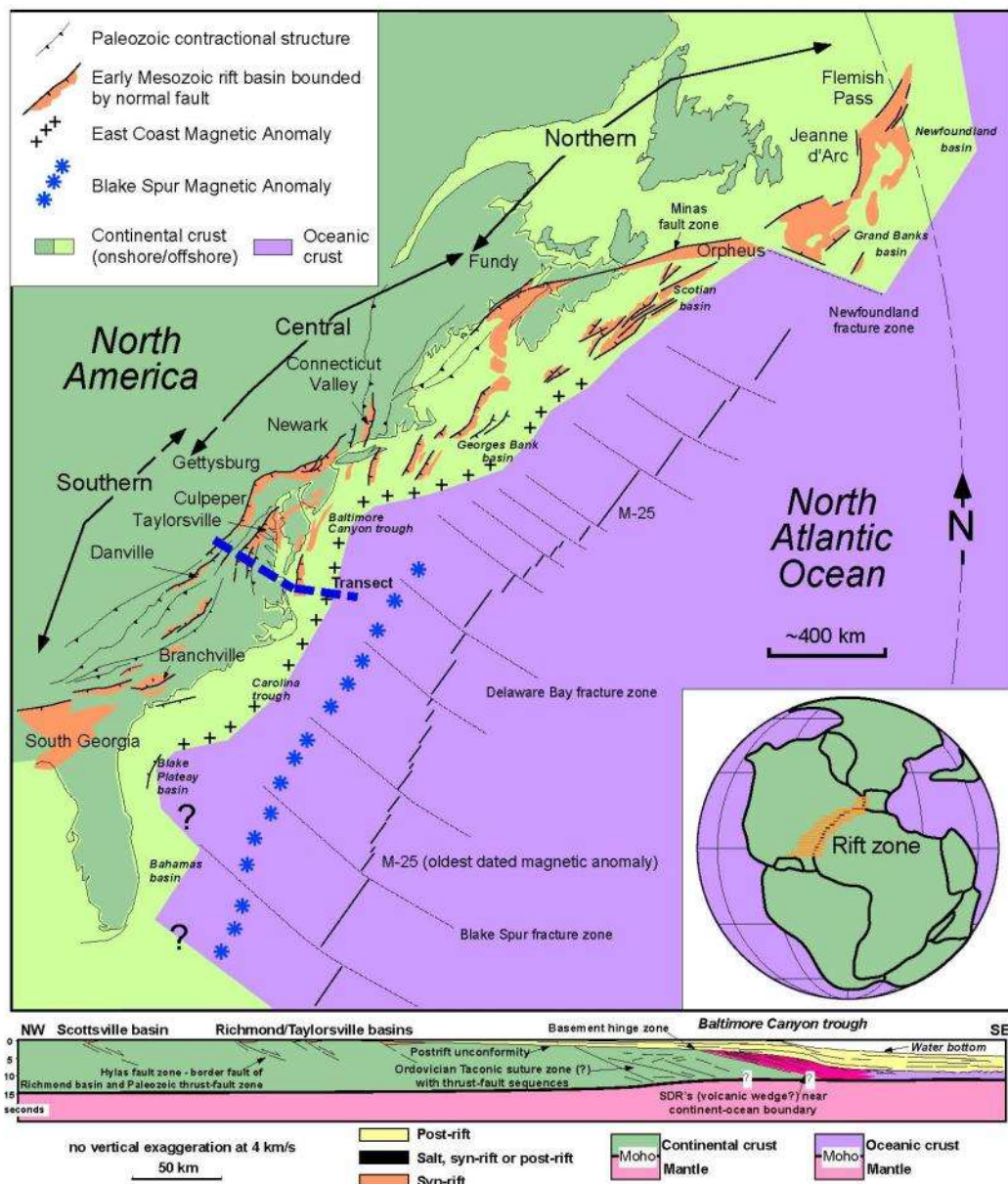


Fig. 5. (previous page). *Mesozoic–Cenozoic tectonic elements of the eastern North American margin. The southern, central, and northern segments exhibit progressively younger ages for the end of rifting and presumably the onset of seafloor spreading. The ECMA approximates the extent of seaward-dipping reflectors at the continent-ocean boundary. The Blake Spur magnetic anomaly may be related to a ridge jump. M-25 is the oldest dated magnetic anomaly; thus, the age of older oceanic crust depends on the inferred spreading rate. Inset shows configuration of the supercontinent Pangea during the Late Triassic (Olsen, 1997), and highlights the rift zone between eastern North America and NW Africa and Iberia. Regional transect through southern segment of margin highlights Paleozoic prerift structures, Triassic-Jurassic rift structures, and Mesozoic/Cenozoic post-rift basins. Modified from Withjack & Schlische (2005).*

After rifting, the rift basins underwent significant erosion (locally >5 km) (e.g., Malinconico, 2010). Much of this erosion occurred soon after breakup, producing a pronounced unconformity between the syn- and post-rift rocks. Additionally, significant deformation occurred after rifting, folding and tilting the synrift strata (e.g., Withjack et al., 1998). Like their post-Rodinian predecessors, low-magnitude, but persistent, seismicity is located along the flanks of the rift basins today (Seeber and Armbruster, 1988; Wheeler, 2006). Rifting rejuvenated the topography and opened up new basins to the east of the foreland which initiated a reversal of Appalachian drainage from formerly towards the foreland (west) to one that was split between the old west-flowing rivers and the newly formed Atlantic slope drainages (Judson, 1975). The formerly low-standing Appalachian basin became a relatively high-standing region and portions of the foreland and Blue Ridge experienced a new pulse of erosion during the Late Jurassic and Early Cretaceous (~140-150 Ma), also recorded by AFT cooling ages (Miller and Duddy, 1989; Roden and Miller, 1989) and delivery of siliciclastic detritus to Atlantic shelf-slope basins (Poag, 1985, 1992; Poag and Sevon, 1989).

The syn- and post-rift geologic, tectonic, and geodynamic development of eastern North America is preserved as a sedimentologic and stratigraphic archive in several shelf-slope basins (Figs. 2 and 5). The long-term depositional and subsidence history of the 400 km long, 100 km wide, and up to 18 km deep Baltimore Canyon trough (BCT) has been particularly well studied (Karner and Watts, 1982; Poag, 1985, 1992; Poag and Sevon, 1989; Steckler et al., 1988, 1999). Collectively, the BCT contains siliciclastic sediment equivalent to about 4 km of rock (Hulver, 1997) removed from an area spanning the modern central and New England Appalachian Atlantic slope. These sediments store a rich record of passive margin forcing mechanisms, such as lithospheric flexure, lower crustal flow, source terrane uplift, basin subsidence, paleoclimate and eustatic sea level changes. Patterns of erosion, transport and deposition evolve through time in response to diverse physical and chemical processes. Gravity-driven sediment transport (e.g., landslides and turbidity flows) destabilize the slope and carry sediment to the deep sea where it may be redistributed by oceanographic processes. The sedimentary section can also be altered chemically via diagenesis, methanogenesis and other processes that are associated with venting of carbon-rich fluids and gasses. Several aspects of the sedimentary wedge make the eastern U.S. ideally suited for passive margin studies. First, because the margin is ‘salt-free’ in many places, many of the processes recorded in the sedimentary wedge can be imaged without limitations

posed by diapiric evaporite bodies common to many other passive margins. Second, sedimentation was nearly continuous and rates were relatively high along the margin providing a robust record of sedimentary environments ranging from glacial-dominated to carbonate.

The geology of North Carolina (Fig. 6) reflects this long history of Proterozoic rifting, Paleozoic assembly, Triassic-Jurassic rifting, and Cretaceous-Cenozoic erosion of eastern North America. The protracted period of late Mesozoic and Cenozoic erosion has etched out distinct physiographic provinces shaped by the underlying rock type and structure (Figs. 1 and 5). From west to east, the Ridge and Valley represents thin-skinned deformation of sedimentary rocks in the eastern part of the Appalachian foreland. The Blue Ridge represents the set of Grenville crystalline thrust sheets with their faulted and folded Paleozoic cover that were emplaced in the eastern part of the foreland. The Piedmont is similar, but underlain by mostly Paleozoic meta-sedimentary and meta-igneous rocks. The topographic break between the Blue Ridge and Piedmont, the Blue Ridge Escarpment, does not correspond to a discrete structure or rock-type change. The Piedmont meets the Coastal Plain along a line of rapids called the Fall line or Fall Zone that has a likely origin in both the change from crystalline rocks to sedimentary cover as well as late Cenozoic epeirogeny. The Coastal Plain has an unsteady and non-uniform subsidence history, recorded as a series of basins and arches visible in the onlapping and allostratigraphic packages of sediment (Owens and Gohn, 1985).

Geophysics of eastern North America (This section is updated and modified from Pazzaglia et al., 2011, EarthScope-GeoPRISMS Conference field trip, Lehigh University, with contribution from Vadim Levin).

Seismological properties of the eastern North American lithosphere have been investigated by numerous researchers, with primary methods being tomographic imaging with different wave types (on regional and global scales), active source studies (COCORP and LITHOPROBE campaigns), studies utilizing shapes of teleseismic body waves (shear-wave birefringence, receiver-function analysis), and investigations using regional seismic activity.

Continent-scale tomographic studies consistently show the lithosphere to be thickest (~300 km) around the Hudson Bay (e.g., Nettles and Dziewonski, 2008; Grand, 1994; van der Lee 2001; van der Lee and Frederiksen, 2005, Darbyshire et al., 2007), and to thin progressively towards the coast (Fig. 7). On a smaller scale, however, this gradual change from continental interior outward is complicated by considerable variability, which is especially pronounced beneath the Appalachian orogen. This variability is further illustrated in contrasting tomographic imaging efforts (van der Lee and Frederiksen, 2005; Nettles and Dziewonski, 2008).

Whereas the intensity of the features varies with technique, there is general agreement on regions of reduced seismic wave speed beneath the coastal plain, likely extending across the entire Appalachian orogen, and also eastward under the ocean. Menke & Levin (2001) used surface waves propagating from the Mid-Atlantic ridge to the coast to show that within the low-

speed zone enclosing Cape Cod (referred to as a “divot” by Fouch et al., 2000) upper mantle rocks are ~3% slower than in the area farther south. Low seismic wave speed in the upper mantle is commonly associated with excess temperature. It is however unclear why, given their largely uniform tectonic history, some regions beneath the Atlantic margin should be considerably warmer than others.

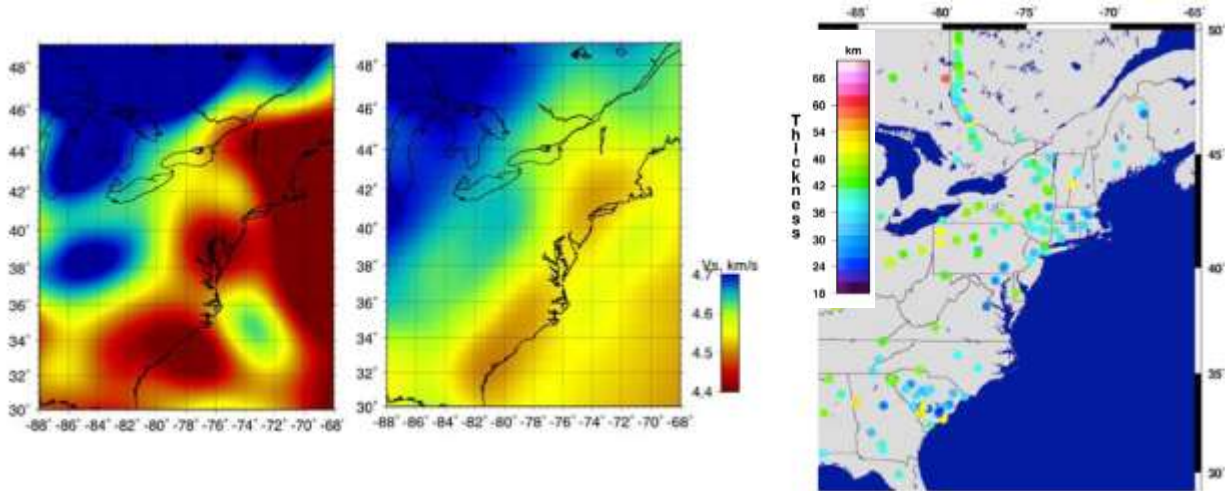


Fig. 7. (Left and center images) Horizontal slices at ~100 km through tomographic models by van der Lee and Frederiksen (2005) (left) and Nettles and Dziewonski (2008) (center) show areas of relatively low seismic wavespeed along the Atlantic passive margin of the North American continent that cut across the strike of the Appalachian orogen. Particularly notable is an area between Cape Cod and the Great Lakes. (Right image), Estimates of crustal thickness using reverberations of teleseismic P-to-S converted waves in the crust (from ears.iris.washington.edu).

Attempts to determine lithospheric thickness beneath the Appalachians yielded a range of outcomes, not all of them compatible. Surface wave tomography (van der Lee, 2002) suggests relatively thin lithosphere (80 km) beneath the Appalachians, although the author noted a difficulty in defining the exact depth where the lithosphere ended. This estimate agrees to a degree with studies of converted-mode body waves by Rychert et al. (2005, 2007), which identified sharp seismic impedance contrasts at depths of 90-110 km, and interpreted them as the base of the lithosphere. On the other hand, compressional-wave tomographic imaging of a small region at the junction of the Grenville and Appalachian terranes (Levin et al., 1995) identified a significant change in the degree of lateral heterogeneity of the upper mantle at ~300 km, which was taken by the authors to represent the transition from the lithosphere to the asthenosphere.

Several studies document seismic anisotropy throughout the eastern U.S. (see Fouch and Rondenay, 2006, for review; Wagner et al., 2012a); however, the relations of the observed anisotropic texture to the present plate motion, asthenospheric processes, and the history of tectonic events are not fully worked out. Various authors explained the observations in terms of a textural remnant from the time of continental accretion (e.g., Barruol et al. 1997), asthenospheric flow modulated by lithosphere shape (Fouch et al., 2000; Forte et al., 2010), and a combination

of both resulting in at least two layers of texture (Levin et al., 1999). In their recent paper, Wagner et al (2012a) use shear wave splits to evaluate the relative contributions of lithospheric deformation and asthenospheric flow to regional anisotropy (Fig. 8). They interpret their results as being consistent with drag-induced asthenospheric flow beneath the older section of the North American continent to vertical or incoherent flow, likely in combination with complex lithospheric anisotropy, beneath the Appalachians and rifted margin. In addition, they find several non-null splitting measurements that are not aligned with the North American plate, but rather with magnetic anomalies that might be a lithospheric record of continental suture or faults.

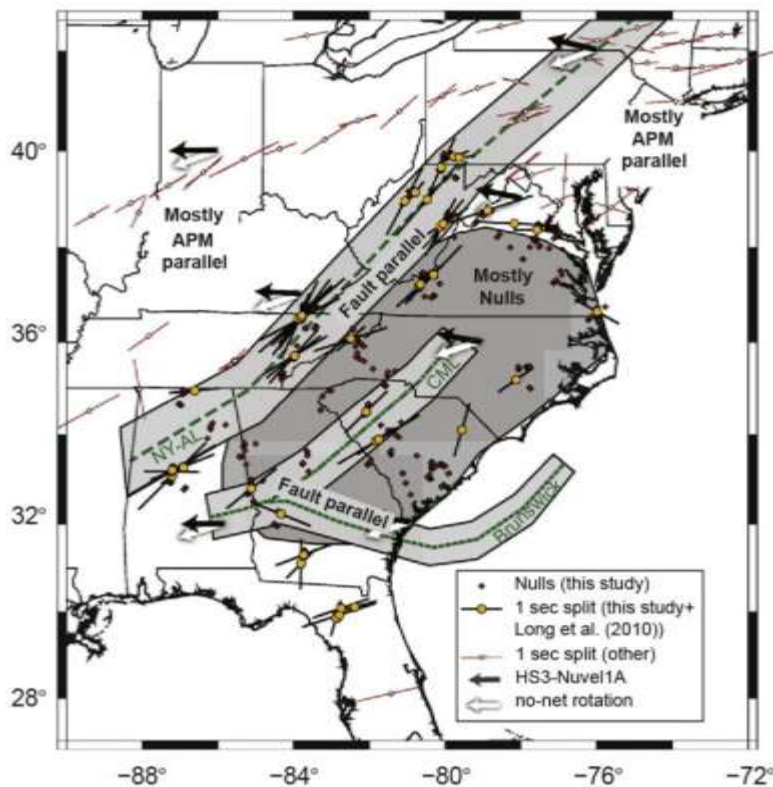


Fig. 8. Interpretive map showing null and non-null shear wave splits of Wagner et al., (2012a) and Long et al. (2010). See Wagner et al., 2012a for study details.

Crustal thickness beneath the Appalachians (Fig. 7) determined by means of active source seismic surveys as well as with teleseismic body wave reverberations varies in the 30-45 km range. A recent study using ambient noise imaging (Bensen et al. 2009) presents uniform maps of crustal thickness throughout the North American continent. A notable feature of this map is a difference of at least 5 km in average crustal thicknesses of northern and southern Appalachians. On the other hand, automated computation of crustal thickness from receiver functions (ears.iris.washington.edu) does not show such a clear difference between these regions.

For the southern Appalachians, crustal thickness beneath the Blue Ridge is interpreted to be about 46 km thick (Wagner et al., 2012a), consistent with some other recent studies (Hawmann, 2008), and thicker than the COCORP interpretations (Cook and Vasudevan, 2006; Fig. 9). More importantly and relevant to investigations that lend themselves well to the

EarthScope facility, receiver functions have been interpreted to be consistent with crustal and lithospheric scale structures responsible for thickening the lithosphere beneath the Blue Ridge (Wagner et al., 2012b; Fig. 10).

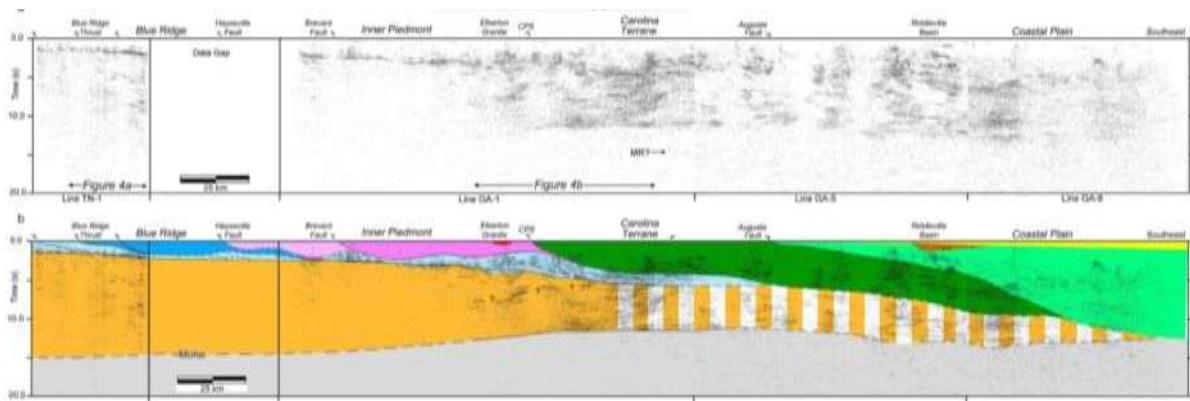
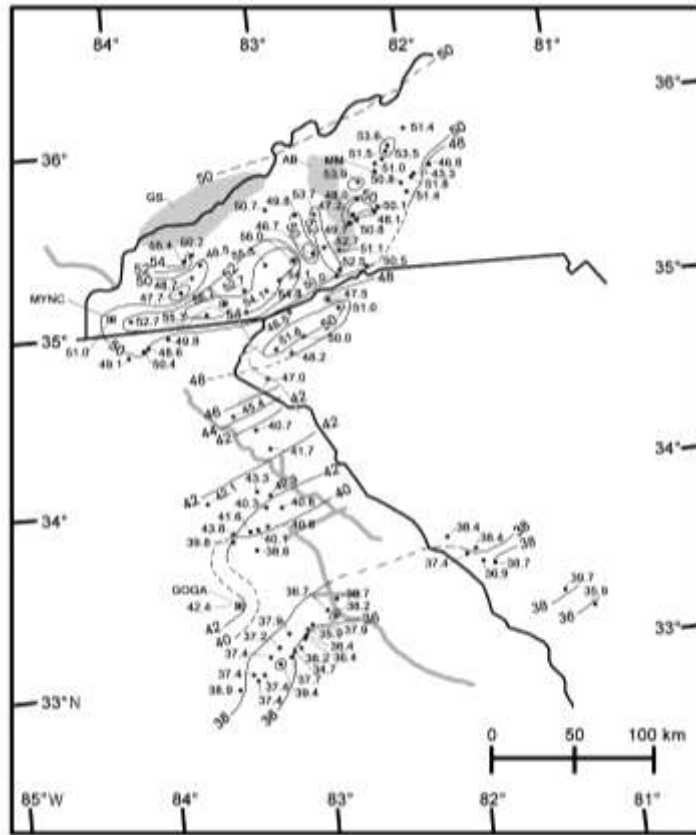


Fig. 9. (top) Crustal thickness in western North Carolina, western South Carolina, and eastern Georgia, showing locations of COCORP lines in gray (see Hawman et al., 2012 for details). (bottom) Migrated, coherency filtered, and uninterrupted data from COCORP lines TN-1, GA-1, GA-5, and GA-8. A portion of the west end on the line GA-5 has been removed due to overlap with line GA-1 (above). Interpretation from Cooke and Vasudevan (2006) (below).

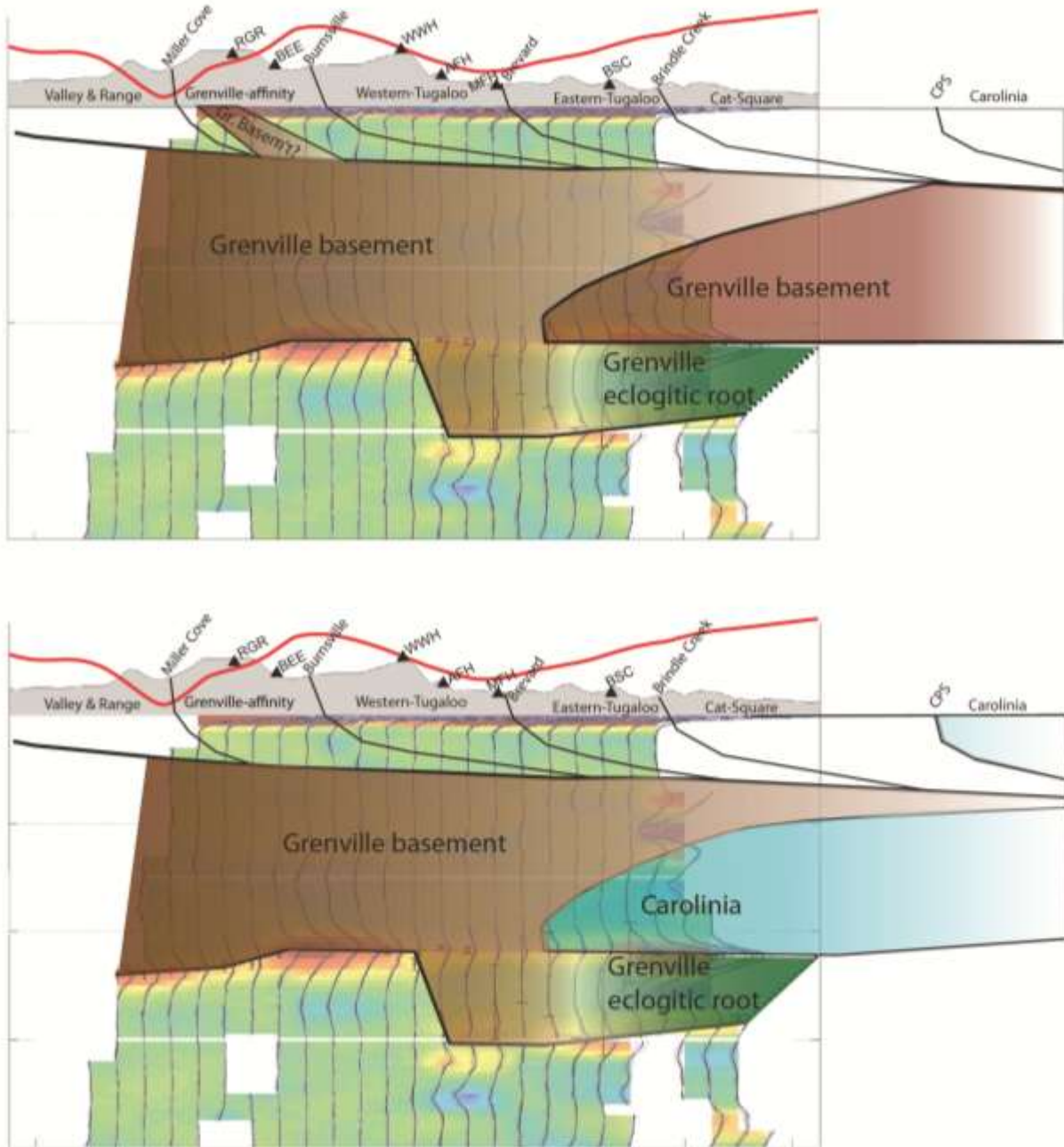


Fig. 10. *Contrasting tectonic interpretations of receiver function data for crustal structure beneath the Blue Ridge of western North Carolina. See Wagner et al., 2012b for study location and details.*

The crust of eastern North America is surprisingly seismically active (Fig. 2). There are also some fairly tight concentrations of seismicity, e.g. in the Montreal-Ottawa area, New York City area, St. Lawrence Valley, and central Virginia. Reactivated features from past tectonic events (e.g., a Ramapo fault system in NJ and NY states) clearly play a role in where earthquakes

take place (e.g., Sykes et al., 2008). In some places (e.g., Adirondacks), occasional earthquakes take place at >30 km depth, which is very unusual for the continental crust (Sbar and Sykes, 1977).

Bouguer and isostatic residual gravity maps of the mid-Atlantic margin (Simpson et al., 1986; Fig. 1) indicate major gravity lows, as in southeastern Pennsylvania where it is interpreted as a particularly deep sedimentary basin above basement (Shanmugam and Lash, 1982) and in the southern Appalachians where it is interpreted as a post-orogenic crustal root (Fischer, 2002). The topography, particularly at shorter wavelengths, does not clearly reflect crustal or lithospheric buoyancy predicted by the gravity data. A similar conclusion has been reached by using wide-angle three-component seismic waves (Hawman et al., 2012).

Framing the problem – Geodynamic predictions of uplift and subsidence for the Eastern U.S.

Collectively, the geomorphic, geologic, and geophysical observables are direct evidence for unsteady ENAM epeirogenic deformation which includes uplift and erosion of the Appalachians and its foreland, and non-uniform subsidence of the passive margin basins. There are several possible causes of this epeirogeny of which isostatic (Fisher, 2002), flexural isostatic (Pazzaglia and Gardner, 1994), and dynamic (Moucha et al., 2008; Flament et al., 2013) mechanisms have already been proposed and modeled. Not tested, but of equal potential importance are effects, both proximal and distal, of the now well-documented Chesapeake impact structure, the subsidence history of which is known to be both long-lived and unsteady (Hayden et al., 2008). In summary all of these geodynamic processes have profound implications for interpretations of eustasy and sequence stratigraphy (Miller et al., 2008). More importantly, ENAM geodynamic research opens the door for exploration of surface processes-lithospheric dynamic interactions that can be tested against geologic and stratigraphic archives.

Dynamic models that make predictions about the uplift, subsidence, and deformation of east coast stratigraphic and geomorphic markers are all rooted in sub-lithospheric mantle heterogeneities that are generally attributed to the progressive subduction of the Farallon slab (Ren et al., 2007; Moucha et al., 2008; Flament et al., 2013). First-order geomorphic markers (Ator et al., 2005; Fig. 11) like the Fall Zone (Pazzaglia and Gardner, 1994), the Blue Ridge escarpment (Spotila et al., 2004), the Orangeburg scarp (Dowsett and Cronin, 1990), and the Cape Fear arch (Owens, 1989) can be used to deconvolve eustatic from epeirogenic processes (Raymo et al., 2011).

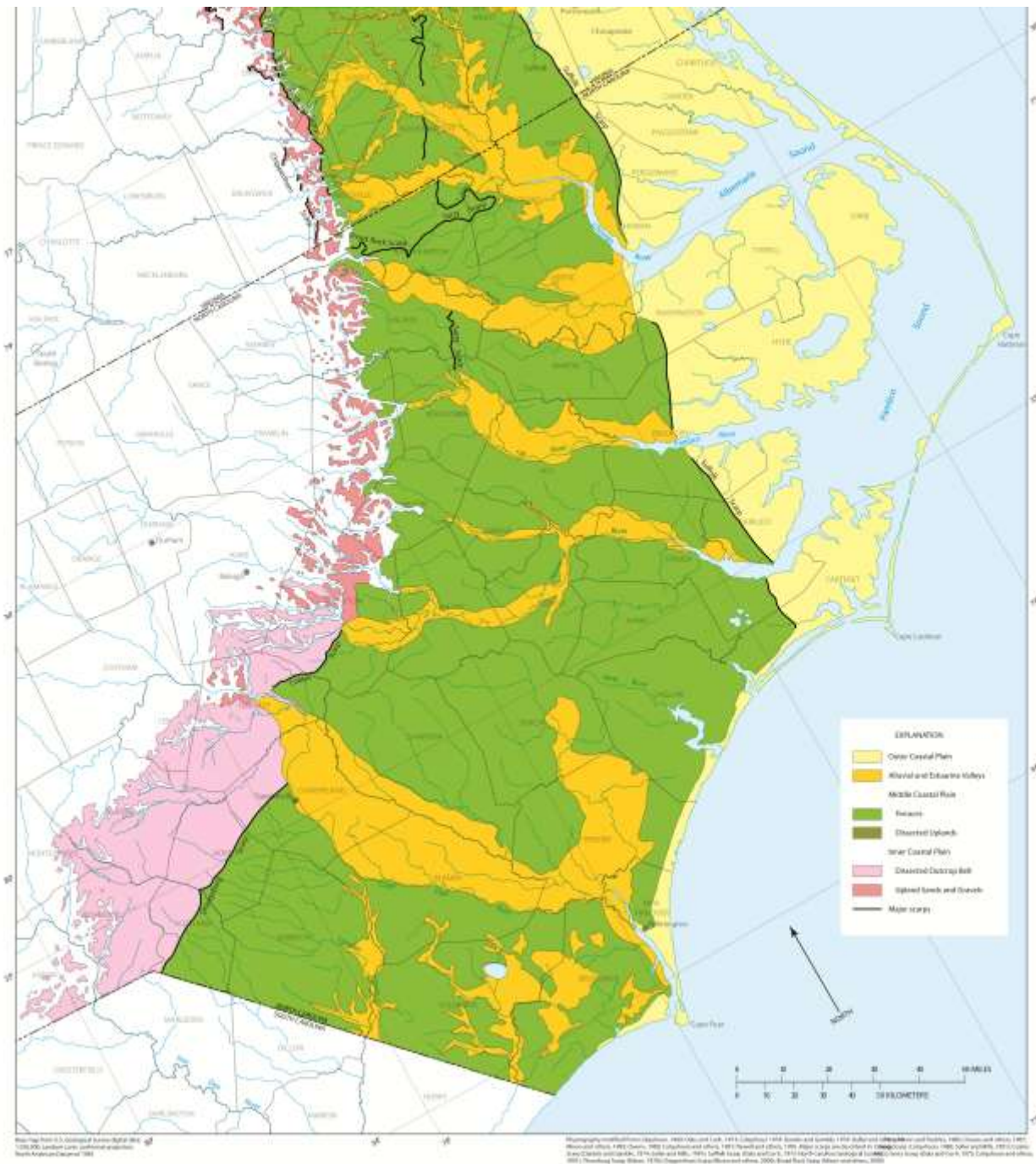


Fig. 11. Southern portion of the Ator et al. 2005 physiographic map of the mid-Atlantic Coastal Plain (USGS Professional Paper 1680) showing upland gravels and prominent inner Coastal Plain scarps.

Landscape evolution observables

Thermochronology

Thermochronology, or the study of how certain mineral phases cool as they are exhumed from deep in the crust to the surface of the Earth, provides a robust method for determining erosion rates over very long periods of time. Although the method can typically only give long-term averages, recent understanding of how cooling rate and mineral composition influence results has allowed for more sophisticated modeling and the inference of erosional (cooling) unsteadiness using these data. One common mineral used in thermochronology is apatite because of its inclusion of radioactive U and Th in the crustal lattice. Through both spontaneous fission (Apatite Fission Track – AFT) and alpha decay (U-Th/He – AHe), apatite provides at least two ways to reconstruct mineral cooling by erosional exhumation. Apatite retains lattice damage generated by fission tracks when the mineral cools below $\sim 100^{\circ}\text{C}$. In contrast, the U-Th/He parent-daughter system becomes closed to rapid diffusion and escape of He when the mineral cools below $\sim 60\text{-}70^{\circ}\text{C}$. Given the gentle geothermal gradient of $15\text{-}20^{\circ}\text{C}/\text{km}$ for the central Appalachians, these temperatures correspond to depths of $\sim 4\text{-}5$ and $3\text{-}4$ km respectively.

Here, we summarize some recent U-Th/He data for the central and southern Appalachians (McKeon et al., in review, GSA Bulletin). U-Th/He cooling ages for Pennsylvania ($n=31$) range from $\sim 60\text{--}240$ Ma and show different pooled ages for the Ridge and Valley (~ 198 Ma), Blue Ridge (~ 98 Ma) and Piedmont (~ 172 Ma) provinces (Fig. 12). The Blue Ridge cooling ages are more tightly clustered and do not have any overlap with the Piedmont or Ridge and Valley samples that exhibit considerable overlap in their ages. Ridge and Valley and Piedmont AHe mean cooling ages are younger than corresponding AFT (Roden and Miller, 1989) and ZFT (Kohn et al., 1993) cooling ages respectively; however, there is overlap in the cooling age ranges for these different thermochronometers. Assuming a uniform geothermal gradient of $\sim 20^{\circ}\text{C}/\text{km}$ for all of Pennsylvania and New Jersey over the cooling history of these samples, it appears that the Ridge and Valley and Piedmont provinces cooled through 70°C in the Early to Middle Jurassic and the erosion rate since that time has averaged ~ 16 m/m.y. (3000 m in 185 m.y.). For a similar geothermal gradient, the Blue Ridge, including presumably South Mountain has been eroded more rapidly at ~ 31 m/m.y. (3000 m in 98 m.y.).

A somewhat more sophisticated treatment of these data, including samples spanning the entire Appalachians is presented in McKeon et al. (in review, GSA Bulletin). Modeling of the cooling history of samples collected from ridge tops and valley bottoms in the southern Appalachians, they find that between 160 and 60 Ma, the ridge tops have eroded at a rate of ~ 20 m/m.y. whereas the valley bottoms eroded more rapidly at a rate of ~ 35 m/m.y. (Fig. 13). The implication is that relief grew in the Appalachian landscape through the Cretaceous and into the early Cenozoic. Since ~ 60 Ma, relief has been slowly decreasing and the mean rate of erosion for both the ridge tops and valleys has been ~ 20 m/m.y. It is not yet known if the McKeon et al. (in review, GSA Bulletin) results could be extended into the South Mountain landscape in describing its long-term erosion history.

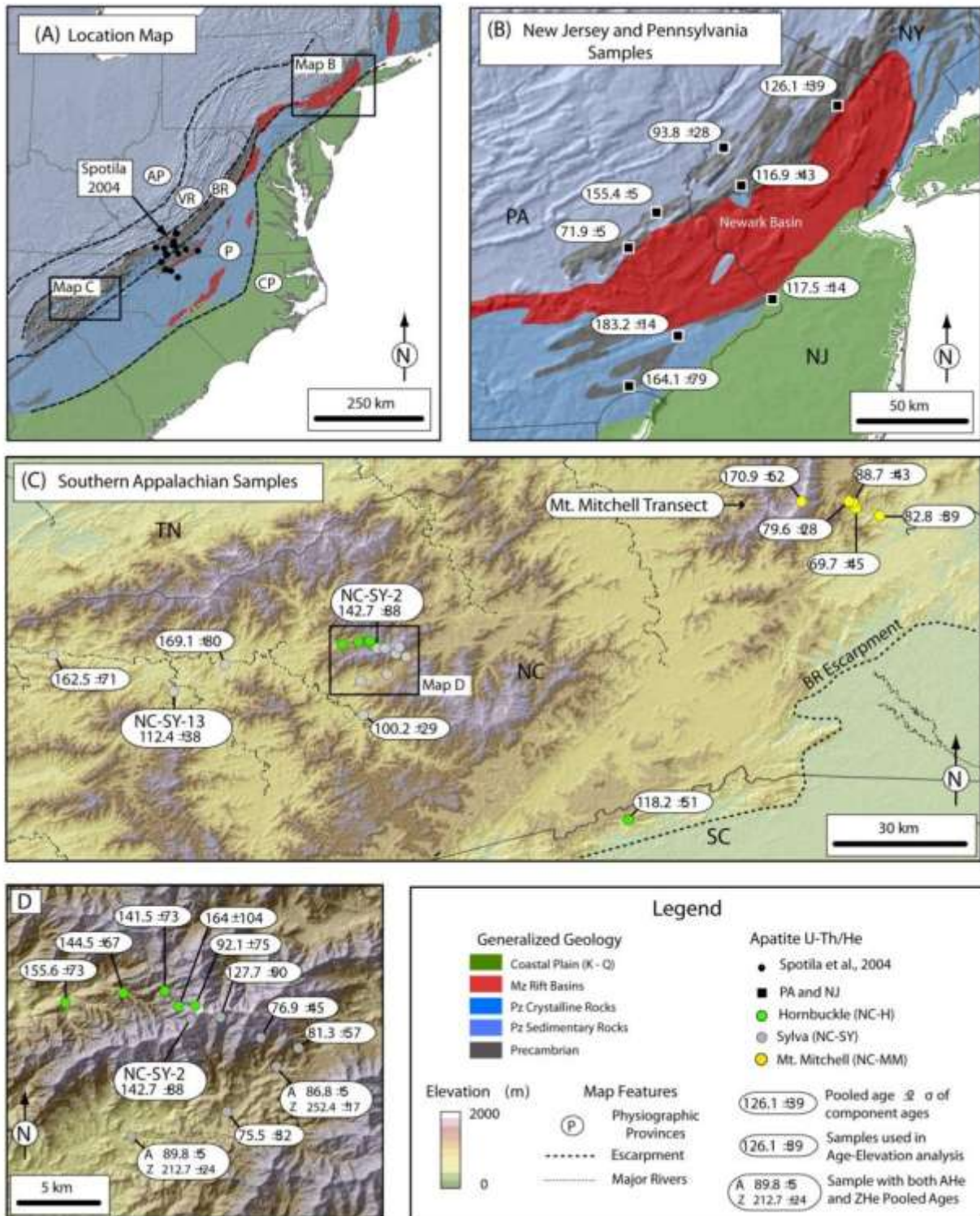


Fig. 12. Summary of AHe thermochronology data for the central and southern Appalachians (from McKeon et al., in review, *GSA Bulletin*).

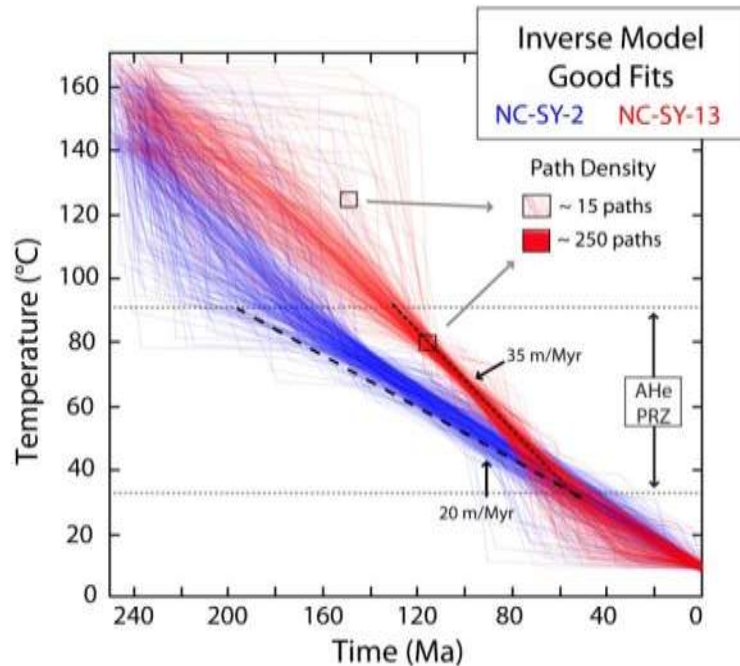


Fig. 13. Comparison of inverse modeling results for AHe samples NC-SY-2 and NC-SY-13 showing only the time-Temperature (tT) paths that produced statistically good fits to the input data from the exponential regressions. The paths are displayed with 90% transparent lines, thus the darker blue (NC-SY-2) and red (NC-SY-13) colors indicate where the random tT paths overlap. The path density examples show the number of paths that pass through an approximately 5 Ma by 5 °C window to produce the color saturation observed. The dark colors within the PRZ for samples illustrate how the good fits follow dominantly linear cooling paths (approximated by the dashed lines) that suggest the different positions in the landscape represented by the two samples experienced different erosion rates. The ~20 °C offset between the samples prior to 120 Ma corroborates our assumption of the assumed geothermal gradient of 20 °C/Km because the samples are vertically offset by 1.2 km.

Rare pre-Neogene deposits preserved in the otherwise erosional Appalachian landscape

The western Piedmont and Great Valley of the Appalachians preserve rare occurrences of deposits that attest to the local antiquity of the landscape in an otherwise erosional setting. The well-known Cretaceous lignite at Pond Bank, Pennsylvania (Pierce, 1965) places some constraints on how much relief, erosion, and chemical dissolution is possible in the Appalachian landscape over tens of millions of years. The lignite contains upper Cretaceous (early Campanian, ~80 Ma) terrestrial pollen, distinctly lacking any marine palynomorphs, encased in part by residuum derived from the host carbonate bedrock. This biostratigraphic age overlaps with the upper part of the Potomac Group lithostratigraphy represented by the Raritan and Magothy formations in New Jersey and eastern Maryland and Patapsco Formation in western Maryland and is at least in part consistent with the lignite common in the Magothy Formation. The highest sea levels in the Cretaceous are Cenomanian in age (~92 Ma) coincident with Raritan-Patapsco Fm deposition, thus it is possible that the Pond Bank lignite represents the fluvial, upstream equivalent of these Coastal Plain deposits. The Pond Bank deposit is unique for the Great Valley in Pennsylvania and it is probably no coincidence that it was preserved in a

carbonate sink hole given that the thermochronology argues for ~ 3 km of average erosion since its deposition. As long as the rate of carbonate dissolution matches the long-term rate of unroofing fixed by the thermochronology, the Pond Bank deposit could have been lowered vertically through carbonate dissolution, escaping removal by mechanical erosion. Measurements of carbonate dissolution in Pennsylvania (White et al., 1984, 2000) range from ~8-30 m/m.y., values consistent with the long term average determined by the thermochronology. Using an insoluble residue content of 10% for the carbonates at Pond Bank, Pierce (1965) inferred that at least 430 m of carbonates have weathered beneath the gravel to produce the residuum below the lignite-bearing beds.

Base level fall, river incision, and knickpoint retreat

River channels in the Appalachian landscape that drain across the Piedmont and Fall Zone, containing concave-up graded reaches with intervening steep, convex reaches containing one or more knickpoints. Similarly, numerous rivers that drain westward off the southern and central Appalachians into the Gulf of Mexico also exhibit upward convexities and knickpoints, in their longitudinal profiles (Gallen et al., 2013; Prince and Spotila, 2013). Where these knickpoints are not fixed by some particularly resistant bedrock, they are presumed to have been caused by unsteady, possibly rapid base level fall since the late Miocene and have been since

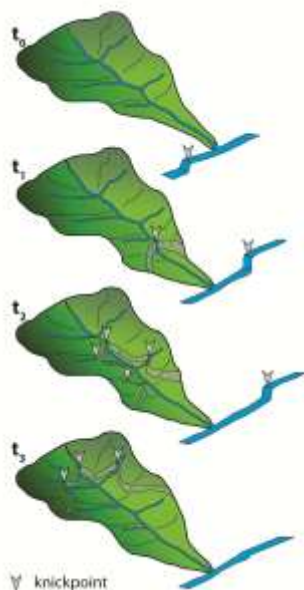


Fig. 14. Illustration of knickpoint migration and a wave of fluvial incision moving up through a drainage basin from a single base level fall.

migrating upstream (Fig. 14). Many knickpoints found along these rivers cut into uniform lithologies and have no clear relation to bedrock structures, indicating they are not related to resistant rock-types but are mobile features propagating headward in the landscape. Such observations are curious implying an external forcing and base level fall is responsible for the generation of these knickpoints. Earlier research has illustrated that base level fall imposed by eustatic fluctuations during Quaternary climate oscillations are likely to be confined to the lower few hundred kilometers of the alluvial Mississippi River (Schumm, 1993) and the Coastal Plain of the Atlantic Slope Rivers (Coleman et al., 1990). It is, therefore, unlikely that changes in sea level can be called upon to explain the genesis of knickpoints now located more than 1500 river kilometers near the Eastern Continental Divide (e.g. Gallen et al., 2011). The two remaining hypotheses for base level fall are: (1) changes in the power of stream incision brought on by a transition to a cooler, stormier and more rapidly fluctuating climate that initiated 3 – 4 Ma and continues to the present-day

(Molnar and England, 1990; Molnar, 2004); or (2) Cenozoic epeirogenic uplift of the southern Appalachians (Pazzaglia and Brandon, 1996; Gallen et al., 2013).

For the central Appalachians, the lower Susquehanna River provides an excellent example of the effects of ~100 m of base level fall in the past ~10 Ma (Pazzaglia and Gardner, 1993, 1994). Correspondingly, river incision in the Pennsylvania Piedmont since the late Miocene is ~10 m/m.y. as a long term average, but can be as rapid as ~250 m/m.y. (Reusser et al., 2005) in the late Pleistocene. At Marietta PA and within the zone evidently impacted by the post-Miocene base level fall outside of the Piedmont, an excellent suite of Pleistocene terraces (Engle et al., 1996) indicates a river incision rate of ~20 m/m.y.

Similarly, for North Carolina and the southern Appalachians, the Cape Fear River is broadly convex at the Fall Zone and through the Piedmont reach (Fig. 15). There is nearly 100 m of incision into the adjacent rolling uplands at Raven Rock State Park.

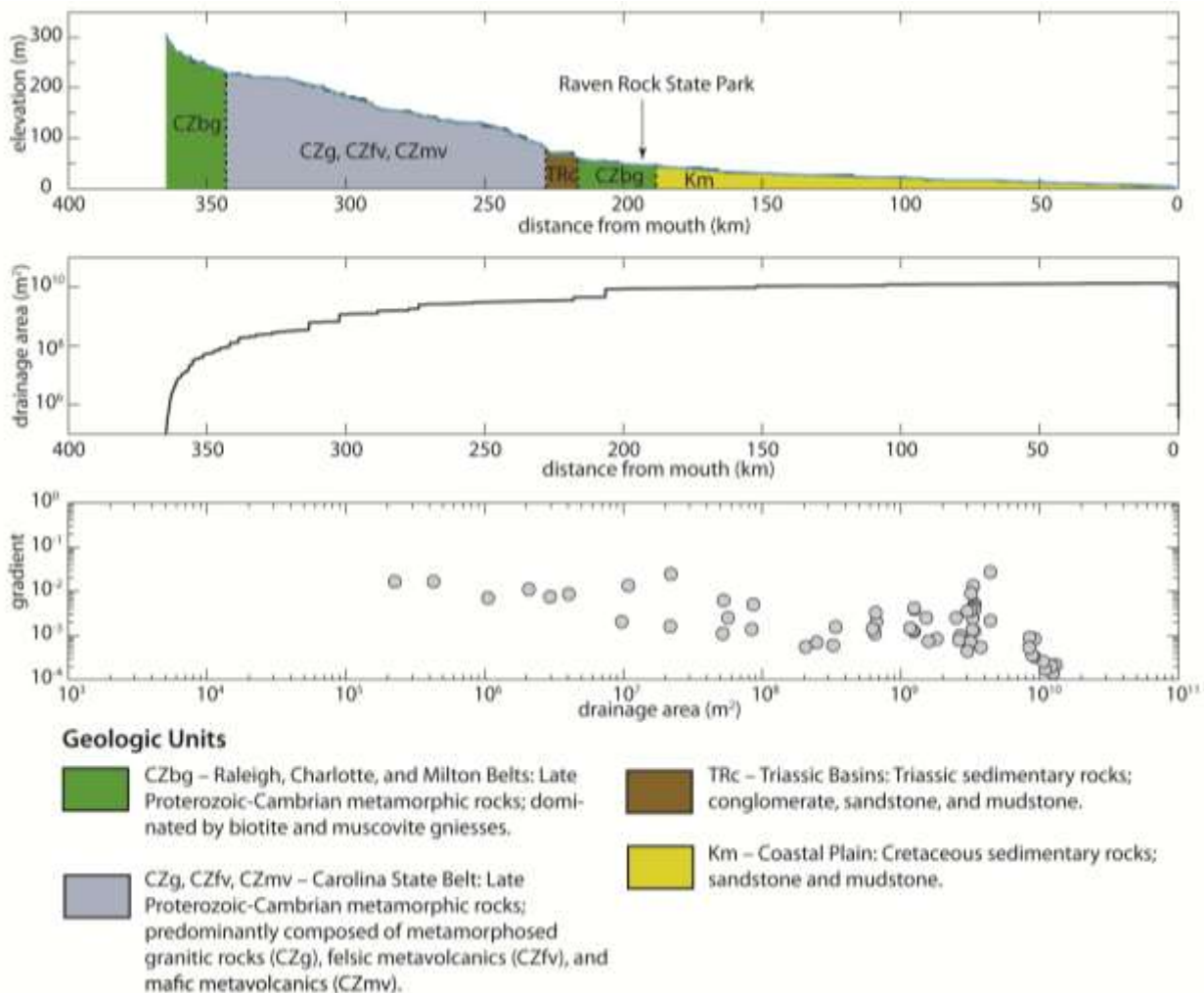


Fig. 15. Longitudinal profile (above), growth of drainage area (middle), and channel slope-drainage area plot for the Cape Far River, North Carolina.

A recent study in the Cullasaja River basin, located in the southern Appalachians of western North Carolina (Fig. 1b), utilized a series of knickpoints to estimate the timing of base level fall in the study catchment and determine the rates of transient landscape evolution in response to such an event (Gallen et al., 2013). Approximately 50 knickpoints are observed in rivers draining the Cullasaja River basin (Fig. 16). Knickpoints cluster in five altitudinal bands indicating that five independent waves of incision are actively propagating upstream (e.g. Niemann et al., 2001). The highest set of knickpoints marks an important transition in the landscape, separating low-relief terrain with thick soils and a low frequency of landslides above the knickpoints from high-relief rugged terrain with a thin soil mantle and frequent landslides below the knickpoints. The Ahnert (1970) relationship, which approximates mean basin erosion rate using mean local relief as a proxy, suggests that erosion rates in the high relief-low altitude portions of the basin are three to four times greater than the low relief-high altitude landscape isolated by the knickpoints. Longitudinal river profile reconstructions illustrate that that base level at the mouth of the Cullasaja basin has dropped ~ 0.5 km and basin relief has increased $\sim 160\%$ since the highest set of knickpoints entered the mouth of the river (Fig. 16).

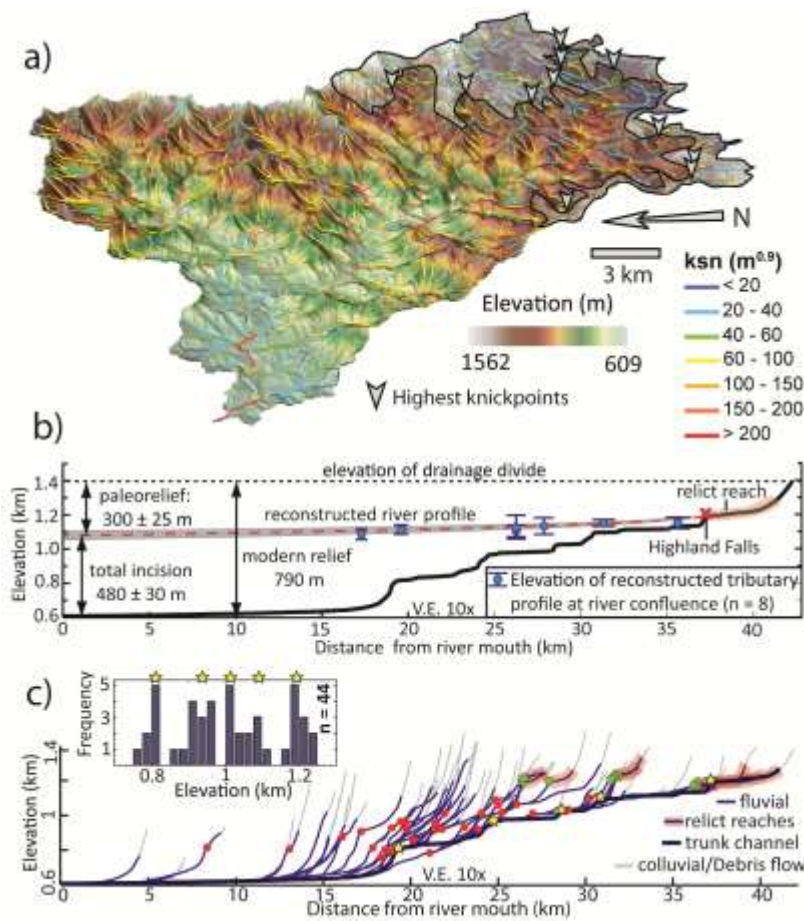


Fig. 16. a. Perspective view of the Cullasaja river basin with the relict landscape highlighted. The normalized channel steepness (k_{sn}) of the fluvial network is overlain on the digital topography. b. Modern and reconstructed paleo-river profiles with the modeled elevations of tributary-trunk channel junctions projected to the paleo profile. The 2σ elevation errors are from the normalized steepness indices and are based on linear regressions through log-log channel slope-drainage area data (Gallen et al., 2013). c. Longitudinal river profiles of 52 streams showing the location of 44 knickpoints, equilibrated and relict reaches, and the approximate transition between fluvial and colluvial/debris flow dominated channels that occurs at drainage areas $\geq 1.25 \times 10^5$ m². Inset is a histogram of the elevations of the 44 knickpoints identified in the Cullasaja Basin showing the defined altitudinal clustering of knickpoints, which is interpreted to represent five pulses of base-level fall propagating through the basin (Gallen et al., 2013).

Using a mass-balance approach Gallen et al. (2013) estimated the amount of time required for the highest set of knickpoint to travel through the Cullasaja basin to be $8.5 \pm 8/4$ Myr. This value represents an extreme minimum estimate for the timing of base level fall as the knickpoints are likely sourced far from the mouth of the Cullasaja River basin. Such time estimates place the origins of the knickpoint into the Miocene, well prior to a transition to Quaternary like climate oscillations that initiated 3-4 Ma. Numerical modeling was used to constrain the rates of knickpoint retreat, where the average knickpoint velocity is $5 - 10 \text{ mm yr}^{-1}$ and present-day speeds are $\sim 1 \text{ mm yr}^{-1}$. Rates of river incision were determined to be comparable to directly measure values from Appalachian draining rivers, between 24 and 140 mm yr^{-1} .

The findings of this study illustrate that a significant phase of topographic rejuvenation initiated in the southern Appalachians in the Miocene. The effects of this transition dictate the pace of landscape evolution in the southern Appalachian to the present-day. Importantly, the timing of this change in topographic form and geomorphic observations in the Cullasaja River basin argue against Quaternary like climate change are the driver of this base level fall, making late Cenozoic (Miocene) epriogeny a more attractive hypothesis.

Cosmogenic and chemical dissolution-determined rates of erosion and landscape change

The recent work of Price et al (2008), Hancock and Kirwan (2007), and Portenga et al (2013) offer excellent summaries of what we know about these background rates of erosion of bedrock exposures in the Appalachians. Coupled with cosmogenic radio nuclide (CRN) determination of basin-wide erosion rates determined from river alluvium (Matmon et al., 2003; Reuter, 2005) these data allow for a landscape view of denudation over $10^5 - 10^6$ yr time scales. Portenga et al (2013) report an average ridge-line erosion rate of exposed bedrock of $\sim 9 \text{ m/my}$ and landscape-scale averages of all bedrock outcrops of 6 m/m.y. which compares favorably to those reported by Hancock and Kirwan (2007). Similarly, geochemical mass balances for watersheds in the Pennsylvania Piedmont of Lancaster County determine that the rate of chemical dissolution of non-carbonate bedrock accompanying its conversion to saprolite is occurring between ~ 4.5 and 6 m/m.y. (Price et al., 2008). These rates account for one-third to one-half of the long-term, watershed-averaged CRN-determined rates of $\sim 10 - 20 \text{ m/m.y.}$ obtained by sampling channel alluvium (Reuter, 2005). The degree of landscape change as measured by the increase or decrease in relief hinges on what is done with the chemically-weathered bedrock once it is produced. If saprolite or other chemically altered bedrock remains on the landscape and weathering profiles thicken through time because of no surface erosion, relief should increase because the rates of river incision are clearly faster than these hillslope and ridgeline chemical weathering rates. Conversely, it is well known that conversion of bedrock to saprolite slows as the thickness of saprolite grows (Heimsath et al., 1997), favoring the establishment of a steady state thickness for the weathering profile. Under these circumstances, the landscape can attain a near steady-state relief where the combination of chemical dissolution of rock and the gravity-driven mechanical erosion of creep/colluvial processes combine to match

the fluvial incision rate. Even under these conditions, transients in the fluvial system like knickpoints will act to grow the relief because channels can respond to base level changes much more readily in the Appalachian landscape in comparison to the hillslopes.

A more direct comparison of hillslope weathering rates for the central Ridge and Valley is offered by other recent work that has emerged from the Shale Hills Critical Zone Observatory in central Pennsylvania (Ma et al., 2010). Here, the relief and sedimentary rock substrate compare reasonably well to South Mountain. Soil production has been calculated using a novel U-series technique (Dosseto et al., 2008) along a ridge to valley bottom catena. The rates of soil production vary from ~45 m/m.y. at the ridge top to ~17 m/m.y. for the toeslope. These rates are not in conflict with the CRN data reported above because they are calculated for soil-mantled, rather than exposed bedrock parts of the landscape. In fact, the faster ridge-line rates compare well to those calculated by Braun (1989) estimated from the volume of fill in colluvial hollows along the periglacial fringe of Pennsylvania. Particularly for those parts of the landscape that are in the latitude or elevation-defined zone impacted by Pleistocene periglacial processes, ridge lines may be lowering faster than the rest of the landscape, resulting in a reduction of relief for these regions.

These findings also reinforce the notion that portions of the eastern United States are experiencing transient forcing due to ongoing adjustments to base level fall (Prince et al., 2011; Gallen et al., 2011, 2013). Gallen et al. (2011) showed that the upstream migration of knickpoints imparts a quantifiable topographic and landscape response. Using airborne LiDAR and detailed landslide inventories (Wooten et al., 2008), these authors observed statistically significant increases in local relief, hillslope steepness, and the frequency of landslides on hillslopes adjacent to stream reaches immediately downstream of fluvial knickpoints (Fig. 14). These data are used to infer that the landscape adjusts in the passing wake of mobile knickpoints. These findings imply that external base level adjustments are important in setting the pace of local and likely regional landscape evolution in the Appalachians.

Inner Coastal Plain and Fall Zone stratigraphy

The Inner Coastal Plain and Fall Zone is mantled by a series of poorly-dated, but post-Cretaceous fluvial and coastal deposits that record changing base level and unsteady delivery of sediment from the Appalachian foreland. Collectively, these deposits are called upland gravels and they represent one of the only means for knowing the stratigraphic or biostratigraphic age of geomorphic markers, retreating knickpoints, and periods of erosion in the Appalachian landscape. Nomenclature and ascribed ages for these upland gravels is complex and a full review of the pertinent literature is beyond the scope of this field guide. Here we summarize a working nomenclature and stratigraphy for the Raven Rock State Park region derived and modified from Daniels et al. (1978), Blackwelder and Ward (1979), the North Carolina Geologic Map (1985), and Owens (1989) (Fig. 17).

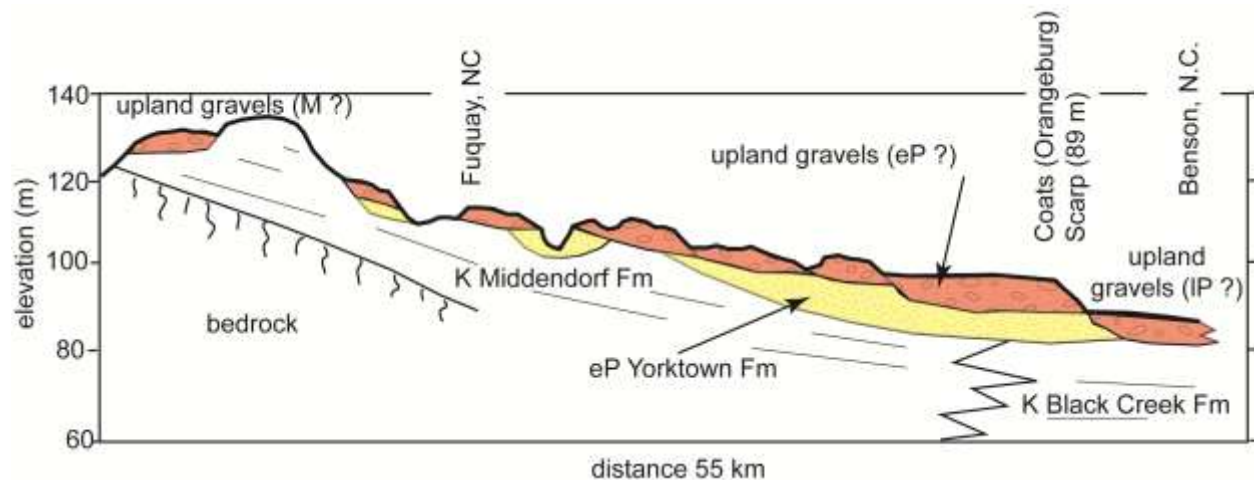


Fig. 17. NW to SE cross section of the Neuse River – Cape Fear River divide between Raleigh and Raven Rock State Park showing the stratigraphic relationship between the marine Inner Coastal Plain deposits and upland gravels. Modified from Daniels et al. (1978).

Basal units of the Inner Coastal Plain include Cretaceous through Eocene neritic sand, silt, clay, minor gravel, and limestone that dip eastward 1-3 degrees. These units onlap weathered Piedmont rocks at the Fall Zone where the contact is marked locally by a well-developed, deep saprolite. Unconformably overlying these Coastal Plain deposits are late Miocene(?) and early Pliocene neritic and shoreface sand of the Yorktown Fm (Raleigh, NC and north) and Duplin Fm (Cape Fear River and south). Thick deposits of Miocene marine deposits are not known from the Cape Fear region, owing perhaps to active Cenozoic arching of the area. In contrast, to the north into the Salisbury embayment and extending into the Chesapeake Bay region, a very thick Miocene sequence of quartz-rich silicilastics of the Chesapeake Group indicate active subsidence of the Coastal Plain and synchronous delivery of sediment from the Appalachian Piedmont, Blue Ridge, and foreland.

At Raven Rock State Park and extending south into South Carolina, the Early Pliocene marine units are deposited seaward of a prominent erosional scarp called the Orangeburg escarpment, typically interpreted as marking the landward extent of early to middle Pliocene sea level. West of the Orangeburg scarp and resting unconformably on Cretaceous and Piedmont rocks are a series of coarse fluvial sand and gravel units mapped as Tertiary (Neogene) terraces on the North Carolina State map and as upland gravels on Owens (1989). These deposits are typically interpreted as the updip, fluvial equivalents of Neogene units in the lower or outer Coastal Plain. Such lithostratigraphic correlations are established for the Salisbury embayment based on sediment provenance (Pazzaglia, 1993) and biostratigraphic data (Pazzaglia et al., 1997). Similar correlations in North and South Carolina are less clear. In some places, such as the Cape Fear-Neuse river divide (Raven Rock State Park), the Orangeburg scarp truncates the upland gravels, indicating that they are Early Pliocene or older. In other places, the gravels appear to cross the scarp and are contiguous with the Duplin Formation, making them Early Pliocene or younger in age. Particularly extensive and thick deposits of these upland gravels are

found near Lilesville, NC. Based on their elevation above the Pee Dee River and mean rates of river incision for the Appalachians compiled by Mills (2000), the age of these deposits has been argued to be ~7-10 Ma (McLean et al., 2012), an age that agrees well with the Bryn Mawr and Brandywine gravels of Maryland and Pennsylvania that are biostratigraphically dated (McCarten et al, 1990; Pazzaglia et al, 1997). In any event, upland gravels on the Fall Zone represent important geomorphic and stratigraphic markers that speak to base level stability and deposition at or near sea level ~3 - 10 million years ago. These gravels are now 100+ m above sea level. Whatever component of that elevation cannot be attributed to eustasy must result from late Cenozoic epeirogeny of eastern North America.

Capping the upland gravels and the Inner Coastal Plain mostly south of the Cape Fear River is another, enigmatic deposit of mostly pure quartz sand called the Pinehurst Formation (Conley, 1962). It is interpreted to contain fluvial, aeolian, and possibly tsunamigenic facies (Farrell and Hoffman, 1994) so its utility as a geomorphic and stratigraphic marker for epeirogeny is limited.

Raven Rock State Park

Raven Rock State Park is situated along the Fall Zone that forms the boundary between the physiographic provinces of the piedmont and coastal plain (Fig. 1). Numerous bedrock exposures throughout the park are composed of highly deformed quartzite, quartzo-feldspatic gneiss, and mica-schist of the Carolina terrane. The Carolina terrane is a volcanic arc exotic to North America (Dennis and Shervais, 1991). Once forming the margin of Gondwana (present-day Africa and South America) before Neoproterozoic rifting, the rocks of the Carolina terrane record docking to the Laurentian (ancestral North America) margin during the mid-to-late Paleozoic (Hibbard et al. 2006). There is a 575 ± 12 Ma date for these rocks that comes from a quarry just south of the nearby Umstead State Park and is assumed to have been collected from the Big Lake-Raven Rock Schist unit (Goldberg, 1994).

The Cape Fear River flows to southeast through the park, forming a series of rapids that were exploited by earlier European settlers for fishing and mill power. These rapids form a broad convexity in the Cape Fear River longitudinal profile (Fig. 15). Contrast in the erodibility of the weathering resistant piedmont rocks and the relatively weaker coastal plain sedimentary units are largely responsible for this convexity. However, steep cliffs along the river in Raven Rock State Park suggest a dynamic landscape. Indeed, along some of the hiking trails in the park, such as the Campbell Creek Trail and Raven rock loop trail, river gravels can be found 10s of meters above the modern river channel (Stewart and Roberson, 2007). The age of such deposits is uncertain, yet they provide evidence of continued river incision in the Piedmont.

Maps and descriptions of the geology, geomorphology, and planned activities at the Park.

This three hour field trip will visit three stops via 3 km of trails in Raven Rock State Park, Lillington, NC (Figs 18 and 19). We will begin by walking east from the parking lot down the southern portion of the Raven Rock Loop trail to Raven rock proper (~1.2 km). Along portions of the trail you will observe rounded upland gravels, perhaps covered by sands of the Pinehurst Formation. While the age of these gravels remains uncertain, based on stratigraphic relationships these deposits are likely have their origins in the late Miocene. One interpretation suggests that they were deposited by the paleo-Cape Fear River.

Stop #0 – Parking lot of Raven Rock State Park. We will assemble near the parking lot of the park, possibly utilizing the pavilion for a brief geological, geophysical, and geomorphologic introduction to the Fall Zone of North Carolina. As we set out on the Raven Rock Loop Trail, we will want to notice the flat upland surface at ~100 m in elevation that is mantled with quartz sand and gravel. This is part of the upland gravel described above and may be late Miocene-early Pliocene in age. These gravels are exposed on the trail as we begin to descend into the dissected part of the topography. Note the contrast between the rounded, fluvial upland gravels and the angular quartz clasts scattered throughout the weathered bedrock (saprolite) as you drop below 100 m in elevation.

Stop #1 – Raven Rock: Here we descend ~30 m (100 ft) in to the Cape Fear River gorge. Note the exposed bedrock or the Carolina terrane on the way down. The gorge itself is a testament to the amount of river incision that has occurred here in geologically recent times. **We will stop here briefly for discussion of the bedrock and the Fall Zone.**

Climbing back up the stairs we will continue westward (to the right) on the Raven Rock Loop trail to the overlook (~0.3 km)

Stop #2 – the overlook: From this vantage point there is a good view of the Cape Fear River and gorge. Looking to the west (left) a small series of rapids can be observed. A portion of these rapids are found at Stop #3. **We will continue our discussion of the Fall Zone and river knickpoints at this location.**

Turning back to the trail, we will walk to the west (right) down the Raven Rock Loop trail to the Fish Traps trail and turn right (~0.6 km). At the intersection of the two trails you can see some more of the upland gravels. The Fish Traps trail traverses down to the river (~0.3 km).

Stop #3 – Cape Fear River rapids: Looking at the river these are a series of small rapids. These would be identified as a convexity in the Cape Fear River longitudinal profile. **We will have an in-depth discussion of the Gallen et al., 2013 paper and its implications for long-term landscape evolution in the Appalachian landscape.**

Follow the Fish Traps trail southward back to the parking lot (~0.6 km)

Raven Rock State Park

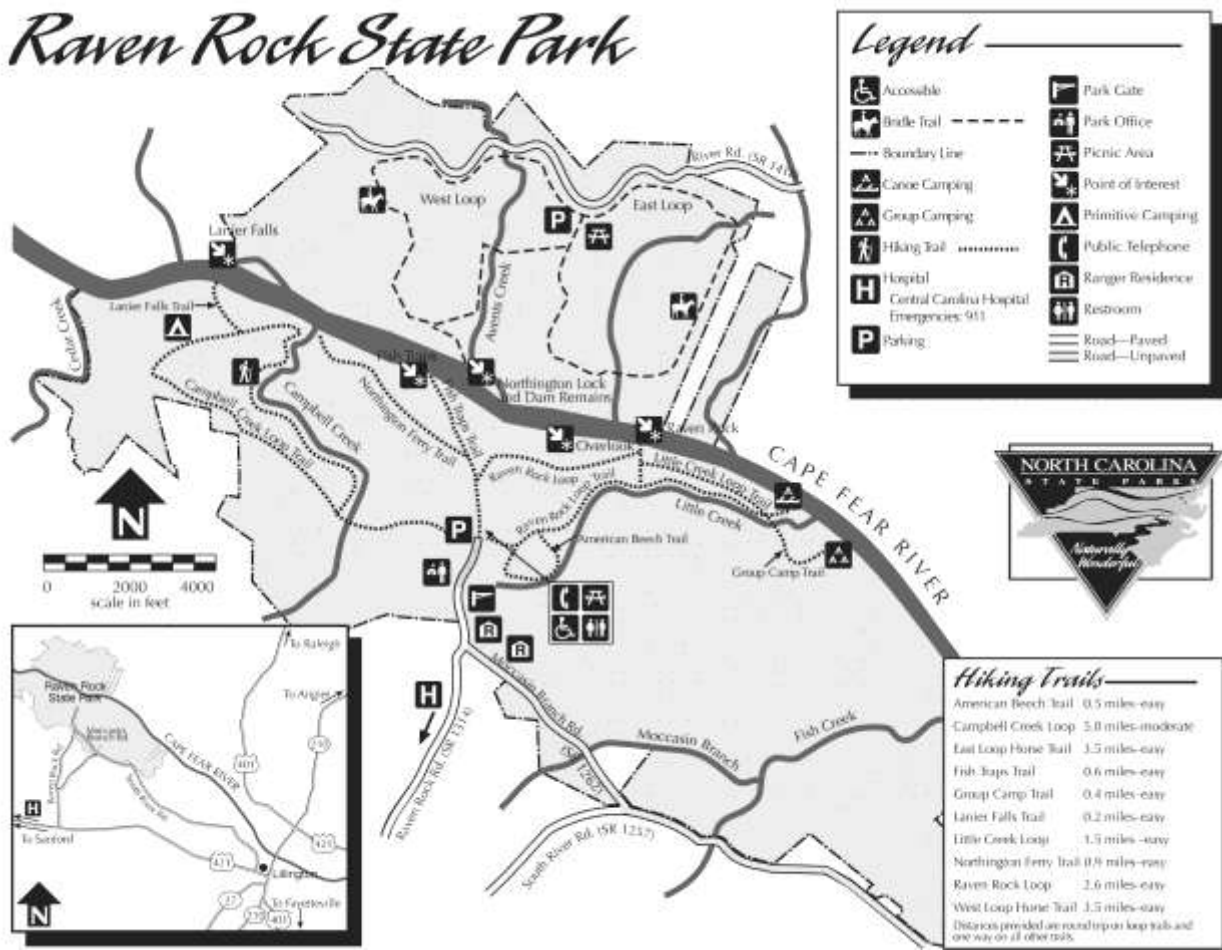


Fig. 18. Map of Raven Rock State Park showing hiking trails and facilities.

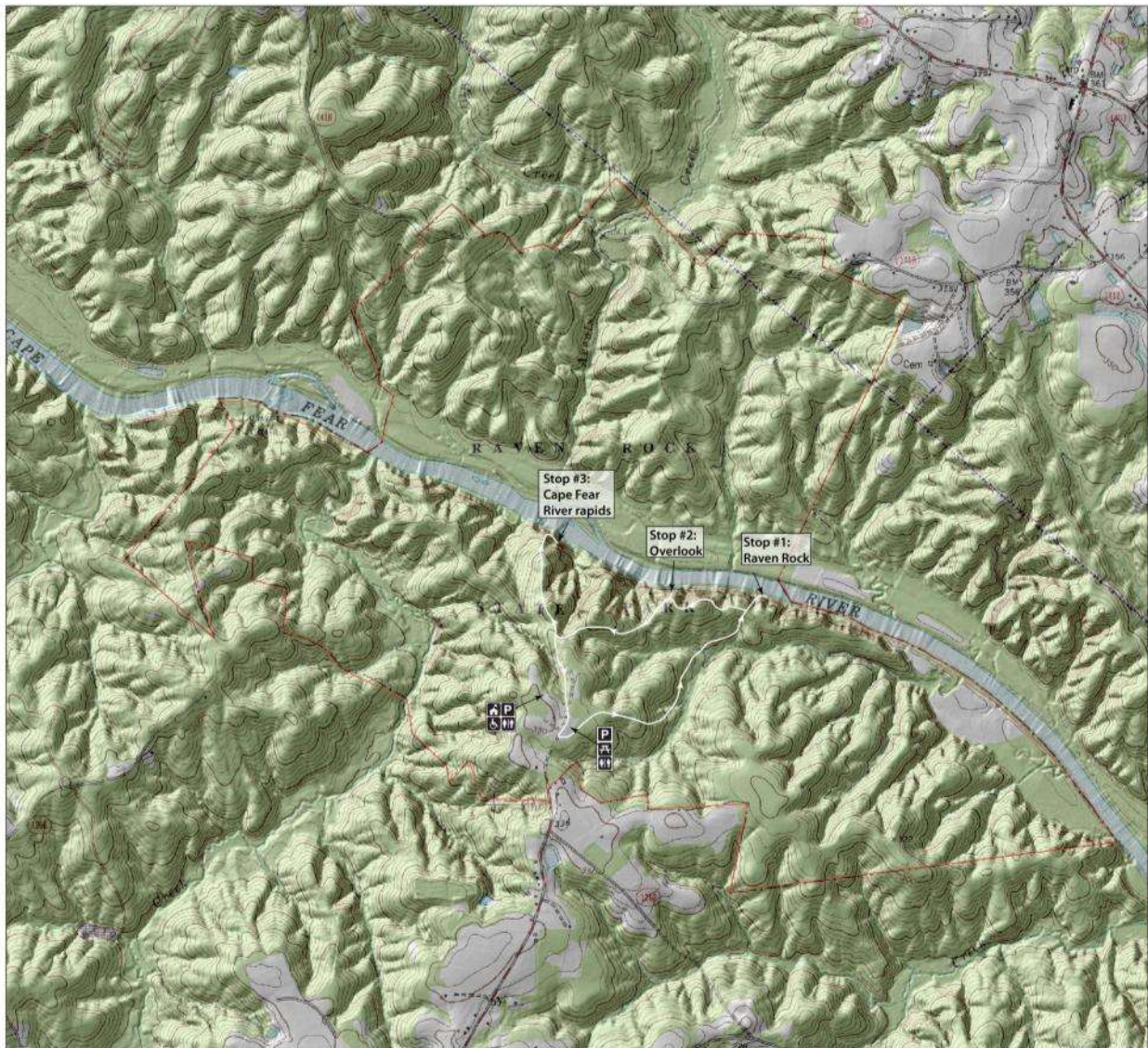


Fig. 19. Marners 7.5 minute topographic quad draped over LiDAR DEM of the Raven Rock State Park area showing field trip route.

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