GeoPRISMS – EarthScope Eastern North America Workshop (ENAM) Workshop
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Post-Meeting Field Trip Guidebook

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Introduction

Eastern North America is an excellent place to study the tectonic processes of rift initiation and evolution to a passive margin in part because rift basins are preserved on the subaerially exposed part of the margin where geology can inform geophysical and geodynamic models. The purpose of this fieldtrip is to visit a part of one of these basins, the Newark basin, exposed in eastern Pennsylvania and New Jersey. The Delaware River corridor between these states affords excellent outcrops where the structure, stratigraphy, and magmatism representative of the rift basins in eastern North America can be studied.

List of stops and stop goals

Stop O, Iacocca Hall Observation Tower, Lehigh University. Overview of greater Lehigh valley geology and major topographic/physiographic features.

STOP 1. Kintnersville, PA. Small normal faults near the northwestern border of the Newark basin.

STOP 2. Ringing Rocks County Park, near Bridgeton,PA. Exposures of Newark basin dolerite (diabase) sill.

STOP 3. Ralph Stover State Park near Tinicum, PA. Exposures of climate-cycle driven, fine-grained deposition of the Lockatong Formation in the Newark basin.

STOP 4. (Optional) Pebble Bluff, Holland Township, NJ. Exposures of fanglomerate facies of the Passaic Formation (Hammer Creek Conglomerate) along the Delaware River.

Appalachian Geology in the Context of the Eastern North American Passive Margin

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 (Figures I-1, I-2). 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., 2002); 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 body of lithosphere that marks the transition from fully continental Precambrian Grenville basement in the west, to fully oceanic Jurassic Atlantic sea floor in the east.

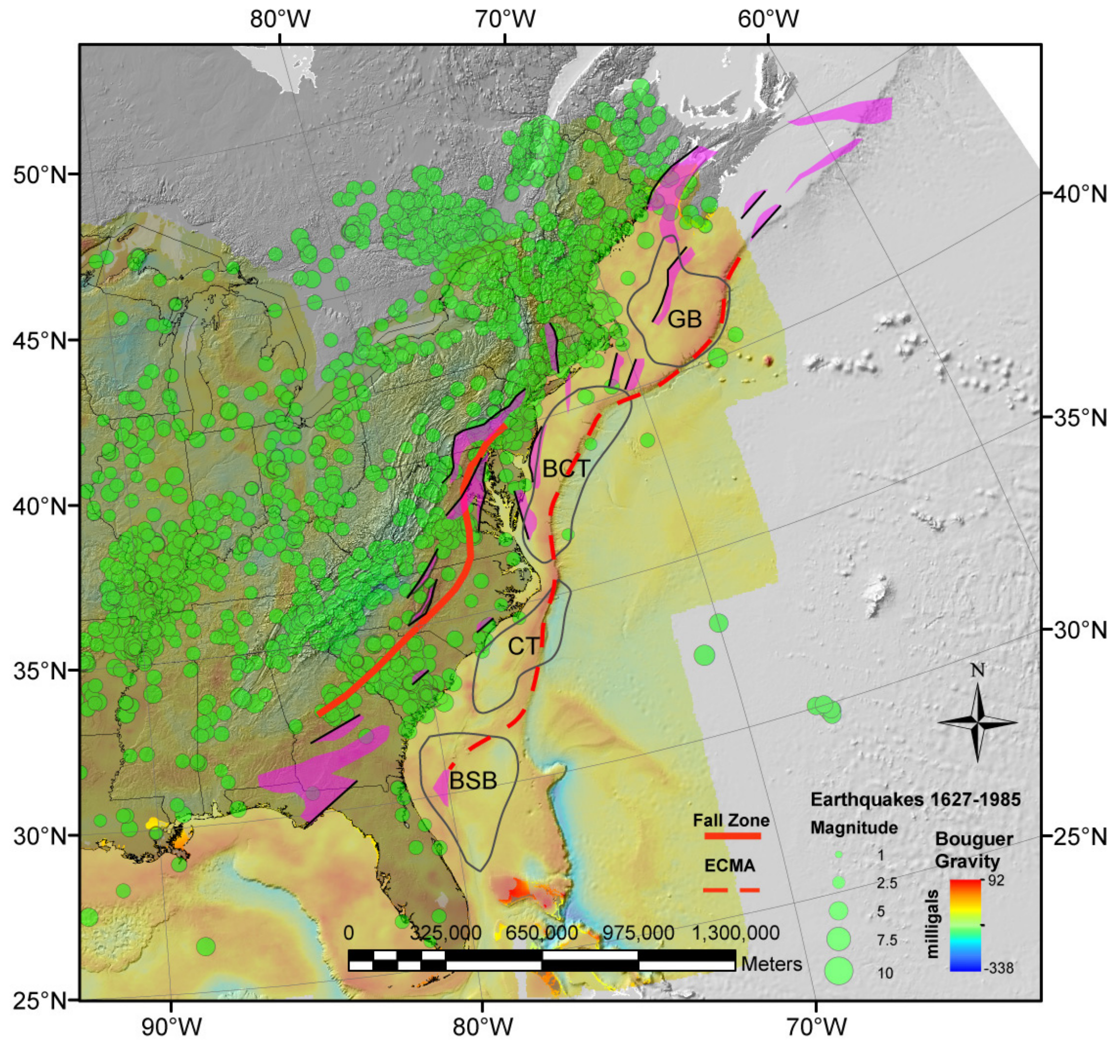


Figure I-1. *Map of ENAM showing the Bouguer gravity anomaly draped on Appalachian topography and Atlantic Ocean bathymetry. Green circles are earthquake locations. Purple polygons are syn-rift basins with border faults in black. Black-outlined polygons are approximate boundaries of shelf-slope basins; 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.*

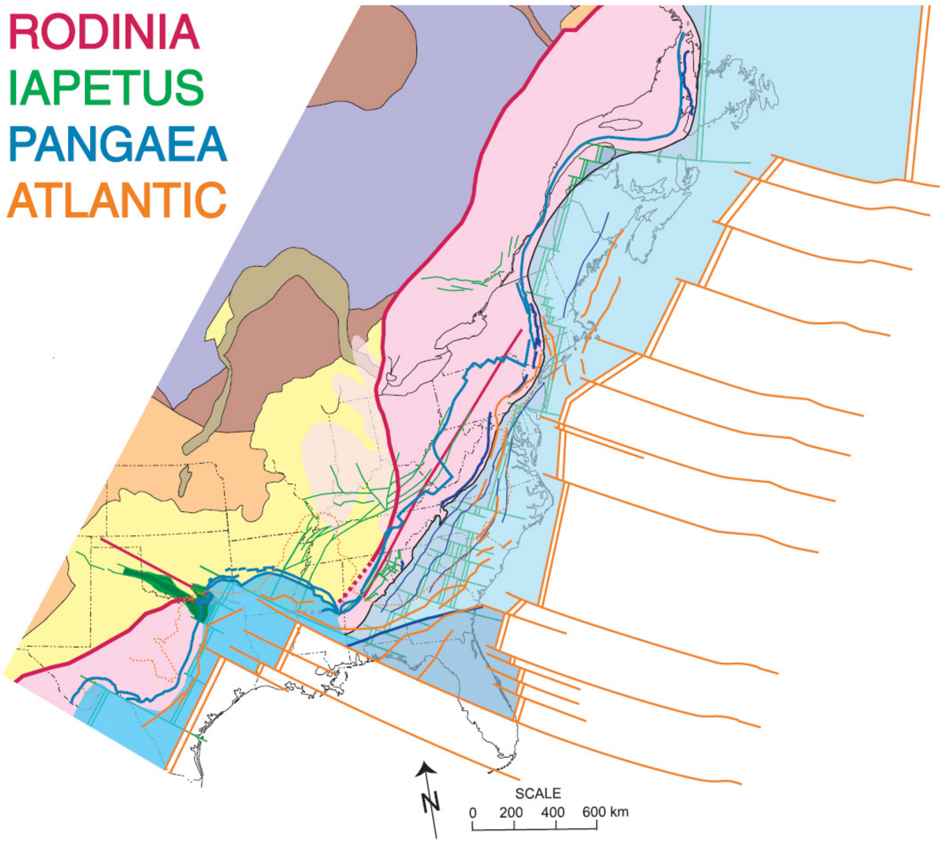


Figure I-2. *ENAM rift margins from the late Proterozoic to the present. (Thomas, 1996).*

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. A latest Proterozoic-early Cambrian passive margin formed by rifting of Rodinia and the opening of the Iapetus Ocean. 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 with the Alleghenian orogeny. The traditional interpretation holds that the three great clastic wedges preserved in the Appalachian foreland are related to three pulses of orogensis during the Paleozoic (Figure I-3). A passive margin existed across eastern North America at the opening of the Cambrian Period. This passive margin collected a thick sequence of siliciclastic and carbonate deposits including the Cambrian shallow-marine facies of the Sauk transgression, represented in Pennsylvania as the Hardyston, Liethsville, and Allentown formations. Late Cambrian to middle Ordovician time saw only carbonate deposition on a shallow marine shelf while east-vergent subduction, growth of an island arc, and closure of Iapetus ensue to the east. Collision of this arc with North America occurred in the late Ordovician and is called the Taconic Orogeny. It destroyed the passive margin, forming a foredeep-foreland along the margin that extended west onto the craton. Detritus from the uplifted Taconic highlands were shed west into this foredeep as a flysch-molasse sequence called the Queenston clastic wedge. Once the Taconic highlands were reduced by erosion, the Appalachian foreland was transformed back to shallow marine conditions accumulating sediment with both margin and cratonic provenance through the middle Devonian. The pattern of collision, uplift, and quiescence repeated two more times before the close of the Paleozoic. In the Devonian, the microcontinent Avalonia collided with North America causing the Acadian orogeny and the shedding of a thick clastic wedge called the Catskill wedge westward into the foreland. Lastly, in the late Carboniferous, Iapetus closed completely with the collision of Africa and Europe with North America driving the Alleghenian orogeny.

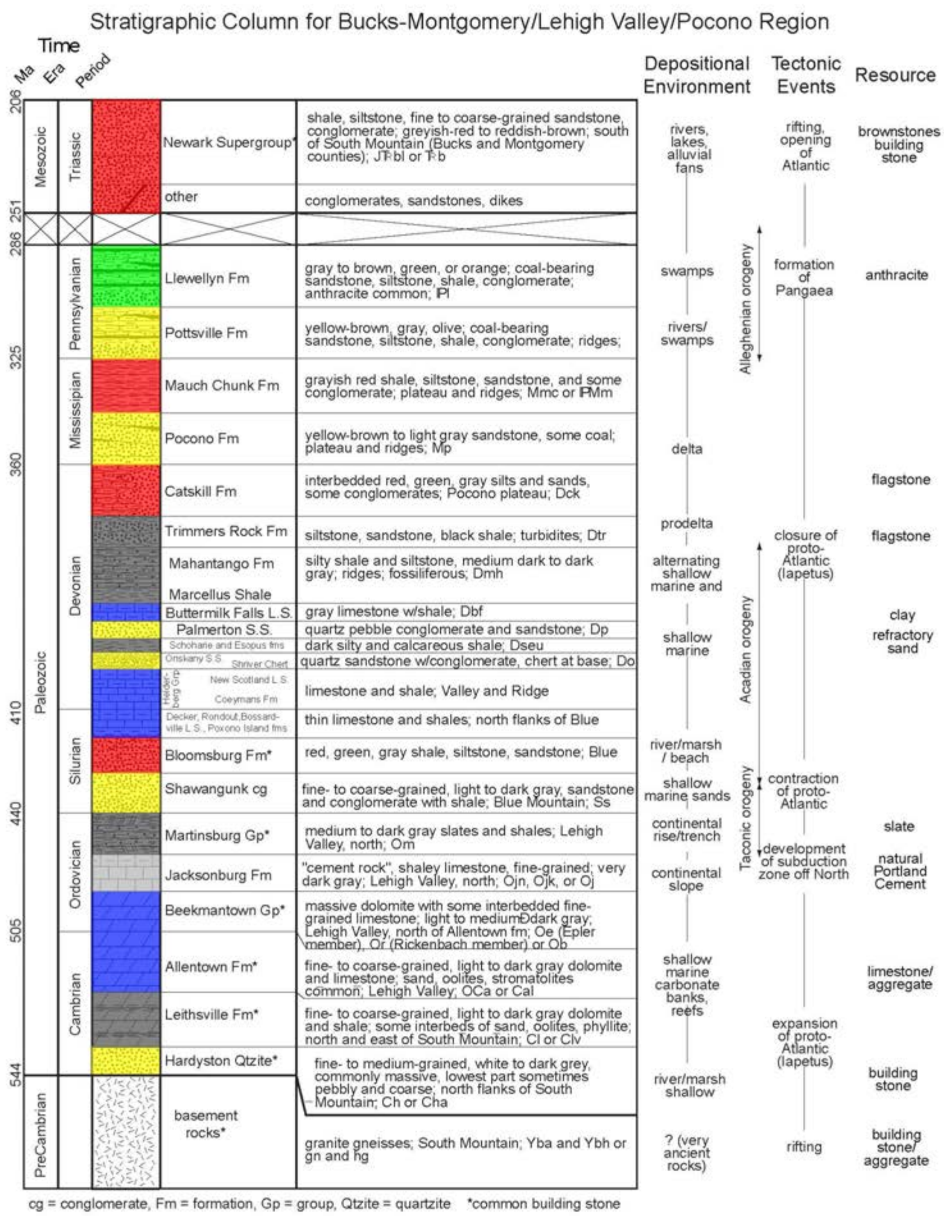


Figure I-3. *Stratigraphic column for eastern Pennsylvania.*

The Early Permian (~280 Ma) 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). Erosion reduced much of Appalachian topography to several hundred meters or less of local relief when rifting began in the Late Triassic.

Extension during theMesozoic reactivated many of the pre-existing Grenvillian 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, 2011; Withjack and Schlische, 2005; Faill, 2003; Figure I-4). 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., 1996a; 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.

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-

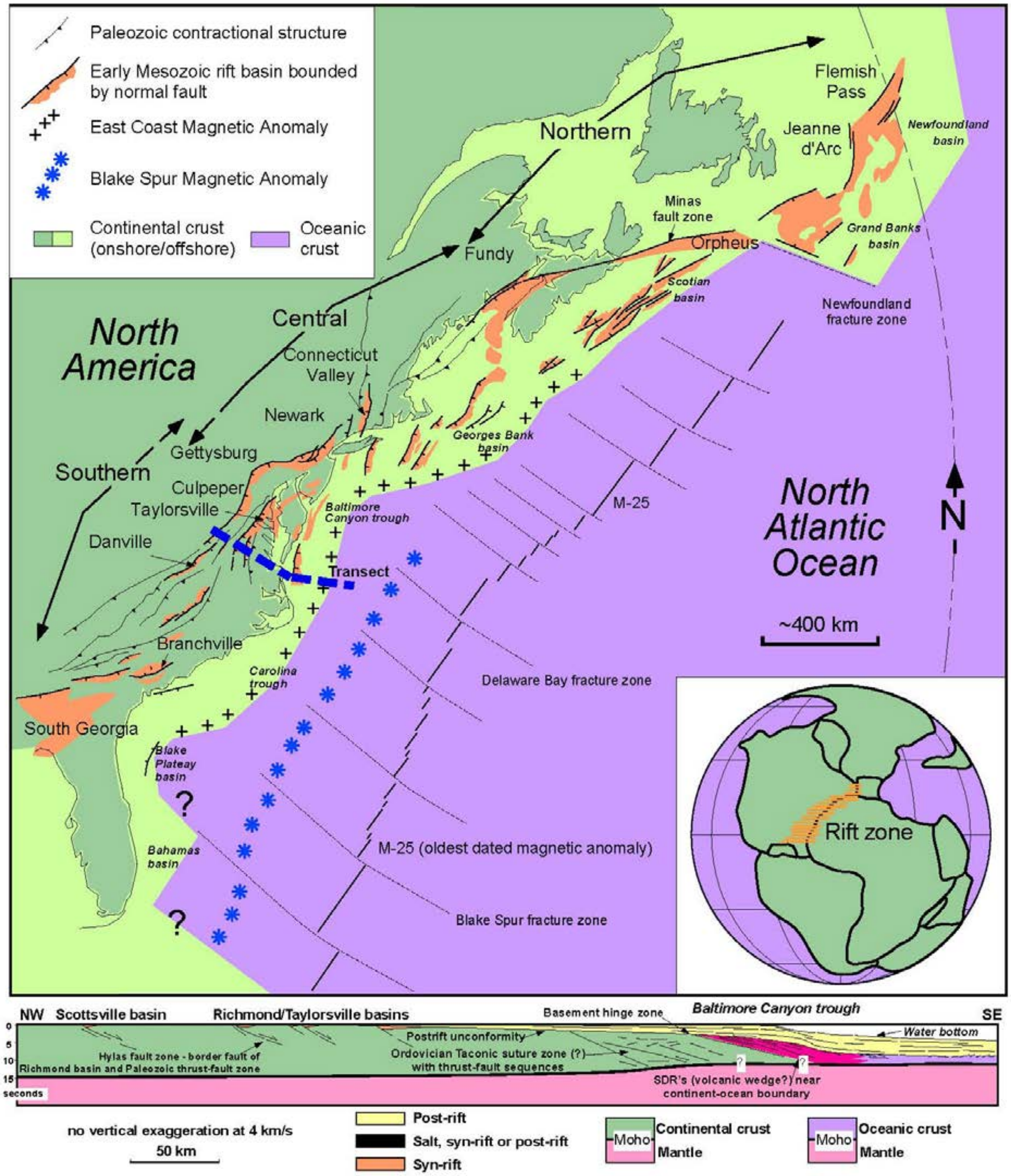


Figure I-4. *Tectonic elements of the eastern North American margin. The southern, central, and northern segments exhibit progressively younger ages for the end of rifting and presumably the onset of seafloor spreading. The East Coast magnetic anomaly approximates the extent of seaward-dipping reflectors 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).*

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 ENAM is preserved as a sedimentologic and stratigraphic archive in several shelf-slope basins (Fig. I-1). 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 ENAM ideally suited for passive margin studies. First, because the ENAM 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.

ENAM Geophysical Background

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 (Figure I-5). 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 wavespeed 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.

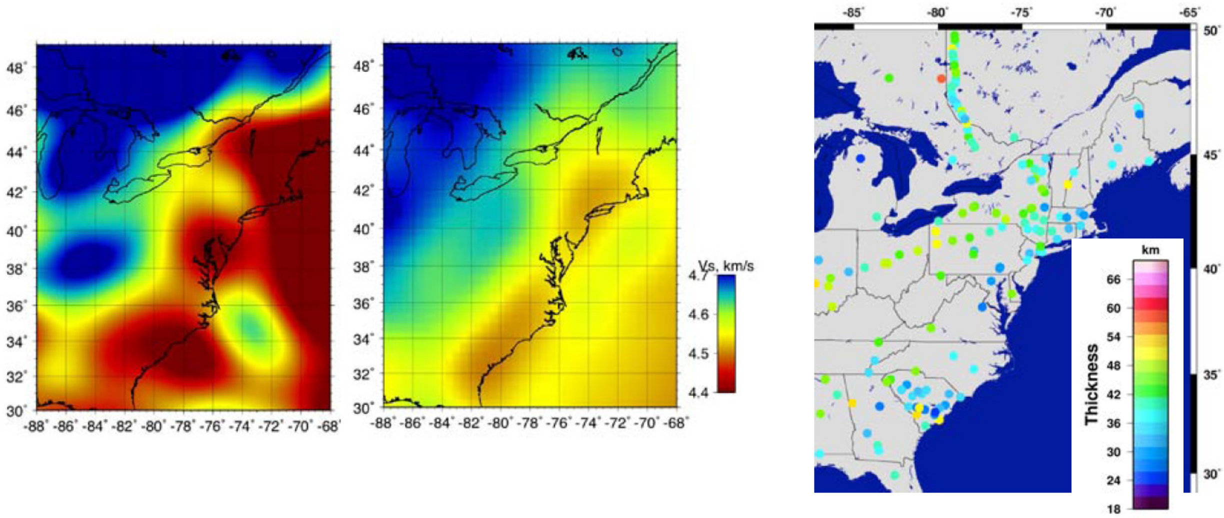


__Figure I-5__. (Left and center images) Horizontal slices at ~100 km through tomographic models by van der Lee and Frederiksen (2005) (left) and Nettles and Dziewonski (2008) (center) show areas of relatively low seismic wavespeed along the Atlantic passive margin of the North American continent that 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).

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.

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.

Numerous studies of seismic anisotropy in the region (see Fouch and Rondenay, 2006, for review) found evidence for significant levels of it throughout the region. However, the relations of the observed anisotropic texture to the present plate motion, asthenospheric processes and the history of tectonic events in the region are not fully worked out. Various authors explained the observations in terms of texture 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).

Crustal thickness beneath the Appalachians, 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.

The crust of eastern North America is surprisingly seismically active (Fig. I-1). There are also some fairly tight concentrations of seismicity, e.g. in the Montreal-Ottawa area, New York City area, St. Lawrence Valley. 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. I-1) indicate a major gravity low in southeastern Pennsylvania interpreted as a particularly deep sedimentary basin above basement (Shanmugam and Lash, 1982). This basin is not evident from surface geology alone and has not been imaged seismically. It represents one of potentially many such Appalachian basins that formed as a result of interactions with non-uniform Grenville basement and were subsequently covered by allochthonous thrust sheets.

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) 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.

Introduction to the Geology of the Newark Rift Basin

The Newark basin is part of a massive rift zone that formed within the Pangean supercontinent during Mesozoic time (Fig. I-1). The fragment of the rift zone preserved on the passive margin of North America, known as the eastern North American rift system, consists of a series of exposed and buried rift basins extending from the southeastern United States to the Grand Banks of Canada (e.g., Manspeizer and Cousminer,1988; Olsen et al., 1989; Schlische, 1993, 2003; Withjack et al., 1998, 2012; Withjack and Schlische, 2005) (Fig. I-1). Withjack and Schlische (2005) divided the eastern North American rift system into three segments based on tectonic history (Fig. I-1). Rifting was underway in all three segments by Late Triassic time. The end of rifting (and presumably the beginning of seafloor spreading), however, was diachronous, occurring first in the southeastern United States (latest Triassic), then in the northeastern United States and southeastern Canada (Early Jurassic), and finally in the Grand Banks (Early Cretaceous) (Withjack et al., 1998, 2012; Withjack and Schlische, 2005; Schettino and Turco, 2009).

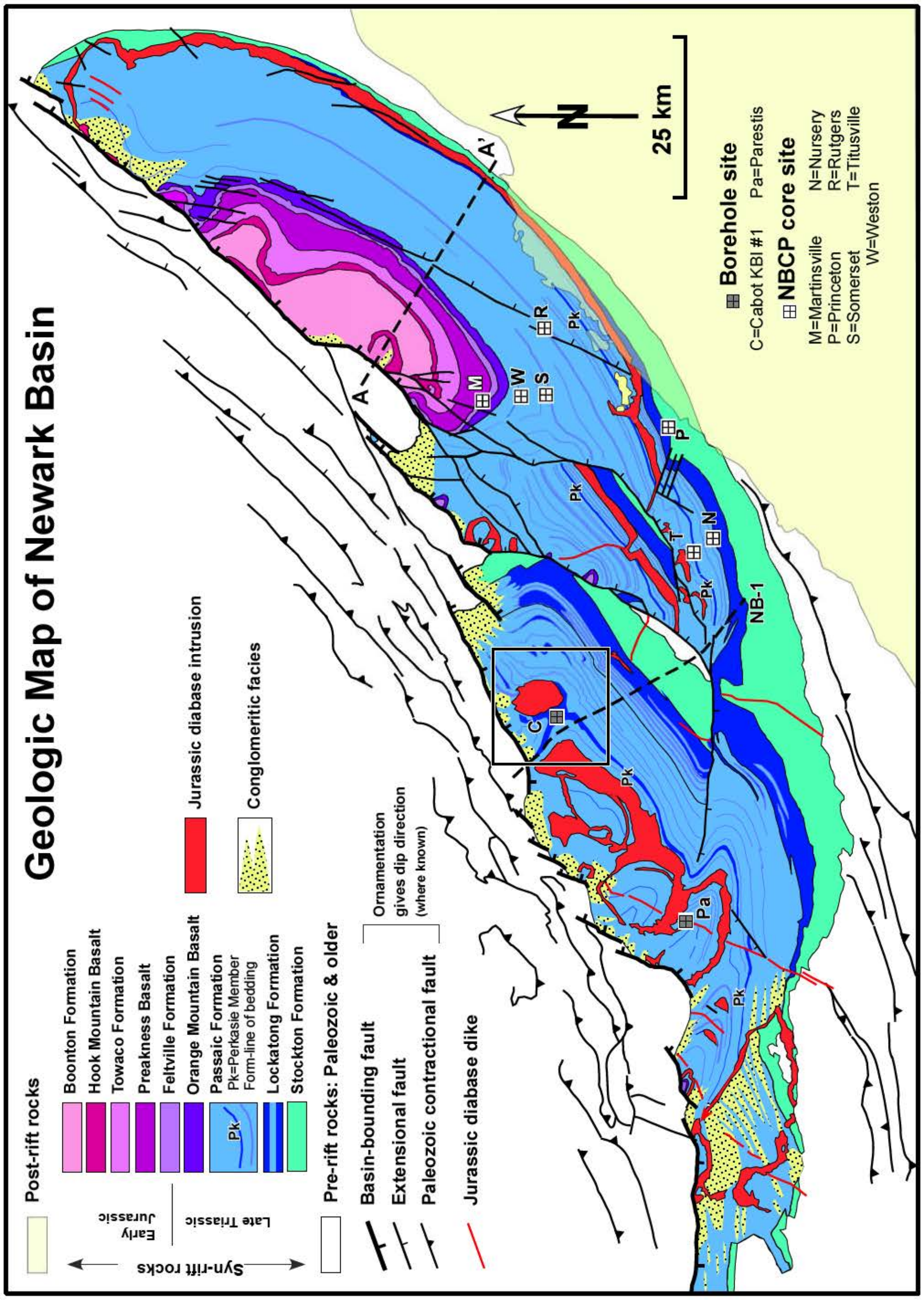
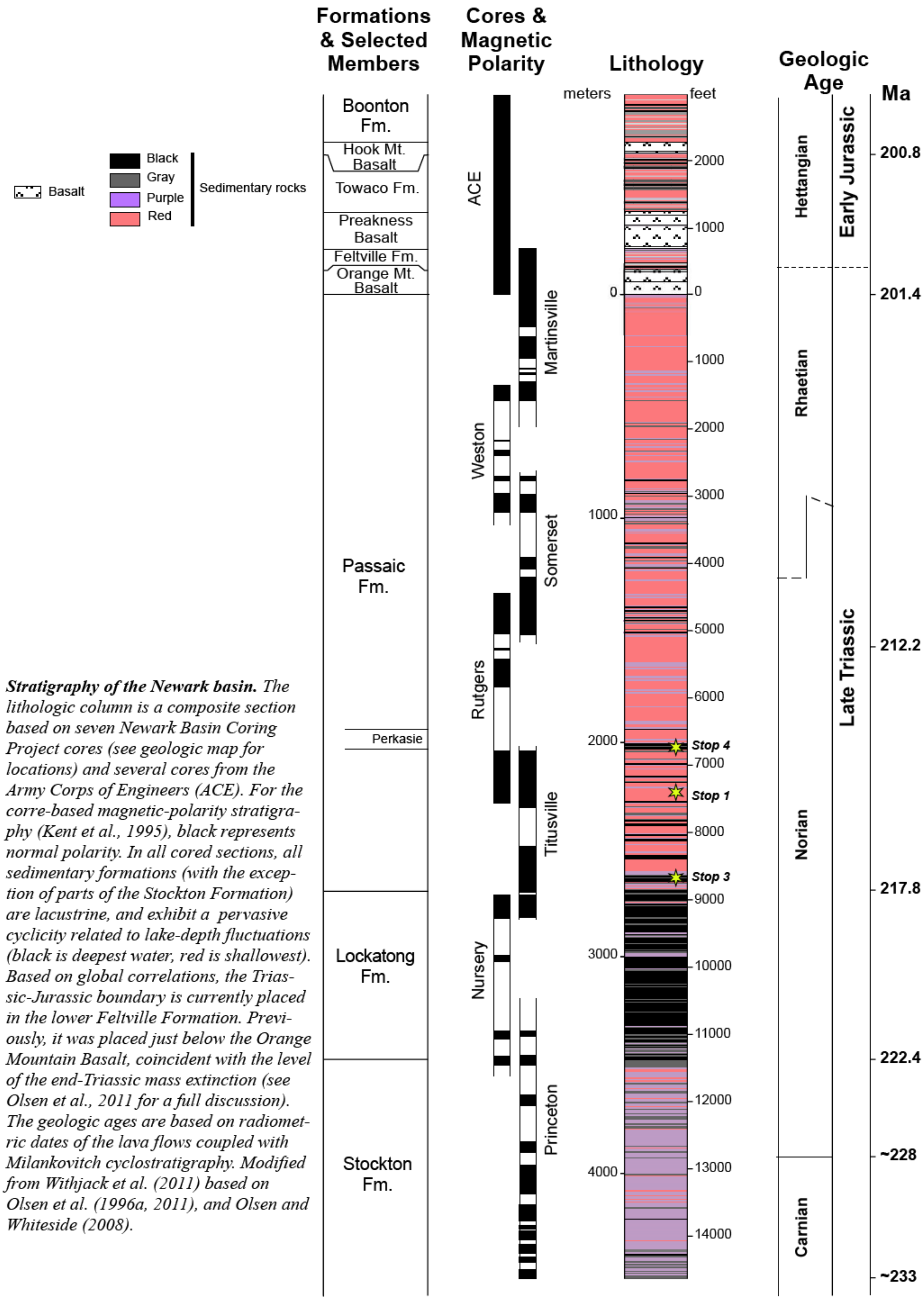


Fig. N-1. *Geologic map of the Newark rift basin showing distribution of sedimentary formations, lava flows, and diabase intrusions; locations of borehole and core sites; seismic-line NB-1; and location of cross-section A-A'. Only the largest Paleozoic contractional faults are shown. Box shows area of field-trip stops. Modified from Withjack et al. (2011) based on Schlische (1992), Olsen et al. (1996a), and Schlische & Withjack (2005).*

A series of NE-striking, SE-dipping, right-stepping faults bound the Newark basin on the northwest (Fig. N-1). The border faults are subparallel to thrust faults present in pre-rift rocks surrounding the basin. Several large intrabasin faults also dissect the basin. Most syn-rift strata dip 10-15° NW toward the border-fault system. Near many of the basin-bounding and intrabasin faults, however, the syn-rift strata are warped into a series of anticlines and synclines whose axes are mostly perpendicular to the adjacent faults (i.e., transverse folds; e.g., Wheeler 1939; Schlische 1992, 1995). The Newark basin, like many other rift basins of the eastern North American rift system, underwent significant post-rift deformation including much of the tilting and folding of the syn-rift strata (e.g., Sanders, 1963; Faill, 1973, 1988; Withjack et al., 1998; Schlische et al., 2003; Withjack et al., 2010). Furthermore, the basin underwent significant erosion (locally more than 6 km) after rifting (e.g., Steckler et al., 1993; Malinconico, 2010).

The stratigraphy of the Newark basin consists of the Stockton, Lockatong, and Passaic formations of Late Triassic age and the overlying basalts and interbedded sedimentary rocks of latest Triassic to Early Jurassic age (i.e., the Orange Mountain Basalt, Feltville Formation, Preakness Basalt, Towaco Formation, Hook Mountain Basalt, and Boonton Formation) (e.g., Olsen et al. 1996a; Fig. N-2). Most syn-rift strata accumulated in a lacustrine setting and exhibit a pervasive cyclicity in sediment fabrics, color, and total organic carbon (from microlaminated black shale to extensively mudcracked and bioturbated red mudstone) (e.g., Olsen, 1986, Olsen et al., 1996a; see Stop 3 for a fuller treatment). Individual members of the stratigraphic units have great lateral extent and continuity and have been traced throughout much of the Newark basin (e.g., McLaughlin, 1948; Olsen, 1988); a prominent example is the Perkasie Member of the Passaic Formation (Fig. N-1). Biostratigraphy indicates that the preserved syn-rift strata in the Newark basin range in age from Carnian (Late Triassic) to Hettangian (Early Jurassic) (e.g., Cornet and Olsen, 1985; Olsen et al., 2011). Igneous rocks (e.g., basaltic lava flows, diabase sheets, and dikes) in the Newark basin are associated with the Central Atlantic Magmatic Province (CAMP), one of the world's largest igneous provinces (e.g., McHone, 1996, 2000; Marzolli et al., 1999; Hames et al., 2003). CAMP-related igneous activity occurred during the very latest Triassic and earliest Jurassic (~ 200 Ma) (see Olsen et al., 2003, 2011 and references therein). We discuss additional aspects of CAMP at Stop 2.

Seismic line NB-1, located near the route of this field trip, images the subsurface geometry of the Newark basin. The seismic line shows that a major SE-dipping fault zone with normal separation bounds the basin on the northwest (Fig. N-3). The fault zone, characterized by a series of high-amplitude reflections, is relatively planar and has a dip magnitude of ~30°. Using core data, Ratcliffe et al. (1986) demonstrated that this fault zone is a mylonitic Paleozoic thrust fault reactivated during rifting; this is consistent with the relatively low-angle dip of the border fault. The seismic data show that the syn-rift strata dip ~10-15° toward the northwest. Furthermore, the Stockton Formation (exposed at the surface) and an unexposed older unit (which onlaps Paleozoic pre-rift strata) thicken toward the border fault, indicating that faulting and deposition were coeval (i.e., these units are growth deposits). Field and core data (see Stops 3 and 4) indicate that the Lockatong and Passaic Formation also exhibit subtle thickening toward the border fault. Furthermore, all sedimentary formations contain conglomeratic facies where present adjacent to the border-fault system (Fig. N-1).



Stratigraphy of the Newark basin. ***The*** *lithologic column is a composite section based on seven Newark Basin Coring Project cores (see geologic map for locations) and several cores from the Army Corps of Engineers (ACE). For the corre-based magnetic-polarity stratigraphy (Kent et al., 1995), black represents normal polarity. In all cored sections, all sedimentary formations (with the exception of parts of the Stockton Formation) are lacustrine, and exhibit a pervasive cyclicity related to lake-depth fluctuations (black is deepest water, red is shallowest). Based on global correlations, the Triassic-Jurassic boundary is currently placed in the lower Feltville Formation. Previously, it was placed just below the Orange Mountain Basalt, coincident with the level of the end-Triassic mass extinction (see Olsen et al., 2011 for a full discussion). The geologic ages are based on radiometric dates of the lava flows coupled with Milankovitch cyclostratigraphy. Modified from Withjack et al. (2011) based on Olsen et al. (1996a, 2011), and Olsen and Whiteside (2008).*

Fig. N-2.

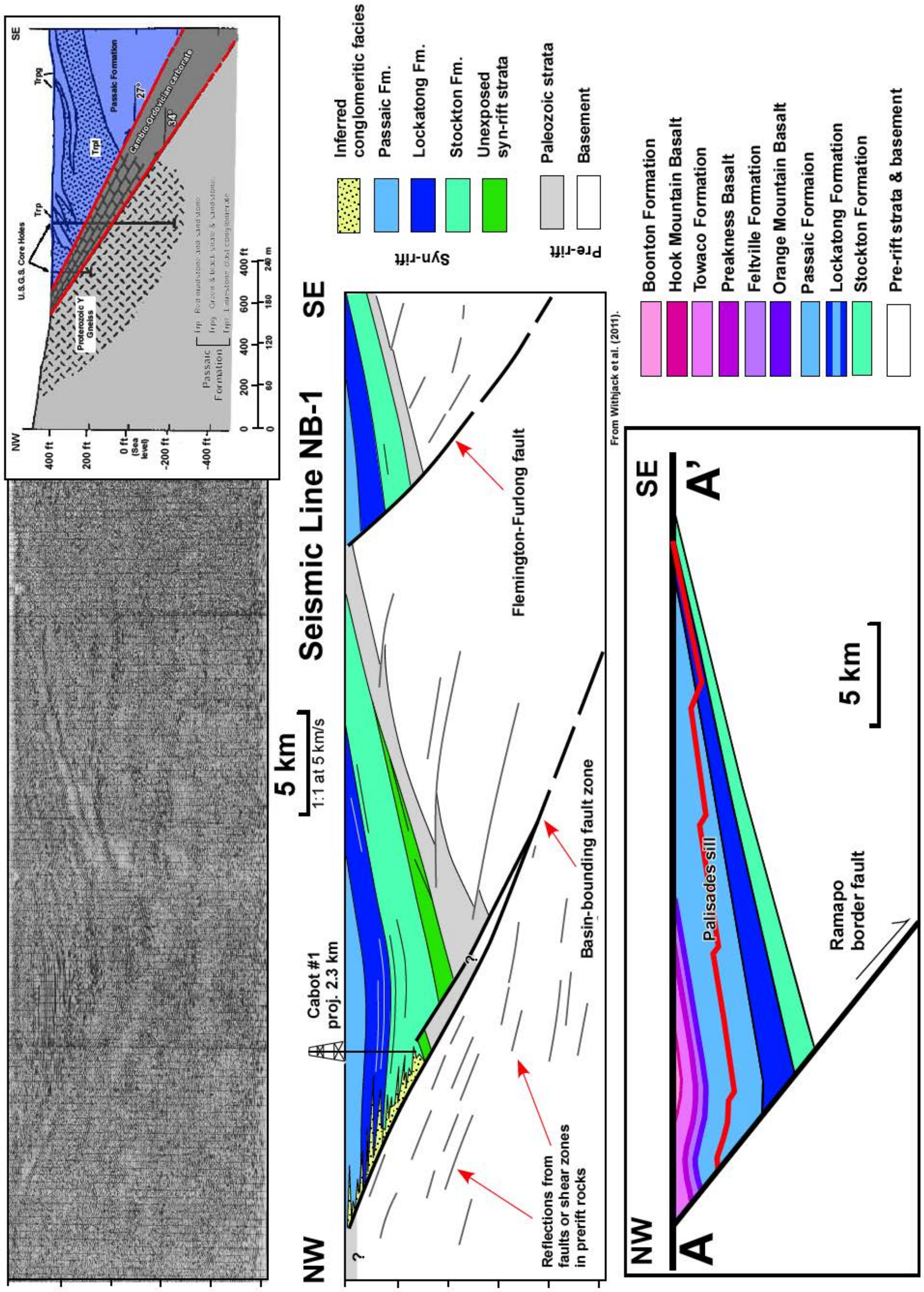


Fig. N-3. *Cross sections of the Newark basin. The interpretation of seismic-line NB-1 utilizes well and outcrop data. Note the shallow dip angle of the border fault, suggesting it originated as a thrust fault and was reactivated during rifting; this interpretation is corroborated by U.S.G.S. coring of the border fault in the vicinity of the seismic line (Ratcliffe et al., 1986).Syn-rift strata generally dip toward the SE-dipping border faults. The oldest syn-rift units (green shades) show substantial thickening toward the fault; younger units exhibit much less obvious thickening. The subsurface geometries of formations and the Palisades sill on section A-A' are inferred. Modified from Withjack et al. (2011) based on Schlische (1992) and Schlische & Withjack (2005). The uninterpreted seismic line is from Bally et al. (1991).*

ROAD LOG and STOPS

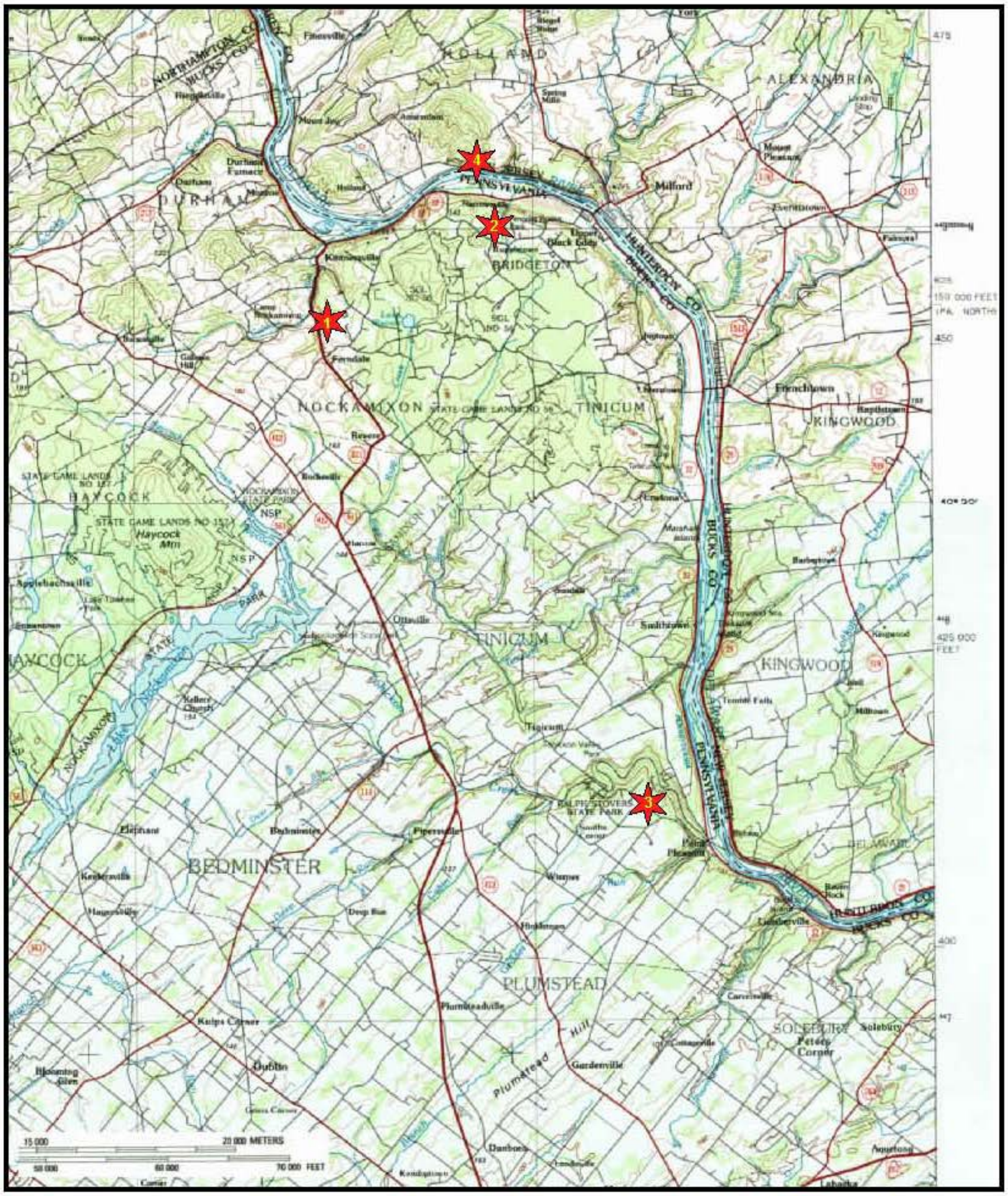


Fig. RL-1A. *Topographic map with stop locations.*

START. STEPS building, Lehigh University. *Proceed east on Packer Ave., turn right on Ryan Street, then quickly left onto Hillside Ave. Proceed to Hayes St, turn right and continue up South Mountain on Mountain Drive N. At the first triangle, bear left; you have the right of way. At the second triangle, stay straight. Turn right into the parking lot in front of Iacocca Hall and park in a visitor's lot parking space.*

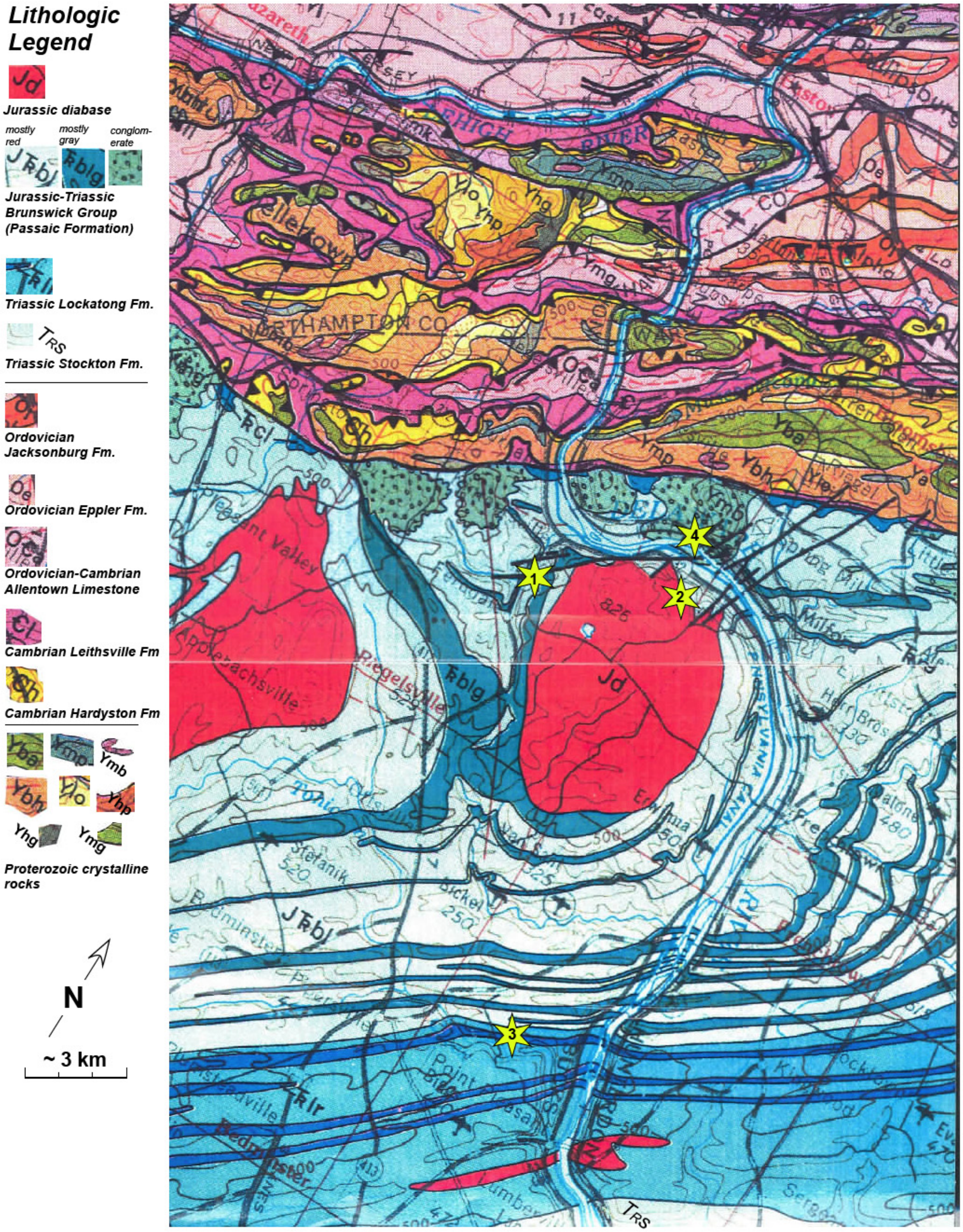


__Fig. RL-1B.__ Geologic map of part of the Newark 1°x2° quadrangle, showing the northwestern Newark basin and adjacent prerift strata and basement in the vicinity of the Delaware River. Stars give field-stop locations. Map is from Lyttle & Epstein (1987). Note that the geology of the Lockatong and Passaic formations differs slightly from Fig. N-1, which reflects updated mapping utilizing the results of the Newark Basin Coring Project.

STOP 0. Iacocca Hall observation tower.

Iacocca tower affords a 360° view of the Lehigh Valley and surrounding region. To the north spreads the Lehigh Valley with Kittatinny Ridge and Ridge and Valley of Pennsylvania beyond. These physiographic provinces are underlain by the Appalachian foreland and contain kilometers of Paleozoic strata that has been folded and faulted by Appalachian orogenesis. To the east and west stretches the Blue Ridge, atop which is built the mountaintop campus of Lehigh University. The Blue Ridge is a vestige of a folded crystalline thrust sheet that was translated westward into the Appalachian foreland during the Alleghenian orogeny. Similar thrust sheets are preserved to the south in the Piedmont, not visible from this location. Visible to the south is the Saucon Valley, underlain by lower Paleozoic carbonates contained within the thrust sheet. Beyond that rise low hills underlain by Newark basin continental strata and syn-rift intrusives.

Depart Iaccoca Hall parking lot and proceed west on Mountain Drive West. At the first triangle, stay left and descend South Mountain. Proceed straight at the stop sign, then stop light and follow the meandering Bingan Road through the Saucon Valley, crossing Saucon Creek, and into Bingen. At the T-intersection with Apples Church Road, make a left and proceed down the hill to Rt. 412. Make a right at the stop sign and follow Rt 412 for ~3 km to its intersection with Rt 212. Turn left onto Rt 212 and proceed through Springtown. Approximately 2 km east of Springtown, Rt 212 stays left (straight) and begins descending down to the Delaware River, following Cooks Run.

The path being followed is a strike valley in Cambrian shale and carbonate, bounded by crystalline rocks of the Blue Ridge thrust sheet on its north and south. All of these rocks are in the footwall of the Newark basin border fault which is exposed a few kilometers to the south.

Rt 212 will intersect Rt 611 at the Delaware River. The remains of the former Delaware Canal (1832-1931) are directly ahead of you. The canal is now administered as part of the Delaware and Lehigh National Heritage Corridor. Turn right onto Rt 611.

The exposures on your right directly after the turn onto Rt 611 expose the Cambrian Hardyston Fm unconformably overlying Blue Ridge Proterozoic crystalline rocks. Larger exposures of the crystalline rocks are present in the outcrops on the west side of Rt 611 as you continue to travel south.

Approximately 1 km south of the Rt 611-Rt 212 intersection you abruptly pass from the Blue Ridge rocks to the rocks of the Newark basin by crossing a large normal fault. The exposure is off limits and can no longer be visited, but we will see synthetic normal faults to this master border fault at stop 1.

Continue south on Rt 611 for ~4 km driving on outwash terraces of the Delaware River to the intersection with Rt 32 at Kintersville. Stay to the right on Rt 611 and begin ascending a hill locally known as Haycock Mountain. Proceed for about 2 km to the large pull-offs of Stop 1.

Stop 1: Normal Faults in Passaic Formation, Route 611, Kintnersville, PA

A set of meter-scale normal faults (Fig. RL-2) cuts red mudstone of the middle Passaic Formation. The normal faults strike ~030° and generally dip 40°-50° SE. Slickenlines are steeply raking, indicating predominantly dip-slip movement. Like most larger intrabasinal faults in the Newark basin, these faults are oblique to the strike of fault segments of the border-fault system (the strike of the nearest segment of the border-fault system is ~060°). This fault pattern is indicative of oblique extension (e.g., Schlische et al., 2002), i.e., the extension direction was not perpendicular to the strike of the border-fault system. The geometry of the intrabasinal faults and Early Jurassic diabase dikes suggest that the extension direction was WNW-ESE.

Fault 2 has the smallest displacement and narrowest zone of breccia and gouge; fault 4 has the highest displacement and widest zone of breccia and gouge. These faults obey an approximately linear scaling relationship between fault-zone thickness and fault displacement (e.g., Hull, 1988; Knott, 1994; Knott et al., 1996). The sequence going from fault 2 to 3 to 1 to 4 likely reflects stages in the evolution of normal faults with increasing displacement; this involves the linkage of originally isolated segments and the widening of the fault zone.

According to Withjack and Olsen (1999), the exposure belongs to the upper part of Member K of the Passaic Formation (nomenclature of Olsen et al., 1996a). Deposition occurred mostly in playas. A prominent bed contains vugs that were once filled with evaporite minerals. Silt bands likely represent slightly deeper or more permanent shallow lakes. Red beds similar to those outcropping here account for more than 80% of the Passaic Formation.

Although this outcrop is only about 3 km from the border fault, the sedimentary rocks are remarkably fine grained. Two possibilities account for this observation: (1) The hanging-wall block and axial sources contributed more sediment than the footwall block (Fig. RL-3), as in many modern rift basins (e.g., Gawthorpe and Leeder, 2000). (2) The present-day boundary fault of the basin was not the outermost border fault during deposition (Fig. RL-3). In Late Triassic time, coarser sediments accumulated in the hanging walls of border faults located to the NW of the present-day border fault. Subsequent erosion removed these coarse-grained strata and exposed the Paleozoic and Precambrian rocks. Vitrinite reflectance data indicated that 4-6 km of erosion occurred in the vicinity of this outcrop (Steckler et al., 1993; Malinconico, 2010; Withjack et al., 2011).

Continue south on Rt 611 to Ferndale. Turn left on Center Hill Road. Proceed for ~0.5 km and turn right on Lake Warren Road. Continue for ~3 km and turn left on Marienstein Road. After a short distance, this road becomes Bridgeton Hill Road. Follow Bridgeton Hill Road for ~ 3 km and turn left on Ringing Rocks Road. Proceed for ~0.5 km and turn right into the park. Follow the trail on foot to the boulder field.

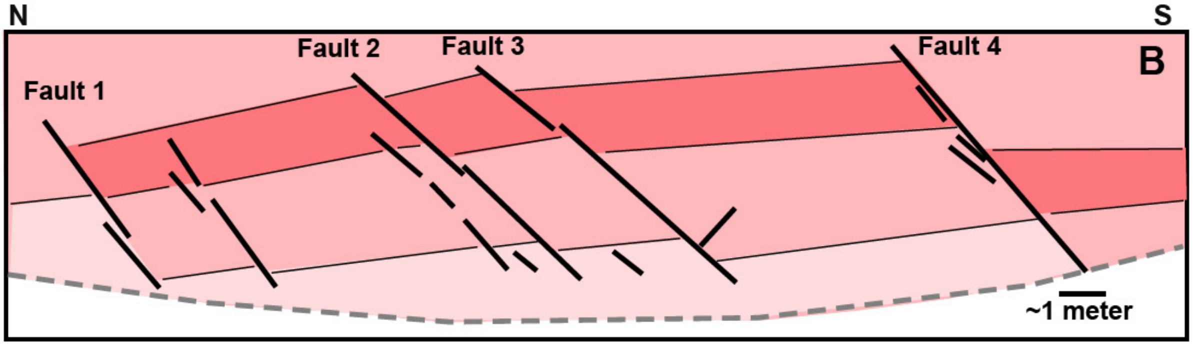
Stop 1: Haycock Mountain, Rt. 611, Kintnersville, PA

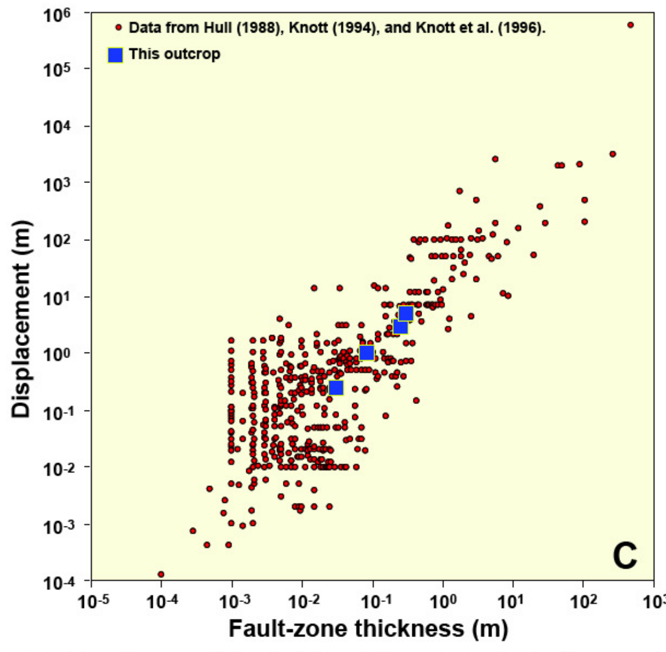
Stratigraphy: Late Triassic Passaic Formation, Member K

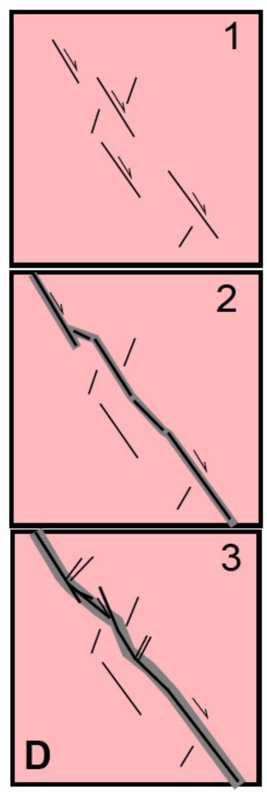
Lithology: Red, mostly massive mudstone with evaporites

Distance from border fault: ~ 3km

Structures: NNE-striking, SE-dipping normal faults with dip-slip displacement; multiple joint sets wth plumose markings

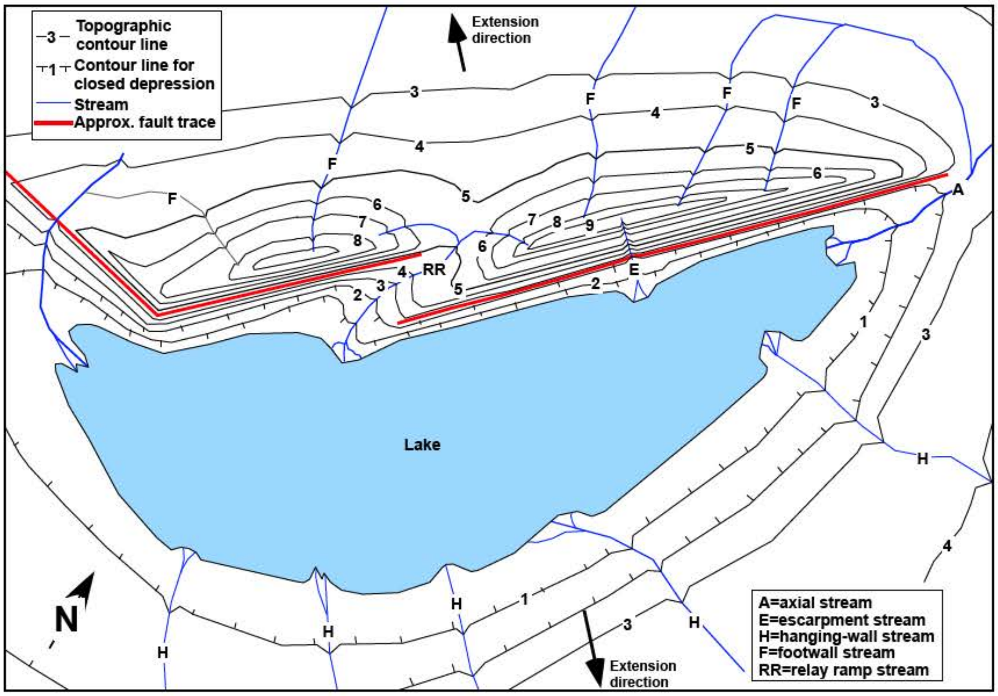




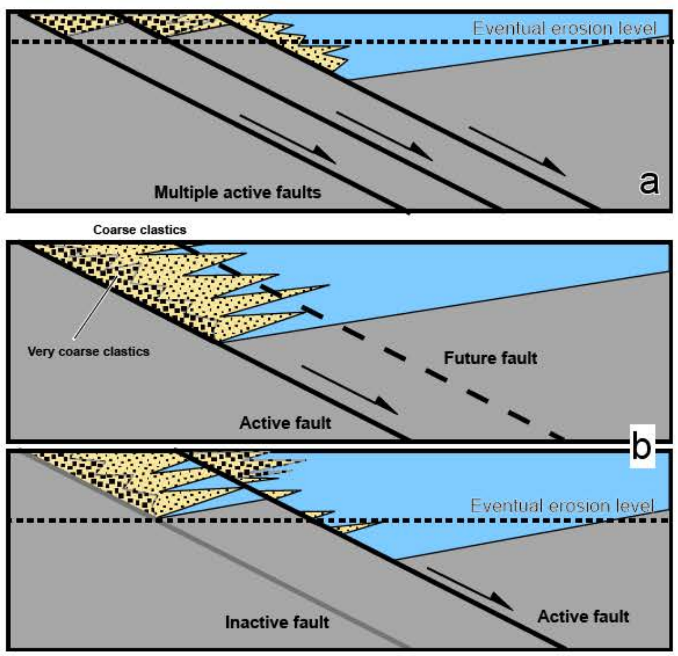


A. Photo of fault 1; Alissa Henza and Martha Withjack for scale. B. Sketch of outcrop cut by multiple normal faults with variable displacements. C. Log-log plot of fault displacement versus fault-zone thickness (breccia and gouge) for faults 2, 3, 1, and 4; the data are comparable to other measurements for faults cutting sedimentary rocks. D. Hypothesized evolution of normal faults in cross section, showing the linkage of originally isolated segments and widening of the fault zone. Modified from Schlische & Withjack (2005).

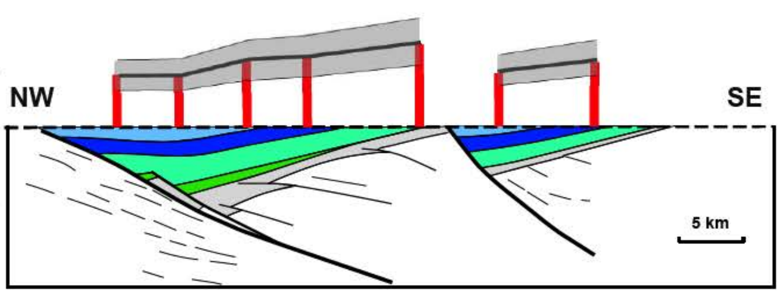
Fig. RL-2.



Inferred topography and fluvial inputs for a rift basin crudely similar to the Newark basin. *Most of the footwall slopes away from the basin; most of the hanging wall slopes toward the basin. Consequently, direct sediment input from the footwall is limited (except at relay ramps between overlapping fault segments). This may account, in part, for the relatively fine-grained strata adjacent to the border-fault system. Modified from Schlische & Withjack (2005).*



Schematic cross sections showing border-fault development during deposition. *(a) Model with multiple active faults. Coarse-grained sediments preferentially accumulate in hanging walls of outer border fault. Later erosion removed these sedimentary rocks. (b) Model with basinward migration of border faults. During the early stages of rifting, the outer border fault was active, and coarse-grained sediments accumulated in its hanging wall. During the later stages of rifting, the inner border fault became active, and coarse-grained sediments accumulated in its hanging wall. Later erosion removed the coarse-grained sedimentary rocks. Thus, the present-day edge of the basin has shifted basinward from its position during deposition of the Passaic Formation. Modified from Schlische & Withjack (2005).*



Estimated amount of post-rift erosion for transect along seismic line NB-1. *Red bars show eroded syn-rift section predicted by vitrinite-reflectance analysis; grey area shows potential error (± 30%). The magntidue of erosion varies from 4 to 6 km. Vitrinite-reflectance data us from Malinconico (2010); figure is modified from Withjack et al. (2011).*

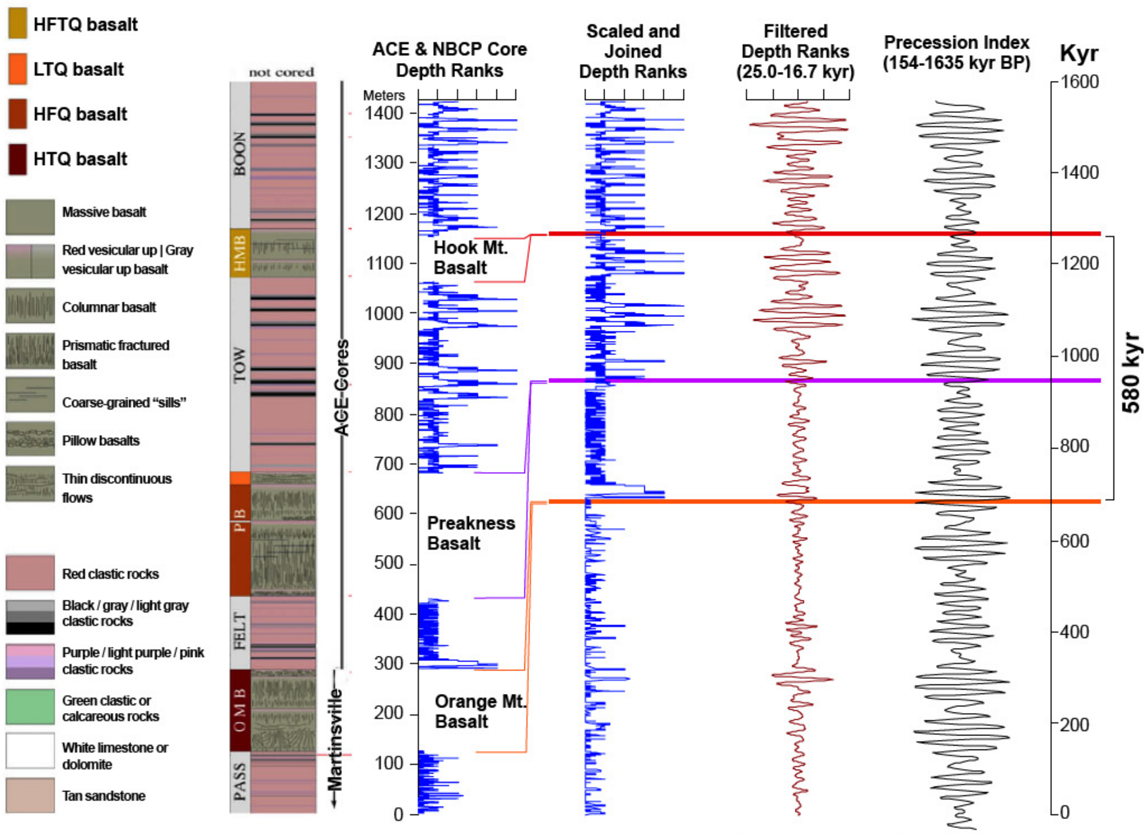
Fig. RL-3.

Stop 2: Ringing Rocks County Park, Bridgeton, PA

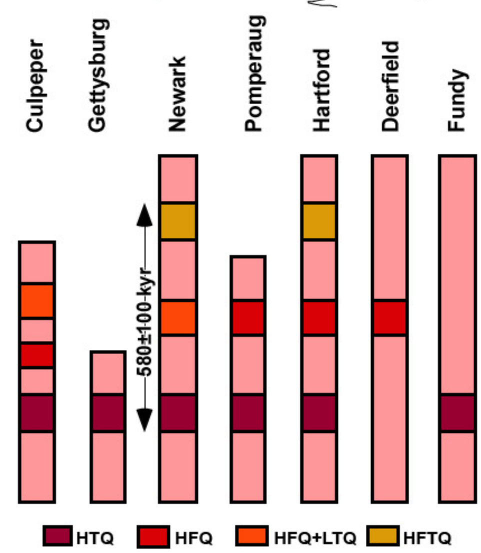
The boulder field in Ringing Rocks County Park contains rocks of the Coffman Hill diabase (dolerite). Boulders exposed to lots of sun do “ring” when struck with a hammer! The Coffman Hill diabase is the erosional remnant of a folded intrusive sheet. The geologic map (Fig. N-1) shows that strata consistently dip toward the diabase, indicating that the diabase sheet and intruded strata define a doubly-plunging syncline (or basin). Contact metamorphic rocks of the Passaic Formation underlying the intrusive sheet are exposed at the waterfall (at the end of the hiking trail). These gray rocks are relatively fine-grained (mostly mudstone with some sandstone). Mudcracks and ripple marks suggest that these strata accumulated in very shallow lakes. Although gray rocks are typically associated with deeper-water deposits, in this case, the gray color is a result of contact metamorphism. The hornfels contains well-developed subvertical joints.

The Newark basin contains many other diabase sheets (Fig. N-1), many of which, like the Coffman Hill diabase, are folded adjacent to the border-fault system. The most famous intrusion is the Palisades sill. At its northern termination, the Palisades intrusion becomes dike-like and feeds one of the lava flows (Ratcliffe, 1988). The Newark basin also contains three extrusive formations (Orange Mountain Basalt, Preakness Basalt, Hook Mountain Basalt), each consisting of multiple flow units (Fig. RL-4). Although each basalt formation has a distinct geochemistry, all extrusive (and intrusive) rocks are quartz-normative tholeiites (Fig. RL-4). Cyclical lacustrine strata with Milankovitch periodicities preceding, interbedded with, and succeeding the extrusive formations indicate that extrusive activity occurred in less than 600,000 years (Olsen et al., 1996b, 2003). This duration likely applies to other Mesozoic rift basins in eastern North America because these basins contain basalts with the same geochemistry as those in the Newark basin (Fig. RL-4). As noted in the introduction, radiometric dates indicate that all igneous activity (intrusive and extrusive) occurred at ~200 Ma (see Olsen et al., 2011, and references therein).

Latest Triassic/earliest Jurassic igneous activity occurred throughout eastern North America. Diabase dikes are present from the southeastern United States to maritime Canada (Fig. RL-5) (e.g., May, 1971; McHone, 2000). Note that dikes in the southeastern U.S. strike NW-SE and N-S, and cut across the rift basins. In the northeastern U.S. and maritime Canada, the dikes generally strike NE-SW, and were intruded while lava flows accumulated in the actively subsiding rift basins (e.g., Schlische et al., 2003). These intrusive and extrusive rocks in eastern North America are part of a vast igneous province (>10,000,000 km^2; Marzolli et al., 1999; McHone, 2000, 2003), known as the Central Atlantic Magmatic Province (CAMP), that also covers parts of South America, Africa, and Europe (Fig. RL-5). Seaward-dipping reflectors (SDR’s) that likely represent a volcanic and/or volcanoclastic wedge at the continent-ocean boundary (and account for the East Coast Magnetic Anomaly) (Fig. RL-6) may be part of CAMP or a second pulse of igneous activity (for a full discussion, see Schlische et al., 2003). Whatever their exact age, these SDR’s are not present along the entire eastern North American continental margin. Thus, parts of the margin are “volcanic” and parts are “non-volcanic.” Nonetheless, igneous rocks of latest Triassic/earliest Jurassic age are present along the entire margin. For example, while the Scotian and Newfoundland segments of the margin lack SDR’s, the Fundy, Orpheus, and Jeanne d’Arc rift basins all contain intrusive sheets and/or basalt flows (e.g., Olsen, 1997; Withjack and Schlische, 2005).



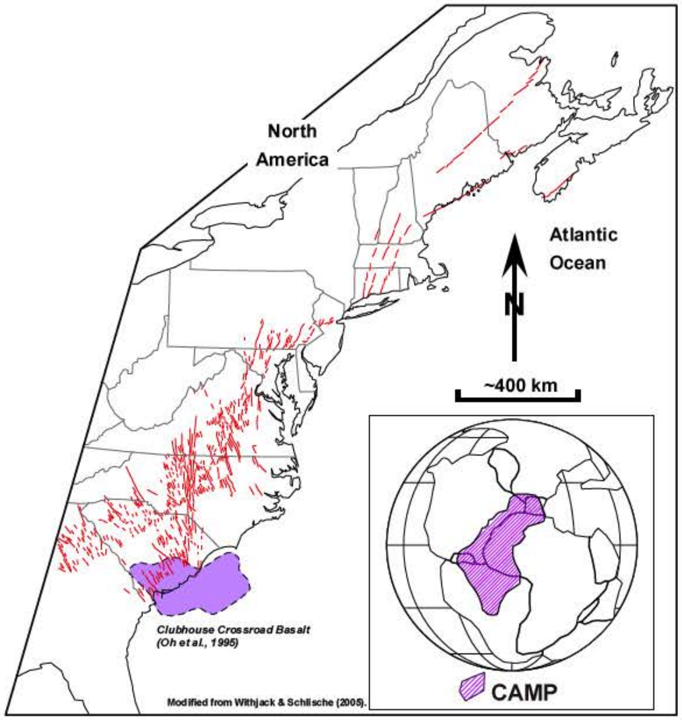
Stratigraphy of extrusive interval in the northern Newark basin based on the Martisville and Army Corps of Engineers (ACE) cores. Cyclostratigraphy of lake deposits in sedimentary strata constrains the duration of the extrusive interval to approximately 600,000 years. Modified from Olsen et al. (1996b) and Whiteside et al. (2007).



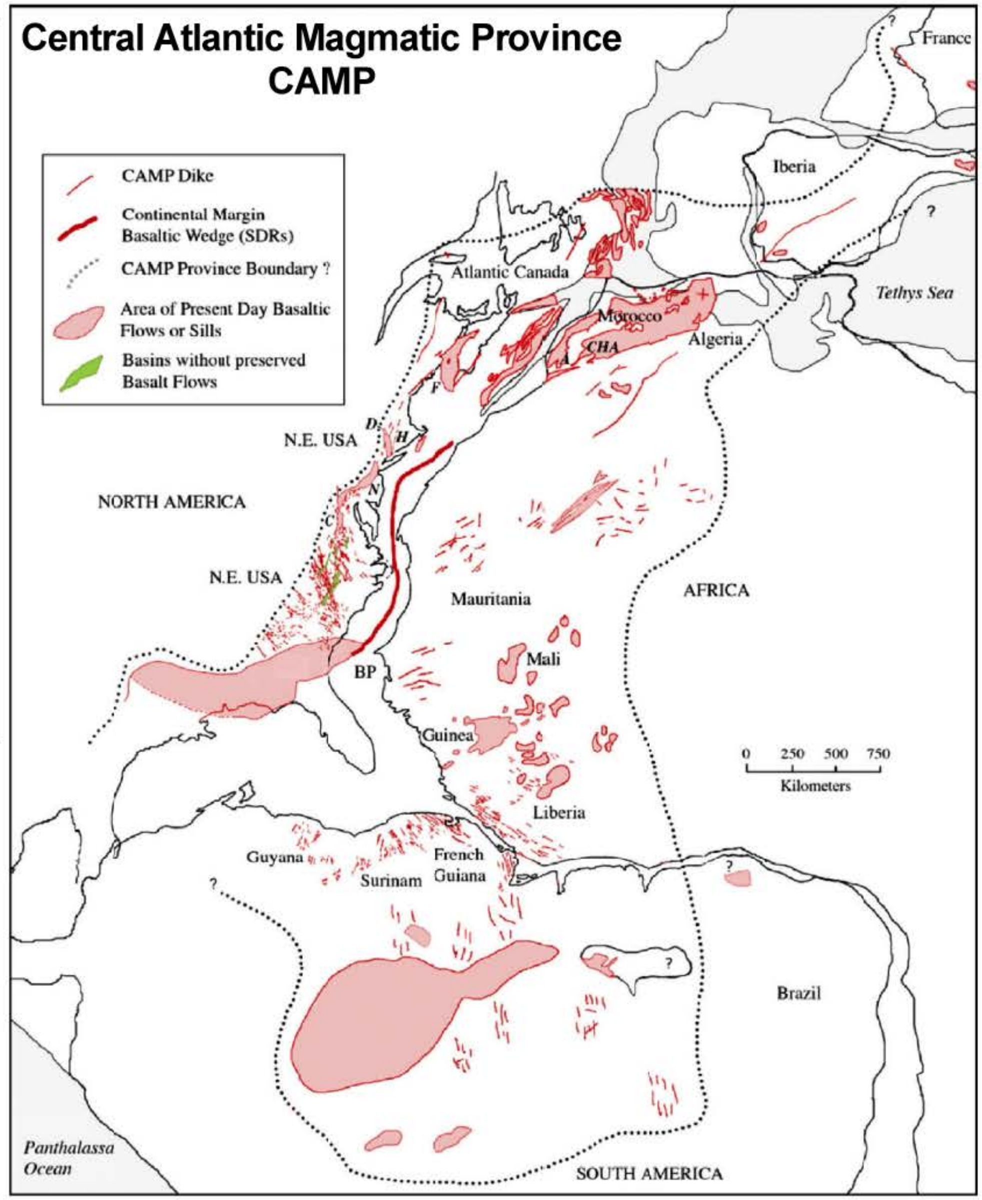
Correlation and geochemistry of extrusive rocks in various Triassic-Jurassic rift basins in eastern North America. Abbreviations are for basalt geochemistry are: HTQ = high-titanium quartz-normative; HFQ = high-iron quartz-normative; LTQ = low-titanium quartz-normative; HFTQ = high-iron & titanium quartz-normative basalt. Modified from Schlische et al. (2003) based on Olsen (1997).

Contact between Orange Mountain Basalt and underlying Passaic Fm. *in quarry in West Paterson, NJ. Photo by Roy Schlische.*

Fig. RL-4. *Extrusive rocks of the Newark basin and other rift basins in eastern North America.*



Map of Early Jurassic diabase dikes in eastern North America. *Dikes in the north trend NE-SW, and were intruded during syn-rift extension. Dikes in the south trend NW-SE and N-S, cut across many rift basins, and likely post-date syn-rift extension. Modified from Withjack & Schlische (2005) based on McHone (2000) and McHone et al. (2004).*



Map of the distribution of CAMP intrusive and extrusive rocks *in North America, South America, Africa, Iberia and Europe plotted on a Pangea reconstruction. Abbreviations are: A=Argana basin; BP=Blake Plateau; C=Culpeper basin; CHA=Central High Atlas; D=Deerfield basin; F=Fundy basin; H=Hartford basin; N=Newark basin. Modified from McHone (2000) and Whiteside et al. (2007).*

Fig RL-5. *Central Atlantic Magmatic Province.*

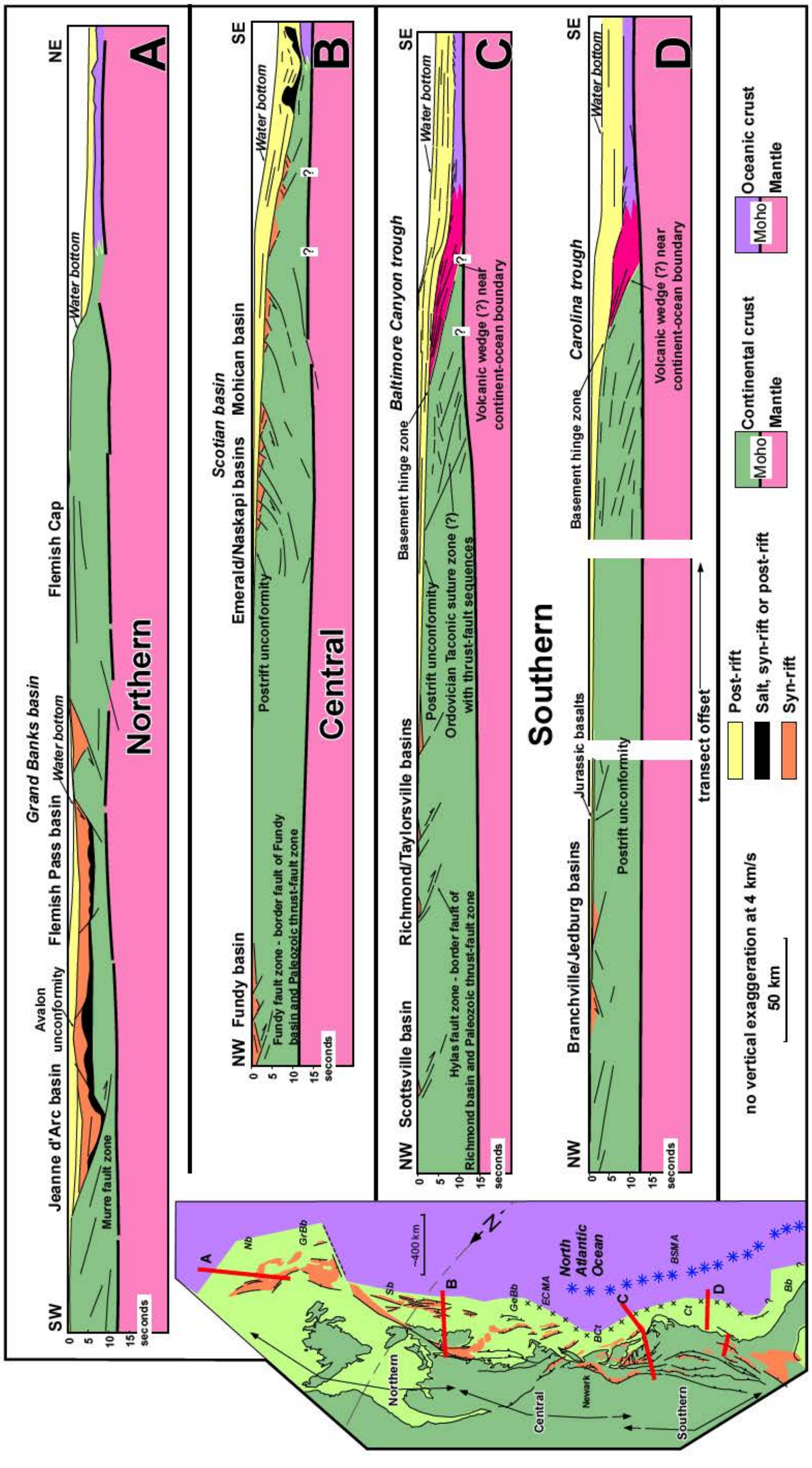


Fig. RL-6. *Regional transects across the northern, central, and southern segments of the eastern North American margin. Although all rift basins contain lava flows and/or intrusives, only the southern and part of the central segments of the margin contain seaward-dipping reflectors. Thus, only part of the margin is considered "volcanic." Abbreviations are: Bb=Bahamas basin, BCt=Baltimore Canyon trough; BSMA=Blake Spur magnetic anomaly; Ct=Carolina trough; ECMA=East Coast magnetic anomaly; GeBb=Georges Bank basin; GrBb=Grand Banks basin; Nb=Newfoundland basin; Sb=Scotian basin. Modified from Withjack & Schlische (2005).*

Exit Ringing Rocks park and turn left on Bridgeton Hill Road. Descend the hill underlain by the diabase sill, cross the canal, and proceed to Rt 32. Turn right on Rt 32 and follow the Delaware River south for about 12 km, passing through Frenchtown and passing Tinicum, Park. Approximately 2.5 km past Tinicum Park, turn right onto East Dark Hollow Road and climb out of the deeply incised portion of the Delaware valley.

Stay straight on E. Dark Hollow Road for ~5 km to the crossroads of Tinicum. Stay right on E. Dark Hollow Road and descend into the Tohickon valley. Make a sharp left to cross Tohickon Creek and pass the restored mill. Continue straight, cross the covered bridge, and then bear left onto Covered Bridge Road. There are exposures of Passaic Formation along the road. Approximately 1 km past the covered bridge turn left onto Stump Road and proceed for 1.5 km to the entrance of Ralph Stover Park. Turn right and park.

STOP 3. Ralph Stover State Park.

Walk from the parking lot to the bridgehead. Here is an excellent opportunity to observe and discuss the deep incision of the Delaware River and its tributaries like Tohickon Creek. Most tributaries like Tohickon Creek are steep, convex, and deeply incised as they reach the Delaware River. This incision is being driven by several knickpoints (Fig. RL-7) that presumably originated on the Delaware and are now moving like waves of erosion up the tributaries, etching out the Piedmont, Newark basin, and the Ridge and Valley. All of the major rivers like the Delaware are incised some 100 m below late Miocene upland gravel deposits at the Fall Zone. So if the incision originated at the Fall Zone and has been going on for the past 5-10 m.y., the long-term incision rate is ~ 10-20 m/m.y. One possible source of the incision is predicted by recent full-coupled dynamic mantle models that argue for 100-200 m of lat Cenozoic Coastal Plain and Piedmont uplift being driven by return flow to the sinking Farallon plate (Moucha et al., 2008).

Cross the bridge and follow the road to the top of the hill past the homes. Look for a yellow-blazed trail that heads off to the right into the woods. This trail will go out to high cliff exposures of fine-grained, lacustrine facies of the Passaic Formation. Watch your step: the trail that descends the cliff is slippery and dangerous.

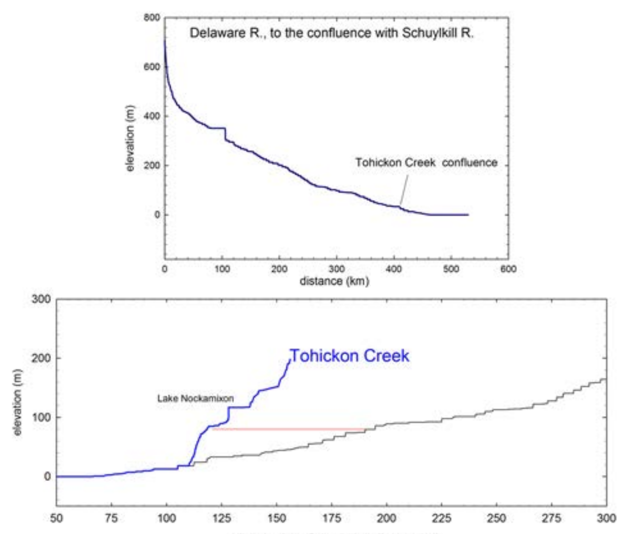


Fig. RL-7. *Longitudinal profile of Tokickon Creek hung plotted on the Delaware River profile.*

Red cyclical lacustrine strata of the Lockatong Fm are exposed in Ralph Stover State Park in Pennsylvania and across the Delaware River at the Devil's Tea Table in New Jersey. A measured section of similar Passaic Fm Member C from the Devil's Tea Table (Fig. RL-8) shows red, purple, gray and black mudstone. The red mudstones are the deposits of very shallow lakes or playas; deeper-water (perennial) lacustrine deposits occur every 5-7 m. The stratigraphic column in outcrop is very similar to that measured in the core from the Titusville drill hole located 24 km to the southeast. However, the outcrop section, located closer to the border fault, is ~11% thicker than the section in the core; this reflects higher subsidence rates along the border-fault margin of the half graben (Schlische, 1992).

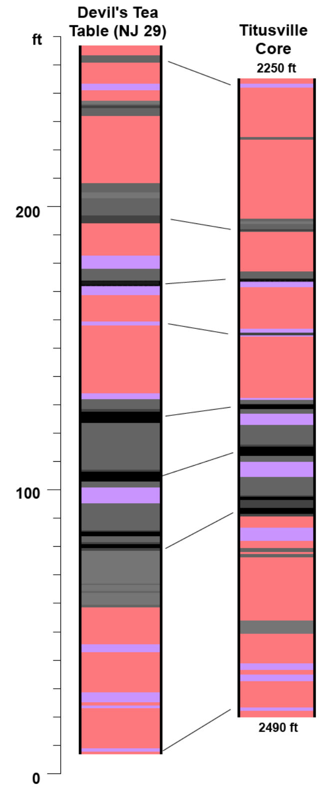
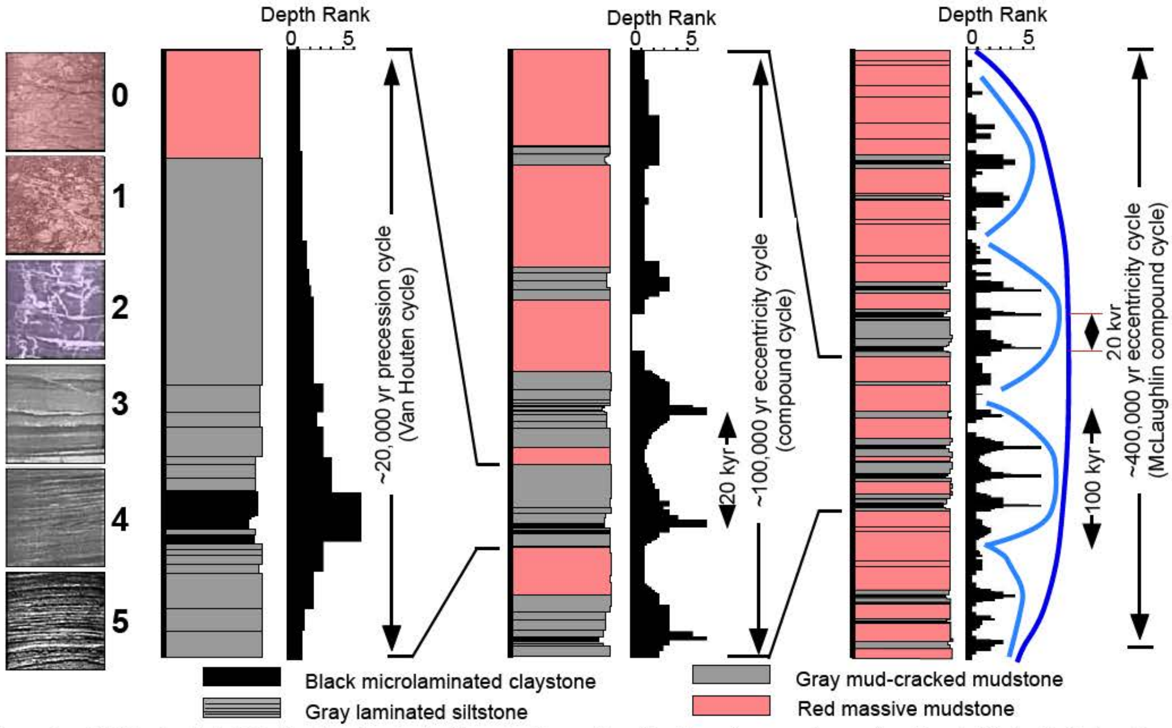


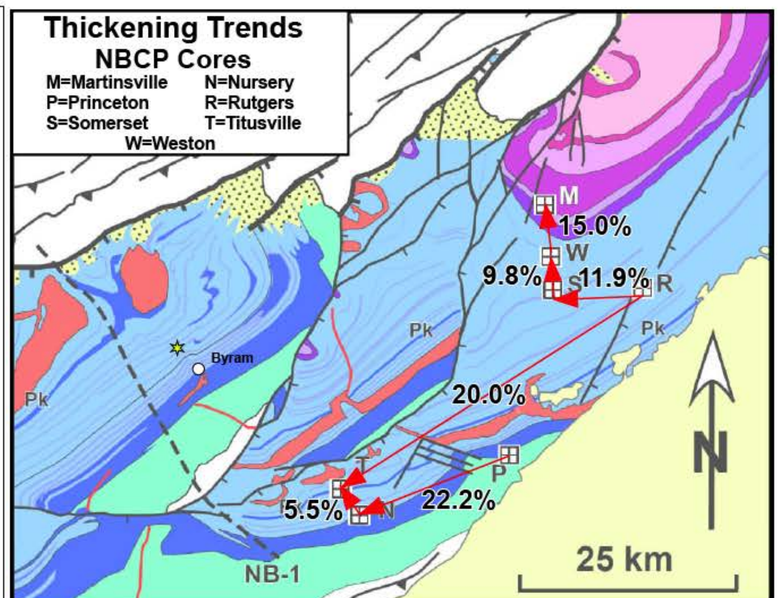
Fig. RL-8. *Measured sections of Member C of the Passaic Formation. The Devil's Tea Table section is located along Route 29 on the New Jersey side of the Delaware River. Note that the outcrop section is thicker than the correlative interval in the Titusville core, located farther from the border-fault system. Modified from Schlische (2004) based on data from P.E. Olsen.*

The strata here and throughout most of the Newark basin section exhibit a pervasive cyclicity in sediment fabrics, color, and total organic carbon (from microlaminated black shale to extensively mudcracked and bioturbated red mudstone) (e.g., Olsen, 1986; Olsen et al., 1996a) (Fig. RL-9). The basic cycles, named Van Houten cycles, represent the deepening and a shoaling of a lake. Van Houten cycles are grouped into at least two types of compound cycles, in which, for example, the deepest lake facies in succeeding Van Houten cycles first becomes deeper and then shallower (Fig. RL-9). Olsen (1986) quantitatively analyzed the cycles' periodicity by depth-ranking stratigraphic successions (i.e., assigning a relative water depth rank to each facies on the basis of color, sedimentary structures, and type and preservation of fossils; Fig. RL-9), subjecting the depth-rank curves to Fourier analysis (see Fig. RL-10) to determine periodicities in thickness, and converting the thickness periodicities to time using varve-calibrated sedimentation rates. Van Houten cycles have periods of ~20 kyr; compound cycles have periods of ~100 and ~400 kyr (Olsen, 1986; Olsen et al., 1989; Olsen and Kent, 1996, 1999). The ratio of Van Houten to compound cycle thickness for a given section always yields the typical Milankovitch ratio (1:5:20, which is equivalent to 20,000:100,000:400,000), even though the thicknesses of cycles commonly vary from section to section. For example, the Skunk Hollow and Tohickon members of the Lockatong Fm. are thicker in outcrop at Byram, NJ, relative to the cored section from the Nursery drill hole (Fig. RL-9). Although the thickness changes appear substantial in Figure RL-9 because of the 100:1 vertical exaggeration, the actual thickness changes are subtle; a slope of <0.5° accounts for the thickness changes in Figure RL-9. Similar thickness changes are observed in all correlative sections of core from adjacent drill sites (Fig. RL-9).

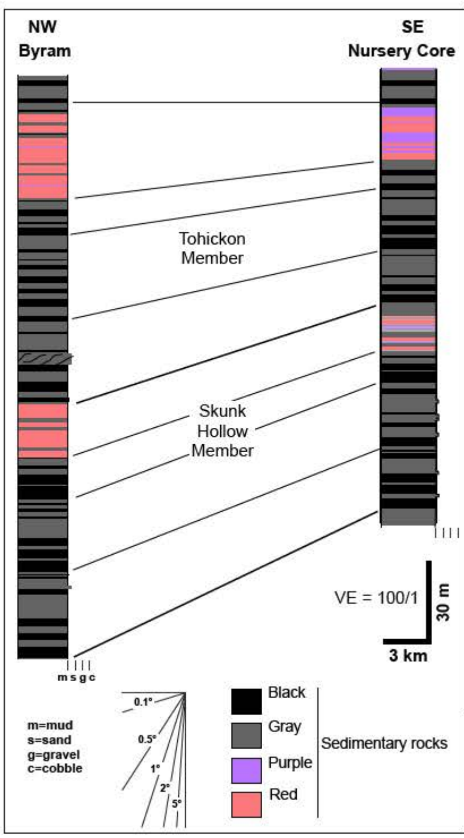
The pervasive cyclicity in lacustrine strata from the Newark basin, confirmed by Fourier analysis of depth-ranked sections (Fig. RL-10), provide absolute ages for much of the section. Figure RL-9 shows the Newark basin stratigraphic section in time rather than depth. Using the radiometric dates for the igneous rocks (202 Ma in RL-10) and counting the number of 400,000-year cycles shows that the rocks in Ralph Stover Park accumulated 219 million years ago.



Hierarchy of Milankovitch lake-level cycles in the Passaic Formation. *Depth rank uses color and sediment fabrics (left) to estimate relative water depth. Modified from Olsen et al. (1996a) and Olsen and Kent (1996).*

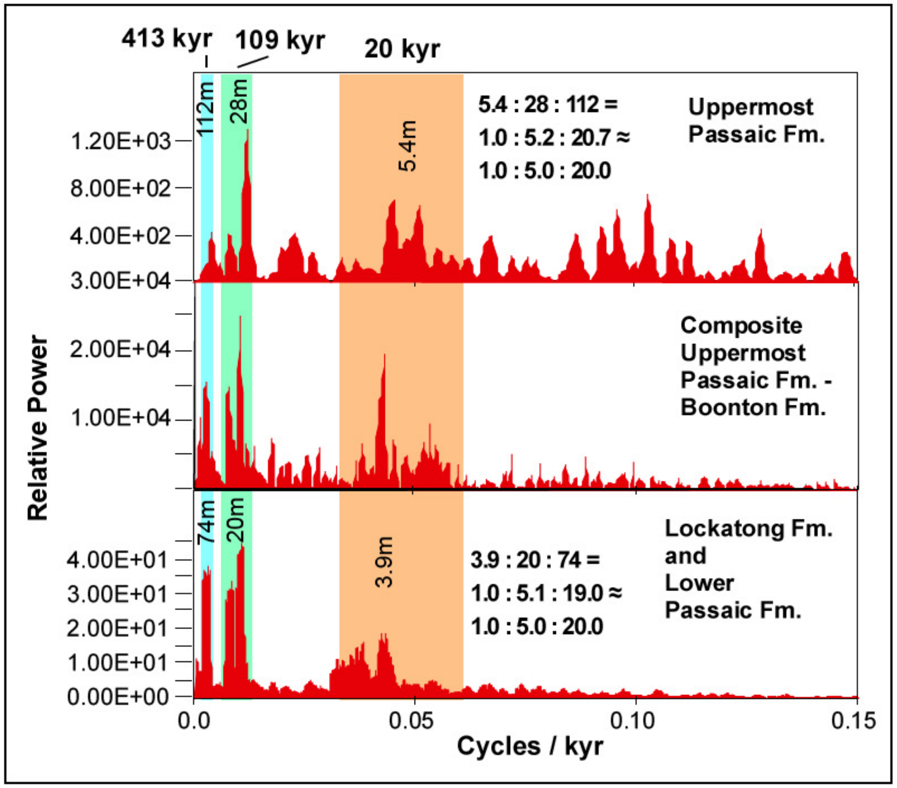


Percent-change in thickness between correlative units of adjacent NBCP cores. *Units generally thicken toward the center of the basin and toward the border-fault system. Modified from Schlische (2003) based on data in Olsen et al. (1996a).*

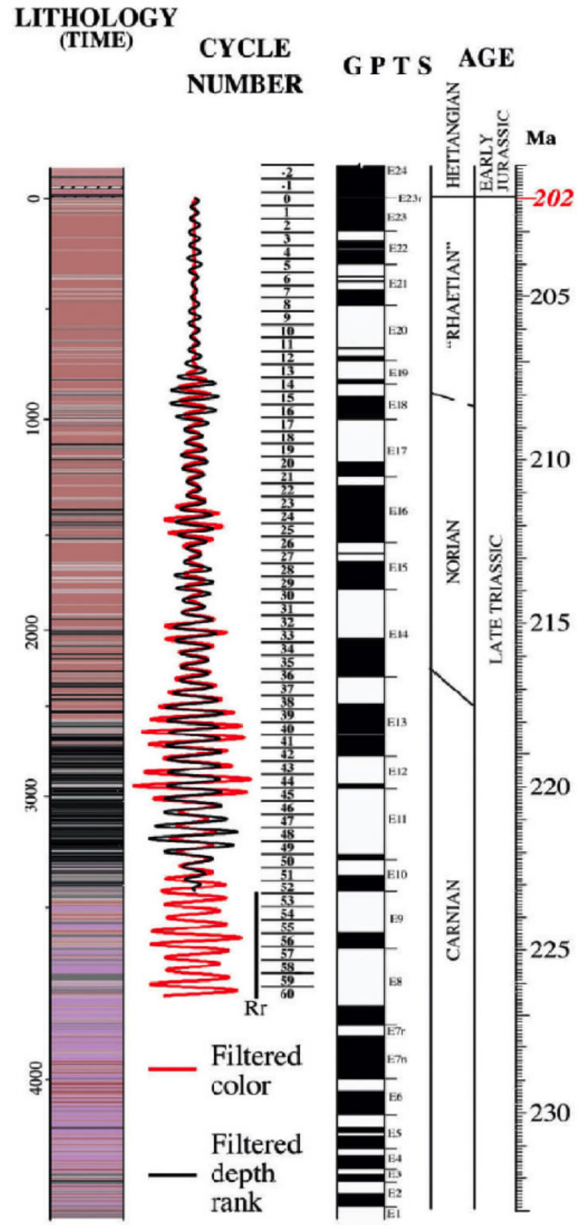


Core-to-outcrop correlation of the two members of the Lockatong Formation. *The units thicken subtly toward the border fault. A slope of <0.5° accounts for the differential thickening. Modified from Withjack et al. (2011) based on Olsen et al. (1996a).*

Fig. RL-9. *Milankovitch cycles in the Newark basin.*

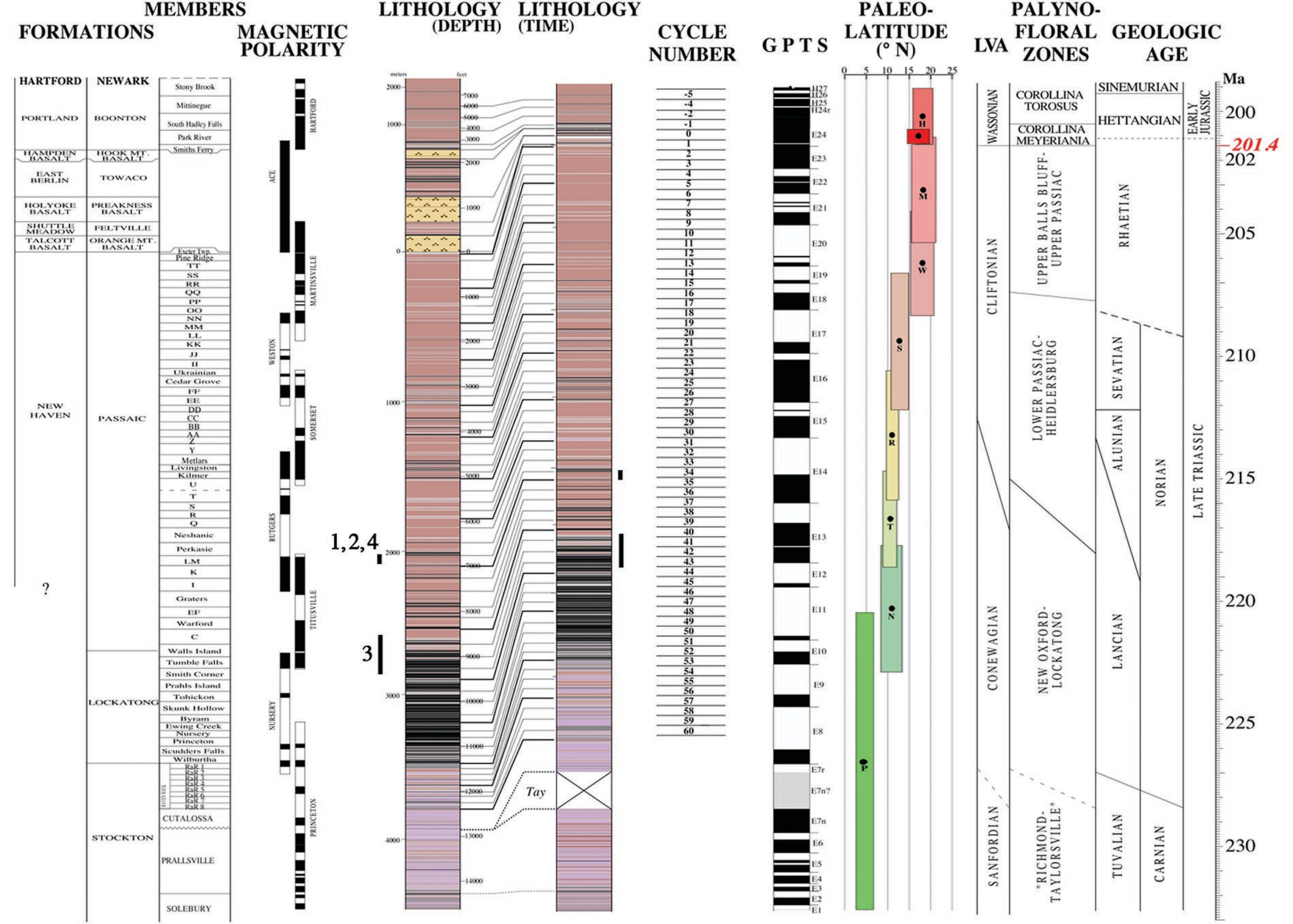


Power spectra derived from Fourier analysis of depth-ranked sectons of lacustrine strata *in the Newark basin. The lower power spectrum shows that cycles have thicknesses of 3.9 m, 20 m, and 74 m. The ratios of these cycles is very similar to the ratios of eccentricity cycles (~400,000 yrs and ~100,000 yerars) to precession cycles (20:5:1). Modified from Olsen et al. (1996b).*



Newark basin Astrochronology and geomagnetic Polarity Time Scale (APTS). *The time-stratigraphic section was produced by tuning the depth-ranked composite core-based stratigraphic section to the period (~400,000 years) of the McLaughlin compound cycles (eccentricity cycles). The spacing between the 1000 meter intervals of the original stratigraphic column are unequal, indicating that the accumulation rate varied through time. From Olsen et al. (2011) based on Kent & Olsen (1999).*

Fig. RL-10. *Fourier analysis and astronomically calibrated timescale.*


FORMATIONS
MEMBERS
MAGNETIC POLARITY
LITHOLOGY (DEPTH)
LITHOLOGY (TIME)
CYCLE NUMBER
G P T S
PALEO-LATITUDE (° N)
LVA
PALYNO-FLORAL ZONES
GEOLOGIC AGE
HARTFORD
NEWARK
PORTLAND
BOONTON
HAMPDEN BASALT
HOOK MT. BASALT
EAST BERLIN
TOWACO
HOLYOKE BASALT
PREAKNESS BASALT
SHUTTLE MEADOW
FELTVILLE
TALCOTT BASALT
ORANGE MT. BASALT
NEW HAVEN
PASSAIC
LOCKATONG
STOCKTON
?
Stony Brook
Mittinegue
South Hadley Falls
Park River
Smiths Ferry
Exeter Twp.
Pine Ridge
Ukrainian
Cedar Grove
Metlars
Livingston
Kilmer
Neshanic
Perkasie
Graters
Warford
Walls Island
Tumble Falls
Smith Corner
Prahls Island
Tohickon
Skunk Hollow
Byram
Ewing Creek
Nursery
Princeton
Scudders Falls
Wilburtha
RAVEN ROCK
CUTALOSSA
PRALLSVILLE
SOLEBURY
HARTFORD
ACE
MARTINSVILLE
WESTON
SOMERSET
RUTGERS
TITUSVILLE
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1,2,4
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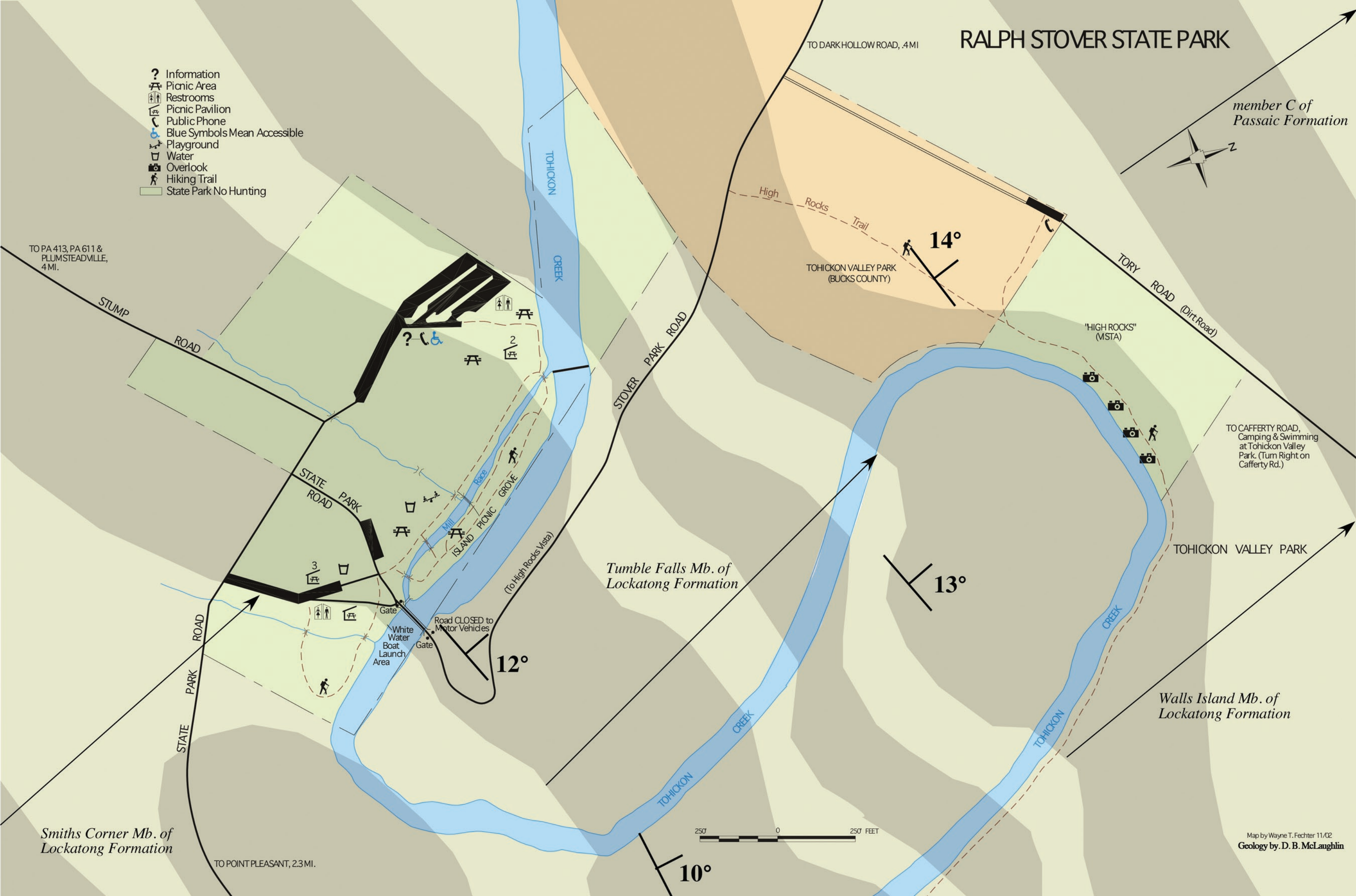
CLIFTONIAN
CONEWAGIAN
SANFORDIAN
COROLLINA TOROSUS
COROLLINA MEYERIANIA
UPPER BALLS BLUFF-UPPER PASSIAC
LOWER PASSIAC-HEIDLERSBURG
NEW OXFORD-LOCKATONG
"RICHMOND-TAYLORSVILLE"
SINEMURIAN
HETTANGIAN
RHAETIAN
SEVATIAN
ALUNIAN
LANCIAN
TUVALIAN
NORIAN
CARNIAN
EARLY JURASSIC
LATE TRIASSIC


RALPH STOVER STATE PARK
Information
Picnic Area
Restrooms
Picnic Pavilion
Public Phone
Blue Symbols Mean Accessible
Playground
Water
Overlook
Hiking Trail
State Park No Hunting
TO PA 413, PA 611 & PLUMSTEADVILLE, 4 MI.
STUMP ROAD
STATE PARK ROAD
TO POINT PLEASANT, 2.3 MI.
TO DARK HOLLOW ROAD, .4 MI
TOHICKON CREEK
Mill Race
ISLAND PICNIC GROVE
Gate
White Water Boat Launch Area
Road CLOSED to Motor Vehicles
(To High Rocks Vista)
STOVER PARK ROAD
High Rocks Trail
TOHICKON VALLEY PARK (BUCKS COUNTY)
TORY ROAD (Dirt Road)
"HIGH ROCKS" (VISTA)
TO CAFFERTY ROAD, Camping & Swimming at Tohickon Valley Park. (Turn Right on Cafferty Rd.)
TOHICKON VALLEY PARK
member C of Passaic Formation
Tumble Falls Mb. of Lockatong Formation
Walls Island Mb. of Lockatong Formation
Smiths Corner Mb. of Lockatong Formation
12°
13°
14°
10°
250' 0 250' FEET
N
Map by Wayne T. Fechter 11/02
Geology by. D. B. McLaughlin

From the cliffs, walk due north to the parking area, the van will be waiting for you there. From the parking lot, turn left onto Tory Road, then left on Cafferty Road, then right onto Smithtown Road. Descend back to the Delaware River and Rt 32. Turn left on Rt 32 and follow the road north back past the Frenchtown bridge all the way back to Upper Black Eddy. Turn right and cross the Delaware River at Upper Black Eddy to Milford, NJ. Cross the railroad tracks, then make the next left onto Spring Garden Street. Follow Spring Garden Street, whiich becomes River Road along the Delaware River in New Jersey. Continue north with the high red-bed cliffs to your right. After about 2 km, you will pass the junction with Spring Garden Road. After this intersection, begin looking to the right for wide pull offs that can be used to park 1 or more vans. The largest pull off is opposite the train mile marker 37. However, there are other pull offs for one or two vans further up the road. Please exercise caution as River Road is very narrow.

Observe the outcrops from the safety of the railroad tracks opposite the guardrail.

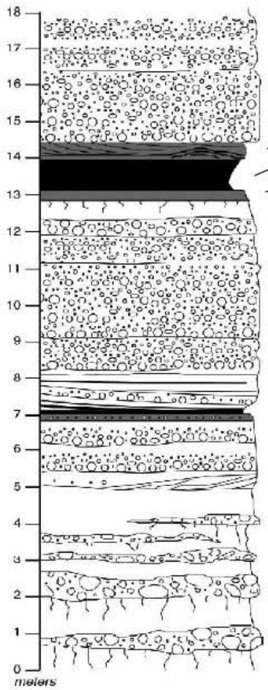
STOP 5: Perkasie Member of Passaic Formation at Pebble Bluff, NJ

These outcrops consist of thick sequences of red conglomerate and sandstone alternating with cyclical black, gray, and red mudstone and sandstone of the Perkasie Member of the Passaic Formation (Fig. RL-11). Depositional environments ranged from deep perennial lakes (black shales) to debris-flow deposits of alluvial fans (red poorly sorted conglomerate) (Olsen et al., 1989). The sequence of strata above the 12-m-mark represents the progressive flooding of the distal toes of an alluvial fan during rising lake levels, followed by lake highstand and regression.

Individual Van Houten cycles of the Perkasie Member in this area average 7 m thick, as compared with a mean of 4.4 m at New Brunswick, NJ, and Pottstown, PA, and 5.3 m in the Titusville core (Fig. RL-11; Olsen et al., 1996a). These thickness trends show that subsidence rates increased from the hinged margin of the half graben toward the border-fault system as well as from the lateral edges toward the center of the basin (Schlische, 1992).

The sedimentologic and stratigraphic features of interest at this outcrop are the coarse pebble conglomerates that represent paleo-alluvial fans sourced in a former footwall escarpment to the northwest. The composition of these conglomerates is enigmatic. On one hand, there are limestone fanglomerate facies that also contain clasts of weathered granite that are clearly indicating a local, Blue Ridge and Lehigh Valley provenance. On the other hand, the majority of the clasts are quartzite and vein quartz that does not have an obvious source in the Blue Ridge, Lehigh Valley, or Ridge and Valley. Overall percentage of limestone and granite clasts increase upsection. One hypothesis that remains to be fully tested is that the quartz-pebble conglomerate is derived from the former Permian molasse that may have still been present at the time of rift initiation. That molasse was locally recycled into these paleo-alluvial fans as the footwall was unroofed.

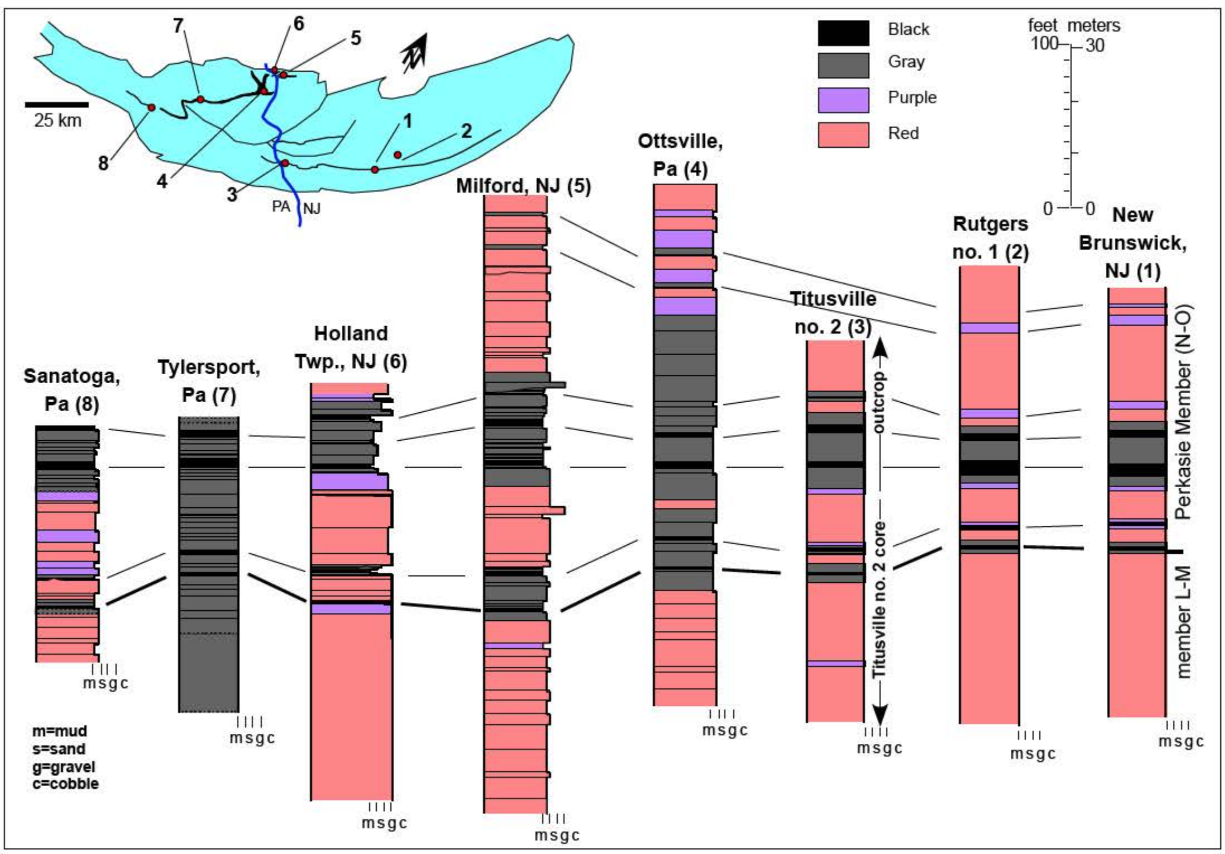
Field trip end. *Continue north on River Road; turn left off of Rt 627 to the town of Riegelsville, PA. Cross the Delaware River at Riegelsville and proceed to the stoplight with Rt 611. Turn left on Rt 611 and continue south to Rt 212. Turn right on Rt 212 and retrace route back to Springtown and Lehigh.*



Measured section of lacustrine deposits interbedded with alluvial-fan deposits. *Modified from Olsen et al. (1989).*

Stop 4: Pebble Bluff near Milford, NJ

Stratigraphy: Late Triassic Passaic Fm., Perkasie Mbr.
Lithology: Mudstone, sandstone, and conglomerate
Distance from border fault: ~ 2 km



Measured sections of the Perkasie Member of the Passaic Formation from across the Newark basin. *The sections generally thicken and coarsen toward the border-fault system and thin toward the lateral edges of the basin. Modified from Olsen et al. (1989).*

Fig. RL-11.

References

Bally, A.W., Withjack, M.O., Meisling, K.E., and Fisher, D.A. 1991. Seismic expression of structural styles. Geological Society of America, Short Course Notes, 1991 National Meeting, 59 p.

Barruol, G., Silver, P.G., Vauchez, A. 1997. Seismic anisotropy in the eastern United States: deep structure of a complex continental plate. Journal of Geophysical Research, 102, 8329-8348.

Bensen, G.D., Ritzwoller, M. H., and Y. Yang, Y. 2009. A 3-D shear velocity model of the crust and uppermost mantle beneath the United States from ambient seismic noise. Geophysical Journal International,177, 1177-1196. doi: 10.1111/j.1365-246X.2009.04125.x.

Cornet, B., and Olsen, P.E. 1985. A summary of the biostratigraphy of the Newark Supergroup of eastern North America, with comments on early Mesozoic provinciality. In: Weber, R., ed., III Congresso Latinoamericano de Paleontologia. Mexico., Simposio Sobre Floras del Triasico Tardio, su Fitografia y Paleoecologia., Memoria., p. 67-81.

Darbyshire, F.A., Eaton, D.W., Frederiksen, A.W., Ertolahti, L., 2007. New insights into the lithosphere beneath the Superior Province from Rayleigh wave dispersion and receiver function analysis. Geophys. J. Int., 169, 1043-1068, doi:10.1111/j.1365-246X.2006.03259.x

Faill, R.T. 1973. Tectonic development of the Triassic Newark-Gettysburg basin in Pennsylvania. Geological Society of America Bulletin, 84, 725-740.

Faill, R.T. 1988. Mesozoic tectonics of the Newark basin, as viewed from the Delaware River. In: Husch, J.M. and Hozik, M.J. (eds) Geology of the Central Newark Basin, Field Guide and Proceedings. Fifth Meeting of the Geological Association of New Jersey, Rider College, Lawrenceville, 19-41.

Faill, R.T., 2003, The early Mesozoic Birdsboro central Atlantic margin basin in the Mid-Atlantic region, eastern United States: geological Society of America Bulletin, 115, 406-421.

Faill, R.T. 1997a. A geologic history of the north-central Appalachians; Part 1, Orogenesis from the Mesoproterozoic through the Taconic Orogeny. American Journal of Science, 297, 551-619.

Faill, R.T. 1997b. A geologic history of the north-central Appalachians; Part 2, The Appalachian Basin from the Silurian through the Carboniferous: American Journal of Science, 297, 729-761.

Fisher, K.M. 2002. Waning buoyancy in the crust roots of old mountains. Nature, 417, 933-936.

Forte, A.M., Mitrovica, J.X., Moucha, R., Simmons, N.A., and Grand, S.P., 2007, Descent of the ancient Farallon slab drives localized mantle flow below the New Madrid seismic zone: Geophysical Research Letters, 34, doi:10.1029/2006GL027895.

Forte, A.M., Moucha, R., Simmons, N.A., Grand, S.P., Mitrovica, J.X. 2010. Deep mantle contributions to the surface dynamics of the North American continent. Tectonophysics, 481(1-4), 3-15.

Fouch, M.J., Fischer, K.M., Parmentier, E.M., Wysession, M.E., Clarke, T.J. 2000. Shear wave splitting, continental keels, and patterns of mantle flow. Journal of Geophysical Research, 105, 6255-6275.

Fouch, M.J., Rondenay, S. 2006. Seismic anisotropy beneath stable continental interiors. Physics of the Earth and Planetary Interiors, 158, 292-320.

Gawthorpe, R.L., and Leeder, M.R., 2000. Tectono-sedimentary evolution of active extensional basins. Basin Research, 12, 195-218.

Grand, S. P., 1994, Mantle shear structure beneath the Americas and surrounding oceans: J. Geophys. Res., 99, 11,591â€“11,621.

Hames, W.E., McHone, J G., Renne, P.R. and Ruppel, C. (eds) 2003. The Central Atlantic Magmatic Province, Insights from Fragments of Pangea. American Geophysical Union, Geophysical Monograph 136, 267 p.

Hames, W.E., Renne, P.R., and Ruppel, C., 2000, New evidence for geologically instantaneous emplacement of earliest Jurassic Central Atlantic magmatic province basalts on the North American margin: Geology, v. 28, p. 859-862.

Haq, B.L., Hardenbol, J., and Vail, P. R. 1987. Chronology of fluctuating sea levels since the Triassic. Science, 235, 1156-1167.

Hatcher, R.D., Thomas, W.A., and Viele, W.A. 1989. The Appalachian-Ouachita Orogen in the United States: The geology of North America, Decade of North American Geology, v. F-2, 767 p.

Hayden, T., Kominz, M., Powars, D.S., Edwards, L.E., Miller, K.G., Browning, J.V., Kulpecz, A.A. 2008. Impact effects and regional tectonic insights: Backstripping the Chesapeake Bay impact structure. Geology, 36, 327-330.

Hegarty, K.A., Foland, S.C., Cook, A.C., Green, P.F., and Duddy, I.R. 2007. Direct measurement of timing; underpinning a reliable petroleum system model for the Mid-Continental Rift System. AAPG Bulletin, 91, 959-

979.
Hull, J. 1988. Thickness-displacement relationships for deformation zones. Journal of Structural Geology, 10, 431-435.
Hulver, M.L. 1997. Post-orogenic evolution of the Appalachian mountains system and its foreland: Chicago, Illinois, University of Chicago, Ph.D. dissertation, 1055 p.
Hutchinson, D.R. 2005. North America; Atlantic margin. In: Selley, R. C., Cocks, R., and Plimer, I.R. (eds) Encyclopedia of geology, 4, 92-108.
Judson, S. 1975. Evolution of Appalachian topography. In: Melhorn, W.N. and Flemal, R.C. (eds) Theories of landform development. Proceedings, 6th Annual Geomorphology Symposia, Binghamton, New York, 1-28.
Karner, G.D. and Watts, A.B. 1982. On isostasy at Atlantic-type continental margins. Journal of Geophysical Research, 87, 2923-2948.
Kent, D.V., and Olsen, P.E. 1999. Astronomically tuned geomagnetic polarity time scale for the Late Triassic. Journal of Geophysical Research, 104, 12,831-12,841.
Kent, D.V., Olsen, P.E., and Witte, W.K. 1995. Late Triassic-Early Jurassic geomagnetic polarity and paleolatitudes from drill cores in the Newark rift basin (Eastern North America). Journal of Geophysical Research, 100 (B8), 14,965-14,998.
Knott, S.D. 1994. Fault zone thickness versus displacement in the Permo-Triassic sandstones of NW England. Journal of the Geological Society, London, 151, 17-25.
Knott, S.D., Beach, A., Brockbank, P.J., Brown, J.L., McCallum, J.E., and Welbon, A.I. 1996. Spatial and mechanical controls on normal fault populations. Journal of Structural Geology, 18, 359-372.
Kohn, B.P., Wagner, M.E., Lutz, T.M., and Organist, G.. 1993. Anomalous thermal regime, central Appalachian Piedmont: evidence from sphene and zircon fission track dating. Journal of Geology, 101, 779-794.
Levin, V., and Park, J. 2000a. Shear zones in the Proterozoic lithosphere of the Arabian Shield and the nature of the Hales discontinuity. Tectonophysics, 323, 131-148.
Levin, V., Park, J., Brandon, M., and Menke, W. 2000b. Thinning of the upper mantle during the late Paleozoic Appalachian orogenesis. Geology, 28, 239-242.
Levin, V., Lerner Lam, A., and Menke, W. 1995. Anomalous mantle structure at the Proterozoic-Paleozoic boundary in northeastern US. Geophysical Research Letters, 22, 121-124.
Levin, V., Menke, W., and Park, J. 1999. Shear wave splitting in the Appalachians and the Urals: A case for multilayered anisotropy, Journal of Geophysical Research, 104, 17975-17987.
Levin, V., Menke, W., and Park, J. 2000. No regional anisotropic domains in the northeastern U.S. Appalachians. Journal of Geophysical Research, 105, 19,029-19042.
Levine, J.R. 1986. Deep burial of coal-bearing strata, anthracite region, Pennsylvania; sedimentation or tectonics? Geology, 14, 577-580.
Lyttle, P.T., and Epstein, J.B. 1987. Geologic map of the Newark 1°x2° quadrangle, New Jersey, Pennsylvania and New York. U.S. Geological Survey, Miscellaneous Investigations Series Map I-1715, 1:250,000.
Malinconico, M. L. 2010. Synrift to early postrift basin-scale groundwater history of the Newark basin based on surface and borehole vitrinite reflectance data. In: Herman, G.C. and Serfes, M.E. (eds) Contributions to the Geology and Hydrogeology of the Newark Basin. New Jersey Geological Survey Bulletin 77, C1-C38.
Manspeizer, W. and Cousminer, H. L. 1988. Late Triassic-Early Jurassic synrift basins of the U.S. Atlantic margin. In: Sheridan, R.E. and Grow, J.A. (eds) The Atlantic continental margin—U.S. Geological Society of America, The Geology of North America, I-2, 197-216.
Marzolli, A., Renne, P.R., Piccirillo, E.M., Ernesto, M., Bellieni, G. and deMin, A. 1999. Extensive 200-million-year-old continental flood basalts of the Central Atlantic Magmatic Province. Science, 284, 616-618.
May, P.R. 1971. Pattern of Triassic-Jurassic diabase dikes around the North Atlantic in the context of the predrift configuration of the continents. Geological Society of America Bulletin, 82, 1285-1292.
McHone, J.G. 1996. Broad-terrane Jurassic flood basalts across northeastern North America. Geology, 24, 319-322.
McHone, J.G. 2000. Non-plume magmatism and rifting during the opening of the central Atlantic Ocean. Tectonophysics, 316, 287-296.
McHone, J.G., 2003.Volatile emissions from Central Atlantic Magmatic Province basalts: Mass assumptions and environmental consequences. In: Hames, W.E., McHone, J.G., Renne, P.R. and Ruppel, C. (eds) The Central Atlantic Magmatic Province, Insights from Fragments of Pangea. American Geophysical Union, Geophysical Monograph 136, p. 241-254.
McLaughlin, D.B. 1948. Continuity of strata in the Newark series. Papers of the Michigan Academy of Science, Arts and Letters, 32, 295-303.
Menke, W., and Levin, V. 2002. Anomalous seaward dip of the lithosphere-asthenosphere boundary beneath

northeastern US detected using differential-array measurements of Rayleigh waves. Geophysical Journal International, 149, 413-421.
Miller, D.S. and Duddy, I.R. 1989. Early Cretaceous uplift and erosion of the northern Appalachian basin New York, based on apatite fission track analyses. Earth and Planetary Science Letters, 93, 35-49.
Miller, K.G., Kominz, M.A., Browning, J.V., Wright, J.D., Mountain, G.S., Katz, M.E., Sugarman, P.J., Cramer, B.S., Critie-Blick, N., and Pekar, S. 2008. The Phanerozoic record of global sea-level change. Science, 310, 1293-1298.
Moucha, R., Forte, A.M., Mitrovica, J.X., Rowley, D.B., Quere, S., Simmons, N.A., and Grand, S.P. 2008. Dynamic topography and long-term sea-level variations: There is no such thing as a stable continental platform. Earth and Planetary Science Letters, 271, 101-108.
Nettles, M., and Dziewonski, A.M. 2008. Radially anisotropic shear velocity structure of the upper mantle globally and beneath North America. Journal of Geophysical Research, 113, B02303, doi:10.1029/2006JB004819.
Oliver, J., Cook, F., and Brown, L. 1983. COCORP and continental crust. Journal of Geophysical Research, 88, 3329-3347.
Olsen, P.E. 1986. A 40-million-year lake record of early Mesozoic climatic forcing. Science, 234, 842-848.
Olsen, P.E. 1988. Continuity of strata in the Newark and Hartford basins of the Newark Supergroup. In: Froelich, A. J. and Robinson, G. R., Jr. (eds) Studies of the Early Mesozoic Basins in the Eastern United States. U.S. Geological Survey Bulletin, 1776, 6-18.
Olsen, P.E. 1997. Stratigraphic record of the early Mesozoic breakup of Pangea in the Laurasia-Gondwana rift system. Annual Reviews of Earth and Planetary Science, 25, 337-401.
Olsen, P.E. and Whiteside, J.H. 2008. Pre-Quaternary Milankovitch cycles and climate variability. In: Gornitz, V. (ed.), Encyclopedia of Paleoclimatology and Ancient Environments, Earth Science Series, Kluwer Academic Publishers, Dordrecht, the Netherlands, 826-835 (ISBN: 978-1-4020-4551-6).
Olsen, P.E., and Kent, D.V. 1996. Milankovitch climate forcing in the tropics of Pangaea during the Late Triassic: Palaeogeography, Palaeoclimatology, Palaeoecology, 122., 1-26.
Olsen, P.E., and Kent, D.V. 1999. Long-period Milankovitch cycles from the Late Triassic and Early Jurassic of eastern North America and their implications for the calibration of the Early Mesozoic time scale and the long-term behaviour of the planets. Philosophical Transactions, Royal Society of London, Series A, 357, 1761-1786.
Olsen, P.E., Kent, D.V. and Whiteside, J. H. 2011. Implications of the Newark Supergroup-based astrochronology and geomagnetic polarity time scale (Newark-APTS) for the tempo and mode of the early diversification of the Dinosauria. Earth and Environmental Science Transactions of the Royal Society of Edinburgh, 101, 201–229.
Olsen, P.E., Kent, D.V., Cornet, B., Witte, W.K. and Schlische, R.W. 1996a. High-resolution stratigraphy of the Newark rift basin (early Mesozoic, eastern North America). Geological Society of America Bulletin, 108, 40-77.
Olsen, P.E., Kent, D.V., Et-Touhami, M. and Puffer, J. 2003. Cyclo-, magneto-, and bio-stratigraphic constraints on the duration of the CAMP event and its relationship to the Triassic-Jurassic boundary. In: Hames, W.E., McHone, J.G., Renne, P.R. and Ruppel, C. (eds) The Central Atlantic Magmatic Province, Insights from Fragments of Pangea. American Geophysical Union, Geophysical Monograph 136, 7-32.
Olsen, P.E., Schlische, R.W., and Fedosh, M.S. 1996b. 580 kyr duration of the Early Jurassic flood basalt event in eastern North America estimated using Milankovitch cyclostratigraphy. In: Morales, M., ed., The Continental Jurassic: Museum of Northern Arizona Bulletin 60, 11-22.
Olsen, P.E., Schlische, R.W. and Gore, P.J.W. (eds). 1989. Tectonic, depositional, and paleoecological history of early Mesozoic rift basins, eastern North America. International Geological Congress Field Trip T-351, American Geophysical Union, 174 p.
Ott, A.N., Takita, C.S., Edwards, R.E., and Bollinger, S.W. 1991. Loads and yields of nutrients and suspended sediment transported in the Susquehanna River basin, 1985-89: Susquehanna River Basin Commission Report, Publication 136, 253 p.
Pavich, M.J., Leo, G.W., Obermeier, S.F., and Estabrook, J.R. 1989. Investigations of the characteristics, origin, and residence time of the upland residual mantle of the Piedmont of Fairfax County, VA. U.S. Gelogical Survey Professional Paper 1352, 114 p.
Pazzaglia, F.J. and Brandon, M.T. 1996. Macrogeomorphic evolution of the post-Triassic Appalachian mountains determined by deconvolution of the offshore basin sedimentary record. Basin Research, 8, 255-278.
Pazzaglia, F.J. and Gardner, T.W. 1994. Late Cenozoic flexural deformation of the middle U.S. Atlantic passive margin. Journal of Geophysical Research, 99 (B6), 12,143-12,157.
Poag, C.W. 1985. Depositional history and reference section for central Baltimore Canyon trough. In: Poag, C.W. (ed) Geologic Evolution of the United States Atlantic margin. New York, Van Nostrand Reinhold, 217-263.

Poag, C.W. 1992. U.S. Middle Atlantic continental rise; provenance, dispersal, and deposition of Jurassic to Quaternary sediments. In: Poag, C.W., and de Graciansky, P.C. (eds) Geologic evolution of Atlantic continental rises. New York, Van Nostrand Reinhold, 100-156.

Poag, C.W. and Sevon, W.D. 1989. A record of Appalachian denudation in postrift Mesozoic and Cenozoic sedimentary deposits of the U.S. Middle Atlantic continental margin. Geomorphology, 2, 119-157.

Rahl, J.M., Reiners, P.W., Campbell, I.H., Nicolescu, S., and Allen, C. M. 2003. Combined single grain (U/Th)/He and U/Pb dating of detrital zircons from the Navajo Sandstone, Utah. Geology, 31, 761-764.

Ratcliffe, N.M. 1988. Reinterpretation of the relationships of the western extension of the Palisades sill to the lava flows at Ladentown, New York, based on new core data. In Froelich, A.J., and Robinson, G.R., Jr. (eds) Studies of the Early Mesozoic Basins of the Eastern United States. U.S. Geological Survey Bulletin 1776, 113-135.

Ratcliffe, N.M., Burton, W.C., D'Angelo, R.M. and Costain, J.K. 1986. Low-angle extensional faulting, reactivated mylonites, and seismic reflection geometry of the Newark basin margin in eastern Pennsylvania. Geology, 14, 766-770.

Reed, J.S., Spotila, J.A., Eriksson, K.A., and Bodnar, R.J. 2005. Burial and exhumation history of Pennsylvania strata, central Appalachian basin: an integrated study. Basin Research, 17, 259-268.

Riggs, N.R., Lehman, T.M., Gehrels, G.E., and Dickinson, W.R. 1996. Detrital zircon link between headwaters and terminus of the Upper Triassic Chinle-Dockum paleoriver system. Science, 273, 97-100.

Roden, M.K. and Miller, D.S. 1989. Apatite fission-track thermochronology of the Pennsylvania Appalachian basin: Geomorphology, 2, 39-51.

Rychert, C.A., Rondenay, S., and Fischer, K.M. 2007. P-to-S and S-to-P imaging of a sharp lithosphere-asthenosphere boundary beneath eastern North America. Journal of Geophysical Research, 112, B08314, doi:10.1029/2006JB004619.

Rychert, C.A., Fischer, K.M., and Rondenay, S., 2005. A sharp lithosphere-asthenosphere boundary imaged beneath eastern North America. Nature, 436, 542-545, doi:10.1038/nature03904.

Sanders, J.E. 1963. Late Triassic tectonic history of northeastern United States. American Journal of Science, 261, 501-524.

Sbar, M., and Sykes, L. 1977. Seismicity and lithospheric stress in New York and adjacent areas. Journal of Geophysical Research, 82(36), 5771-5786.

Schettino, A. and Turco, E. 2009. Tectonic history of the western Tethys since the Late Triassic. Geological Society of America Bulletin, 123, 89-105. DOI: 10.1130/B30064.1

Schlische, R.W. 1992. Structural and stratigraphic development of the Newark extensional basin, eastern North America; Implications for the growth of the basin and its bounding structures. Geological Society of America Bulletin, 104, 1246-1263.

Schlische, R.W. 1993. Anatomy and evolution of the Triassic-Jurassic continental rift system, eastern North America. Tectonics, 12, 1026-1042.

Schlische, R.W. 1995. Geometry and origin of fault-related folds in extensional settings. AAPG Bulletin, 79, 1661-1678.

Schlische, R.W. 2003. Progress in understanding the structural geology, basin evolution, and tectonic history of the eastern North American rift system. In: LeTourneau, P.M. and Olsen, P.E. (eds) The Great Rift Valleys of Pangea in Eastern North America, Volume 1, Tectonics, Structure, and Volcanism. Columbia University Press, New York, 21-64.

Schlische, R.W. and Withjack, M. O. 2005. The early Mesozoic Birdsboro central Atlantic margin basin in the Mid-Atlantic region, eastern United States: Discussion. Geological Society of America Bulletin, 117, 823-828.

Schlische, R.W., 2004, Structural and Climatic Controls on Sedimentation in the Newark Rift Basin--Field Guide to Classic Outcrops Along and Near the Delaware River, NJ and PA: National Association of Geoscience Teachers Eastern Section Meeting, Newark, NJ.

Schlische, R.W., Withjack, M. O. and Olsen, P.E. 2003. Relative timing of CAMP, rifting, continental breakup, and inversion: tectonic significance. In: Hames, W.E., McHone, J.G., Renne, P.R. and Ruppel, C. (eds) The Central Atlantic Magmatic Province, Insights from Fragments of Pangea. American Geophysical Union, Geophysical Monograph 136, 33-59.

Schlische, R.W., Withjack, M.O., and Eisenstadt, G. 2002. An experimental study of the secondary deformation produced by oblique-slip normal faulting. AAPG Bulletin, 86, 885-906.

Schmandt, B., and Humphreys, E. 2010. Seismic heterogeneity and small-scale convection in the southern California upper mantle. Geochem. Geophys. Geosyst., 11, Q05004, doi:10.1029/2010GC003042.

Seeber, L. and Armbruster, J. 1988. Seismicity along the Atlantic seaboard of the U.S.; intraplate neotectonics and earthquake hazard. In: Sheridan, R.E. and Grow, J.A. (eds) The Atlantic Continental Margin, Geological Society

of America, The Geology of North America, I-2, 565-582.
Shanmugam, G. and Lash, G.G. 1982. Analogous tectonic evolution of the Ordovician foredeeps, southern and central Appalachians. Geology, 10, 562-566.
Sheridan, R.E. and Grow, J.A. 1988. The Atlantic continental margin; U.S., Geological Society of America, The Geology of North America v. I-2, 527 p.
Simpson, R.W., Jachens, R.C., Blakely, R.J., and Saltus, R.W. 1986. A new isostatic residual gravity map of the conterminous United States with a discussion on the significance of isostatic residual anomalies. Journal of Geophysical Research, 91, 8348-8372.
Steckler, M.S., Mountain, G.S., Miller, K.G., and Christie-Blick, N. 1999. Reconstruction of Tertiary progradation and clinoform development on the New Jersey passive margin by 2-D backstripping. Marine Geology, 154, 399-420.
Steckler, M.S., Watts, A.B., and Thorne, J.A. 1988. Subsidence and basin modeling at the U.S. Atlantic passive margin. In: Sheridan, R.E. and Grow, J.A. (eds) The Atlantic continental margin; U.S. Geological Society of America, The Geology of North American I-2, 399-416.
Steckler, M.S., Omar, G.I., Karner, G.D. and Kohn, B.P. 1993. Pattern of hydrothermal circulation with the Newark basin from fission-track analysis. Geology, 21, 735-738.
Sykes, L.R., Armbruster, J.G., Kim, W., and Seeber, L. 2008. Observations and tectonic setting of historic and instrumentally located earthquakes in the greater New York City - Philadelphia area. Bulletin of the Seismological Society of America, 98, 1696-1719, doi: 10.1785/0120070167.
Talwani, M. and Abreu, V. 2000. Inferences regarding initiation of oceanic crust formation from the U.S. east coast margin and conjugate south Atlantic margins. In: Mohriak, W. and Talwani, M. (eds) Atlantic rifts and continental margins. American Geophysical Union.
Thomas, W.A. 2006. Tectonic inheritance at continental margins. GSAToday, 16, doi: 10.1130/1052-5173(2006)016<4:TIAACM>2.0.CO;2
Turcott D. L. and Schubert, G. 2001. Geodynamics, second edition. Cambridge, Cambridge University Press, 528 p.
van der Lee, S., 2001. Deep below North America. Science, 294, 1297-1298.
van der Lee, S. 2002. High-resolution estimates of lithospheric thickness from Missouri to Massachusetts, USA. Earth and Planetary Science Letters, 203, 15-23.
van der Lee, S., and Frederiksen, A. 2005. Surface wave tomography applied to the North American upper mantle. In: Nolet, G., and Levander, A. (eds) Seismic Data Analysis and Imaging With Global and Local Arrays. Geophysical Monograph, 157, 67-80, AGU, Washington, DC, USA.
Wheeler, G. 1939. Triassic fault-line deflections and associated warping. Journal of Geology, 47, 337-370.
Wheeler, R.L. 2006. Quaternary tectonic faulting in the Eastern United States. Engineering Geology, 82, 165-186.
Whiteside, J.H., Olsen, P.E., Kent, D.V., Fowell, S.J. and Et-Touhami, M. 2007. Synchrony between the CAMP and the Triassic-Jurassic mass-extinction event? Palaeogeography, Palaeoclimatology, and Palaeoecology, 244, 345-367.
Whiteside, J.H., Olsen, P.E., Kent, D.V., Fowell, S.J., and Et-Touhami, M. 2008. Synchrony between the Central Atlantic magmatic province and the Triassic–Jurassic mass-extinction event? Reply to comment of Marzoli et al. Palaeogeography, Palaeoclimatology, Palaeoecology, 262(3-4), 194-198
Wilson, J.T. 1966. Did the Atlantic Close and then Re-Open? Nature, 211, 676-681.
Withjack, M.O., Baum, M.S., and Schlische, R.W. 2010. Influence of preexisting fault fabric on inversion-related deformation: A case study of the inverted Fundy rift basin, southeastern Canada. Tectonics, 29, TC6004, 22 pages, doi:10.1029/2010TC002744.
Withjack, M.O. and Olsen, P.E. 1999. Rift Basins and Passive Margins Tectonics Seminar, Mobil Technology Company, July 14-24, 1999.
Withjack, M.O. and Schlische, R.W. 2005. A review of tectonic events on the passive margin of eastern North America. In: Post, P. (ed) Petroleum Systems of Divergent Continental Margin Basins. 25th Bob S. Perkins Research Conference, Gulf Coast Section of SEPM, 203-235.
Withjack, M.O., Schlische, R.W. and Olsen, P.E. 1998. Diachronous rifting, drifting, and inversion on the passive margin of central eastern North America: An analog for other passive margins. AAPG Bulletin, 82, 817-835.
Withjack, M.O., Schlische, R.W., Malinconico, M.L., and Olsen, P.E. 2011. Rift-basin development—lessons from the Triassic-Jurassic Newark basin of eastern North America. In: Post, P.J., Brown, D.E., and Tari, G.C. (eds) Conjugate Divergent Margins, Geological Society (London) Special Publication, in press.
Withjack, M.O., Schlische, R.W. and Olsen, P.E. 2012. Development of the passive margin of eastern North America: Mesozoic rifting, igneous activity, and breakup. In: Bally, A.W. and Roberts, D.G. (eds) Principles of Phanerozoic Regional Geology, Volume 1. Elsevier, Amsterdam, in press.