2.5.1.1.4.1.1 Late Proterozoic and Paleozoic Plate Tectonic History

Although details about the kinematics, provenance, and histories of lithostratigraphic units within the Appalachian orogenic belt continue to be debated and reclassified (e.g., (Hatcher, 1989) (Horton, 1991) (Glover, 1995b) (Hibbard, 2006)), it is well accepted that plate boundary deformation has occurred repeatedly in the site region since ~~late Precambrian~~middle Proterozoic time. Two complete Wilson cycles, the paired large-scale events of suturing of continents to form supercontinents and rifting to breakup the supercontinents and form ocean basins, occurred twice during this time period (see Fig. 2.5-8). Numerous studies have been published reviewing in detail the individual tectonic events that comprised these two Wilson cycles (e.g., Faill, 1997a; Faill, 1997b, 1998; Hatcher et al., 2007; Thomas, 2006; Whitmeyer and Karlstrom, 2007). The largest-scale events that comprised these Wilson cycles are:

- The Grenville orogeny: The Grenville orogeny marked the beginning of the first Wilson cycle with the suturing of numerous tectonic blocks to Laurentia forming the supercontinent Rodinia. The orogeny occurred over a prolonged period of time extending from approximately 1.3 to 1.0 Ga.
- Rodinia breakup and opening of Iapetus Ocean: This stage of rifting marks the completion of the first Wilson cycle. Extension began as early as approximately 760 to 650 Ma with major rifting occurring around 620 to 550 Ma. The final stages of minor rifting are thought to have been completed by approximately 530 Ma.
- The Appalachian orogeny: The Appalachian orogeny is a broad term used to describe the successive collisional episodes that mark the beginning of the second Wilson cycle and resulted in the formation of Pangea. Three main compressive, orogenic episodes led to the formation of Pangea: Taconic (Ordovician-Silurian), Acadian (Devonian-Mississippian), and Alleghanian (Mississippian-Permian). However, some researchers also explicitly identify the Avalonian (Late Proterozoic-Cambrian), Potomac (pre-Early Ordovician) and Penobscot (Cambrian-Ordovician) orogenies and periods of subduction as key compressional events in the formation of Pangea.
- Pangea breakup and opening of the Atlantic Ocean: The breakup of Pangea during Jurassic time marks the end of the second Wilson cycle.

RAI 130 02.05.01-38

~~Suturing events that mark the welding of continents to form supercontinents and rifting events that mark the breakup of supercontinents to form ocean basins have each occurred twice during this interval. Foreland~~Evidence for most of these compressive tectonic events are preserved in the geologic record based on foreland strata, deformation structures, and metamorphism ~~associated with the Grenville (Middle Proterozoic) and Allegheny (Late Paleozoic) orogenies record the closing of ocean basins and welding of continents to form the supercontinents Rodinia and Pangaea, respectively~~ (Figure 2.5-8). Synrift basins, normal faults, and postrift strata associated with the opening of the Iapetus ~~(Late Proterozoic to Early Cambrian)~~ and Atlantic ~~(Early Mesozoic)~~ Ocean basins record the break-up of the supercontinents. The principal structures that formed during the major events are relevant~~salient~~ to the current seismic hazards in that: (1) they penetrate the seismogenic crust, (2) they subdivide different crustal elements that may have contrasting seismogenic potential, and (3) their associated lithostratigraphic units make up the North American continental crust that underlies most of the site region. Many of the principal structures are inherited faults that have been reactivated repeatedly through time. Some are spatially associated with current zones of concentrated seismic activity and historical large earthquakes. For example, the 1811-1812 New Madrid earthquake sequence ruptured a failed Late Proterozoic rift that also may have been active in the Mesozoic (Ervin, 1975).

RAI 130 02.05.01-38

During the interval between opening of the Iapetus Ocean and opening of the Atlantic Ocean, the eastern margin of the ancestral North America continent was alternately (1) an active rift margin accommodating lithospheric extension with crustal rift basins and synrift strata and volcanism; (2) a passive continental margin accumulating terrestrial and shallow marine facies strata; and (3) an active collisional margin with accretion of microcontinents, island arcs, and eventually the African continent. Major Paleozoic mountain building episodes associated with the collision and accretion events included the Taconic, Acadian, and Allegheny Orogenies. More localized collisional events in the site region include the Avalon, Virgilina and Potomac (Penobscot) orogenies (Hatcher, 1987) (Hatcher, 1989) (Glover, 1995b) (Hibbard, 1995) (Drake, 1999) (Figure 2.5-8). The geologic histories of these orogenies are described in Section 2.5.1.1.2.

Tectonic structures developed during the interval between the Late Proterozoic and Triassic Periods are variable in sense of slip and geometry. Late Proterozoic and early Cambrian rifting associated with the breakup of Rodinia and development of the Iapetus Ocean formed east-dipping normal faults through Laurentian (proto-North American) crust (Figure 2.5-16 and Figure 2.5-17). Late Proterozoic extended crust of the Iapetan margin probably underlies the Appalachian fold belt southeastward to beneath much of the Piedmont Province (Wheeler, 1996). Paleozoic compressional events associated with the Taconic, Acadian, and Allegheny orogenies formed predominantly west-vergent structures that include (1) Valley and Ridge Province shallow folding and thrusting within predominantly passive margin strata, (2) Blue Ridge Province nappes of Laurentian crust overlain by Iapetan continental margin deposits, (3) Piedmont Province thrust-bounded exotic and suspect terranes including island arc and accretionary complexes interpreted to originate in the Iapetan Ocean, and (4) Piedmont Province and sub-Coastal Plain Province east-dipping thrust, oblique, and reverse fault zones that collectively are interpreted to penetrate much of the crust and represent major sutures that juxtapose crustal elements (Hatcher, 1987) (Horton, 1991) (Glover, 1995b) (Hibbard, 2006) (Figure 2.5-16 and Figure 2.5-17). Many investigators recognize significant transpressional components to major faults bounding lithostratigraphic units (Hatcher, 1987) (Glover, 1995b) (Hibbard, 2006) (Figure 2.5-8 and Figure 2.5-16).

2.5.1.1.4.1.2 Mesozoic and Cenozoic Passive Margin Evolution

RAI 130 02.05.01-54

At the time of the EPRI (1986) study much was published about the structure and crustal elements of the Mesozoic to Cenozoic Atlantic passive margin (e.g., (Klitgord, 1979)). However, it was not until the Geological Society of America's Decade of North American Geology (DNAG) volume on the U.S. Atlantic continental margin (Sheridan, 1988), seminal papers within it (e.g., (Klitgord, 1988)), and later summary publications (e.g., (Klitgord, 1995) (Withjack, 1998) (Schlische, 2003) (Withjack, 2005)) that the current understanding of the margin structure and tectonic history was formulated comprehensively.

RAI 130 02.05.01-54

The current Atlantic passive continental margin has evolved since rifting initiated in the Early Triassic. The progression from active continental rifting to sea-floor spreading and a passive continental margin included: (1) initial rifting and hot-spot plume development, (2) thinning of warm, buoyant crust with northwest-southeast extension, normal faulting and deposition of synrift sedimentary and volcanic rocks, and (3) cooling and subsidence of thinned crust and deposition of postrift sediments on the coastal plain and continental shelf, slope, and rise (Klitgord, 1988) (Klitgord, 1995). The transition between the second (rifting) and third (drifting) phases during the Late Triassic and Early Jurassic marked the initiation of a passive margin setting in the site region, in which active spreading migrated east away from the margin (Withjack, 1998) (Withjack 2005). As the thinned crust of the continental margin cooled and migrated away from the warm, buoyant crust at the mid-Atlantic spreading center, horizontal northwest-southeast tension changed to horizontal compression as gravitational potential energy from the spreading ridge exerted a lateral "ridge push" force on the oceanic crust. Northwest-southeast-directed postrift shortening, manifested in Mesozoic basin inversion RAI 130 02.05.01-54 structures, provides the clearest indication of this change in stress regime (Withjack, 1998) (Schlische, 2003). The present-day direction of maximum horizontal compression-east-northeast to west-southwest-is rotated from this hypothesized initial postrift direction.

The crustal structure of the passive continental margin includes areas of continental crust, (Iapetan-extended crust (Wheeler, 1996)), rifted continental crust, rift-stage (transitional) crust, marginal oceanic crust, and oceanic crust (Klitgord, 1995) (Figure 2.5-18 and Figure 2.5-19). Rifted continental crust is crust that has been extended, faulted, and thinned slightly. In the site region, rifted-continental crust extends from the western border faults of the exposed synrift Danville, Scottsville, Culpeper, Gettysburg, and Newark basins to the basement hinge zone, approximately coincident with the seaward edge of the continental shelf (Klitgord, 1995) (Figure 2.5-12 and Figure 2.5-19). Rifted crust also includes exposed and buried Upper Triassic to Lower Jurassic basins within the eastern Piedmont and Coastal Plain Provinces, including the RAI 130 02.05.01-54 Richmond, Taylorsville, and Norfolk basins (LeTourneau, 2003) (Schlische, 2003) (Figure 2.5-10). Several additional basins with poorly defined extent also underlie the Coastal Plain and Continental Shelf and are shown directly east and northeast of the site (Figure 2.5-10). Buried synrift basins are delineated based on sparse drillhole data, magnetic and gravity anomalies, and seismic reflection data (e.g., (Benson, 1992)). Figure 2.5-19 shows east-dipping basin-bounding faults that penetrate the seismogenic crust and have listric geometries at RAI 71 02.05.01-6 depth. Many of the synrift normal faults within the site region are interpreted as Paleozoic thrust faults reactivated during Mesozoic rifting. The Mesozoic basins are discussed further in Section 2.5.1.1.4.4.3 as well as the hypothesized Queen Anne basin shown as lying beneath the RAI 130 02.05.01-54 site (Figure 2.5-10) as one alternate interpretation of basement lithology.

Rift-stage (transitional) crust is extended continental crust intruded by mafic magmatic material during rifting. In the site region, this crustal type coincides with the basement hinge zone and postrift Baltimore Canyon Trough (Klitgord, 1995) (Figure 2.5-12). The basement hinge zone is defined where pre-Late Jurassic basement abruptly deepens seaward from about 1 to 2.5 mi (1.6 to 4 km) to more than 5 mi (8 km). Overlying this lower crustal unit seaward of the basement hinge zone is the Jurassic volcanic wedge, representing a period of excess volcanism and is greater than 65 mi (105 km) wide and 1 to 5 mi (1.5 to 8 km) thick. The wedge is identified on seismic reflection lines as a prominent sequence of seaward-dipping reflectors. The East Coast magnetic anomaly (ECMA) coincides with the seaward edge of the wedge (Figure 2.5-18) (Section 2.5.1.1.4.3.2).

The last transitional crustal unit between continental and oceanic crust is marginal oceanic crust (Klitgord, 1995) (Figure 2.5-18). Marginal oceanic crust is located east of the ECMA where the Jurassic volcanic wedge merges with the landward edge of oceanic crust. Here, the transition from rifting to sea-floor spreading created a thicker than normal oceanic crust with possible magmatic underplating.

A postrift unconformity separates synrift from postrift deposits and represents the change in tectonic regime in the Middle Jurassic from continental rifting to the establishment of the passive margin ("drifting"). Sedimentary rocks below the unconformity are cut by numerous faults. In contrast, the rocks and strata above the unconformity accumulated within the environment of a broadly subsiding passive margin and are sparsely faulted. Sediments shed from the faulted blocks of the rifting phase and from the core of the Allegheny orogen accumulated on the coastal plain, continental shelf, slope, and rise above the postrift unconformity and contributed to subsidence of the cooling postrift crust by tectonic loading.

Postrift deformation is recorded in synrift basins and within postrift strata as normal faults seaward of the basement hinge zone and as contractional features landward of the basement hinge zone. Extensive normal faulting penetrates the postrift strata (and upper strata of the volcanic wedge) of the marginal basin overlying the volcanic wedge (Figure 2.5-18 and Figure 2.5-19). This set of faults is thought to have been caused by sediment loading on the outer edge of the margin due to differential compaction of the slope-rise deposits relative to adjacent carbonate platform deposits (Poag, 1991) (Klitgord, 1995). These faults are interpreted as margin-parallel structures that bound large mega-slump blocks and are not considered active tectonic features (Poag, 1991).

RAI 130 02.05.01-54

RAI 130 02.05.01-54

RAI 130 02.05.01-54

The~~Schlische (2003) summarizes~~ evidence for postrift shortening and positive basin inversion (defined as extension within basins followed by contraction) is well documented in several Atlantic margin basins, including the Newark, Taylorsville, and Richmond basins in the site region (LeTourneau, 2003) (Schlische, 2003) (Withjack, 2005) (Figure 2.5-10). Contractional postrift deformation is interpreted to record the change in stress regime from horizontal maximum extension during rifting to horizontal maximum compression during passive margin drifting. The hypothesis that the change in stress regime following rifting was recorded in reverse and strike slip faulting and folding was known prior to the 1986 EPRI study (e.g., (Sanders, 1963) (Swanson, 1982) (Wentworth, 1983)), but significant advances in the documentation and characterization of the rift to drift transition and postrift deformation has occurred since the mid-1980s (Withjack, 1998) (Schlische, 2003). Based on structural analysis and age control of widespread approximately 200 Ma basaltic dikes and faulting, much of the site region was under a state of northwest-southeast maximum compression by Late Triassic and ~~e~~Early~~iest~~ Jurassic time (Withjack, 1998) (Schlische, 2003a) (Withjack, 2005). This deformation regime may have persisted locally into the Cenozoic based on the recognized early Cenozoic contractional growth faulting associated with the northeast-striking Brandywine fault system (Jacobeen, 1972) (Wilson, 1990), Port Royal fault zone (Mixon, 1984) (Mixon, 2000) and Skinkers Neck anticline (Mixon, 1984) (Mixon, 2000) (Section 2.5.1.1.4.4.4). The present-day stress field of east-northeast to west-southwest maximum horizontal compression (Zoback, 1989a) is rotated from the hypothesized Jurassic and Cretaceous northwest-southeast orientation. The east-northeast to west-southwest maximum horizontal stress direction is consistent with resolved dextral transpressive slip locally documented on the northeast-striking Stafford fault system (Mixon, 2000), a recognized Tertiary tectonic feature (Section 2.5.1.1.4.4.4.1).

2.5.1.1.4.1.3 Cenozoic Passive Margin Flexural Tectonics

Tectonic processes along the Atlantic passive continental margin in the Cenozoic Era include vertical tectonics associated with lithospheric flexure. Vertical tectonics are dominated by: (1) cooling of the extended continental, transitional, and oceanic crust as the spreading center migrates eastward, and (2) the transfer of mass from the Appalachian core to the Coastal Plain and Continental Shelf, Slope, and Rise via erosion. Erosion and exhumation of the Allegheny crustal root of the Piedmont, Blue Ridge, Valley and Ridge, and Appalachian Plateau Provinces has been balanced by deposition on and loading of the Coastal Plain and offshore provinces by fluvial, fluvial-deltaic, and marine sediment transport. Margin-parallel variations in the amount of uplift and subsidence have created arches (e.g. South New Jersey and Norfolk Arches) and basins or embayments (e.g. Salisbury Embayment) along the Coastal Plain and Continental Shelf (Figure 2.5-12).

Flexural zones show both passive-margin-normal and passive-margin-parallel trends. Flexure normal to the passive margin is clearly recorded in the basement hinge zone (Figure 2.5-19). The vertical relief across the offshore basement hinge zone accounts for a change in postrift sediment thickness from 1 to 2.5 mi (1.6 to 4 km) to over 5 mi (8 km) and indicates lateral changes in tectonic loading (Klitgord, 1995). It has been proposed that the downwarping of the margin in the vicinity of the main depocenter of the Baltimore Canyon Trough led to the flexural uplift of the Coastal Plain units to the west (Watts, 1982). However, more recent studies show that sea-level variations since the Cretaceous are compatible with the present elevations of exposed Coastal Plain strata and thus do not support flexural uplift of the Coastal Plain (e.g., (Pazzaglia, 1993)).

A simple elastic model of Cenozoic flexural deformation across the Atlantic passive margin has been used to approximate the response of rifted continental crust to surface erosion of the Piedmont and deposition on the Coastal Plain and Continental Shelf (Pazzaglia, 1994) (Figure 2.5-12 and Figure 2.5-19). The boundary between areas of net Cenozoic erosion and deposition, the Fall Line, marks the flexural hinge between uplift and downwarping. Geologic correlation and longitudinal profiles of Miocene to Quaternary river terraces on the Piedmont with deltaic and marine equivalent strata on the Coastal Plain provide data for model validation (Pazzaglia, 1993). A one-dimensional elastic plate model replicates the form of the profiles and maintenance of the Fall Line with flexure driven by exhumation of the Piedmont and adjacent Appalachian provinces coupled with sediment loading in the Salisbury Embayment and Baltimore Canyon Trough (Pazzaglia, 1994). Model results suggest a long-term denudation rate of approximately 33 ft (10 m) per million years and about 115 to 426 ft (35 to 130 m) of upwarping of the Piedmont in the last 15 million years.

The flexural hinge zones (Fall Line and basement hinge zone) do not appear to be seismogenic. The spatial association between the Fall Line and observed Cenozoic faults such as the Stafford and Brandywine fault systems is commonly attributed to the fact that those faults are recognizable where Cenozoic cover is thin and there is greater exposure of bedrock compared to areas farther east toward the coast (e.g., (Wentworth, 1983)). It is suggested (Pazzaglia, 1994) that low rates of contractional deformation on or near the hinge zone documented on Cenozoic faults may be a second-order response to vertical flexure and horizontal compressive stresses. Neither the Fall Line nor basement hinge zone was considered a potential tectonic feature by EPRI (1986). They were considered zones where ground amplification could be affected. It is also suggested (Weems, 1998) that multiple fall lines (i.e., alignments of anomalously steep river gradients) located near or within the Fall Line may be of neo-tectonic origin. Subsequent studies performed during the North Anna ESP study demonstrates that the fall lines (Weems, 1998) are erosional features and not capable tectonic sources (NRC, 2005) (Section 2.5.1.1.4.4.5.1) Post-EPRI seismicity also shows no spatial patterns suggestive of seismicity aligned with either the basement hinge zone or Fall Line. Crone and Wheeler (Crone, 2000) and Wheeler (Wheeler, 2005) (Wheeler, 2006) also do not list these as potentially Quaternary active features. Accordingly, it is concluded that these features are not capable tectonic sources. Post-EPRI seismicity also shows no spatial patterns suggestive of seismicity aligned with either the basement hinge zone or Fall Line (Section 2.5.2).

RAI 71
02.05.01-6

Along-strike variations in the amount of epeirogenic movement along the Atlantic continental margin has resulted in a series of arches and embayments identified based on variations in thickness of Coastal Plain strata from Late Cretaceous through Pleistocene time. The Salisbury Embayment is a prominent, broad depocenter in the site region, and coincides with Chesapeake Bay and Delaware Bay (Figure 2.5-12). In general, it appears that downwarping associated with the Salisbury Embayment (Figure 2.5-12) began early in the Cretaceous and continued intermittently throughout the Cretaceous and Tertiary periods. Deposition apparently kept pace, resulting in a fluvial-deltaic environment. Biostratigraphic data from test wells on the west side of Chesapeake Bay indicate that Upper Cretaceous sediments reach maximum thickness in Anne Arundel County and show progressive thinning to the south. This appears to reflect deposition within the downwarping, northwest-trending Salisbury

RAI 71 02.05.01-6

Embayment during the Cretaceous (Hansen, 1978). At the margins of the Salisbury Embayment are the South New Jersey Arch to the northeast and the Norfolk Arch to the south. Both arches are broad anticlinal warps reflected in the top of basement and overlying sediments. Thinning and overlapping within the Upper Cretaceous interval suggests that the northern flank of the Norfolk Arch was tectonically active during late Cretaceous time (Hansen, 1978) (Figure 2.5-12). The processes that form and maintain the arches and embayments are poorly understood, and there has been little advancement in the thinking about these features since publication of the EPRI study report (EPRI, 1986). Poag (2004), however, uses new basement data obtained from seismic reflection profiles and exploratory boreholes in the region of the main Chesapeake Bay impact crater to show that the Norfolk Arch is not as well expressed as originally interpreted by earlier authors (Brown, 1972) using limited data. Previous elevation differences cited as evidence for the basement arch appear to be due to subsidence differential between the impact crater and the adjacent deposits (Poag, 2004) (Section 2.5.1.1.4.4.4). Regardless, no published hypothesis was found suggesting causality between epeirogenic processes maintaining these specific arches and the embayment and potentially seismogenic structures, and there is no spatial association of seismicity with the basement arches. Thus, it is concluded that these features are not capable tectonic sources.

2.5.1.1.4.2 Tectonic Stress in the Mid-Continent Region

Expert teams that participated in the 1986 EPRI evaluation of intra-plate stress generally concluded that tectonic stress in the CEUS region is characterized by northeast-southwest-directed horizontal compression. In general, the expert teams concluded that the most likely source of tectonic stress in the mid-continent region was ridge-push force associated with the Mid-Atlantic ridge, transmitted to the interior of the North American plate by the elastic strength of the lithosphere. Other potential forces acting on the North American plate were judged to be less significant in contributing to the magnitude and orientation of the maximum compressive principal stress. Some of the expert teams noted local~~that~~ deviations from ~~the~~this regional trend.~~northeast-southwest trend of principal stress may be present along the east coast of North America and in the New Madrid region.~~ They assessed the quality of stress indicator data and discussed various hypotheses to account for their~~what were~~ interpreted ~~as~~ variations in the regional stress trajectories.

RAI 71 02.05.01-9

Since 1986, an international effort to collate and evaluate stress indicator data has resulted in publication of a new world stress map (Zoback, 1989a) (Zoback, 1989b). Data for this map are ranked in terms of quality, and plate-scale trends in the orientations of principal stresses are assessed qualitatively based on analysis of high-quality data (Zoback, 1992). Subsequent statistical analyses of stress indicators confirmed that the trajectory of the maximum compressive principal stress is uniform across broad continental regions at a high level of statistical confidence. In particular, the northeast-southwest orientation of principal stress in the CEUS inferred by the EPRI experts is statistically robust, and is consistent with the theoretical trend of compressive forces acting on the North American plate from the mid-Atlantic ridge (Coblentz and Richardson, 1995). However, local variations in the regional stress field similar to those recognized by the EPRI teams are also present in the more recent datasets (e.g., Kim, 2005; Reinecker, 2008)

RAI 71 02.05.01-9

~~More recent assessments of lithospheric stress do not support inferences by some EPRI expert teams that the orientation of the principal stress may be locally perturbed in the New England area, along the east coast of the United States, or in the New Madrid region. A variety of data was summarized (Zoback, 1989a), including well-bore breakouts, results of hydraulic fracturing studies, and newly calculated focal mechanisms, which indicate that the New England and eastern seaboard regions of the U.S. are characterized by horizontal northeast-southwest to east-west compression. Similar trends are present in the expanded set of stress indicators for~~

RAI 71 02.05.01-9

~~the New Madrid region. Zoback and Zoback (Zoback, 1989a) grouped all of these regions, along with a large area of eastern Canada, with the CEUS in an expanded "Mid-Plate" stress province characterized by northeast-southwest directed horizontal compression.~~

In addition to better documenting the orientation of stress, research conducted since 1986 has addressed quantitatively the relative contributions of various forces that may be acting on the North American plate to the total stress within the plate. Richardson and Reding (Richardson, 1991) performed numerical modeling of stress in the continental U.S. interior, and considered the contribution to total tectonic stress to be from three classes of forces:

- Horizontal stresses that arise from gravitational body forces acting on lateral variations in lithospheric density. These forces commonly are called buoyancy forces. Richardson and Reding emphasize that what is commonly called ridge-push force is an example of this class of force. Rather than a line-force that acts outwardly from the axis of a spreading ridge, ridge-push arises from the pressure exerted by positively buoyant, young oceanic lithosphere near the ridge against older, cooler, denser, less buoyant lithosphere in the deeper ocean basins (Turcotte, 2002). The force is an integrated effect over oceanic lithosphere ranging in age from about 0 to 100 million years (Dahlen, 1981). The ridge-push force is transmitted as stress to the interior of continents by the elastic strength of the lithosphere.
- Shear and compressive stresses transmitted across major plate boundaries (strike-slip faults and subduction zones).
- Shear tractions acting on the base of the lithosphere from relative flow of the underlying asthenospheric mantle.

Richardson and Reding (Richardson, 1991) concluded that the observed northeast-southwest trend of principal stress in the CEUS dominantly reflects ridge-push body forces. They estimated the magnitude of these forces to be about 2 to 3 X 1012 N/m (i.e., the total vertically integrated force acting on a column of lithosphere 1 m wide), which corresponds to average equivalent stresses of about 40 to 60 MPa distributed across a 30 mi (50 km) thick elastic plate. The fit of the model stress trajectories to data was improved by the addition of compressive stress (about 5 to 10 MPa) acting on the San Andreas Fault and Caribbean plate boundary structures. The fit of the modeled stresses to the data further suggested that shear stresses acting on these plate boundary structures is in the range of 5 to 10 MPa.

Richardson and Reding (Richardson, 1991) noted that the general northeast-southwest orientation of principal stress in the CEUS also could be reproduced in numerical models that assume a shear stress, or traction, acting on the base of the North American plate. Richardson and Reding (Richardson, 1991) and Zoback and Zoback (Zoback, 1989) do not favor this as a significant contributor to total stress in the mid-continent region. A basal traction predicts or requires that the horizontal compressive stress in the lithosphere increases by an order of magnitude moving east to west, from the eastern seaboard to the Great Plains. Zoback and Zoback (Zoback, 1989) noted that the state of stress in the southern Great Plains is characterized by north-northeast to south-southwest extension, which is contrary to this prediction. They further observed that the level of background seismic activity is generally higher in the eastern United States than in the Great Plains, which is not consistent with the prediction of the basal traction model that compressive stresses (and presumably rates of seismic activity) should be higher in the middle parts of the continent than along the eastern margin.

To summarize, analyses of regional tectonic stress in the CEUS since EPRI (1986) have not significantly altered the characterization of the northeast-southwest orientation of the maximum compressive principal stress. The orientation of a planar tectonic structure relative to the principal stress direction determines the magnitude of shear stress resolved onto the structure. Given that the current interpretation of the orientation of principal stress is similar to that adopted in EPRI (1986), a new evaluation of the seismic potential of tectonic features based on a favorable or unfavorable orientation to the stress field would yield similar results. Thus, there is no significant change in the understanding of the static stress in the CEUS since the publication of the EPRI source models in 1986, and there are no significant implications for existing characterizations of potential activity of tectonic structures.

2.5.1.1.4.3 Gravity and Magnetic Data and Features of the Site Region and Site Vicinity

Gravity and magnetic anomaly datasets of the site region have been published following the 1986 EPRI study. Significant datasets include regional maps of the gravity and magnetic fields in North America by the Geological Society of America (GSA), as part of the Society's DNAG project (Tanner, 1987) (Hinze, 1987). The DNAG datasets are widely available in digital form via the internet (Hittelman, 1994). A magnetic anomaly map of North America was published in 2002 that featured improved reprocessing of existing data and compilation of a new and more complete database (Bankey, 2002) (Figure 2.5-20).

These maps present the potential field data at 1:5,000,000-scale, and thus are useful for identifying and assessing gravity and magnetic anomalies with wavelengths on the order of tens of kilometers or greater (Bankey, 2000) (Hittelman, 1994). Regional gravity anomaly maps are based on Bouguer gravity anomalies onshore and free-air gravity anomalies offshore. The primary sources of magnetic data reviewed for this CCNPP Unit 3 study are from aeromagnetic surveys onshore and offshore (Bankey, 2002), and the DNAG datasets available digitally from the internet (Hittelman, 1994).

Most of the contributed gravity and magnetic data that went into the regional compilations were collected prior to the 1986 EPRI study; thus, most of the basic data were available for interpretation at local and regional scales. Large-scale compilations (1:2,500,000-scale) of the free-air anomalies offshore and Bouguer anomalies onshore were published in 1982 by the Society of Exploration Geophysicists (Lyons, 1982) (Sheridan, 1988). The DNAG magnetic anomaly maps were based on a prior analog map of magnetic anomalies of the U.S. published in the early 1980's (Zietz, 1982) (Behrendt, 1983) (Sheridan, 1988).

In addition, the DNAG Continent-Ocean transect program published a synthesis of gravity and magnetic data with seismic and geologic data (Klitgord, 1995). Transect E-3, which crosses the site region, is presented in Figure 2.5-16 and Figure 2.5-17. Much of the seismic and geophysical data through the Piedmont region was reanalyzed from a geophysical survey conducted along Interstate I-64 in Virginia that was published prior to release of the 1986 EPRI study (e.g., (Harris, 1982)).

In summary, the gravity and magnetic data published since 1986 do not reveal any new anomalies related to geologic structures that were not identified prior to the 1986 EPRI study. Rather, post-EPRI publications have refined the characteristics and tectonic interpretation of the anomalies. Discussion of the gravity and magnetic anomalies is presented in the following sections.

2.5.1.1.4.3.1 Gravity Data and Features

Gravity data compiled at 1:5,000,000-scale for the DNAG project provide documentation of previous observations that the gravity field in the site region is characterized by a long-wavelength, east-to-west gradient in the Bouguer gravity anomaly over the continental margin (Harris, 1982) (Hittelman, 1994) (Figure 2.5-21). The free-air gravity anomaly shows broad gravity lows over offshore oceanic crust near the continental margin and over the broad marginal embayments. Offshore marginal platforms are marked by shorter-wavelength, higher-amplitude gravity highs and lows. The present shelf edge is marked by a prominent free-air gravity anomaly that also corresponds to the continent-ocean boundary (Sheridan, 1988) (Klitgord, 1995).

Bouguer gravity values increase eastward from about -80 milligals (mgal) in the Valley and Ridge Province of western Virginia to about +10 mgal in the Coastal Plain Province, corresponding to an approximately 90 mgal regional anomaly across the Appalachian Orogen (Figure 2.5-17 and Figure 2.5-21). This regional gradient is called the "Piedmont gravity gradient" (Harris, 1982), and is interpreted to reflect the eastward thinning of the North American continental crust and the associated positive relief on the Moho discontinuity with proximity to the Atlantic margin.

The Piedmont gravity gradient is punctuated by several smaller positive anomalies with wavelengths ranging from about 15 to 50 mi (25 to 80 km) , and amplitudes of about 10 to 20 mgal. Most of these anomalies are associated with accreted Taconic terranes such as the Carolina/Chopawamsic terrane (Figure 2.5-17). Collectively, they form a gravity high superimposed on the regional Piedmont gradient that can be traced northeast-southwest on the 1:5,000,000-scale DNAG map relatively continuously along the trend of the Appalachian orogenic belt through North Carolina, Virginia, and Maryland (Figure 2.5-21). The continuity of this positive anomaly diminishes to the southwest in South Carolina, and the trend of the anomaly is deflected eastward in Maryland, Pennsylvania, and Delaware.

The short-wavelength anomalies and possible associations with upper crustal structure are illustrated by combining gravity profiles with seismic reflection data and geologic data (Harris, 1982) (Glover, 1995b). In some cases, short-wavelength positive anomalies are associated with antiformal culminations in Appalachian thrust sheets. For example, there is a positive anomaly associated with an anticline at the western edge of the Blue Ridge nappe along the Interstate I-64 transect (Harris, 1982) (Figure 2.5-17). The anomaly is presumably due to the presence of denser rocks transported from depth and thickened by antiformal folding in the hanging wall of the thrust.

The Salisbury geophysical anomaly (SGA) is a paired Bouguer gravity anomaly and magnetic high that is located along the west side of the Salisbury Embayment (Klitgord, 1995) (Figure 2.5-17, Figure 2.5-18, Figure 2.5-20, and Figure 2.5-21). The SGA is located about 10 mi (16 km) west of the CCNPP site (Figure 2.5-22). The anomaly is expressed most clearly as a magnetic lineation that separates a zone of short-wavelength, high-amplitude magnetic lineations to the west from a zone of low-amplitude, long-wavelength anomalies to the east. The gravity data show the SGA to form the western margin of a broad gravity low that extends seaward to the basement hinge zone. The anomaly takes the form of a north-northeast-trending gravity high having about 30 mgal relief (Johnson, 1973). The anomaly has also been named the Sussex-Curioman Bay trend (Levan, 1963) or the Sussex-Leonardtown anomaly (Daniels, 1985), and is believed to reflect an east-dipping mafic rock body associated with a suture zone buried beneath coastal plain sediments (Figure 2.5-17). The SGA is interpreted (Klitgord, 1995) to mark the likely location of the Taconic suture that separates the Goochland terrane on the west from a zone of island arc and oceanic metavolcanics formed in the Iapetus Ocean on the east. The SGA is shown (Horton, 1991) to be associated with the buried Sussex terrane is a probable mafic mélange that was interpreted by Lefort and Max (Lefort, 1989) to mark the Alleghenian "Chesapeake Bay suture" (Figure 2.5-16).

The offshore portions of the site region contain a prominent, long-wavelength free-air gravity anomaly associated with the transition from continental to oceanic crust (Sheridan, 1988) (Klitgord, 1995) (Figure 2.5-19). This anomaly is large (75 to 150 mgal peak to trough) and is 45 to 80 mi (72 to 129 km) wide. Variations in the amplitude and shape of the anomaly along the Atlantic margin are due to seafloor relief, horizontal density variations in the crust, and relief on the crust-mantle boundary (Sheridan, 1988) (Klitgord, 1995).

In summary, gravity data published since the mid-1980s confirm and provide additional documentation of previous observations of a gradual "piedmont gravity gradient" across the Blue Ridge and Piedmont Provinces of Virginia and a prominent gravity anomaly at the seaward margin of the continental shelf. Shorter-wavelength anomalies such as the SGA also are recognized in the data. All anomalies were known at the time of the 1986 EPRI study. The "piedmont gravity gradient" is interpreted to reflect eastward thinning of the North American crust and lithosphere. The free-air anomaly at the outer shelf edge is interpreted as reflecting the transition between continental and oceanic crust. Second-order features in the regional field, such as the Salisbury geophysical anomaly and the short discontinuous northeast-trending anomaly east of the site, primarily reflect density variations in the upper crust associated with the boundaries and geometries of Appalachian thrust sheets and accreted terranes.

2.5.1.1.4.3.2 Magnetic Data and Features

Magnetic data compiled for the 2002 Magnetic Anomaly Map of North America reveal numerous northeast-southwest-trending magnetic anomalies, generally parallel to the structural features of the Appalachian orogenic belt (Bankey, 2002) (Figure 2.5-20). Unlike the gravity field, the magnetic field is not characterized by a regional, long-wavelength gradient that spans the east-west extent of the site region. A magnetic profile along Interstate-64 published to accompany a seismic reflection profile (Harris, 1982) shows anomalies with wavelengths of about 6 to 30 mi (10 to 48 km). It has been concluded (Harris, 1982) that anomalies in the magnetic field primarily are associated with upper-crustal variations in magnetic susceptibility and, unlike the gravity data, do not provide information on crustal-scale features in the lithosphere.

Prominent north- to northeast-trending magnetic anomalies in the CCNPP site region include the interior New York-Alabama, Ocoee, and Clingman lineaments, the Coastal Plain Salisbury geophysical anomaly and near shore Brunswick magnetic anomaly, and the offshore East Coast magnetic anomaly (King, 1978) (Klitgord, 1988) (Klitgord, 1995) (Bankey, 2002) (Figure 2.5-20). The offshore Blake Spur magnetic anomaly is outside the site region.

King and Zietz (1978) identified a 1,000 mi (1,600 km) long lineament in aeromagnetic maps of the eastern U.S. that they referred to as the "New York-Alabama lineament" (NYAL) (Figure 2.5-20). The NYAL primarily is defined by a series of northeast-southwest-trending linear magnetic anomalies in the Valley and Ridge province of the Appalachian fold belt that systematically intersect and truncate other magnetic anomalies. The NYAL is located about 160 mi (257 km) northwest of the CCNPP site.

The Clingman lineament is an approximately 750 mi (1,200 km) long, northeast-trending aeromagnetic lineament that passes through parts of the Blue Ridge and eastern Valley and Ridge provinces from Alabama to Pennsylvania (Nelson, 1981). The Ocoee lineament splays southwest from the Clingman lineament at about latitude 36°N (Johnston, 1985a). The Clingman-Ocoee lineaments are sub-parallel to and located about 30 to 60 mi (48 to 97 km) east of the NYAL. These lineaments are located about 60 mi northwest of the CCNPP site.

King and Zietz (King, 1978) interpreted the NYAL to be a major strike-slip fault in the Precambrian basement beneath the thin-skinned fold-and-thrust structures of the Valley and Ridge province, and suggested that it may separate rocks on the northwest that acted as a mechanical buttress from the intensely deformed Appalachian fold belt to the southeast. Shumaker (Shumaker, 2000) interpreted the NYAL to be a right-lateral strike-slip fault that formed during an initial phase of Late Proterozoic continental rifting that eventually led to the opening of the Iapetus Ocean.

The Clingman lineament also is interpreted to arise from a source or sources in the Precambrian basement beneath the accreted and transported Appalachian terranes (Nelson, 1981). Johnston (Johnston, 1985a) observed that the "preponderance of southern Appalachian seismicity" occurs within the "Ocoee block", a Precambrian basement block bounded by the NYAL and Clingman-Ocoee lineaments (the Ocoee block was previously defined by (Johnston, 1985b)). Based on the orientations of nodal planes from focal mechanisms of small earthquakes, it was noted (Johnston, 1985) that most events within the Ocoee block occurred by strike-slip displacement on north-south and east-west striking faults, Johnston (Johnston, 1985a) did not favor the interpretation of seismicity occurring on a single, through-going northeast-southwest-trending structure parallel to the Ocoee block boundaries.

The Ocoee block lies within a zone defined by Wheeler (Wheeler, 1995) (Wheeler, 1996) as extended continental crust of the Late Proterozoic to Cambrian Iapetan terrane. Synthesizing geologic and geophysical data, Wheeler (Wheeler, 1995) mapped the northwest extent of the Iapetan normal faults in the subsurface below the Appalachian detachment, and proposed that earthquakes within the region defined by Johnston and Reinbold (Johnston, 1985b) as the Ocoee block may be the result of reactivation of Iapetan normal faults as reverse or strike-slip faults in the modern tectonic setting.

The East Coast magnetic anomaly (ECMA) is a prominent, linear, segmented magnetic high that extends the length of the Atlantic continental margin from the Carolinas to New England (Figure 2.5-20). The anomaly is about 65 mi (105 mi) wide and has an amplitude of about 500 nT. This anomaly approximately coincides with the seaward edge of the continental shelf, and has been considered to mark the transition from continental to oceanic crust. Klitgord et al. (1995) note that the anomaly is situated above the seaward edge of the thick Jurassic volcanic wedge and lower crustal zone of magmatic under plating along the boundary between rift-stage and marginal oceanic crust (Figure 2.5-18 and Figure 2.5-19). The ECMA is not directly associated with a ~~fault or~~capable tectonic feature, and thus is not considered as a potential seismic source.

RAI 71 02.05.01-10

The Brunswick magnetic anomaly (BMA) is located along the basement hinge zone offshore of the Carolinas, at the southern portion of the site region about 200 mi (322 km) from the CCNPP site (Figure 2.5-20). The lineament is narrower and has less amplitude than the ECMA (Klitgord, 1995). The BMA may continue northward along the hinge zone of the Baltimore Canyon Trough, but the magnetic field there is much lower in amplitude and the lineament is diffuse. The BMA is not directly related to a fault or other tectonic structure, and thus is not a potential seismic source.

The Blake Spur magnetic anomaly (BSMA) is located east of the site region above oceanic crust, about 290 mi (465 km) from the CCNPP site (Figure 2.5-20). The BSMA is a low-amplitude magnetic anomaly that lies subparallel to the East Coast magnetic anomaly (Klitgord et al., 1995). The BSMA probably formed during the Middle Jurassic as the midocean ridge spreading center shifted to the east. The BSMA coincides with a fault-bounded, west-side-down scarp in oceanic basement. Since its formation, the BSMA has been a passive feature in the Atlantic crust, and thus is not a potential seismic source.

The Salisbury geophysical anomaly (SGA), as mentioned above, is a paired Bouguer gravity and magnetic anomaly along the west side of the Salisbury embayment that is located about 10 mi (16 km) of the CCNPP site (Figure 2.5-22). The anomaly is expressed in the magnetic data as a lineament separating short-wavelength, high-amplitude magnetic lineations to the west from a zone of low-amplitude, long-wavelength anomalies to the east. The contrast in magnetic signature is related to the juxtaposition of terranes of contrasting affinity beneath coastal plain sediments, and in particular the mafic to ultramafic rocks and mélange termed the Sussex terrane by Horton et al. (1991) and believed to represent alternatively a Taconic (Glover, 1995b) or Alleghenian (Lefort, 1989) suture (Figure 2.5-16). Lower intensities to the west are associated with the Goochland terrane, which represents continental basement (Figure 2.5-17).

Discrete magnetic lows associated with the Richmond and Culpeper basins are discernible on the 2002 North America magnetic anomaly map (Bankey, 2002) (Figure 2.5-22). Basaltic and diabase dikes and sills are a component of the synrift fill of the exposed basins in the Piedmont and of the Taylorsville basin (Schlische, 2003) (Klitgord, 1995). The distinctive, elongate magnetic anomalies associated with these igneous bodies within the synrift basins of the Piedmont are also used beneath the Coastal Plain to delineate the Taylorsville, Queen Anne, and other synrift basins (e.g., (Benson, 1992)). The elongate magnetic anomalies are less prevalent in the magnetic field east of the Salisbury geophysical anomaly. Either the eastern rift basins do not contain as much volcanic material as the western set of rift basins or the depth to this volcanic material is considerably greater (Klitgord, 1995). Small, circular magnetic highs across the coastal plain have been interpreted as intrusive bodies (Horton, 1991) (Klitgord, 1995).

Approximately 5 to 7 mi (8 to 11 km) east of the CCNPP site is an unnamed short, discontinuous weak to moderate northeast-trending magnetic anomaly that aligns subparallel to the SGA (Figure 2.5-22). Similar features to the south have been interpreted as granitic intrusive anomalies, whereas Benson (1992) interprets the feature as being bound by a Mesozoic basin (Figure 2.5-10). A deep borehole (SM-DF-84, Figure 2.5-11) drilled near the southern margin of this feature encountered Jurassic (?) volcanic rocks (dated at 169 ± 8 million years old) related to Mesozoic rifting, or perhaps basic metavolcanic rocks accreted to North America as part of the Brunswick Terrane (Hansen, 1986).

RAI 71 02.05.01-12

A magnetic profile along an approximately west-northwest to east-southeast transect through central Pennsylvania (Glover, 1995b) (Figure 2.5-17) indicates that paired high and low magnetic anomalies are associated with the ~~western~~ margins of crustal units.~~, truncated by thrust faults.~~ Many of these anomalies have very high amplitudes and short wavelengths. For example, there is a 400-600 nT anomaly associated with the western margin of the Blue Ridge thrust nappe. Similarly, along a continuing transect line through Virginia, Glover and Klitgord (Glover, 1995a) show a 1500-2000 nT anomaly associated with the western edge of the Potomac mélange. This transect crosses the Salisbury geophysical anomaly where it is expressed as an 600 nT anomaly (Figure 2.5-17). In summary, magnetic data published since the mid-1980's confirm and provide additional documentation of previous observations (i.e., pre-EPRI) across this region of eastern North America, and do not reveal any new anomalies related to geologic structures previously unknown to EPRI (EPRI, 1986).

2.5.1.1.4.4 Principal Tectonic Structures

Research since the EPRI study (EPRI, 1986) has advanced the understanding of the character and timing of the crustal architecture and tectonic history of the Atlantic continental margin.

The research has explained the significance of many geophysical anomalies and has clarified the timing and kinematics of tectonic processes from the Late Precambrian through the Cenozoic. Since the EPRI study (EPRI, 1986) was completed, new Cenozoic tectonic features have been proposed and described in the site region, and previously described features have since been characterized in more detail. New features identified since the EPRI study (EPRI, 1986) in the CCNPP site region area include gentle folds and a hypothesized minor fault on the western shore of Chesapeake Bay directly south of the CCNPP site (Kidwell, 1997). Also, new geologic data collected since 1986 has clarified the geometry and location of the Port Royal fault zone and Skinkers Neck anticline, and tectonic features representing the southern continuation of the Brandywine fault system, all of which are discussed further in the following sections. Tectonic features suggested by poorly constrained data include an unnamed fault underlying the upper Chesapeake Bay inferred by Pazzaglia (Pazzaglia, 1993), a series of warps beneath the lower Patuxent River and Chesapeake Bay near the CCNPP site hypothesized by McCartan (McCartan, 1995), and a hypothesized Stafford fault system by Marple and Talwani (Marple, 2004b) that is significantly longer and more active than previously recognized (Mixon, 2000). An additional geologic feature discovered since EPRI (1986) in the site region is the Eocene Chesapeake Bay impact crater (Figure 2.5-5 and Figure 2.5-6) (King, 1974) (Schruben, 1994). Based on the absence of published literature documenting Quaternary tectonic deformation and spatially associated with seismicity, we conclude that this feature is not a capable tectonic source (Section 2.5.1.1.4.4.4).

In the sections below, specific tectonic features and their evidence for activity published since the EPRI (1986) study are discussed. We find that no new information has been published since 1986 on any tectonic feature within the CCNPP site region that would cause a significant change in the EPRI seismic source model.

We divide principal tectonic structures within the 200 mi (322 km) CCNPP site region into five categories based on their age of formation or most recent reactivation. These categories include Late Proterozoic, Paleozoic, Mesozoic, Tertiary, and Quaternary. Late Proterozoic, Paleozoic, and Mesozoic structures are related to major plate tectonic events and generally are mapped regionally on the basis of geological and/or geophysical data. Late Proterozoic structures include normal faults active during post-Grenville orogeny rifting and formation of the Iapetan passive margin. Paleozoic structures include thrust and reverse faults active during Taconic, Acadian, Alleghenian, and other contractional orogenic events. Mesozoic structures include normal faults active during break-up of Pangaea and formation of the Atlantic passive margin.

Tertiary and Quaternary structures within the CCNPP site region are related to the tectonic environment of the Atlantic passive margin. This passive margin environment is characterized by southwest- to northeast-oriented, horizontal principal compressive stress, and vertical crustal motions. The vertical crustal motions associated with loading of the coastal plain and offshore sedimentary basins and erosion and exhumation of the Piedmont and westward provinces of the Appalachians. Commonly, these structures are localized, and represent reactivated portions of older bedrock structures. Zones of seismicity not clearly associated with a tectonic feature are discussed separately in Section 2.5.1.1.4.5.

2.5.1.1.4.4.1 Late Proterozoic Tectonic Structures

Extensional structures related to Late Proterozoic-Early Cambrian rifting of the former supercontinent Rhodinia and formation of the Iapetan Ocean basin are located along a northeast-trending belt between Alabama and Labrador, Canada, and along east-west-trending branches cratonward (Wheeler, 1995) (Johnston, 1994) (Figure 2.5-23). Major structures along this northeast-trending belt include the Reelfoot rift, the causative tectonic feature of the 1811 1812 New Madrid earthquake sequence. Within the 200 mi (322 km) site region, a discrete Late Proterozoic feature includes the New York-Alabama lineament (King, 1978) (Shumaker, 2000). The Rome Trough (Ervin and McGinnis, 1975) is located directly outside the 200-mile (322 km) site region. Within the eastern Piedmont physiographic province, ~~E~~extended crust of the Iapetan passive margin extends eastward beneath the Appalachian thrust front approximately to the eastern edge of Paleozoic crust extended during the Mesozoic ~~extended crust within the eastern Piedmont physiographic province~~ (Johnston, 1994; Wheeler, 1996) (Figure 2.5-15). This marks the western boundary of major Paleozoic sutures that juxtapose Laurentian crust against exotic crust amalgamated during the Paleozoic orogenies (Wheeler, 1996) (Figure 2.5-16 and Figure 2.5-17). At its closest approach, the area of largely intact and slightly extended Iapetan crust is located about 70 mi (113 km) northwest of the CCNPP site (Figure 2.5-23).

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The earthquake potential of Iapetan normal faults was recognized by the EPRI team members due to the association between the Reelfoot rift and the 1811 to 1812 New Madrid earthquake sequence (EPRI, 1986). Seismic zones in eastern North America spatially associated with Iapetan normal faults include the Giles County seismic zone of western Virginia, and the Charlevoix, Quebec seismic zone, both of which are located outside the CCNPP site region (Wheeler, 1995) (Figure 2.5-23). Because the Iapetan structures are buried beneath Paleozoic thrust sheets and/or strata, their dimensions are poorly known except in isolated, well studied cases.

Although published literature since the EPRI study (EPRI, 1986) has made major advances in showing the association between local seismic sources and Late Proterozoic structures (Wheeler, 1992) (Wheeler, 1995) and has highlighted the extent of extended Iapetan passive margin crust (Wheeler, 1995) (Wheeler, 1996), no new information has been published since 1986 on any Late Proterozoic feature within the CCNPP site region that would cause a significant change in the EPRI study (EPRI, 1986) seismic source model.

2.5.1.1.4.4.2 Paleozoic Tectonic Structures

The central and western portions of the CCNPP site region encompass portions of the Piedmont, Blue Ridge, Valley and Ridge, and Appalachian Plateau physiographic provinces (Figure 2.5-1). Structures within these provinces are associated with thrust sheets, shear zones, and sutures that formed during convergent and transpressional Appalachian orogenic events of the Paleozoic Era. Tectonic structures of this affinity exist beneath the sedimentary cover of the Coastal Plain and Continental Shelf Provinces. Paleozoic structures shown on Figure 2.5-23 include: 1) sutures juxtaposing allochthonous (tectonically transported) rocks against proto-North American crust, 2) regionally extensive Appalachian thrust faults and oblique-slip shear zones, and 3) a multitude of smaller structures that accommodated Paleozoic deformation within individual blocks or terranes (Figure 2.5-16, Figure 2.5-17, and Figure 2.5-18). The majority of these structures dip eastward and sole into either a low angle thrust or~~one or more levels of~~ the low angle, basal Appalachian decollement (Figure 2.5-17). Below the decollement are rocks that form the North American basement complex (Grenville or Laurentian crust).

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Researchers have observed that much of the sparse seismicity in eastern North America occurs within the North American basement below the basal decollement. Therefore, seismicity within the Appalachians may be unrelated to the abundant, shallow thrust sheets mapped at the surface (Wheeler, 1995). For example, seismicity in the Giles County seismic zone, located in the Valley and Ridge Province, is occurring at depths ranging from 3 to 16 mi (5 to 25 km) (Chapman, 1994), which is generally below the Appalachian thrust sheets and basal decollement (Bollinger, 1988).

2.5.1.1.4.4.2.1 Appalachian Structures

Paleozoic faults within 200 mi (322 km) of the CCNPP site and catalog seismicity are shown on Figure 2.5-23 and Figure 2.5-24 (see section 2.5.2 for a complete discussion on seismicity). Paleozoic faults with tectonostratigraphic units are shown on Figure 2.5-16, Figure 2.5-17, and Figure 2.5-18. Faults mapped within the Appalachian provinces (Piedmont, Blue Ridge, Valley and Ridge) are discussed in this section along with postulated Paleozoic faults in the Coastal Plain that are buried by Cenozoic strata. No new information has been published since 1986 on any Paleozoic fault in the site region that would cause a significant change in the EPRI study (EPRI, 1986) seismic source model. Paleozoic faults are discussed below from west to east across the CCNPP site region.

Major Paleozoic tectonic structures of the Appalachian Mountains within 200 mi (322 km) of the site include the Little North Mountain-Yellow Breeches fault zone, the Hylas shear zone, the Mountain Run-Pleasant Grove fault system, the Brookneal shear zone, and the Central Piedmont shear zone (including the Spotsylvania fault) (Figure 2.5-23). These structures bound lithotectonic units as defined in recent literature (Horton, 1991) (Glover, 1995b) (Hibbard, 2006) (Hibbard, 2007).

The northeast-striking Little North Mountain fault zone is located within the eastern Valley and Ridge Physiographic Province of western Virginia, eastern Maryland, and southern Pennsylvania (Figure 2.5-16 and Figure 2.5-23). The fault zone forms the tip of an upper level thrust sheet that attenuated Paleozoic shelf deposits of the Laurentian continental margin during the Alleghenian Orogeny (Hibbard, 2006). The east-dipping Little North Mountain thrust sheet soles into a decollement shown as a couple miles deep (Figure 2.5-17). This decollement represents an upper-level detachment above a deeper decollement about 5 mi (8 km) deep (Glover, 1995b) (Figure 2.5-17). The Little North Mountain fault and Yellow Breeches fault to the northeast mark the approximate location of the westernmost thrusts that daylight within the Valley and Ridge Province (Figure 2.5-23). Farther west, thrust ramps branching from the deeper decollement rarely break the surface and overlying fault-related folds control the morphology of the Valley and Ridge Province.

The Little North Mountain-Yellow Breeches fault zone is not considered a capable tectonic source. The decollement associated with the Little North Mountain thrust is within a couple miles of the surface, suggesting the fault probably does not penetrate to seismogenic depths. No seismicity is attributed to the Little North Mountain-Yellow Breeches fault zone and published literature does not indicate that it offsets late Cenozoic deposits or exhibits geomorphic expression indicative of Quaternary deformation. Therefore, this Paleozoic fault is not considered to be a capable tectonic source.

RAI 130 02.05.01-41

The Hylas shear zone, active between 330 and 220 million years ago during the Alleghenian orogeny, comprises a 1.5 mi (2.4 km) wide zone of ductile shear fabric and mylonites located 71 mi (115 km) southwest of the site (Bobyarchick, 1979; Gates, 1989). The Hylas shear zone also locally borders the Mesozoic Richmond and Taylorsville basins and appears to have been reactivated during Mesozoic extension to accommodate growth of the basin (Figure 2.5-10) (LaTourneau, 2003; Hibard, 2006). Discussions of the post-Paleozoic reactivation of the Hylas shear zone are presented in Section 2.4.1.1.4.4.3, Mesozoic Tectonic Structures, and in Section 2.4.1.1.4.4.4, Tertiary Tectonic Structures. Based on review of published literature and historical seismicity, there is no reported geomorphic expression, historical seismicity, or Quaternary deformation along the Hylas shear zone, and thus this feature is not considered to be a capable tectonic source.

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The Mountain Run-Pleasant Grove fault system is located within the Piedmont Physiographic Province in Virginia and Maryland and may extend to near Newark, New Jersey (Hibbard~~ et al.~~, ~~1995~~2006) (Figure 2.5-17 and Figure 2.5-23). This fault system extends across the entire site region and juxtaposes multiple-tectonized, allochthonous rocks and terranes to the east against the passive margin rocks of North American affinity to the west. Included in this fault systm are portions of the Bowens Creek fault, the Mountain Run fault zone, the Pleasant Grove fault, and the Huntingdon Valley fault (Horton, 1991; Mixon, 2000; Hibbard, 2006). ~~The f~~Fault zones along this fault system exhibit~~s~~ mylonitic textures, indicative of the ductile conditions in which it formed during the Paleozoic Era. Locally the allochthonous rocks are the Potomac composite terrane (Horton~~ et al.~~, 1991), which consists of a stack of thrust sheets containing tectonic mélange deposits that include ophiolites, volcanic arc rocks, and turbidites. This east-dipping thrust probably shallows to a decollement a couple miles below ground surface, and is shown to be truncated by the Brookneal shear zone (Figure 2.5-17) (Glover, 1995b). In the site region, the southeastern boundary of the Mesozoic Culpeper basin locally is bounded by the Mountain Run fault zone (Mixon, 2000)~~overlies the Mountain Run-Pleasant Grove fault system~~, suggesting that portions of the Paleozoic thrust fault system may have been reactivated ~~as normal faults in~~since the ~~Triassic~~Paleozoic (Figure 2.5-10). Discussions of the Culpeper basin and local reactivation of portions of the Mountain Run-Pleasant Grove fault system are in Section 2.5.1.1.4.4.3.

RAI 130 02.05.01-41

Within the Mountain Run-Pleasant Grove fault system, only local portions of the Mountain Run fault zone have been identified with possible late Cenozoic tectonic activity (Cron, 2000; Wheeler, 2006). These portions of the Mountain Run fault zone are discussed in Section 2.5.1.1.4.4.5.2. For other faults within the Mountain Run-Pleasant Grove fault system, published literature does not indicate that it offsets late Cenozoic deposits or exhibits geomorphic expression indicative of Quaternary deformation, and no seismicity has been attributed to it. Therefore, these faults are not considered to be capable tectonic sources.~~In northern Virginia, about 70 mi (113 km) west of the site, the Everona fault was identified within Tertiary, and possibly early Quaternary, debris flow deposits (Pavlides, 1983) (Pavlides, 1986). Subsequent studies performed during the North Anna ESP (Dominion, 2004a) on the activity of the Everona-Mountain Run fault system indicate that this fault system is not a capable tectonic source (Section 2.5.1.1.4.4.5.2).~~

RAI 130 02.05.01-41

The Brookneal shear zone is located within the Piedmont in Virginia and probably extends beneath the Coastal Plain across Virginia and Maryland to within about 50 mi (80 km) of the site (Figure 2.5-16 and Figure 2.5-23). The dextral-reverse shear zone is the northern continuation of the Brevard zone, a major terrane boundary extending from Alabama to North Carolina (Hibbard, 2002). The Brookneal shear zone juxtaposes magmatic and volcaniclastic rocks of the Chopawamsic volcanic arc to the east against the Potomac mélange to the west. This east-dipping thrust possibly truncates the Mountain Run fault at about 2.5 mi (4 km) depth, then flattens to a decollement at about 4 to 5 mi (6 to 8 km) depth that dips gently eastward beneath the surface trace of the Spotsylvania fault (Figure 2.5-17) (Glover, 1995b). Southwest of the site region, the Mesozoic Danville basin locally ~~coincides with~~overlies the Brookneal shear zone. The depositional contact defining the southeasterrn margin of the Danville basin crosses the Brookneal shear zone and is unfaulted, suggesting that ~~portions of~~ the Paleozoic fault ~~may have been~~was not reactivated as a normal fault~~s in the Triassic Period~~ during Triassic rifting.. The Brookneal shear zone is not considered a capable tectonic source. No seismicity is attributed to it and published literature does not indicate that it offsets late Cenozoic deposits or exhibits geomorphic expression indicative of Quaternary deformation. Therefore, this Paleozoic fault is not considered to be capable tectonic source.

RAI 130 02.05.01-41

The northeast-striking ~~Central Piedmont shear zone~~ Spotsylvania fault has been mapped in the Virginia piedmont as far north as Fredericksburg and beneath the Coastal Plain in eastern Virginia and Maryland (Hibbard, 2006) (Horton, 1991) (Glover, 1995b) (Figure 2.5-16, Figure 2.5-17 and Figure 2.5-23). At its closest approach, the fault is about 40 mi (64 km) northwest of the site (Figure 2.5-16). The fault juxtaposes terranes of different affinity, placing Proterozoic continental rocks of the Goochland terrane to the east against Early Paleozoic (Ordovician) volcanic arc rocks of the Chopawamsic terrane to the west (Glover, 1995b; Hibbard, 2006) (Figure 2.5-9). The Spotsylvania fault is a Late Paleozoic dextral-reverse fault active during the Alleghanian orogeny (Pratt, 1988; Bailey, 2004). The fault~~that~~ is ~~part~~the norther continuation of the Central Piedmont shear zone, a zone of ductile and brittle shear that accommodated thrust and right-lateral movement of various exotic volcanic arc terranes to the east against rocks of the Piedmont domain (including the Chopawamsic terrane) to the west (Hibbard, 1998; Hibbard, 2000; Bailey, 2004;~~(~~Hibbard, 2006). The Hycoshear zone, the part of the Central Piedmont shear zone located direly southeast of the Spotsylvania fault (Hibbard, 1998; Bailey, 2004), is partially located within the 200-mile site region (Figure 2.5-9 and Figure 2.5-23).~~The fault juxtaposes terranes of different affinity, placing continental rocks of the Goochland terrane to the east against volcanic arc rocks of the Chopawamsic terrane to the west.~~ The east-dipping Spotsylvania fault and Hyco shear zone likely penetrate~~s~~ the crust at gentle to intermediate angles (Hibbard, 1998; Pratt, 1988; Glover, 1995b), and the Spotsylvania fault may truncate~~s~~ the basal Appalachian decollement and higher decollement of the Brookneal shear zone (Figure 2.5-17) (Glover, 1995b).

RAI 130 02.05.01-41

RAI 130 02.05.01-41

RAI 130 02.05.01-41

The Spotsylvania fault and the Hyco shear zone are~~is~~ not considered ~~a~~capable tectonic sources. Specific studies of ~~this~~the Spotsylvania fault ~~feature~~ by Dames and Moore (DM, 1977b) demonstrate that ~~the Spotsylvania thrust fault~~it exhibits negligible vertical deformation of a pre- to early-Cretaceous erosion surface and is not related to Tertiary faulting along the younger Stafford fault zone (Section 2.5.1.1.4.4.4). The fault was determined by the NRC (AEC) to be not capable within the definition of 10 CFR 100, Appendix A (CFR, 2006). No subsequent evidence has been published since the Dames and Moore (DM, 1977b) study to indicate potential Quaternary activity on the Spotsylvania fault. Additionally, no geomorphic, geologic, or seismic evidence has been identified that indicates that the Hyco shear zone (the portion of the Central Piedmont shear zone within the 200-mile site region) has been active in Quaternary time. The Hyco shear zone is not considered a capable tectonic source.

RAI 130 02.05.01-41

RAI 130 02.05.01-41

2.5.1.1.4.4.2.2 Coastal Plain Structures

Major Paleozoic tectonic structures beneath the Coastal Plain in the 25 mi (40 km) CCNPP site vicinity include faults bounding the Sussex terrane west of the site and unnamed faults mapped seaward of the CCNPP site by Glover and Klitgord (Glover, 1995a) (Figure 2.5-16, Figure 2.5-17 and Figure 2.5-23). These fault zones, cited here as the western and eastern zones, are interpreted to dip steeply east, penetrate the crust, and juxtapose lithostratigraphic terranes.

The western fault zone coincides with the margins of the Sussex Terrane of Horton (Horton, 1991) (Figure 2.5-16 and Figure 2.5-17). The narrow Sussex Terrane and potential bounding faults are delimited in part by the Salisbury geophysical anomaly, a positive gravity and magnetic high described in Section 2.5.1.1.4.3. The eastern fault zone is shown to extend from coastal North Carolina to southern Delaware, trending north along the eastern part of southern Chesapeake Bay before branching into two splays that trend northeast across the Delmarva Peninsula (Figure 2.5-16 and Figure 2.5-23). The regional crustal cross section shows the fault zone as dipping east at moderate to steep angles (Figure 2.5-17).

No seismicity is attributed to the buried Paleozoic faults and published literature does not indicate that these faults offset late Cenozoic deposits or exhibit geomorphic expression indicative of Quaternary deformation. Therefore, the Paleozoic structures (faults bounding the Sussex terrane west of the site and unnamed faults mapped seaward of the CCNPP site by Glover and Klitgord (Glover, 1995a) in the site vicinity are not considered to be capable tectonic sources.

Other Paleozoic faults mapped by Hibbard (Hibbard, 2006) within the 200 mi (322 km) site region are smaller features that typically are associated with larger Paleozoic structures and accommodate internal deformation within the intervening structural blocks (Figure 2.5-23). No seismicity is attributed to these faults and published literature does not indicate that any of these faults offset late Cenozoic deposits or exhibit geomorphic expression indicative of Quaternary deformation. Therefore, these Paleozoic structures in the site region are not considered to be capable tectonic sources

2.5.1.1.4.4.3 Mesozoic Tectonic Structures

RAI 130 02.05.01-43

Mesozoic basins have long been considered potential sources for earthquakes along the eastern seaboard and were considered by most of the EPRI teams in their definition of seismic sources (EPRI, 1986). A series of elongate rift basins of early Mesozoic age are exposed in a belt extending from Nova Scotia to South Carolina and define ~~the~~an area of crust extended during the Mesozoic~~ crust~~ (Figure 2.5-10)(Benson, 1992). These Mesozoic rift basins, also commonly referred to as Triassic basins, exhibit a high degree of parallelism with the surrounding structural grain of the Appalachian orogenic belt. The parallelism generally reflects reactivation of pre-existing Paleozoic structures (Ratcliffe, 1986a; Schlische, 2003; LeTourneau, 2003; Schlische, 2003a). The rift basins formed during extension and thinning of the crust as Africa and North America rifted apart to form the modern Atlantic Ocean (Section 2.5.1.1.4.1.2) (Withjack, 2005).

RAI 130 02.05.01-43

RAI 71 02.05.01-7

Generally, the rift basins are asymmetric half-grabens with the primary rift-bounding faults on the western margin of the basin (Figure 2.5-10, Figure 2.5-18 and Figure 2.5-19) (Benson, 1992; Schlische, 1990; Withjack, 1998, Schlische, 2003). The rift-bounding normal faults are interpreted by some authors to be listric at depth and merge into Paleozoic low -angle detachments (Crespi, 1988) (Harris, 1982) (Manspeizer, 1988). Other authors interpret rift-bounding faults to penetrate deep into the crust following deep crustal fault zones (Wentworth, 1983) (Pratt, 1988) (Klitgord, 1995) (Figure 2.5-19).

RAI 130 02.05.01-43

Within the 200 mi (322 km) CCNPP site region, rift basins with rift-bounding faults on the western margin include the exposed Danville, Richmond, Culpeper, Gettysburg, and Newark basins, and the buried Taylorsville, Norfolk, hypothesized Queen Anne, and other smaller basins (Figure 2.5-10). As discussed below,~~In~~ most of the above-mentioned basins~~,~~ are bound by~~the basin-bounding~~ reactivated ~~normal fault is located in close proximity to a~~ Paleozoic thrust or reverse faults (e.g., ~~the Culpeper basin and the Paleozoic Mountain Run fault zone;~~ the Richmond basin and the Paleozoic Hylas shear zone) (Figure 2.5-10 and Figure 2.5-23). Field data also indicate that the Ramapo Fault was reactivated with both strike-slip and dip-slip displacement during Paleozoic orogenies and Mesozoic extension (Ratcliff, 1971). The principal basins within the site region are discussed below in further detail~~The rift-bounding normal faults are interpreted by some authors to be listric at depth and merge into Paleozoic low angle basal décollement (Manspeizer, 1989). Other authors interpret rift-bounding faults to penetrate deep into the crust following deep crustal fault zones (Figure 2.5-19)~~.

RAI 130 02.05.01-43

The Culpepper, Gettysburg, and Newark basins (i.e. the composite Birdsboro basin of Faill [2003]) form an east- to northeast-trending band of mostly exposed Mesozoic basins located 60 to 125 miles west, northwest, and north of the CCNPP Unit 3 site (Figure 2.5-10). These basins are asymmetric half-grabens bounded on the west or northwest by a series of interconnected east- to southeast-dipping fault zones (Lindholm, 1978) (Hibbard, 2006). The fault bounding the western margin of the Culpeper basin was observed to follow a well-developed foliation in metamorphic rocks by Lindholm (1978), indicating to him that the Mesozoic faulting was controlled by Paleozoic structure. However, a named Paleozoic fault zone associated with the western margin of the Culpeper basin is not clearly identified in the published literature. The southeast margin of the Culpeper basin is locally in fault contact with the Paleozoic Mountain Run fault zone (Mixon, 2000) (Hibbard, 2006) (Figure 2.5-10). This southeast-dipping fault contact probably represents post-Triassic, east-side up movement, although the total post-Triassic throw on the fault is limited and does not seem to strongly influence the basin architecture (Mixon, 2000). The Mountain Run fault zone is discussed further in FSAR sections 2.5.1.1.4.4.2.1 and 2.5.1.1.4.4.5.2.

The Gettysburg and Newark basins are bounded on their northwestern margins by southeast-dipping faults with a recognized Paleozoic history. The Gettysburg basin is bounded by the Shippenburg and Carbaugh-Marsh Creek faults (Root, 1989). The Newark basin is at least partially bounded by the Ramapo Fault zone (Ratcliffe, 1985; 1986a) (Schlische, 1992). Detailed studies of these basin-bounding faults confirm they formed as a result of reactivation of Paleozoic faults or metamorphic structures (Ratcliffe, 1985) (Root, 1989) (Schlische, 1993) (Swanson, 1986). None of these basinbounding faults have demonstrable associated Quaternary seismic activity or conclusive evidence for recent fault activity (Section 2.5.1.1.4.4.5). The northeast-striking, narrow Danville basin (also grouped with the larger Dan River-Danville basin) is located about 170 miles southwest of the CCNPP Unit 3 site (Figure 2.5-10). The primary basin-bounding fault is located on the northwest margin of the basin and dips southeast (Benson, 1992) (Hibbard, 2006), creating a highly asymmetric cross-section (Schlische, 2003). Swanson (1986) summarizes evidence suggesting the main basin-bounding fault reactivated ductile Paleozoic faults, specifically the Stony Ridge fault zone, a probable northern extension of the Paleozoic Chatham fault. The Danville basin and the basin-bounding Chatham fault separates the Smith River Terrane on the northwest against the Milton terrane on the southeast within the central portion of the basin, but farther northeast the fault and basin are located within the Potomac terrane as mapped by Horton (1991).

RAI 130
02.05.01-43

The northeast-striking Richmond Taylorsville basins are located about 80 miles and 30 miles west and southwest of the CCNPP Unit 3 site, respectively within central Virginia and Maryland (Figure 2.5-10). The Richmond basin is subaerially exposed and its extent is well defined by mapping. In contrast, the Taylorsville basin is mainly buried beneath the coastal plain and its extent is constrained by limited geologic mapping, multiple seismic lines, boreholes, and interpretation of gravity and aeromagnetic data (Milici, 1995) (LeTourneau, 2003). The extent of the buried portions of the Taylorsville basin is well-defined in Virginia, but poorly constrained within Maryland based on limited subsurface data (Jacobeen, 1972) and a lack of seismic lines.

Where exposed, both the Taylorsville and Richmond basins are bounded on the west by the northeast-striking, southeast-dipping Paleozoic Hylas shear zone (Section 2.5.1.1.4.4.2.1) (Figure 2.5-10 and Figure 2.5-23). Bobyarchick and Glover (Bobyarchick,1979) argue that the Hylas shear zone was reactivated as an extensional fault to accommodate the growth of the Richmond and Taylorsville basins during Mesozoic rifting based on a 220 million year old phase of brittle extensional deformation mapped throughout the fault zone. Evidence for later Mesozoic and early Tertiary inversion of the Taylorsville basin is based on interpretation of seismic reflection profiles (LeTourneau, 2003) and the coincidence of the eastern margin of the Taylorsville basin with contractional structures that disrupt the Cretaceous and early Tertiary coastal plain sediments (i.e. Skinker's Neck anticline, Port Royal fault zone, and Brandywine fault zone) (Section 2.5.1.1.4.4.4) (Figure 2.5-25).

The extension of the basin bounding fault of the Taylorsville basin (Hylas shear zone) beneath the CCNPP site can be hypothesized based on a range of possible down-dip geometries. The northwestern boundary of the Taylorsville basin is approximately 27 to 30 miles (44 to 48 km) northwest of the CCNPP site (Figure 2.5-10) (Schlische, 1990)(Benson, 1992). Available crustal-scale cross sections provide a range of dip angles from 20 degrees (Withjack 1998) (Schlische, 2003a) to 25 degrees (Glover, 1995) (Klitgord, 1995) (Figure 2.5-17 and Figure 2.5-19) to 30 degrees (Pratt, 1988). Based on this range in dip angle the Hylas shear zone would be 10-11 mi (16-18 km), 12-14 mi (20-22 km), and 15-17 mi (25-28 km) beneath the CCNPP site within crystalline bedrock. The thickness of the seismogenic upper crust (i.e. depth to the Moho) is variable in these cross sections and is typically depicted as either 9 mi (15 km) thick (Schlische, 1990)(Schlische, 2003a) or 18-25 mi (30-40 km thick). The 9 mi (15 km) thick model suggests that the Hylas shear zone should sole into the Moho before the fault extends beneath the CCNPP site.

The geometry and continuity of the buried Queen Anne basin and other smaller rift basins beneath the Coastal Plain and Continental Shelf are not clear, but the recognition and interpretation of these basins have expanded since the EPRI (1986) study (Figure 2.5-10).

Data constraining the location of the buried Queen Anne basin with respect to the CCNPP Unit 3 Site are sparse and thus the geometry and continuity of the basin are unclear. Seismic reflection studies (Hansen, 1988)(Benson, 1992), borehole data (Hansen, 1978) (Figure 2.5-11), and gravity and magnetic signatures (Benson, 1992)(Hansen, 1988)(Figure 2.5-23) were used to characterize the limits of the Queen Anne basin. These data permit multiple interpretations of the location of a basin at or near the CCNPP Site (Klitgord, 1988) (Schlische, 1990) (Horton, 1991) (Bensen, 1992) (Klitgord 1995) (Withjack 1998) (LeTourneau, 2003) (Figure 2.5-10, Figure 2.5-12, Figure 2.5-16, and Figure 2.5-22).

RAI 130 02.05.01-43

The delineation of the Queen Anne basin by Benson (1992) (shown on Figure 2.5-10) is derived from a seismic reflection profile (Hansen, 1988) approximately 40 mi northeast of the site, "extensive proprietary seismic reflection profiling" data south of CCNPP, a borehole located about 13 miles southwest of the site, and aeromagnetic and gravity data. The Queen Anne basin first named and imaged by Hansen (1988) in the TXC-10C Vibroseis profile located 40 mi northeast of the CCNPP site. This seismic line crosses the eastern boundary of the basin imaging west-dipping Triassic basin deposits above high-angle west-side-down faults offsetting crystalline basement (Hansen, 1988), but does not cross the western boundary of the basin. The Coastal Plain section is not deformed by the underlying faults. As discussed below, Benson (1992) extends the Queen Anne basin to the south based on the presence of proprietary seismic lines. Although Benson (1992) did not review the data, he inferred, based on the local concentration of these proprietary seismic lines, that they were acquired to better image a known Tertiary basin. A borehole located about 13 miles southwest of the CCNPP Unit 3 site encountered a diabase dike at depth (Benson, 1992). Although suggestive, Benson (1992) acknowledges that the diabase dike may or may not be associated with a Mesozoic basin. Benson (1992) summarizes: "The areas of inferred buried rift basins/synrift rocks shown in this map might best be considered as areas where efforts should be concentrated to verify their presence or absence." To convey this uncertainty, Benson (1992) shows the southern extension of the Queen Anne basin with a dashed and queried boundary, whereas to the north-northeast of the site the basin boundary is depicted as a solid line where geophysical data are available (and verifiable). Subsequent authors have relied upon and modified Benson (1992), yet no new published information is available near the CCNPP site to better constrain the presence or absence of a Triassic basin beneath the site. The Hillville fault (Hansen,1986) may represent a fault along the western margin of the

Queen Anne basin or the eastern margin of the Taylorsville basin reactivated during Cretaceous and early Tertiary time. The geometry of this fault discussed in Section 2.5.1.1.4.4.5—is poorly constrained in the vibroseis line by Hansen (1978), which illustrates offset crystalline basement. There are limited data to constrain its length and no data to constrain its down-dip geometry (Hansen, 1986). In addition, there is no evidence for Quaternary activity of the Hillville fault or any other structure associated with the hypothesized Queen Anne basin.

RAI 130 02.05.01-43 ~~The geometry and continuity of buried rift basins beneath the Coastal Plain and Continental Shelf is not clear, but the recognition and interpretation of these basins have expanded since the EPRI (1986) study. In addition to the identification of new basins since 1986, several alternative geometries have been proposed for the site region (Figure 2.5-10 and Figure 2.5-16) (Horton, 1991) (Benson, 1992) (Klitgord, 1995) (Withjack, 1998) (LeTourneau, 2003). Interpretations are constrained loosely based on sparse borehole, seismic, and aeromagnetic anomaly data (Benson, 1992). Some authors show the Queen Anne basin located beneath the CCNPP site (e.g., in Figure 2.5-10(Benson, 1992) and in Figure 2.5-16 (Horton, 1991)). More recent compilations of rift basins do not show the CCNPP site overlying a Mesozoic basin (e.g., in Figure 2.5-10 (Withjack, 1998) and in Figure 2.5-16 (Glover, 1995b)).~~

~~Reactivation of faults bordering or within Triassic basins in the Cenozoic as reverse faults is recognized in several basins within the site region and is discussed in Section 2.5.1.1.4.1.2. (e.g., (Schlische, 2003)). For example, the buried Taylorsville basin coincides with numerous postrift contractional structures of Cretaceous and Tertiary age including the Brandywine, Port Royal, Skinkers Neck, and Hillville faults (Section 2.5.1.1.4.4.4).~~

RAI 71 02.05.01-17 In summary~~Aside from the global finding of Johnston et al. (1994) that areas of Mesozoic extended crust are correlated with large magnitude earthquakes within stable continental regions (i.e., New Madrid seismic zone)~~, there are no specific Mesozoic basin-bounding faults within the site region that have demonstrable associated seismic activity or evidence ~~for~~of recent fault activity (Figure 2.5-10~~ and Figure 2.5-24~~). The major postulated basins closest to the site (Taylorsville and Queen Anne) were considered during the 1980s to exist and several were incorporated into seismic sources by the different EPRI teams. Seismicity potentially associated with reactivation of faults bordering or beneath the Mesozoic basins is captured in the existing EPRI seismic source model. No new data have been developed to demonstrate that any of the Mesozoic basins are currently active, and Crone and Wheeler (Crone, 2000), Wheeler (Wheeler, 2005) and Wheeler (Wheeler, 2006) do not recognize any basin-margin faults that have been reactivated during the Quaternary in the site region. No Mesozoic basin in the site region is associated with a known capable tectonic source, and no new information has been developed since 1986 that would require a significant revision to the EPRI seismic source model.

2.5.1.1.4.4.4 Tertiary Tectonic Structures

Several faults were active during the Tertiary Period within the 200 mi (322 km) CCNPP site region (Figure 2.5-25). These faults have been recognized in the western part of the Coastal Plain Province where Tertiary strata crop out in river valleys and where the faults have bee investigated using seismic and borehole data. These faults include the relatively well characterized Stafford fault system in Virginia, the Brandywine fault system in Maryland, and the National Zoo/Rock Creek faults in Washington, D.C. Additional faults and fault-related folds defined by seismic and borehole data include the Port Royal fault zone and Skinkers Neck anticline in Virginia, and the Hillville fault in Maryland. Tertiary structures that have been proposed but are poorly constrained by data include east-facing monoclines along the western shore of Chesapeake Bay (McCartan, 1995) and a northeast-striking fault in the upper Chesapeake Bay (Pazzaglia, 1993). In addition, Kidwell (Kidwell, 1997) uses detailed stratigraphic analysis of the Calvert Cliffs area to postulate the existence of several broad folds developed in Miocene strata as well as a poorly constrained postulated fault. All of these structures are located within about 50 mi (80 km) of the site, and the proposed east-facing monoclines of McCartan (McCartan, 1995) are within a few miles of the CCNPP site. Within 25 mi (40 km) of the site, the only fault with documented Tertiary displacement is the Hillville fault (Hansen, 1978) (Hansen, 1986) (Figure 2.5-25).

RAI 130 02.05.01-44

Several faults associated with the Eocene Chesapeake Bay impact crater have been identified near the mouth of the Chesapeake Bay about 60 mi (97 km) south of the site (Powars, 1999) (Figure 2.5-5). The impact crater formed on a paleo-continental shelf when the Eocene sea in this location was approximately 1,000 ft (305 m) deep. The Chesapeake Bay impact crater was discovered in 1993, and thus post-dates the EPRI study (EPRI, 1986). The 35-million year old Chesapeake Bay impact crater is a 56 mi (90 km) wide, complex peak-ring structure.~~, defined by a series of inner and outer ring faults, some of which penetrate the Proterozoic and Paleozoic crystalline basement rocks~~Fault styles observed within the impact include a series of inner and outer ring, post-impact, compaction related growth faults, sin-impact faults that offset Proterozoic and Paleozoic crystalline basement rocks, and syn-impact faults related to secondary craters (Powars, 1999; Poag, 2004; Poag, 2005)). These faults and others within the outer and inner ring include normal-faulted slump blocks and compaction faults that extend up-section into upper Miocene and possibly younger deposits. Published literature does not indicate that any faults related to the impact crater are seismogenic or offset Quaternary deposits.

Multiple, fault-bounded secondary craters of Eocene age also have been interpreted from multichannel seismic profiles previously collected by Texaco along the Potomac River and Chesapeake Bay 20 and 40 mi (32 and 64 km) north and northwest of the main Chesapeake Bay impact crater (Poag, 2004). The secondary impact craters have diameters ranging from 0.25 to 2.9 mi (0.4 to 4.7 km). Faults associated with the secondary craters occasionally penetrate Proterozoic and Paleozoic crystalline basement rocks (Poag, 2004). Primarily middle Miocene to Quaternary sediments thicken and sag into the primary and secondary craters. Faults associated with the impact crater are not considered capable tectonic sources and are not discussed further in this section.

Faults and folds mapped within the 200 mi (322 km) CCNPP site region that displace Tertiary Coastal Plain deposits are described below. These structures include the Stafford fault system, Brandywine fault system, National Zoo/Rock Creek faults, Port Royal fault zone, Skinkers Neck anticline, and the Hillville fault. Additional hypothesized Tertiary structures for which compelling geologic or geophysical evidence is lacking are then described. These structures include hypothesized east-facing monoclines along the western shore of Chesapeake Bay near the CCNPP site described by McCartan (McCartan, 1995), a hypothesized fault in the upper Chesapeake Bay mapped by Pazzaglia (Pazzaglia, 1993), and structures interpreted in Calvert Cliffs by Kidwell (Kidwell, 1997).

2.5.1.1.4.4.4.1 Stafford Fault of Mixon, et al.

The Stafford fault (#10 on Figure 2.5-31) approaches within 47 mi (76 km) southwest of the site (Figure 2.5-25). The 42 mi (68 km) long fault system strikes approximately N35°E (Newell, 1976). The fault system consists of several northeast-striking, northwest-dipping, high-angle reverse to reverse oblique faults including, from north to south, the Dumfries, Fall Hill, Brooke, Tank Creek, Hazel Run, and an unnamed fault (Mixon et al., 2000). Two additional northeast-striking, southeast-side-down faults, the Ladysmith and the Acadia faults, are included here as part of the Stafford fault system. These individual faults are 10 to 25 mi (16 to 40 km) long and are separated by 1.2 to 3 mi (2 to 5 km) wide en echelon, left step-overs. The left-stepping pattern and horizontal slickensides found on the Dumfries fault suggest a component of dextral shear on the fault system (Mixon, 2000).

RAI 130 02.05.01-45 Locally, the Stafford fault system coincides with the Fall Line and a northeast-trending portion of the Potomac River (Figure 2.5-25). Mixon and Newell (Mixon, 1977) suggest that the Fall Line and river deflection may be tectonically controlled. ~~Detailed d~~Drilling, trenching, and mapping in the Fredericksburg region ~~by Dames and Moore (DM, 1973)~~ showed that ~~the youngest identifiable~~most fault movement on any of the four primary faults comprising the Stafford fault system was pre-middle Miocene in age. Mixon, 1978; 1982). Mesozoic and Tertiary movement is documented by displacement of Ordovician bedrock over lower Cretaceous strata along the Dumfries fault and abrupt thinning of the Paleocene Aquia Formation across multiple strands of the fault system (Mixon, 1977). Minor late Tertiary activity of the fault system is documented by an 11-14-inch (28-36 cm) displacement by the Fall Hill fault of a Pliocene terrace deposit along the Rappahannock River (Mixon, 1978) (Mixon, 1982)(Mixon, 2000) and an 18 in (46 cm) displacement near the Hazel Run fault of upland gravels of Miocene or Pliocene age (Mixon, 1978). Both offsets suggest southeast-side-down displacement (Mixon, 1978).

RAI 130 02.05.01-45 Subsequent studies of the Stafford fault system better document the timing of displacement, mostly by refining the age of units. For example, the Rappahannock River terrace deposit was originally cited as Late Pliocene or early Pleistocene. However, later work has revealed that the deposit is Pliocene in age (Mixon et al., 2000). Similarly, the Miocene or Pliocene upland gravels offset 18" are now interpreted as the Pliocene sand and gravel unit, Tps (Mixon et al., 2000).. ~~Mesozoic and Tertiary movement is documented by displacement of Ordovician bedrock over lower Cretaceous strata along the Dumfries fault and abrupt thinning of the Paleocene Aquia Formation across multiple strands of the fault system (Mixon, 2000). Minor late Tertiary activity of the fault system is documented by an 11 inch displacement by the Fall Hill fault of a Pliocene terrace deposit along the Rappahannock River (Mixon, 1978) (Mixon, 2000) and an 18 in (46 cm) displacement by the Hazel Run fault of upland gravels of Miocene or Pliocene age (Mixon, 1978). Both offsets suggest southeast-side-down displacement (Mixon, 1978).~~

Recent geologic and geomorphic analysis of the Stafford fault system for the application of North Anna Early Site Permit (ESP) to the NRC provides additional constraints on the age of deformation (Dominion, 2004a). Geomorphic analyses (structure contour maps and topographic profiles) of upland surfaces capped by Neogene marine deposits and topographic profiles of Pliocene and Quaternary fluvial terraces of the Rappahannock River near Fredericksburg, Virginia, indicate that these surfaces are not visibly deformed across the Stafford fault system (Dominion, 2004a). In addition, field and aerial reconnaissance of these features during the North Anna ESP, and as part of this CCNPP Unit 3 study, indicate that there are no distinct scarps or anomalous breaks in topography on the terrace surfaces associated with the mapped fault traces. The NRC (2005) agreed with the findings of the subsequent study for the North Anna ESP, and stated: "Based on the evidence cited by the applicant, in particular the applicant's examination of the topography profiles that cross the fault system, the staff concludes that the applicant accurately characterized the Stafford fault system as being inactive during the Quaternary Period." Collectively, this information indicates that the Stafford fault system is not a capable tectonic source as defined in Appendix A of Regulatory Guide 1.165 (NRC, 1997).

Marple (Marple, 2004a) recently proposed a significantly longer Stafford fault system that extends from Fredericksburg, Virginia to New York City as part of a northeastern extension of

RAI 134 02.05.01-58

the postulated East Coast fault system (ECFS), (Figure 2.5-31) (Section 2.5.1.1.4.4.5.~~15~~14). The proposed northern extension of the Stafford fault system is based on: (1) aligned apparent right-lateral deflections of the Potomac (22 mi (35 km) deflection), Susquehanna (31 mi (50 km) deflection) and Delaware Rivers (65 mi (105 km) deflection) (collectively these are named the "river bend trend"), (2) upstream incision along the Fall Line directly west of the deflections, and (3) limited geophysical and geomorphic data. Marple and Talwani (Marple, 2004b) proposed that the expanded Stafford fault system of Marple (Marple, 2004a) was a northeast extension of the ECFS of Marple and Talwani (Marple, 2000). Marple and Talwani (Marple, 2004b) further speculate that the ECFS and the Stafford fault system were once a laterally continuous and through-going fault, but subsequently were decoupled to the northwest and southeast, respectively, during events associated with the Appalachian orogeny.

Data supporting the extended Stafford fault system of Marple (Marple, 2004a) is limited. Marple and Talwani (Marple, 2004b) suggest that poorly located historical earthquakes that occurred in the early 1870's and 1970's lie close to the southwestern bend in the Delaware River and concluded an association between historical seismicity and the postulated northern extension of the Stafford fault system. Review of seismicity data available both before and after the EPRI study (EPRI, 1986) indicates a poor correlation in detail between earthquake epicenters and the expanded Stafford fault system (Figure 2.5-25). Geophysical, borehole and trench data collected by McLaughlin (McLaughlin, 2002), near the Delaware River across the trace of the postulated expanded Stafford fault system of Marple (Marple, 2004a), provide direct evidence for the absence of Quaternary deformation. Collectively, there is little geologic and seismologic evidence to support this extension of the fault system beyond that mapped by Mixon (Mixon, 2000).

In summary, all significant information on timing of displacement for the Stafford fault system was available prior to 1986 and incorporated into the EPRI (1986) seismic source models. New significant information published since 1986 regarding the activity of the Stafford fault system includes the geomorphic and geologic analysis performed for the North Anna ESP that concluded the fault system was not active (Dominion, 2004a). Field and aerial reconnaissance performed for the North Anna ESP and this CCNPP COL application also did not reveal any geologic or geomorphic features indicative of potential Quaternary activity along the fault system. Therefore, on the basis of a review of existing geologic literature, the Stafford fault system is not considered a capable tectonic source, and there is no new information that would require a significant revision to the EPRI (1986) seismic source model.

2.5.1.1.4.4.4.2 Brandywine Fault System

The Brandywine fault system is located approximately 30 mi (48 km) west of the site and north of the Potomac River (Figure 2.5-25). The 12 to 30 mi (19 to 48 km) long Brandywine fault system consists of a series of en echelon northeast-trending, southeast-dipping reverse faults with east-side-up vertical displacement. Jacobeen (Jacobeen, 1972) and Dames and Moore (DM, 1973) first described the fault system from Vibroseis™ profiles and a compilation of borehole data as part of a study for a proposed nuclear power plant at Douglas Point along the Potomac River. The fault system is composed of the Cheltenham and Danville faults, which are 4 mi and 8 mi (6 to 13 km) long, respectively. These two faults are separated by a 0.6 to 1 mi (1 to 1.6 km) wide left step-over (Jacobeen, 1972). Later work by Wilson and Fleck (Wilson, 1990) interpret one continuous 20 to 30 mi (32 to 48 km) long fault that transitions into a west-dipping flexure to the south near the Potomac River. The mapped trace of the

RAI 130 02.05.01-46

RAI 130 02.05.01-46

Brandywine fault system is generally coincident with (within 1.0 to 2.5 miles (2 to 4 km)) and parallel to the aeromagnetic and gravity anomalies used to define ~~coincides with~~ the western boundary~~margin~~ of the Taylorsville basin but they do not precisely coincide (Mixon, 1977) (Hansen, 1986) (Wilson, 1990) (Benson, 1992). This observation lead Mixon and Newell (Mixon,

1977) to speculate the origin of the Brandywine fault system may be related to the reversal of a pre-existing zone of crustal weakness (i.e., Taylorsville Basin border fault).

The Brandywine fault system was active in the Early Mesozoic and reactivated during late Eocene and possibly middle Miocene time (Jacobeen, 1972) (Wilson, 1990). Basement rocks have a maximum vertical displacement of approximately 250 ft (76 m) across the fault (Jacobeen, 1972). Also, the Cretaceous Potomac Formation is 150 ft (46 m) thinner on the east (up-thrown) side of the fault indicating syndepositional activity of the fault. The faulting is interpreted to extend upward into the Eocene Nanjemoy Formation (70 ft (21 m) offset) (Wilson, 1990), and die out as a subtle flexure developed within the Miocene Calvert Formation (8 ft (2.4 km) flexure) (Jacobeen, 1972).

Wilson and Fleck (Wilson, 1990) speculate that the fault system continues northeast toward the previously mapped Upper Marlboro faults, near Marlboro, Maryland (Figure 2.5-25). Dryden (Dryden, 1932) reported several feet of reverse faulting in Pliocene Upland deposits in a railroad cut near Upper Marlboro, Maryland (Prowell, 1983). However, these faults are not observed beyond this exposure. Wheeler (Wheeler, 2006) suggests that the Upper Marlboro faults have a surficial origin (i.e., landsliding) based on the presence of very low dips and geometric relations inconsistent with tectonic faulting. Field reconnaissance conducted as part of this CCNPP Unit 3 study used outcrop location descriptions from Prowell (Prowell, 1983) but failed to identify any relevant exposures associated with the faults of Dryden (Dryden, 1932). Wheeler's (Wheeler, 2006) assessment of the Upper Marlboro fault appears to be consistent with the outcrop described by Dryden (Dryden, 1932) as not being associated with the Brandywine fault system.

Geologic information indicates that the Brandywine fault system was last active during the Miocene. All geologic information on the timing of displacement on the Brandywine fault system was available and incorporated into the EPRI seismic source models in 1986. The post-EPRI study by Wilson and Fleck (Wilson, 1990) extended the fault north and south as an anticline, but offers no new information about the timing of the deformation. There is no pre-EPRI or post-EPRI seismicity associated with this fault system. This fault system is identified only in the subsurface and geologic mapping along the surface projection of the fault zone does not show a fault (DM, 1973) (McCartan, 1989a) (McCartan, 1989b). Field and aerial reconnaissance performed as part of this CCNPP Unit 3 study, coupled with interpretation of Light Detection and Ranging (LiDAR) data (see Section 2.5.3.1 for additional information regarding the general methodology), revealed no anomalous geomorphic features indicative of potential Quaternary activity. The Brandywine fault system, therefore, is not a capable tectonic source and there is no new information developed since 1986 that would require a significant revision to the EPRI seismic source model.

2.5.1.1.4.4.4.3 Port Royal Fault Zone and Skinkers Neck Anticline

The Port Royal fault zone and Skinkers Neck anticline are located about 32 mi (51 km) west of the CCNPP site, south of the Potomac River (Figure 2.5-25). First described by Mixon and Powars (Mixon, 1984), these structures have been identified within the subsurface by: (1) contouring the top of the Paleocene Potomac Formation, (2) developing isopach maps of the Lower Eocene Nanjemoy Formation, and (3) interpreting seismic lines collected in northern Virginia (Milici, 1991) (Mixon, 1992) (Mixon, 2000). The fault and anticline are not exposed in surface outcrop. The Port Royal fault zone is located about 4 to 6 mi (6 to 10 km) east and strikes subparallel to the Skinkers Neck anticline and the Brandywine fault system. In our discussion, we consider the Skinkers Neck anticline to consist of a combined anticline and fault zone, following previous authors.

Mixon and Newell (Mixon, 1977) first hypothesized that a buried fault zone existed beneath Coastal Plain sediments and connected the Taylorsville basin in the north to the Richmond basin in the south along a fault zone coincident with the Brandywine fault zone of Jacobeen (Jacobeen, 1972). The inferred fault of Mixon and Newell (Mixon, 1977) coincides with a gravity gradient used to target exploration studies that led to the discovery of the Port Royal fault and Skinkers Neck anticline in 1984 (Mixon, 1984) (Mixon, 1992).

The Port Royal fault zone consists of a 32 mi (51 km) long, north to northeast-striking fault zone that delineates a shallow graben structure that trends parallel to a listric normal fault bounding the Taylorsville basin (Mixon, 2000) (Milici, 1991). In map view, the fault zone makes a short left-step to the Brandywine fault system (Figure 2.5-25). Along the northern part of the fault zone, near the town of Port Royal, Virginia, the fault is expressed in the subsurface as a 3 mi (5 km) wide zone of warping with a west-side-up sense of displacement. Water well and seismic reflection data show an apparent west-side-up vertical component for the southwestern part of the structure also (Mixon, 1992) (Mixon, 2000) (Milici, 1991).

The Skinkers Neck anticline is located directly west of the Port Royal fault zone and southwest of the mapped terminus of the Brandywine fault system (Figure 2.5-25). The north- to northeast-striking structure is 30-mi (48 km) long and 3 to 5 mi (5 to 8 km) wide, and is defined as an asymmetric, low-amplitude, north-plunging anticline with a west-bounding fault (Mixon, 2000). Locally, Mixon (Mixon, 2000) map the feature as two separate, closely-spaced anticlines. Along the west side of the structure, a fault zone strikes north-to-northeast and is interpreted as a fault-bounded, down-dropped block. The Skinkers Neck anticline is not mapped north of the Potomac River by Mixon (Mixon, 1992) (Mixon, 2000). However, McCartan (McCartan, 1989a) shows two folds north of the Potomac River, west of the Brandywine fault system, and along trend with the Skinkers Neck anticline as mapped by Mixon (Mixon, 2000).

The Port Royal fault zone and Skinkers Neck anticline likely are associated with Paleozoic structures that were reactivated in the Early Mesozoic, Paleocene, and possibly middle Miocene (Mixon, 1992) (Mixon, 2000) (McCartan, 1989c). Similar to the Brandywine fault system, these structures closely coincide with the Mesozoic Taylorsville basin (Mixon, 1992) (Milici, 1991). This apparent coincidence with a Mesozoic basin suggests that the Port Royal fault zone and the Skinkers Neck anticline represent possible pre-existing zones of crustal weakness. Post-Mesozoic deformation includes as much as 30 to 33 ft (9 to 10 km) of Paleocene offset, and less than 25 ft (7.6 m) of displacement across the basal Eocene Nanjemoy Formation. Deformation on the order of 5 to 10 ft (1.5 to 3 m) is interpreted to extend upward into the Middle Miocene Calvert and Choptank Formations (Mixon, 1992). The overlying Late Miocene Eastover Formation is undeformed across both the Port Royal fault zone and Skinkers Neck anticline, constraining the timing of most recent activity (Mixon, 1992) (Mixon, 2000).

Although the Port Royal fault zone and Skinkers Neck anticline where characterized after the EPRI study (EPRI, 1986), geological information available to the EPRI teams regarding the pre-Quaternary activity of the structures was available (Mixon, 1984). Both of these structures are mapped in the subsurface as offsetting Tertiary or older geologic units (Mixon, 2000). Field and aerial (inspection by plane) reconnaissance, coupled with interpretation of aerial photography (review and inspection of features preserved in aerial photos) and LiDAR data (see Section 2.5.3.1 for additional information regarding the general methodology), conducted during this CCNPP Unit 3 study shows that there are no geomorphic features indicative of potential Quaternary activity along the surface-projection of the fault zone (i.e., along the northern banks of the Potomac River and directly northeast of the fault zone). Also, there is no pre-EPRI or post-EPRI (EPRI, 1986) seismicity spatially associated with the Port Royal fault zone or the Skinkers Neck anticline. In summary, the Port Royal fault zone and Skinkers Neck anticline are not considered capable tectonic sources, there is no new information developed since 1986 that would require revision to the EPRI seismic source model regarding these features.

2.5.1.1.4.4.4.4 National Zoo Faults

The National Zoo faults in Washington D.C. approach to within 47 mi (76 km) of the site (Figure 2.5-25). The National Zoo faults are primarily low-angle to high-angle, northwest-striking, southwest-dipping thrust faults that occur within a 1.0 to 1.5 mi (1.6 to 2.4 km) long, north to northeast-trending fault zone (Prowell, 1983) (McCartan, 1990) (Fleming, 1994) (Froelich, 1975). The mapped surface traces of these faults range from 500 to 2000 ft (152 to 610 km) with up to 20 ft (6 m) of post-Cretaceous reverse displacement visible in outcrops at the National Zoo (Fleming, 1994). The faults were first identified by Darton (Darton, 1950) in exposures along Rock Creek in historic excavations between the National Zoo and Massachusetts Avenue in Washington D.C.

The National Zoo faults were active during the Early Mesozoic with probable reactivation during the Pliocene (Darton, 1950) (McCartan, 1990) (Fleming, 1994). This fault zone is coincident with the mapped trace of the Early Paleozoic Rock Creek shear zone, which led several researches to infer that the National Zoo faults are related to reversal of a pre-existing zone of crustal weakness (McCartan, 1990) (Fleming, 1994). Combined with the Rock Creek fault zone, the National Zoo faults could be up to 16 mi (26 km) long. Differential offset across basement and Potomac Group contacts also suggests Paleozoic fault reactivation (Fleming, 1994). The Cretaceous Potomac formation offsets are primarily less than 50 ft (15 m) and isopach maps show a thickening of Coastal Plain sediments east of these faults (Fleming, 1994) (Darton, 1950). The youngest two faults juxtapose basement rocks over Pliocene Upland gravels (Fleming, 1994) (McCartan, 1990). One exposure of these two faults is still preserved along Adams Mill road as a special monument (Prowell, 1983). Based on our field reconnaissance with USGS researchers, future additional investigations are planned by the USGS to further investigate the age of the gravels and lateral continuity of the National Zoo faults.

All information on timing of displacement of the National Zoo faults was available and incorporated into the EPRI seismic source models in 1986. Although later detailed mapping of these thrust faults with the Rock Creek shear zone was published after completion of the EPRI study (EPRI, 1986), Darton (Darton, 1950) and Prowell (Prowell, 1983) identified these faults as active during Cenozoic time. In addition, there is no pre-EPRI or post-EPRI seismicity spatially associated with this fault zone. Therefore, the conclusion is that the National Zoo faults are not a capable tectonic source. There also is no new published geologic information developed since 1986 that would require a significant revision to the EPRI seismic source model.

2.5.1.1.4.4.4.5 Hillville Fault Zone

The Hillville fault zone of Hansen (1978) approaches to within 5 mi (8 km) of the site in the subsurface (Figure 2.5-25, Figure 2.5-26, and Figure 2.5-27). The 26 mi (42 km) long, northeast-striking fault zone is composed of steep southeast-dipping reverse faults that align with the east side of the north-to northeast-trending Sussex-Currioman Bay aeromagnetic anomaly (i.e. SGA, Figure 2.5-22). Based on seismic reflection data, collected about 9 mi (15 km) west-southwest of the site, the fault zone consists of a narrow zone of discontinuities that vertically separate basement by as much as 250 ft (76 m) (Hansen, 1978).

The Hillville fault zone delineates a possible Paleozoic suture zone reactivated in the Mesozoic and Early Tertiary. The fault zone is interpreted as a lithotectonic terrane boundary that separates basement rocks associated with Triassic rift basins on the west from low-grade metamorphic basement on the east (i.e., Sussex Terrane/Taconic suture of Glover and Klitgord, (Glover, 1995a) (Figure 2.5-17) (Hansen, 1986). The apparent juxtaposition of the Hillville fault zone with the Sussex-Currioman Bay aeromagnetic anomaly suggests that the south flank of the Salisbury Embayment may be a zone of crustal instability that was reactivated during the Mesozoic and Tertiary. Cretaceous activity is inferred by Hansen (Hansen, 1978) who extends the fault up into the Cretaceous Potomac Group. The resolution of the geophysical data does not allow an interpretation for the upward projection of the fault into younger overlying Coastal Plain deposits (Hansen, 1978). Hansen (Hansen, 1978), however, used stratigraphic correlations of Coastal Plain deposits from borehole data to speculate that the Hillville fault may have been active during the Early Paleocene.

There is no geologic data to suggest that the Hillville fault is a capable tectonic source. Field and aerial reconnaissance, coupled with interpretation of aerial photography and LiDAR data (see Section 2.5.3.1 for additional information regarding the general methodology), conducted during this COL study shows that there are no geomorphic features indicative of potential Quaternary activity along the surface-projection of the Hillville fault zone. A review of geologic cross sections (McCartan, 1989a) (McCartan, 1989b) (Glaser, 2003b) (Glaser, 2003c) show south-dipping Lower to Middle Miocene Calvert Formation and no faulting along projection with the Hillville fault zone. Furthermore Quaternary terraces mapped by McCartan (McCartan, 1989b) and Glaser (Glaser, 2003b) (Glaser, 2003c) bordering the Patuxent and Potomac Rivers were evaluated for features suggestive of tectonic deformation by interpreting LiDAR data and aerial reconnaissance (Figure 2.5-26 and Figure 2.5-27). No northeast-trending linear features coincident with the zone of faulting were observed where the surface projection of the fault intersects these Quaternary surfaces. Aerial reconnaissance of this fault zone also demonstrated the absence of linear features coincident or aligned with the fault zone. Lastly, interpretation of the detailed stratigraphic profiles collected along Calvert Cliffs and the western side of Chesapeake Bay provide geologic evidence for no expression of the fault where the projected fault would intersect the Miocene-aged deposits (Kidwell, 1997; see Section 2.5.3 for further explanation). Therefore, we conclude that the Hillville fault zone is not a capable tectonic source, and there is no new information developed since 1986 that would require a significant revision to the EPRI model.

2.5.1.1.4.4.4.6 Unnamed Fault beneath Northern Chesapeake Bay, Cecil County, Maryland

Pazzaglia (1993) proposed a fault in northern Chesapeake Bay that comes to within 70 mi (113 km) north of the site (Figure 2.5-25). On the basis of geologic data and assuming that the bay is structurally controlled, Pazzaglia (1993) infers a 14 mi (23 km) long, northeast-striking fault with
RAI 71 02.05.01-19 a ~~southwest-side~~southeast-side up sense of displacement. Near the mouth of the Susquehanna River, in Maryland, the unnamed fault is interpreted to vertically separate Pleistocene Turkey Point gravels of the Quaternary Pennsauken Formation on the east at elevations higher than a similar gravel deposit mapped on the west side of the Chesapeake Bay. The amount of apparent vertical separation is unconstrained because the base of the gravel unit is not exposed west of the bay; however, estimates of the exposed section provide a minimum of 26 ft (8 m) of vertical separation of the Pleistocene Turkey Point gravels (Pazzaglia, 1993).

This fault is unconfirmed based on the lack of direct supporting evidence. First, the fault has not been observed as a local discontinuity on land. Second, the correlation of gravels is permissible based on the data, but has not been confirmed by detailed stratigraphic or chronologic studies. Geologic mapping of the area (Higgins, 1986) shows Miocene Upland gravels along the northeast mouth of the Susquehanna River where Pazzaglia (Pazzaglia, 1993) maps the Quaternary Pennsauken Formation.

There is no geologic data to suggest that this unnamed fault zone is a capable tectonic source. There is no pre-EPRI or post-EPRI seismicity spatially associated with this fault zone. Field and aerial reconnaissance conducted to support CCNPP Unit 3 shows that there are no geomorphic features indicative of potential Quaternary activity along the surface-projection of the unnamed fault; therefore, this fault is not a capable tectonic source.

2.5.1.1.4.4.4.7 Unnamed Monocline beneath Chesapeake Bay

McCartan (McCartan, 1995) show east-facing monoclinal structures bounding the western margin of Chesapeake Bay 1.8 and 10 mi (2.9 and 16 km) east and southeast, respectively, of the site (Figure 2.5-25). Also, McCartan (McCartan, 1995) interprets an east-facing monocline about 10 mi (16 km) west of the site. The three monoclinal structures are depicted on two cross sections as warping Lower Paleocene to Upper Miocene strata with approximately 60 to 300 ft (18 to 91 m) of relief. The monoclines exhibit a west-side up sense of structural relief that projects upward into the Miocene Choptank Formation (McCartan, 1995). The overlying Late Miocene St. Marys Formation is not shown as warped. Boreholes shown with the cross sections accompanying the McCartan (MaCartan, 1995) map provide the only direct control on cross section construction. The boreholes are widely spaced and do not appear to provide a constraint on the existence and location of the warps. No borehole data is available directly west of the cliffs and within the bay to substantiate the presence of the warp. No surface trace or surface projection of the warps is indicated on the accompanying geologic map. Based on text accompanying the map and cross sections, we infer that the cross sections imply two approximately north- to northeast-striking, west-side up structures, of presumed tectonic origin.

McCartan (McCartan, 1995) interpret the existence of the monocline based on three observations in the local landscape. Firstly, the north to northeast-trending western shore of Chesapeake Bay within Calvert County is somewhat linear and is suggestive of structural control (McCartan, 1995). Secondly, land elevation differences west and east of Chesapeake Bay are on the order of 90 ft (27 m), with the west side being significantly higher in elevation, more fluvially dissected, and composed of older material compared to the east side of Chesapeake Bay. On the west side of the bay, the landscape has surface elevations of 100 to 130 ft (30 to 40 m) msl and drainages are incised into the Pliocene Upland Deposits and Miocene-aged deposits of the St. Mary's, Choptank, and Calvert Formations. Along the eastern shoreline of the Delmarva Peninsula, surface elevations are less than 20 to 30 ft (6 to 9 m) msl and the surface exhibits minor incision and a more flat-lying topographic surface. These eastern shore deposits are mapped as Quaternary estuarine and deltaic deposits. Thirdly, variations in unit thickness within Tertiary deposits between Calvert Cliffs and Delmarva Peninsula are used to infer the presence of a warp. Based on these physiographic, geomorphic and geologic observations, McCartan (McCartan, 1995) infer the presence of a fold along the western shore of Chesapeake Bay (Figure 2.5-25).

Based on the paucity of geologic data constraining the cross sections of McCartan (McCartan, 1995), the existence of the monocline is speculative. The borehole data that constrain the location of the monocline are approximately 18 to 21 mi (29 to 34 km) apart and permit, but do not require the existence of a monocline. McCartan (McCartan, 1995) do not present additional data that are inconsistent with the interpretation of flat-lying, gently east-dipping Miocene strata shown in prior published cross sections north and south of this portion of Chesapeake Bay (Cleaves et al., 1968; Milici, et al., 1995) and within Charles and St. Mary's Counties, Maryland (McCartan, 1989a) (McCartan, 1989b) (DM, 1973). No geophysical data are presented as supporting evidence for this feature. In contrast, shallow, high-resolution geophysical data collected along the length of Chesapeake Bay to evaluate the ancient courses of the submerged and buried Susquehanna River provide limited evidence strongly indicating that

Tertiary strata are flat lying and undeformed along the western shore of Chesapeake Bay (Colman, 1990) (Figure 2.5-29).

Alternatively, the change in physiographic elevation and geomorphic surfaces between the western and eastern shores of Chesapeake Bay can be explained by erosional processes directly related to the former course of the Susquehanna River, coupled with eustatic sea level fluctuations during the Quaternary (Colman, 1990) (Owens, 1979). Colman and Halka (Colman, 1989) also provide a submarine geologic map of Chesapeake Bay at and near the site which depicts Tertiary and Pleistocene deposits interpreted from high-resolution geophysical profiles. No folding or warping or faulting is depicted on the Colman and Halka (Colman, 1989) map which encompasses the warp of McCartan (McCartan, 1995). Colman (Colman, 1990) utilize the same geophysical data to track the former courses of the Susquehanna River between northern Chesapeake Bay and the southern Delmarva Peninsula. Paleo-river profiles developed from the geophysical surveys that imaged the depth and width of the paleochannels show that the Eastville (150 ka) and Exmore (200 to 400 ka) paleochannels show no distinct elevation changes within the region of the Hillville fault and McCartan (McCartan, 1995) features.

RAI 71 02.05.01-20

~~Field reconnaissance along much of the western shoreline shows that the north- to northeast-trending linear coastline could be controlled locally, in part, by a weak, poorly-developed, sub-vertical joint set oriented subparallel to the coast (Section 2.5.1.2.4). The observation that the west side of Chesapeake Bay is elevated and dissected, and that approximately 37 ka estuarine deposits are approximately 6 feet above sea level is compelling evidence for recent (late Quaternary) uplift. Similar elevated, dissected topography and approximately 37 ka estuarine deposits are observed over broad portions of the Coastal Plain along the eastern seaboard east and west of Chesapeake Bay. These surfaces of apparent anomalous elevations have recently been attributed to the presence of a glacial fore-bulge developed outboard of the Laurentide ice sheet (Scott, 2006).~~

There is no geologic data to suggest that the postulated monocline along the western margin of Chesapeake Bay of McCartan (McCartan, 1995), if present, is a capable tectonic source. Field and aerial reconnaissance, coupled with interpretation of aerial photography and LiDAR data (see Section 2.5.3.1 for additional information regarding the general methodology), conducted during this COL study, shows that there are no geomorphic features indicative of folding directly along the western shores of Chesapeake Bay. There is no pre-EPRI or post-EPRI seismicity spatially associated with this structure. These data indicate that the McCartan (McCartan, 1995) warps, if present, most likely do not deform Pliocene to Quaternary deposits, and thus are not capable tectonic sources that would require a revision to the EPRI (1986) seismic source model.

2.5.1.1.4.4.4.8 Unnamed Folds and Postulated Fault within Calvert Cliffs, Western Chesapeake Bay, Calvert County, Maryland

The Calvert Cliffs along the west side of Chesapeake Bay provide a 25 mile (40 km) long nearly continuous exposure of Miocene, Pliocene and Quaternary deposits (Figure 2.5-26). Kidwell (1988 and 1997) prepared over 300 comprehensive lithostratigraphic columns along a 25 mi (40 km) long stretch of Calvert Cliffs (Figure 2.5-30). Because of the orientation of the western shore of Chesapeake Bay, the cliffs intersect any previously potential structures (i.e., Hillville fault) trending northeast or subparallel to the overall structural trend of the Appalachians. The cliff exposures provide a 230 ft (70 m) thick section of Cenozoic deposits that span at least 10 million years of geologic time.

On the basis of the stratigraphic profiles, Kidwell (Kidwell, 1997) develops a chronostratigraphic sequence of the exposed Coastal Plain deposits and provides information on regional dip and lateral continuity. The Miocene Choptank Formation is subdivided into two units and is unconformably overlain by the St .Marys Formation. The St. Marys Formation is subdivided into three subunits each of which is bound by a disconformity. The youngest subunit is unconformably overlain by the Pliocene Brandywine Formation (i.e., Pliocene Upland gravels). The exposed Coastal Plain deposits strike northeast and dip south-southeast between 1 and 2 degrees. The southerly dip of the strata is disrupted occasionally by several low amplitude broad undulations in the Choptank Formation, and decrease in amplitude upward into the St Marys Formation (Figure 2.5-30). Kidwell (Kidwell, 1997) interprets the undulations as monoclines and asymmetrical anticlines. The undulations typically represent erosional contacts that have wavelengths on the order of 2.5 to 5 mi (4 to 8 km) and amplitudes of 10 to 11 ft (about 3 m). Any inferred folding of the overlying Pliocene and Quaternary fluvial strata is very poorly constrained or obscured because of highly undulatory unconformities within these younger sand and gravel deposits. For instance, the inferred folding of the overlying Pliocene and Quaternary channelized sedimentary deposits consist of intertidal sand and mud-flats, tidal channels and tidally-influenced rivers exhibit as much as 40 ft (12 m) of erosional elevation change (Figure 2.5-30).

~~Near Moran Landing, a~~About 1.2 mi (1.9 km) south of the site, Kidwell (Kidwell, 1997) interprets an apparent 6 to 10 ft (2 to 3 m) elevation change in Miocene strata by extrapolating unit contacts across the approximatley 0.6 mile wide (1 km) gap at Moran Landing (Figure 2.5-25 and Figure 2.5-30). Kidwell (Kidwell, 1997) also interprets~~, and~~ a 3 to 12 ft (1 to 3.6 m) elevation change in younger ~~Pliocene and~~ (Quaternary(?)) fluvial material across the same gap. Because of the lack of cliff exposures at Moran Landing (only the valley margins), no direct observations of these elevation changes can be made.~~(Figure 2.5-26 and Figure 2.5-30).~~ Kidwell (Kidwell,1997) ~~infers the presence of a fault to~~ explains the differences in elevation ~~of strata across~~of the Miocene-Quaternary stratigraphy by hypothesizing the existence of a fault at Moran Landing that strikes northeast and accomodates a north-side down sense of separation. However, ~~T~~the postulated fault of Kidwell (Kidwell, 1997) is not shown on any of ~~the~~ Kidwell's (Kidwell, 1997) cross-sections, or any published geologic map(e.g., Glaser, 2003b and 2003c). In addition, Hansen (Hansen, 1978) does not describe faulting in seismic reflection lin St. M-2 that intersects the inferred southwest projection of the hypothesized Kidwell (Kidwell, 1997) fault
RAI 71 02.05.01-27 (Figure 2.5-27).~~; however, the inferred location is approximately 1.2 mi (1.9 km) south of the CCNPP site. The hypothesized fault is not exposed in the cliff face and is based entirely on a change in elevation and bedding dip of Miocene stratigraphic boundaries projected across the fluvial valley of Moran Landing. Kidwell (Kidwell, 1997) postulates that the fault strikes northeast and exhibits a north-side down sense of separation across all the geologic units (Miocene through Quaternary). With regard to the apparent elevation changes for the Pliocene and Quaternary unconformities, these can be readily explained by channeling and highly irregular erosional surfaces (Figure 2.5-30).~~

~~LiDAR data was reviewed for the possible presence of northeast-striking lineaments in the region of Moran Landing and to the southeast along the Patuxent River. Field and aerial reconnaissance, coupled with interpretation of aerial photography and LiDAR data (see Section 2.5.3.1 for additional information regarding the general methodology), conducted during the CCNPP Unit 3 investigation shows that there are no geomorphic features indicative of potential Quaternary activity developed in the Pliocene-Quaternary surfaces along a southeast projection from Chesapeake Bay across the Patuxent and Potomac Rivers (Figure 2.5-26). The features also do not coincide with magnetic and gravity anomalies, and thus are not rooted, and more likely are surficial in origin. There is no pre-EPRI or post-EPRI (1986) seismicity spatially associated with the Kidwell (Kidwell, 1997) features, nor are there direct geologic data to indicate that the features proposed by Kidwell (Kidwell, 1997) are capable tectonic sources (Section 2.5.3.2.3).~~

The observations of offset younger gravels do not provide any evidence for the existence of a fault because the surface on which the gravels are deposited is an erosional unconformity with extensive variable relief (Kidwell, 1997). Observations made during field reconnaissance, as part of the FSAR preparation, confirmed that this contact was an erosional unconformity with significant topography north and south of Moran Landing consistent with stratigraphic representations in Kidwell (1997) profiles. The observations of several feet of elevation change in the Miocene units over several thousands of feet of horizontal distance is at best weak evidence for faulting within the Miocene deposits. For example, subtle elevation variations in Miocene strata characterized along a near-continuous exposure south of Moran Landing contain similar vertical and lateral dimensions as to the inferred elevation change across Moran
RAI 71 02.05.01-27 Landing; however, the features are interpreted as subtle warps and not faults by Kidwell (1997). On the basis of association with similar features to the south and the lack of a continuous exposure, there is little to no evidence to support a fault across Moran Landing. The lack of evidence for Quaternary faulting within the observations made by Kidwell (Kidwell, 1997), and the results of the studies undertaken as part of the CCNPP Unit 3 COLA effort (field and aerial reconnaissance, air photo and LiDAR analysis) (see FSAR Section 2.5.3.1), collectively support the conclusion that the hypothesized fault of Kidwell (Kidwell, 1997) is not a capable fault.

2.5.1.1.4.4.5 Quaternary Tectonic Features

In an effort to provide a comprehensive database of Quaternary tectonic features, Crone and Wheeler (Crone, 2000), Wheeler (Wheeler, 2005), and Wheeler (Wheeler, 2006) compiled geological information on Quaternary faults, liquefaction features, and possible tectonic features in the CEUS. Crone and Wheeler (Crone, 2000) and Wheeler (Wheeler, 2005) evaluated and classified these features into one of four categories (Classes A, B, C, and D; see Table 2.5-1 for definitions (Crone, 2000) (Wheeler, 2005)) based on strength of evidence for Quaternary activity.

Within a 200 mi (322 km) radius of the CCNPP site, Crone and Wheeler (Crone, 2000), Wheeler (Wheeler, 2005) and Wheeler (Wheeler, 2006) identified 17 potential Quaternary features (Figure 2.5-31). Work performed as part of the CCNPP Unit 3 investigation, including literature review, interviews with experts, and geologic reconnaissance, did not identify any additional potential Quaternary tectonic features within the CCNPP site region, other than those previously mentioned (McCartan, 1995) (Kidwell, 1997). Within approximately 200 mi (322 km) of the site, Crone and Wheeler (Crone, 2000) found only one feature described in the literature that exhibited potential evidence for Quaternary activity (Figure 2.5-31). This feature (shown as number 12) is the paleo-liquefaction features within the Central Virginia seismic zone.

The following sections provide descriptions of 15 of the 17 potential Quaternary features identified by Crone and Wheeler (Crone, 2000), Wheeler (Wheeler, 2005) (Wheeler, 2006), and of the postulated East Coast fault system of Marple and Talwani (Marple, 2004). Note that the Central Virgina and Lancaster seismic zones are discussed in Section 2.5.1.1.4.5 and Section 2.5.2. Out of the 17 features evaluated for this CCNPP Unit 3 study, nearly all are classified as Class C features, with the exception of the Central Virginia seismic zone (Class A).

The features are labeled with the reference numbers utilized in Figure 2.5-31:

1. Fall lines of Weems (1998) (Class C)
2. Ramapo fault system (Class C)
3. Kingston fault (Class C)
4. New York Bight fault (offshore) (Class C)

5. Cacoosing Valley earthquake (Class C)
6. Lancaster seismic zone (Class C)
7. New Castle County faults (Class C)
8. Upper Marlboro faults (Class C)

RAI 130 02.05.01-51

9. Everona-fault and Mountain Run fault zone (Class C)
10. Stafford fault of Mixon et al. (Class C)
11. Lebanon Church fault (Class C)
12. Central Virginia seismic zone (Class A)
13. Hopewell fault (Class C)
14. Old Hickory faults (Class C)
15. Stanleytown-Villa Heights faults (Class C)
16. (The Stafford fault system of Marple is included in (17), i.e. the East Coast fault system)
17. East Coast fault system (Class C)

RAI 130 02.05.01-51

~~The Everona-Mountain Run fault zone and Stafford fault of Mixon (Mixon, 2000) also are discussed in detail in previous Section 2.5.1.1.4.4.2 and Section 2.5.1.1.4.4.4.1.~~

2.5.1.1.4.4.5.1 Fall Lines of Weems (1998)

In 1998, Weems defined seven fall lines across the Piedmont and Blue Ridge Provinces of North Carolina and Virginia (Figure 2.5-31). The eastern fall line is located approximately 47 mi (76 km) west of the CCNPP site. The fall lines, not to be confused with the Fall Line separating the Piedmont and Coastal Plain provinces, are based on the alignment of short stream segments with anomalously steep gradients. Weems (1998) explores possible ages and origins (rock hardness, climatic, and tectonic) of the fall lines and "based on limited available evidence favors a neo-tectonic origin" for these geomorphic features during the Quaternary. Weems (1998) interprets longitudinal profiles for major drainages flowing primarily southeast and northwest across the Piedmont and Blue Ridge Provinces to assess the presence and origin of the "fall zones".

A critical evaluation of Weems' (1998) study, as part of the North Anna ESP, demonstrates that there are inconsistencies and ambiguities in Weems' (1998) correlations and alignment of steep reaches of streams used to define continuous fall lines (Dominion, 2004b). The North Anna ESP study concludes that that the individual fall zones of Weems (1998) may not be as laterally continuous as previously interpreted. For instance, stratigraphic, structural and geomorphic relations across and adjacent to the Weems (1998) fall zones can be readily explained by differential erosion due to variable bedrock hardness rather than Quaternary tectonism (Dominion, 2004b). Furthermore, there is no geomorphic expression of recent tectonism, such as the presence of escarpments, along the trend of the fall lines between drainages where one would expect to find better preservation of tectonic geomorphic features. Similarly, Wheeler (2005) notes that the Weems (1998) fall zones are not reproducible and are subjective, thus tectonic faulting is not yet demonstrated as an origin, and the fall lines are designated as a Class C feature. In the Safety Evaluation Report for the North Anna ESP site study, the NRC staff agrees with the assessment that the fall lines of Weems (1998) are nontectonic features (NRC, 2005). In summary, based on review of published literature, field reconnaissance, and geologic and geomorphic analysis performed previously for the North Anna ESP application, the fall lines of Weems (1998) are erosional features related to contrasting erosional resistances of adjacent rock types, and are not tectonic in origin, and thus are not capable tectonic sources.

RAI 130 02.05.01-51

2.5.1.1.4.4.5.2 Everona-Fault and Mountain Run Fault Zone

The Mountain Run fault zone is located ~~along the eastern margin of the Culpeper Basin and lies~~ approximately 71 mi (114 km) southwest of the site (~~Figure~~ Figure 2.5-9~~2.5-17 and Figure 2.5-31~~). The 75 mi (121 m) long, northeast-striking fault zone is mapped from the southeastern margin of the Triassic Culpeper Basin near the Rappahannock River southwestward to near Charlottesville, in the western Piedmont of Virginia (Pavlides, 1986) (Horton, 1991). The fault zone consists of a broad zone of sheared rocks, mylonites, breccias, phyllonites, and phyllites ~~of variable width~~up to 2.5 to 3 mi (4 to 5 km) wide (Pavlides, 1989) (Crone, 2000) (Mixon, 2000). Within this broad fault zone are three features that have been identified by Crone and Wheeler (Crone, 2000) as having possible Quaternary tectonic activity. From northeast to southwest, these are: (1) the northwest-facing, 1-mi- (1.6-km-) long Kelly's Ford scarp, (2) the northwest-facing, 7-mi- (11-km-) long Mountain Run scarp, and (3) the northwest-dipping fault exposed near the town of Everona, Virginia, named informally the Everona fault (Pavlides, 1983) (Pavlides, 1986) (Pavlides, 1994) (Crone, 2000) (Mixon, 2000) (Figure 2.5-31)..

RAI 130 02.05.01-51

The Mountain Run fault zone is interpreted to have formed initially as a thrust fault upon which back-arc basin rocks (mélange deposits) of the Mine Run Complex were accreted onto ancestral North America at the end of the Ordovician (Pavlides, 1989). This major structure separates the Blue Ridge and Piedmont terranes (Pavlides, 1983) (Figure 2.5-9, Figure 2.5-16, and Figure 2.5-17). Subsequent reactivation of the fault during the Paleozoic and/or Mesozoic produced strike-slip and dip-slip movements. Horizontal slickenside~~s~~ lineations within phyllite found in borehole samples beneath the alluvium-filled valley of Mountain Run~~and at several places near the base of the Mountain Run scarp~~ suggest strike-slip movement, whereas small scale folds in the uplands near the scarp suggest an oblique dextral sense of slip (Pavlides, 2000). The timing of the reverse and strike-slip histories of the fault zone, and associated mylonitization and brecciation, is constrained to be pre-Early Jurassic, based on the presence of undeformed Early Jurassic diabase dikes that cut rocks of the Mountain Run fault zone (Pavlides, 2000). The northern portion of the Mountain Run fault zone bounds the southeastern margin of the Culpeper basin (Mixon, 2000) (Figure 2.5-9 and Figure 2.5-10), indicating that the fault locally has been active since the Triassic (Crone, 2000) (Section 2.5.1.1.4.4.3).

RAI 130 02.05.01-51

RAI 130 02.05.01-51

~~The~~Two features within the northeast-striking Mountain Run fault zone ~~is~~are moderately to well-expressed geomorphically (Pavlides, 2000). Two northwest-facing scarps occur along the fault zone, including: (1) the 1 mi (1.6 km) long Kelly's Ford scarp located directly northeast of the Rappahannock River and; (2) the 7 mi (11 km) long Mountain Run scarp located along the southeast margin of the linear Mountain Run drainage. The presence of these two locally c~~C~~onspicuous bedrock scarps in the Piedmont, an area characterized by deep weathering and subdued topography, has led some experts to suggest that the scarps formed due to~~fault has experienced~~ a Late Cenozoic phase of movement within the Mountain Run fault zone (Pavlides, 2000) (Pavlides, 1983).

RAI 130 02.05.01-51

RAI 130 02.05.01-51

~~Near Everona, Virginia, a small reverse fault, found in an excavation, vertically displaces "probable Late Tertiary" gravels by 5 ft (1.5 m) (Pavlides, 1983). The fault strikes northeast, dips 20 degrees northwest, and based on kinematic indicators is an oblique strike-slip fault. More recently others have estimated that the offset colluvial gravels are Pleistocene age (Manspeizer et. al, 1989). The Everona fault is located about 0.5 mi (0.8 km) west of the Mountain Run fault zone. Due to the close proximity of these two faults and their shared similar orientation and sense of slip, the Everona and Mountain Run faults are considered to be part of the same fault~~

RAI 130 02.05.01-51

~~zone, hence the Everona-Mountain Run fault zone (Crone, 2000). Crone and Wheeler (Crone, 2000) assessed that the faulting at Everona is likely to be of Quaternary age, but because the likelihood has not been tested by detailed paleo-seismological or other investigations, this feature was assigned to Class C.~~

RAI 130 02.05.01-51

RAI 130 02.05.01-51

RAI 130 02.05.01-51

Field and aerial reconnaissance, and geomorphic analysis of deposits and features associated with the fault zone, recently performed for the North Anna ESP provide new information on the ~~absence of Quaternary faulting~~Mountain Run and Kelly's Ford scarps in particular, and ~~along~~ the ~~Everona-~~Mountain Run fault zone in general (Dominion, 2004a). In response to NRC comments for the North Anna ESP, geologic cross sections and topographic profiles were prepared along the Mountain Run fault zone across and between the Mountain Run and Kelly's Ford scarps to further evaluate the inferred tectonic geomorphology coincident with the fault zone first proposed by Pavlides (1986). The results of the additional analysis were presented in the response to an NRC Request for Additional Information (RAI) (Dominion, 2004a) and demonstrated that the Mountain Run and Kelly's Ford scarps are probably a result of a differential erosion and not late Cenozoic tectonic activity. Three main findings from the Dominion (2004a) study are summarized below:

- There is no consistent expression of a scarp along the Mountain Run fault zone in the vicinity of the Rappahannock River. The northwest-facing Kelly's Ford scarp is similar to a northwest-facing scarp along the southeastern valley margin of Mountain Run; both scarps were formed by streams that preferentially undercut the southeastern valley walls, creating asymmetric valley profiles.

RAI 130 02.05.01-51

- There is no northwest-facing scarp associated with the 10 mile (16 km) long portion of the Mountain Run fault zone between the Rappahannock and Rapidan Rivers (i.e., between the Kelly's Ford and Mountain Run scarps). Undeformed late Neogene colluvial deposits bury the Mountain Run fault zone in this region, demonstrating the absence of Quaternary fault activity.

RAI 130 02.05.01-51

- The northwest-facing "Mountain Run" scarp southwest of the Rappahannock River alternates with a southeast-facing scarp on the opposite side of Mountain Run valley; both sets of scarps have formed by the stream impinging on the edge of the valley.

RAI 130 02.05.01-51

Near Everona, Virginia, a small reverse fault, found in an excavation, vertically displaces a "probable Late Tertiary" or "Pleistocene" gravel layer by 5 ft (1.5 m) (Pavlides, 1983) (Manspeizer, 1989) (Crone, 2000). The fault strikes northeast and dips between about 55 to 20 degrees northwest, shallowing up-dip (Manspeizer, 1989) (Crone, 2000). This isolated fault exposure, called the Everona fault by Crone and Wheeler (Crone, 2000), is located about 0.4 mi (0.6 km) northwest of the Mountain Run scarp and is within but near the northwest margin of the Mountain Run fault zone (Pavlides, 1983) (Mixon, 2000). There is no geomorphic expression associated with the exposure (Crone, 2000). The CCNPP Unit 3 investigation did not reveal additional investigations of the Everona fault since the initial exposure was documented in 1983 (Pavlides, 1983). Crone and Wheeler (Crone, 2000) assessed that the faulting at Everona is likely to be of Quaternary age, but because the likelihood has not been tested by detailed paleoseismological or other investigations, this feature was assigned to Class C.

All of the basic information on the style and timing of displacement of the Everona~~Mountain Run~~ fault ~~zone and associated faults~~ was available ~~and incorporated in~~to the EPRI ~~seismic source models~~SOG team in 1986. Significant new information developed since 1986 includes the work performed for the North Ana ESP that shows the Mountain Run fault zone in the vicinity of the Kelly's Ford and Mountain Run scarps has not been active during the Quaternary. In addition, the NRC staff agrees that the scarps along the Mountain Run Fault zone were not produced by Cenozoic fault activity (NRC, 2005). Similarly, Crone and Wheeler (Crone, 2000) do

RAI 130 02.05.01-51
RAI 130 02.05.01-51

not show the Mountain Run fault zone as a known Quaternary structure in their compilation of active tectonic features in the CEUS, having assigned it to Class C. Based on the lack of new information on the Everona fault and the findings of the previous studies performed for the North Anna ESP (Dominion, 2004a)~~and approval by the Nuclear Regulatory Commission (NRC, 2005)~~, it is concluded that the Everona-Mountain Run fault zone is not a capable tectonic source. No new information has been developed since 1986 that would require a significant revision to the EPRI seismic source model.

2.5.1.1.4.4.5.3 Stafford Fault of Mixon, et al.

RAI 134 02.05.01-58

The Stafford fault (#10 on Figure 2.5-31) approaches within 47 mi southwest of the site (Figure 2.5-25). The Stafford fault (Mixon, 2000) is discussed in more detail in Section 2.5.1.1.4.4.4.1 (Stafford Fault System). The northern extension of the Stafford fault system as proposed by Marple (#16 on Figure 2.5-31) is discussed in Section 2.5.1.1.4.4.4.1 and Section 2.5.1.1.4.4.5.14~~2.5.1.1.4.4.5.15~~. The 42 mile (68 km) long fault system strikes approximately N35°E and was identified and described first by Newell (Newell, 1976). The fault system consists of a series of five northeast-striking, northwest-dipping, high-angle reverse faults including, from north to south, the Dumfries, Fall Hill, Hazel Run, and Brooke faults, and an unnamed fault. The Brooke fault also includes the Tank Creek fault located northeast of the Brooke fault (Mixon, 2000).

No new significant information has been developed since 1986 regarding the activity of the Stafford fault system with the exception of the response to an NRC RAI for the North Anna ESP (Dominion, 2004a). Field reconnaissance performed for the CCNPP Unit 3 study also did not reveal any geologic or geomorphic features indicative of potential Quaternary activity along the fault system. In addition, near the site and along the portion of the Stafford fault mapped by Mixon et al. (2000) no seismicity is attributed to the Stafford fault. Similarly, Wheeler (Wheeler, 2005) does not show the Stafford fault system as a Quaternary structure in his compilation of active tectonic features in the CEUS. The NRC (NRC, 2005) agreed with the findings of the subsequent study for the North Anna ESP, and stated: "Based on the evidence cited by the applicant, in particular the applicant's examination of the topography profiles that cross the fault system, the staff concludes that the applicant accurately characterized the Stafford fault system as being inactive during the Quaternary Period." Based on a review of existing information for the Stafford fault system, including the response to the NRC RAI for the North Anna ESP, the Stafford fault system is not a capable tectonic source and there is no new information developed since 1986 that would require a significant revision to the EPRI seismic source model.

2.5.1.1.4.4.5.4 Ramapo Fault System

RAI 134 02.05.01-61

~~The Ramapo fault system is located in northern New Jersey and southern New York State, approximately 130 mi (209 km) north-northeast of the CCNPP site (Figure 2.5-31). This fault system consists of northeast-striking, southeast-dipping, normal faults that bound the northwest side of the Mesozoic Newark basin that to the northeast become a single 40 mi (64 km) long northeast-striking fault (Ratcliffe, 1971) (Schlische, 1992) (Drake, 1996) (Figure 2.5-10). Bedrock mapping by Drake (Drake, 1996) shows primarily northwest-dipping Lower Jurassic and Upper Triassic Newark Supergroup rock in the hanging wall and tightly folded and faulted Paleozoic basement rocks in the footwall of the fault. The Ramapo fault splays into several fault strands southwest of Bernardsville and merges with the Flemington Fault zone. This fault zone also splays into several northeast- to east-trending faults in Rockland and Westchester Counties, New York.~~

~~The Ramapo fault system has been considered a potentially active tectonic feature because the fault: (1) exhibits repeated reactivation during the Paleozoic, (2) bounds the Mesozoic Newark~~

~~basin (i.e. the region is composed of extended crust), and (3) aligns with earthquake epicenters (Wheeler, 2006) (Aggarwal, 1978). In cross section and map view, the seismicity data and focal mechanisms illustrate a 60° to 65° southeast-dipping fault zone that projects upward to the mapped trace of the Ramapo fault. In addition, 14 focal mechanism solutions have orientations that are consistent with the present-day stress field and suggest reverse reactivation of the Ramapo fault. Collectively, these data led Aggarwal and Sykes (Aggarwal, 1978) to conclude that the Ramapo fault is likely active.~~

~~Many of the assumptions and conclusions made by Aggarwal and Sykes (Aggarwal, 1978) were later reevaluated with alternative interpretations suggesting the fault probably has not been active during the Quaternary. Subsequent fault activity studies included several types of geophysical and geologic techniques. First, a modified velocity model and a carefully re-evaluated earthquake catalog refined the location of the earthquakes previously inferred as aligned with the Ramapo fault, and demonstrated that approximately half of the reported earthquakes occur near the margins of the Newark Basin, far from the Ramapo fault, but still within the Ramapo fault system proper (Kafka, 1985) (Thurber, 1985) (Wheeler, 2006). In addition, a reassessment of the eastern U.S. stress field demonstrated that the present-day stress field is oriented east-southeast (Zoback, 1989a), which would be inconsistent with the previously inferred reverse reactivation of the fault. Kinematic analysis of fault zone samples collected from deep exploratory boreholes provides evidence that the latest style of deformation probably included extensional faulting during the Mesozoic (Ratcliffe, 1980) (Ratcliffe, 1982) (Burton, 1985) (Ratcliffe, 1990). The borehole data also confirm that the dip of the Ramapo fault is 10° to 15° shallower than inferred by Aggarwal and Sykes (Aggarwal, 1978).~~

RAI 134 02.05.01-61

~~In summary, several papers infer that evidence for Quaternary deformation exists near the Ramapo fault zone (Nelson, 1980) (Newman, 1983) (Newman, 1987) (Kafka, 1989); however, Crone and Wheeler (Crone, 2000) and Wheeler (Wheeler, 2006) argue convincingly that none of the data used to infer seismic slip can be used to differentiate seismic from asesimic slip. Additionally, trenches excavated across the up-dip projection of the fault zone revealed no evidence for Quaternary faulting (Stone, 1984) (Ratcliffe, 1990). Besides the presence of microseismicity within the vicinity of the Ramapo fault zone, there is no clear evidence of Quaternary tectonic faulting (Crone, 2000) (Wheeler, 2006), thus the Ramapo fault system is assigned a Class C designation by Crone and Wheeler (Crone, 2000). The Ramapo fault zone was a known structure for the EPRI study (EPRI, 1986). Based on the review of post-EPRI literature and seismicity, there is no new information developed since 1986 that would require a significant change to the EPRI seismic source model.~~ The Ramapo fault is located in northern New Jersey and southern New York State, approximately 200 mi (320 km) north-northeast of the CCNPP site (Figure 2.5-31, Figure 2.5-216). The Ramapo fault is one segment of a system of northeast-striking, southeast-dipping, normal faults that bound the northwest side of the Mesozoic Newark basin (Figure 2.5-10, Figure 2.5-216), (Drake et al., 1996; Ratcliffe, 1971; Schlische, 1992). Bedrock mapping by Drake et al. (Drake et al., 1996) shows primarily northwest-dipping Lower Jurassic and Upper Triassic Newark Supergroup rocks in the hanging wall and tightly folded and faulted Paleozoic basement rocks in the footwall of the fault. The Ramapo fault proper extends for 50 mi (80 km) from Peakpack, NJ to the Hudson River (Ratcliffe, 1971). To the south the Ramapo fault splays into several fault strands and merges with the Flemington Fault zone. On the north side of the Hudson River the fault splays into several northeast- to east-trending faults in Rockland and Westchester Counties, New York.

The Ramapo fault first received significant attention as a potentially capable fault during the licensing process for the Indian Point nuclear power plant in the late 1970s (Aggarwal and Sykes, 1978). Due to the close proximity of a proposed strand of this fault to the plant (several miles at most) and the questions raised regarding the capability of the fault during the licensing process, a considerable amount of research has been conducted to addresses the potential capability of the fault. The vast majority of the research was conducted prior to the development of the EPRI-SOG source characterizations (EPRI, 1986-1989) that are used as the base source model for the CCNPP Unit 3 COLA (see discussion in Section 2.5.2). Therefore, much of this information was known to the EPRI-SOG teams and considered by them in the development of the existing source characterizations for the Ramapo fault (see Section 2.5.2.2.1). Of the information on the Ramapo fault that has been published since the EPRI-SOG study (e.g., Kafka and Miller, 1996; Kafka et al., 1989; Newman et al., 1987; Ratcliffe et al., 1990; Sykes et al., 2008), none has presented any information or data that requires updating of the EPRI-SOG model. The primary basis for this conclusion is the observation that none of the more recent publications provide conclusive evidence that the Ramapo and related faults are capable structures.

Interest in the Ramapo fault as a potential seismogenic fault was initially driven by the work of seismologists at what is now referred to as the Lamont Doherty Earth Observatory in New Jersey. Largely based on earthquake locations generated from local network data, these researchers noticed a spatial association between earthquakes and the Ramapo fault (e.g., Aggarwal and Sykes, 1978; Kafka et al., 1985; Page et al., 1968). The study of Page et al. (1968) used the locations of four earthquakes that they located near the Ramapo fault as the basis for concluding that the earthquakes were occurring on the Ramapo fault, and, therefore, the Ramapo was experiencing small slip events. In a later study, Aggarwal and Sykes (1978) located 33 earthquakes with magnitudes less than or equal to mb 3.3 that occurred between 1962 and 1977 within the New York - New Jersey region surrounding the Ramapo fault. Based on the locations of these earthquakes, Aggarwal and Sykes (1978) also noted a spatial association between the locations of the earthquakes and the Ramapo and related faults. Aggarwal and Sykes (1978) described this association as "leav[ing] little doubt that earthquakes in this area occur along preexisting faults" (page 426) (Aggarwal and Sykes, 1978). In particular, Aggarwal and Sykes (1978) focused on the Ramapo fault: (1) noting that over half of the 32 events plot along the Ramapo fault, and (2) concluding that that Ramapo fault is an active fault with the capability of generating large earthquakes. Aggarwal and Sykes (1978) based this conclusion on: (1) the spatial association of seismicity; (2) earthquake focal mechanisms near the Ramapo fault that show high-angle thrust faulting along roughly northeast trending faults, implying a northwest maximum compressive stress direction; and (3) earthquake hypocenters from within 10 km of the Ramapo fault surface trace that align with a dip of approximately 60°.

RAI 134 02.05.01-61

Despite the strong insistence from earlier authors that there was little doubt the Ramapo fault is active, numerous studies (e.g., Kafka et al., 1985; Quittmeyer et al., 1985; Seborowski et al., 1982; Thurber and Caruso, 1985) post-dating those of Aggarwal and Sykes (1978) and Page et al. (1968) presented revised analyses of the seismicity that contradict the earlier work and clearly demonstrate that there is considerable uncertainty as to whether or not slip on the Ramapo and related faults is causing the recorded seismicity. Seborowski et al. (1982) analyzed a sequence of aftershocks in 1980 near the northern end of the Ramapo fault close to Annesville, NY (Figure 2.5-216). Seborowski et al. (1982) demonstrated that the alignment of these earthquakes and their composite focal mechanism suggest thrusting on a north-northwest trending fault plane. This observation led Seborowski et al. (1982) to conclude that their observations are not consistent with the conclusion of Aggarwal and Sykes (1978) that the Ramapo fault is active because their slip direction, and corresponding maximum compressive stress direction, is perpendicular to that hypothesized by Aggarwal and Sykes (1978).

Quittmeyer et al. (1985) analyzed another earthquake sequence that occurred in 1983 approximately 7 miles from the sequence analyzed by Seborowski et al. (1982) and also reanalyzed one of the earthquakes used by Aggarwal and Sykes (1978) explicitly to address the discrepancy between the expected slip directions, and thus maximum compressive stress directions, of the Aggarwal and Sykes (1978) and Seborowski et al. (1982) studies. Quittmeyer et al. (1985) demonstrated two main points: (1) a composite fault plane solution for the 1983 earthquake sequence indicates thrust faulting along faults striking northwest with a maximum compressive stress direction oriented to the northeast; and (2) the earthquake analyzed by Aggarwal and Sykes (1978) has a non-unique fault plane solution that could be consistent with either the results of Aggarwal and Sykes (1978) or consistent with the fault plane solution for the 1983 earthquake sequence. Based on these observations, Quittmeyer et al. (1985) hypothesized the maximum compressive stress direction is directed roughly northeasterly and implied that the Ramapo fault is not likely a source of earthquakes within the region.

Kafka et al. (1985) presented a revised and extended seismicity catalog for the New York - New Jersey area surrounding the Ramapo fault region extending from 1974 to 1983. Kafka et al. (1985) described this compilation as an improvement over previous catalogs because the increased robustness of the network during that timeframe provides more accurate earthquake locations and uniform magnitude estimates. During this time period, Kafka et al. (1985) recorded a total of 61 earthquakes, all with magnitudes less than or equal to mbLg 3.0. Assuming that their earthquake catalog is complete down to magnitudes of mbLg > 2.0, Kafka et al. (1985) reported that 7 out of 15 earthquakes occur within 10 mi (6 km) of the Ramapo fault. Kafka et al. (1985) describe the remaining earthquakes as occurring around the outside of the Newark basin. Importantly, Kafka et al. (1985) concluded that while "much emphasis was placed on the significance of the Ramapo fault and its relationship to seismicity" (page 1279), the other seismicity occurring throughout the region suggests that "the geologic structures associated with most (if not all) earthquakes in this region are still unknown" (page 1285). In a later publication in which Kafka and Miller (1996) analyze updated seismicity with respect to geologic structures, Kafka and Miller (1996) further discredit the association between seismicity and the Ramapo fault by saying, "...the currently available evidence is sufficient to rule out ... a concentration of earthquake activity along the Ramapo fault" (page 83). RAI 134 02.05.01-61

Thurber and Caruso (1985) derived new, one- and three-dimensional crustal velocity models of the upper crust in the region of the northern Ramapo fault to provide better earthquake locations in that area. These new velocity models were considered improvements over those used in previous studies (e.g., Aggarwal and Sykes, 1978). The new models resulted in some changes in depths for the 15 earthquakes examined by Thurber and Caruso (1985). Based on their work, Thurber and Caruso (1985) concluded that: (1) there are significant lateral velocity variations within the region surrounding the Ramapo fault that can impact earthquake locations made using simple velocity models; and (2) "the Ramapo fault proper is not such a salient seismic feature in New York State, unlike the findings of Aggarwal and Sykes" (page 151). As with the Quittmeyer et al. (1985), Seborowski et al. (1982), and Kafka et al. (1985) studies, these conclusions of Thurber and Caruso (1985) indicate that there is considerable uncertainty surrounding the potential activity of the Ramapo fault.

Primarily triggered by the seismological suggestions that the Ramapo fault is active, geological investigations were also conducted to look for evidence of Quaternary slip on the Ramapo fault. The primary researcher involved in these efforts was Nicholas Ratcliffe of the U.S. Geological Survey. Ratcliffe and his colleagues' work consisted of detailed geologic mapping, seismic reflection profiling, petrographic analysis, borings and core analysis along much of the Ramapo fault and its corollary northern and southern extension (e.g., Ratcliffe, 1980, 1983, 1992; Ratcliffe and Burton, 1985, 1988; Ratcliffe et al., 1986a; Ratcliffe et al., 1986b; Ratcliffe and Costain, 1985). Much of Ratcliffe's work was explicitly focused on investigating the potential relationship between the Ramapo fault and the seismicity that had been noted in the surrounding region (e.g., Aggarwal and Sykes, 1978). The primary conclusions of the cumulative work of Ratcliffe and his colleagues' with respect to the potential for Quaternary slip on the Ramapo fault are:

- The most recent episodes of slip along the Ramapo fault, as determined from rock core samples taken across the fault, were in a normal sense with some along-strike slip motion (i.e., oblique normal faulting). Ratcliffe and others concluded that the evidence for extension across the fault as the most recent slip and the lack of compression (i.e., thrust faulting), as would be required in the modern day stress field (Zoback and Zoback, 1980; Zoback and Zoback, 1989), is evidence that the Ramapo fault has not been reactivated since the latest episode of extension in the Mesozoic.
- The Ramapo fault generally has a dip that is less than that inferred from the earthquake epicenters of Aggarwal and Sykes (1978), with the exception of the northernmost end of the fault where the dip measured from borings is approximately 70°. The implication of this observation is that earthquakes near the Ramapo fault hypothesized as being due to slip on the Ramapo fault are more likely occurring within the Proterozoic footwall rocks of the Ramapo fault.

Ratcliffe and his colleagues' results are additional evidence of the uncertainty with respect to the potential activity of the Ramapo fault because they found positive evidence for a lack of slip along the fault since the Mesozoic.

RAI 134 02.05.01-61

Most, if not all, of this geologic and seismologic information was known at the time of the EPRI-SOG study (EPRI, 1986-1989) when the seismic source characterizations that are used as the base model for CCNPP Unit 3 were developed (see Section 2.5.2). As such, the EPRI-SOG characterizations take into account uncertainty in the potential for the Ramapo fault to be a capable fault. For example, some of the EPRI-SOG Earth Science Teams explicitly characterized the Ramapo fault, and the probability of activity for the Ramapo fault given by those teams is less than 1.0 (see Section 2.5.2.2.1).

Since the research pre-dating the ERPI-SOG study, there has been some additional research on the Ramapo fault. However, none of this research has provided any additional certainty with respect to the potential for activity of the Ramapo fault. For example, a fieldtrip guidebook of Kafka et al. (1989) for the New York region briefly discusses geomorphic evidence of the Ramapo fault including valley tilting, concentrations of terraces on only one valley side, and tributary offsets as evidence of Quaternary activity along the Ramapo fault. The use of these observations of Kafka et al. (1989) as evidence supporting Quaternary activity of the Ramapo fault should be treated cautiously because:

- Kafka et al. (1989) present no data or evidence supporting these observations;
- Some of the noted geomorphic features may be older than Quaternary in age; and
- The observations themselves are not necessarily positive evidence of seismogenic, Quaternary faulting.

Newman et al. (1987; 1983) also presents observations that they interpret as evidence of Quaternary activity along the Ramapo fault. In their studies, Newman et al. (1987; 1983) constructed marine transgression curves based on radiocarbon dating of peat deposits for a series of tidal marsh sites along the Hudson River where it crosses the Ramapo fault. A total of eleven sites were investigated by Newman et al. (1987), six of which were within the Ramapo fault zone as it crosses Hudson River. Of the six sites within the Ramapo fault zone, Newman et al. (1987) report that three of the sites show a discontinuity in transgression curves that they conclude reflects Holocene normal faulting within the Ramapo fault zone. These observations and conclusions of Newman et al. (1987; 1983) are questionable with respect to the argument for Quaternary faulting along the Ramapo fault because:

- There is considerable uncertainty in the radiocarbon and elevation data used to develop the transgression curves that was not clearly taken into account in testing the faulting or no faulting hypotheses;
- The sense of motion indicated by the transgression curves (normal faulting) is contrary to the current state of stress (reverse faulting is expected);
- Trenching studies across the Ramapo fault have not revealed any evidence of Quaternary faulting (Ratcliffe et al., 1990; Stone and Ratcliffe, 1984); and
- If the inferred offsets within the transgression curves are from fault movement, there is no evidence that the movement was accommodated seismically (e.g., could have been accumulated through aseismic slip).

Finally, in an abstract for a regional Geological Society of America meeting, Nelson (1980) reported the results of pollen analysis taken from a core adjacent to the Ramapo fault near Ladentown, NY (Figure 2.5-216). In the brief abstract Nelson (1980) reports that the pollen history can be interpreted as either a "continuous, complete Holocene pollen profile suggesting an absence of postglacial seismicity along the fault" or as a pollen profile with a reversal, potentially suggesting a disruption of the infilling process caused by faulting. In summarizing his work, Nelson (1980) concludes that, "the pollen evidence is equivocal but certainly not strongly suggestive of seismicity."

RAI 134 02.05.01-61

More recently, another reanalysis of the seismicity within the region surrounding the Ramapo fault has been conducted by Sykes et al. (2008). Sykes et al. (2008) compiled a seismicity catalog extending from 1677 through 2006 for the greater New York city - Philadelphia area. This catalog contains 383 earthquakes occurring within parts of New York, Connecticut, Pennsylvania, and New Jersey (Figure 2.5-216). Of these 383 earthquakes, those occurring since 1974 are thought to have the best constraints on location due to the establishment of a more robust seismograph network at that time. Sykes et al. (2008) claim that one of the striking characteristics of their seismicity catalog is the concentration of seismicity within what they refer to as the Ramapo Seismic Zone (RSZ), a zone of seismicity approximately 7.5 mi (12 km) wide extending from the Ramapo fault to the west and from northern New Jersey north to approximately the Hudson River (Figure 2.5-216). The RSZ defined by Sykes et al. (2008) is approximately 200 mi (320 km) from the CCNPP site. All of the instrumentally located earthquakes within the RSZ have magnitudes less than mb 3.0 (Sykes et al., 2008). The only earthquake with mb > 3.0 is the historical mb 4.3 earthquake of 30 October 1783. However, uncertainty in the location of this earthquake is thought to be as much as 100 km (62 mi) (Sykes et al., 2008) raising significant suspicion as to whether the event occurred within the RSZ given the small extent of the RSZ relative to the location uncertainty.

From analyzing cross sections of the earthquakes, Sykes et al. (2008) concluded that the earthquakes within the RSZ occur within the highly deformed middle Proterozoic to early Paleozoic rocks to the west of the Mesozoic Newark basin and not the Ramapo fault proper. Figure 2.5-217 shows the Sykes et al. (2008) seismicity from the box in Figure 2.5-216 plotted along a cross section perpendicular to the Ramapo fault with the range of expected dips for the Ramapo fault (approximately 45° near the south end and 70° near the north end) (Ratcliffe, 1980; Ratcliffe and Burton, 1985). Sykes et al. (2008) specifically noted that, with the exception of three earthquakes with magnitudes less than or equal to mb 1.0 that are poorly located, earthquake hypocenters are almost vertically aligned beneath the surface trace of the Ramapo fault and not aligned with the Ramapo fault at depth (Figure 2.5-217). Instead of associating the earthquakes with the Ramapo fault. Sykes et al. (2008) attributed the observed seismicity within the RSZ to minor slip events on numerous small faults within the RSZ. However, neither Sykes et al. (2008), nor any other researchers (e.g., Kafka et al., 1985; Wheeler, 2005, 2006, 2008; Wheeler and Crone, 2001), have identified distinct faults on which they believe the earthquakes may be occurring thus preventing the characterization of any potentially active faults. Also, Sykes et al. (2008) only vaguely described the geometry of the RSZ and did not provide robust constraints on the geometry of the zone, the orientation of the potentially active faults they interpret to exist within the zone, or the maximum expected magnitude of earthquakes within the zone. As such, the Sykes et al. (2008) study presents no new information that suggests changes to the EPRI-SOG model are required to adequately represent the potential capability of the Ramapo fault or the Ramapo seismic zone.

RAI 134 02.05.01-61

A good summary of the current state of knowledge concerning the capability of the Ramapo fault is provided by Wheeler (2006). While the Wheeler (2006) paper did not consider the results of the Sykes et al. (2008) study, Wheeler's (2006) comments accurately describe the current state of knowledge concerning the capability of the Ramapo fault of RSZ. Wheeler (2006) states that: "No available arguments or evidence can preclude the possibility of occasional small earthquakes on the Ramapo fault or other strands of the fault system, or of rarer large earthquakes whose geologic record has not been recognized. Nonetheless, there is no clear evidence of Quaternary tectonic faulting on the fault system aside from the small earthquakes scattered within and outside the Ramapo fault system". The implication for the CCNPP Unit 3 site is that there is no new information to suggest that the EPRI-SOG (EPRI, 1986-1989) characterizations for the Ramapo fault do not adequately capture the current technical opinion with respect to the seismic hazard posed by the Ramapo fault or RSZ.

2.5.1.1.4.4.5.5 Kingston Fault

The Kingston fault is located in central New Jersey, approximately 175 mi (282 km) northeast of the CCNPP site (Figure 2.5-31). The Kingston fault is a 7 mi (11 km) long north to northeast-striking fault that offsets Mesozoic basement and is overlain by Coastal Plain sediments (Owens, 1998). Stanford (Stanford, 1995) use borehole and geophysical data to interpret a thickening of as much as 80 ft (24 m) of Pliocene Pennauken Formation across the surface projection of the Kingston fault. Stanford (Stanford, 1995) interprets the thickening of the Pennauken Formation gravel as a result of faulting rather than fluvial processes. Geologic cross sections prepared by Stanford (Stanford, 2002) do not show that the bedrock-Pennauken contact is vertically offset across the Kingston fault. Therefore, it seems reasonable to conclude that faulting of the Pennauken Formation is not required and that apparent thickening of the Pliocene gravels may represent a channel-fill from an ancient pre-Pliocene channel. Furthermore, Pleistocene glaciofluvial gravels that overlie the fault trace are not offset, thus indicating the fault is not a capable tectonic source (Stanford, 1995). Wheeler (Wheeler, 2006) reports that the available geologic evidence does not exclusively support a fault versus a fluvial origin for the apparent thickening of the Pennauken Formation. Wheeler (Wheeler, 2005) assigns the Kingston fault as a Class C feature based on a lack of evidence for Quaternary deformation. Given the absence of evidence for Quaternary faulting and the presence of undeformed Pleistocene glaciofluvial gravels overlying the fault trace, we conclude that the fault is not a capable tectonic feature.

2.5.1.1.4.4.5.6 New York Bight Fault

On the basis of seismic surveys, the New York Bight fault is characterized as an approximately 31 mile (50 km) long, north-northeast-striking fault, located offshore of Long Island, New York (Hutchinson, 1985) (Schwab, 1997a) (Schwab, 1997b) (Figure 2.5-31). The fault is located about 208 mi (335 km) northeast of the CCNPP site. Seismic reflection profiles indicate that the fault originated during the Cretaceous and continued intermittently with activity until at least the

RAI 130 02.05.01-52

RAI 130 02.05.01-52

Eocene. The sense of displacement is northwest-side down and displaces bedrock as much as ~~280~~357 ft (~~85~~109 m), and Upper Cretaceous deposits about ~~150~~236 ft (~~46~~72 m) (~~PSEG, 2002~~Hutchinson, 1985). High-resolution seismic reflection profiles that intersect the surface projection of the fault indicate that middle and late Quaternary sediments are undeformed within a resolution of 3 ft (1 m) (Hutchinson, 1985) (Schwab, 1997a) (Schwab, 1997b).

RAI 130 02.05.01-52

The Mesozoic New York Bight basin is located immediately east of the New York Bight fault (Hutchinson et al., 1986) (FSAR Figure 2.5-10). On the basis of seismic reflection data, Hutchinson (1986) interpret the basin to be structurally controlled by block faulting in the crystalline basement accompanied by syn-rift Mesozoic sedimentation. There is no evidence reported by Hutchinson (1986) that the basin bounding faults extend into the overlying Cretaceous sediments. Although not explicitly stated in the published literature (Hutchinson,1985)(Schwab, 1997a) (1997b), the association of the New York Bight fault along the western edge of the New York Bight basin suggests late Cretaceous through Eocene reactivation of the early Mesozoic basin bounding fault.

RAI 130 02.05.01-52

RAI 130 02.05.01-52

Only a few, poorly located earthquakes are spatially assoiciated within the vicinity of the New York Bight fault (Wheeler, 2006) (Figure 2.5-31). Wheeler (Wheeler, 2006) defines the fault as a feature having insufficient evidence to demonstrate that faulting is Quaternary and assigns the New York Bight fault as a Class C feature. Based on the seismic reflection surveys of Schwab (Schwab, 1997a) (Schwab, 1997b) and Hutchinson (1985) and the absence of Quaternary deformation, we conclude that the New York Bight fault is not a capable tectonic source.

2.5.1.1.4.4.5.7 Cacoosing Valley Earthquake Sequence

The 1993 to 1997 Cacoosing Valley earthquake sequence occurred along the eastern margin of the Lancaster seismic zone with the main shock occurring on January 16, 1994, near Reading, Pennsylvania, about 135 mi (217 km) north of the CCNPP site (Seeber, 1998) (Figure 2.5-31). This earthquake sequence also is discussed as part of the Lancaster seismic zone discussion (Section 2.5.1.1.4.5.2). The maximum magnitude earthquake associated with this sequence is an event of mbLg 4.6 (Seeber, 1998). Focal mechanisms associated with the main shock and aftershocks define a shallow subsurface rupture plane confined to the upper 1.5 mi (2.4 km) of the crust. It appears that the earthquakes occurred on a pre-existing structure striking N45°W in contrast to the typical north-trending alignment of microseismicity that delineates the Lancaster seismic zone. Seeber (Seeber, 1998) use the seismicity data, as well as the shallow depth of focal mechanisms, to demonstrate that the Cacoosing Valley earthquakes likely were caused by anthropogenic changes to a large rock quarry. Wheeler (Wheeler, 2006) defines the fault as a feature having insufficient evidence to demonstrate that faulting is Quaternary and assigns the Cacoosing Valley earthquake sequence as a Class C feature. Based on the findings of Seeber (Seeber, 1998), we interpret this earthquake sequence to be unrelated to a capable tectonic source.

2.5.1.1.4.4.5.8 New Castle County Faults

RAI 130 02.05.01-52

RAI 130 02.05.01-52

RAI 130 02.05.01-52

The New Castle faults are interpreted~~characterized~~ as 3 to 4 mi (4.8 to 6.4 km) long buried north and northeast-striking basement faults (Spoljaric, 1972) (Spolijaric, 1973). The faults are interpreted from structural contours of the top of Precambrian to Paleozoic crystalline basement derived from geophysical and borehole data, and define a 1 mi (1.6 km) wide, N25°E-trending graben in basement rock (Spoljaric, 1973)~~that displace an unconformable contact between Precambrian to Paleozoic bedrock and overlying Cretaceous deposits~~. The faults are located in northern Delaware, near New Castle, about 97 mi (156 km) northeast of the CCNPP site (Figure 2.5-31). ~~Spoljaric (Spoljaric, 1972) (Spoljaric, 1973) interprets the presence of the New Castle County faults using structural contours for the top of basement. On the basis of geophysical and borehole data, coupled with Vibroseis™ profiles, Spolijaric (Spolijaric, 1973)~~

RAI 130 02.05.01-52

~~(Spolojaric, 1974) interprets a 1 mi (1.6 km) wide, N25°E-trending graben in basement rock.~~ The graben is bounded by faults ~~having~~that displace~~ments~~ the basement surface on the order of 32 to 98 ft (10 to 30 m) ~~across the basement-Cretaceous boundary~~ (Spoljaric, 1972). ~~Also, there is a~~Spoljaric (1973) suggests~~ion~~ that the overlying Cretaceous deposits are tilted in a direction consistent with fault deformation; however, ~~there is~~ no direct evidence is reported to indicate that the faults extend into the Cretaceous~~these~~ sediments ~~are displaced~~. Sbar (Sbar, 1975) evaluates a 1973 M3.8 earthquake and its associated aftershocks, and note that the microseismicity defines a causal fault striking northeast and parallel to the northeast-striking graben of Spoljaric (Spoljaric, 1973). Subsequently, subsurface exploration by the Delaware Geological Survey (McLaughlin, 2002), that included acquisition of high resolution seismic reflection profiles, borehole transects, and paleoseismic trenching, provides evidence for the absence of Quaternary faulting on the New Castle faults. Wheeler (Wheeler, 2005) characterizes the New Castle County faults as a Class C feature. Based on McLaughlin (McLaughlin, 2002) there is strong evidence to suggest that the New Castle County faults as mapped by Spolijaric (Spolijaric, 1972) are not a capable tectonic source.

2.5.1.1.4.4.5.9 Upper Marlboro Faults

The Upper Marlboro faults are located in Prince George's County, Maryland, approximately 36 mi (58 km) northwest of the CCNPP site (Figure 2.5-31). These faults were first shown by Dryden (Dryden, 1932) as a series of faults offsetting Coastal Plain sediments. The faults were apparently exposed in a road cut on Crain Highway at 3.3 mi (5.3 km) south of the railroad crossing in Upper Marlboro, Maryland (Prowell, 1983). Two faults displace Miocene and Eocene sediments and a third fault is shown offsetting a Pleistocene unit. These faults are not observed beyond this exposure. No geomorphic expression has been reported or was noticed during field reconnaissance for the CCNPP Unit 3 study. Based on a critical review of available literature, Wheeler (Wheeler, 2006) re-interprets the Upper Marlboro faults as likely related to surficial landsliding because of the very low dips and concavity of the fault planes. The Marlboro faults are classified by Crone and Wheeler (Crone, 2000) and Wheeler (Wheeler, 2006), as a Class C feature based on a lack of evidence for Quaternary faulting. Given the absence of seismicity along the fault, lack of published literature documenting Quaternary faulting, coupled with the interpretation of Crone and Wheeler (Crone, 2000) and Wheeler (Wheeler, 2006), we conclude that the Upper Marlboro faults are not a capable tectonic source.

2.5.1.1.4.4.5.10 Lebanon Church Fault

The Lebanon Church fault is a poorly-known northeast-striking reverse fault located in the Appalachian Mountains of Virginia, near Waynesboro, about 119 mi (192 km) southwest of the CCNPP site (Prowell, 1983) (Figure 2.5-31). The fault is exposed in a single road cut along U.S. Route 250 as a small reverse fault that offsets Miocene-Pliocene terrace gravels up to as about 5 ft (1.5 m) (Prowell, 1983). The terrace gravels overlie Precambrian metamorphic rocks of the Blue Ridge Province. An early author (Nelson, 1962) considered the gravels to be Pleistocene, whereas Prowell (1983) interprets the gravel to be Miocene to Pliocene. Wheeler (Wheeler, 2006) classifies the Lebanon Church fault as a Class C feature having insufficient evidence to demonstrate that faulting is Quaternary. As part of this CCNPP Unit 3 study, inquiries with representatives with the Virginia Geological Survey and United States Geological Survey indicate that there is no new additional geologic information on this fault. Based on literature review, discussion with representatives with Virginia Geological Survey, as well as the absence of seismicity spatially associated with the feature, we conclude that the Lebanon Church fault is not a capable tectonic source.

2.5.1.1.4.4.5.11 Hopewell Fault

The Hopewell fault is located in central Virginia, approximately 89 mi (143 km) southwest of the CCNPP site (Figure 2.5-31). The Hopewell fault is a 30 mi (48 km) long, north-striking, steeply east-dipping reverse fault (Mixon, 1989) (Dischinger, 1987). The fault was originally named the Dutch Gap fault by Dischinger (Dischinger, 1987), and was renamed the Hopewell fault by Mixon (Mixon, 1989). The fault displaces a Paleocene-Cretaceous contact and is inferred to offset the Pliocene Yorktown Formation (Dischinger, 1987). Mixon (Mixon, 1989) extend the mapping of Dischinger (Dischinger, 1987), but include conflicting data regarding fault activity. For instance, a cross section presented by Mixon (Mixon, 1989) shows the Hopewell fault displacing undivided upper Tertiary and Quaternary units, whereas the geologic map used to produce the section depicts the fault buried beneath these units. A written communication from Newell (Wheeler, 2006) explains that the Hopewell fault was not observed offsetting Quaternary deposits and the representation of the fault in the Mixon (Mixon, 1989) cross section is an error. Thus, the Hopewell fault zone is assigned as a Class C feature because no evidence is available to demonstrate Quaternary surface deformation. Based on the written communication of Newell (Wheeler, 2006), an absence of published literature documenting Quaternary faulting, and an absence of seismicity spatially associated with the feature, we conclude that the Hopewell fault is not a capable tectonic source.

2.5.1.1.4.4.5.12 Old Hickory Faults

The Old Hickory faults are located near the Fall Line in southeastern Virginia, approximately 115 mi (185 km) south-southwest of the CCNPP site (Figure 2.5-31). Based on mining exposures of the Old Hickory Heavy Mineral deposit, the Old Hickory faults consist of a series of five northwest-striking reverse faults that offset Paleozoic basement and Pliocene Coastal Plain sediments. The northwest-striking reverse faults juxtapose Paleozoic Eastern Slate Belt diorite over the Pliocene Yorktown Formation (Berquist, 1999). Strike lengths range between 330 to 490 ft (100 to 150 m) and are spaced about 164 ft (50 m) apart. Berquist and Bailey (Berquist, 1999) report up to 20 ft (6 m) of oblique dip-slip movement on individual faults, and suggest that the faults may be reactivated Mesozoic structures. There is no stratigraphic or geomorphic evidence of Quaternary or Holocene activity of the Old Hickory faults (Berquist, 1999). Crone and Wheeler (Crone, 2000) and Wheeler (Wheeler, 2006) conclude that "no Quaternary fault is documented" and assign a Class C designation to the Old Hickory faults. Based on the absence of published literature documenting the presence of Quaternary deformation, and the absence of seismicity spatially associated with this feature, we conclude that the Old Hickory faults are not a capable tectonic source.

2.5.1.1.4.4.5.13 Stanleytown-Villa Heights Faults

The postulated Stanleytown-Villa Heights faults are located in the Piedmont of southern Virginia, approximately 223 mi (359 km) southwest of the CCNPP site (Figure 2.5-31). The approximately 660 ft long (201 m long) faults juxtapose Quaternary alluvium against rocks of Cambrian age, and reflect an east-side-down sense of displacement (Crone, 2000). No other faults are mapped nearby (Crone, 2000). Geologic and geomorphic evidence suggests the "faults" are likely the result of landsliding. Crone and Wheeler (Crone, 2000) classify the Stanleytown-Villa Heights faults as a Class C feature based on lack of evidence for Quaternary faulting. Based on the absence of published literature documenting the presence of Quaternary faulting, and the absence of seismicity spatially associated with this feature, we conclude that the Stanleytown-Villa Heights faults are not a capable tectonic source.

2.5.1.1.4.4.5.14 East Coast Fault System

The postulated East Coast fault system (ECFS) of Marple and Talwani (2000) trends N34°E and is located approximately 70 mi (113 km) southwest of the site (Figure 2.5-31). The 370 mi (595 km) long fault system consists of three approximately 125 mi (201 km) long segments extending from the Charleston area in South Carolina northeastward to near the James River in Virginia (Figure 2.5-31). The three segments were initially referred to as the southern, central, and northern zones of river anomalies (ZRA-S, ZRA-C, ZRA-N) and are herein referred to as the southern, central and northern segments of the ECFS. The southern segment is located in South Carolina; the central segment is located primarily in North Carolina. The northern segment, buried beneath Coastal Plain deposits, extends from northeastern North Carolina to southeastern Virginia, about 70 mi (113 km) southwest of the CCNPP site. Marple and Talwani (Marple, 2000) map the northern terminus of the ECFS between the Blackwater River and James River, southeast of Richmond. Identification of the ECFS is based on the alignment of geomorphic features along Coastal Plain rivers, areas suggestive of uplift, and regions of local faulting. The right-stepping character of the three segments, coupled with the northeast orientation of the fault system relative to the present day stress field, suggests a right-lateral strike-slip motion for the postulated ECFS (Marple and Talwani, 2000).

The southern segment of the fault system, first identified by Marple and Talwani (1993) as an approximately 125 mi (201 km) long and 6 to 9 mi (10 to 14.5 km) wide zone of river anomalies, has been attributed to the presence of a buried fault zone. The southern end of this segment is associated with the Woodstock fault, a structure defined by fault-plane solutions of microearthquakes and thought to be the causative source of the 1886 Charleston earthquake (Marple, 2000). The southern segment is geomorphically the most well-defined segment of the fault system and is associated with micro-seismicity at its southern end. This segment was included as an alternative geometry to the areal source for the 1886 Charleston earthquake in the 2002 USGS hazard model (Section 2.5.2) for the National Seismic Hazard Mapping Project (Frankel, 2002).

Crone and Wheeler (Crone, 2000) do not include the central and northern segments of the ECFS in their compilation of potentially active Quaternary faults. The segments also were not presented in workshops or included in models for the Trial Implementation Project (TIP), a study that characterized seismic sources and ground motion attenuation models at two nuclear power plant sites in the southeastern United States (Savy, 2002). As a member of both the USGS and TIP workshops, Talwani did not propose the northern and central segments of the fault system for consideration as a potential source of seismic activity. There is no pre-EPRI or post-EPRI seismicity spatially associated with the northern and central segments of the fault system.

Recent geologic and geomorphic analysis of stream profiles across sections of the ECFS, and critical evaluation of Marple and Talwani (Marple, 2000) for the North Anna ESP, provides compelling evidence that the northern segment of the ECFS, which lies nearest to the CCNPP site, has a very low probability of existence (Dominion, 2004b). Wheeler (Wheeler, 2005) states that although the evidence for a southern section of the ECFS is good, there is less evidence supporting Quaternary tectonism along the more northerly sections of the ECFS, and designates the northern portion of the fault system as a Class C feature.

In the Safety Evaluation Report for the North Anna ESP site, the NRC staff agreed with the assessment of the northern segment of the East Coast Fault System (ECFS-N) presented by the North Anna applicant (NRC, 2005). Based on their independent review, the NRC staff concluded that:

- "Geologic, seismologic, and geomorphic evidence presented by Marple and Talwani is questionable."
- "The majority of the geologic data cited by Marple and Talwani in support of their postulated ECFS apply only to the central and southern segments."
- There are "no Cenozoic faults or structure contour maps indicating uplift along the ECFS-N."

- "The existence and recent activity of the northern segment of the ECFS is low."

Despite the statements above, the NRC concluded that the ECFS-N could still be a contributor to the seismic hazard at the North Anna site and should be included in the ground motion modeling to determine the Safe Shutdown Earthquake. The NRC agreed with the 10% probability of existence and activity proposed in the North Anna ESP application. The results of the revised ground motion calculations indicate that the ECFS-N does not contribute to the seismic hazard at the North Anna ESP site. The CCNPP site is approximately 70 mi (113 km) northeast of the ECFS-N, or 7 mi (11 km) further away than the North Anna site is from the ECFS-N. Based on the above discussion and the large distance between the site and the ECFS-N, this fault is not considered a contributing seismic source and need not be included in the seismic hazard calculations for the CCNPP site.

Marple and Talwani (Marple, 2004) suggest a northeast extension of the ECFS of Marple and Talwani (Marple, 2000), based on existing limited geologic, geophysical and geomorphic data. Marple and Talwani (Marple, 2004) postulate that the northern ECFS may step left (northwest) to the Stafford fault system near northern Virginia and southern Maryland (Figure 2.5-31) and thus extending the ECFS along the Stafford fault up to New York. As stated in Section 2.5.1.1.4.4.4.1, the NRC (NRC, 2005) agreed with an analysis of the Stafford fault performed as part of the North Anna ESP application and states: "Based on the evidence cited by the applicant, in particular the applicant's examination of the topography profiles that cross the fault system, the staff concludes that the applicant accurately characterized the Stafford fault system as being inactive during the Quaternary Period."

In summary, the ECFS in its entirety represents a new postulated tectonic feature that was not known to the EPRI Earth Science Teams in 1986. The 1986 EPRI models include areal sources to model the Charleston seismic source; therefore, the southern segment of the East Coast fault system is in essence covered by the different Charleston sources zone geometries. A review of the seismic sources that contribute 99% of the seismic hazard to the CCNPP shows that the Charleston source is not a contributor. The central and northern segments of the ECFS represent a new tectonic feature in the Coastal Plain that postdates the EPRI studies. The closest approach of the northern segment to the site is approximately 77 mi (124 km) as described above. Although the postulated ECFS represents a potentially new tectonic feature in the Coastal Plain of Virginia and North Carolina (Marple, 2000), current interpretations of the ECFS based on existing data indicate that the fault zone probably does not exist (especially the northern segment) and, if it does exists, has a very low probability of activity and does not contribute to hazard at the site.

2.5.1.1.4.5 Seismic Sources Defined by Regional Seismicity

Within 200 mi (322 km) of the CCNP site, two potential seismic sources are defined by a concentration of small to moderate earthquakes. These two seismic sources include the Central Virginia seismic zone in Virginia and the Lancaster seismic zone in southeast Pennsylvania, both of which are discussed below (Figure 2.5-31).

2.5.1.1.4.5.1 Central Virginia Seismic Zone

The Central Virginia seismic zone is an area of persistent, low level seismicity in the Piedmont Province (Figure 2.5-24 and Figure 2.5-31). The zone extends about 75 mi (121 km) in a north-south direction and about 90 mi (145 km) in an east-west direction from Richmond to Lynchburg and is coincident with the James River (Bollinger, 1985). The CCNPP site is located 47 to 62 mi (76 to 100 km) northeast of the northern boundary of the Central Virginia seismic zone. The largest historical earthquake to occur in the Central Virginia seismic zone was the body-wave magnitude (mb) 5.0 Goochland County event on December 23, 1875 (Bollinger, 1985). The maximum intensity estimated for this event was Modified Mercalli Intensity (MMI) VII in the epicentral region. More recently, an mb 4.5 earthquake (two closely-spaced events that when combined = Mw 4.1) occurred on December 9, 2003 within the Central Virginia seismic zone (Kim and Chapman, 2005). The December 9, 2003 earthquake occurred close to the Spotsylvania fault, but due to the uncertainty in the location of the epicenter (3.7 to 5 mi (6 to 8 km)), no attempt could be made to locate the epicenter with a specific fault or geologic lineament in the CVSZ (Kim, 2005).

RAI 130 02.05.01-53

Seismicity in the Central Virginia seismic zone ranges in depth from about 2 to 8 mi (3 to 13 km) (Wheeler, 1992). It is suggested (Coruh, 1988) that seismicity in the central and western parts of the zone may be associated with west-dipping reflectors that form the roof of a detached antiform, while seismicity in the eastern part of the zone near Richmond may be related to a near-vertical diabase dike swarm of Mesozoic age. However, given the depth distribution of 2 to 8 mi (3 to 13 km) (Wheeler, 1992) and broad spatial distribution, it is difficult to uniquely attribute the seismicity to any known geologic structure and it appears that the seismicity ~~extends both~~is generally above ~~and below~~ the Appalachian detachment.

No capable tectonic sources have been identified within the Central Virginia seismic zone, but two paleo-liquefaction sites have been identified within the seismic zone (Crone, 2000) (Obermier, 1998). The presence of these paleo-liquefaction features on the James and Rivanna Rivers shows that the Central Virginia seismic zone reflects both an area of paleo-seismicity as well as observed historical seismicity. Based on the absence of widespread paleo-liquefaction, however, it was concluded (Obermier, 1998) that an earthquake of magnitude 7 or larger has not occurred within the seismic zone in the last 2,000 to 3,000 years, or in the eastern portion of the seismic zone for the last 5,000 years. It was also conclude that the geologic record of one or more magnitude 6 or 7 earthquakes might be concealed between streams, but that such events could not have been abundant in the seismic zone. In addition, these isolated locations of paleo-liquefaction may have been produced by local shallow moderate magnitude earthquakes of M 5 to 6.

The paleo-liquefaction sites reflect pre-historical occurrences of seismicity within the Central Virginia seismic zone, and do not indicate the presence of a capable tectonic source. Recently, Wheeler (Wheeler, 2006) hypothesizes that there may be two causative faults for the small dikes of Obermier and McNulty (Obermier, 1998), and that earthquakes larger than those represented by historic seismicity are possible; whereas Marple and Talwani (Marple, 2004) interpret seismicity data to infer the presence of a hypothesized northwest-trending basement fault (Shenandoah fault) that coincides with the Norfolk fracture zone (Marple, 2004). However, no definitive causative fault or faults have been identified within the Central Virginia seismic zone (Wheeler, 2006).

RAI 72 02.05.02-4

RAI 72 02.05.02-4

The 1986 EPRI source model includes various source geometries and parameters to capture the seismicity of the Central Virginia seismic zone. Subsequent hazard studies have used maximum magnitude (Mmax) values that are within the range of maximum magnitudes used by the six EPRI models. Collectively, upper-bound maximum values of Mmax used by the EPRI teams range from mb 6.6 to 7.2 (Section 2.5.2.2). More recently, Bollinger (Bollonger, 1992) has estimated a Mmax of mb 6.4 for the Central Virginia seismic source. Also, Chapman and Krimgold (Chapman, 1994) have used a Mmax of Mw 7.53 (mb 7.25) for the Central Virginia seismic source zone based on the estimated magnitude of the 1886 Charleston earthquake.~~and most other sources in their seismic hazard analysis of Virginia.~~ ~~This m~~More recent estimates of the 1886 earthquake magnitude are lower (Bakun and Hopper, 2004); Johnston, 1996) indicating that the Mmax of Chapman and Krimgold (Chapman, 1994) should also be lowered.~~Mmax is similar to the Mmax values used in the 1986 EPRI studies.~~ These more recent

RAI 72 02.05.02-4 estimates of Mmax values for the Central Virginia seismic zone are within the range of the Mmax values used in the 1985 EPRI studies (Section 2.5.2.2.1.7). ~~Similarly~~Also, the distribution and rate of seismicity in the Central Virginia seismic source have not changed since the 1986 EPRI study (Section 2.5.2.2.8). Thus, there is no new information or data that motivates modifying ~~change to~~ the source geometry, ~~or~~ rate of seismicity, or Mmax values for the Central Virginia seismic zone in the EPRI-SOG model. ~~In 2005, the NRC agreed with the findings of the North Anna ESP application's assessment of the Central Virginia seismic zone (NRC, 2005). Therefore, the conclusion is that no new information has been developed since 1986 that would require a significant revision to the EPRI seismic source model.~~ The same conclusion was reached in the North Anna ESP application, and in 2005 the NRC agreed with this conclusion (NRC, 2005).

RAI 72 02.05.02-4

2.5.1.1.4.5.2 Lancaster Seismic Zone

The Lancaster seismic zone, as defined by Armbruster and Seeber (Armbruster, 1987), of southeast Pennsylvania has been a persistent source of seismicity for at least two centuries. The seismic zone is about 80 mi (129 km) long and 80 mi (129 km) wide and spans a belt of allochthonous Appalachian crystalline rocks between the Great Valley and Martic Line about 111 mi (179 km) northwest of the CCNPP site (Figure 2.5-31). The Lancaster seismic zone crosses exposed Piedmont rocks that include thrust faults and folds associated with Paleozoic collisional orogenies. It also crosses the Newark-Gettysburg Triassic rift basin which consists of extensional faults associated with Mesozoic rifting. Most well-located epicenters in the Lancaster seismic zone lie directly outside the Gettysburg-Newark basin (Scharnberger, 2006). The epicenters of 11 events with magnitudes 3.04 to 4.61 rmb from 1889 to 1994 from the western part of Lancaster seismic zone define a north-south trend that intersects the juncture between the Gettysburg and Newark sub-basins. This juncture is a hinge around which the two sub-basins subsided, resulting in east-west oriented tensile stress. Numerous north-south trending fractures and diabase dikes are consistent with this hypothesis. It is likely that seismicity in at least the western part of the Lancaster seismic zone is due to present-day northeast-southwest compressional stress which is activating the Mesozoic fractures, with dikes perhaps serving as stress concentrators (Armbruster, 1987).

It also is probable that some recent earthquakes in the Lancaster seismic zone have been triggered by surface mining. For instance, the 16 January 1994 Cacoosing earthquake (mb 4.6) is the largest instrumented earthquake occurring in the Lancaster seismic zone (Section 2.5.1.1.4.4.5.7). This event was part of a shallow (depths generally less than 1.5 mi (2.4 km)) earthquake sequence linked to quarry activity (Seeber, 1998). The earthquake sequence that culminated in the January 16 event initiated after a quarry was shut down and the quarry began to fill with water. Seeber (Seeber, 1998) interprets the reverse-left lateral oblique earthquake sequence to be due to a decrease in normal stress caused by quarrying followed by an increase in pore fluid pressure (and decrease in effective normal stress) when the pumps were turned off and the water level increased.

Prior to the Cacoosing earthquake sequence, the 23 April 1984 Martic earthquake (mb 4.1) was the largest instrumented earthquake in the seismic zone and resembles pre-instrumental
RAI 130 02.05.01-53 historical events dating back to the middle 18th century (Armbruster, 1987). The 1984 earthquake sequence appears centered at about 2.8 mi (4.5 km) in depth and may have ruptured a steeply east-dipping, north-to northeast-striking fault aligned subparallel to Jurassic dikes with a reverse-right lateral oblique movement, consistent with east-northeast horizontal maximum compression. These dikes are associated with many brittle faults and large planes of weakness suggesting that they too have an effect on the amount of seismicity in the Lancaster
RAI 130 02.05.01-53 seismic zone (Armbruster, 1987). Most of the seismicity in the Lancaster seismic zone is occurring on secondary faults at high angles to the main structures of the Appalachians

RAI 130
02.05.01-53

(Armbruster, 1987) (Seeber, 1998). The EPRI study (EPRI, 1986) source models do not identify the Lancaster seismic zone as a separate seismic source. However, the 5.3 to 7.2 Mb maximum magnitude distributions of EPRI source zones are significantly greater than any reported earthquake in this Lancaster seismic zone. Thus, the EPRI study (EPRI, 1986) models adequately characterized this region and no significant update is required.

2.5.1.2 Site Geology

Sections 2.5.1.2.1 through 2.5.1.2.6 are added as a supplement to the U.S. EPR FSAR.

2.5.1.2.1 Site Area Physiography and Geomorphology

The CCNPP site area is located within the Western Shore Uplands of the Atlantic Coastal Plain Physiographic Province and is bordered by the Chesapeake Bay to the east and the Patuxent River to the west (Figure 2.5-4 and Figure 2.5-7).

The site vicinity geologic map (Figure 2.5-27 and Figure 2.5-28), compiled from the work of several investigators, indicates that the counties due east from the CCNPP site across Chesapeake Bay are underlain by Pleistocene to Recent sands. Most of the site vicinity is underlain by Tertiary Coastal Plain deposits. Quaternary to Recent alluvium beach deposits and terrace deposits are mapped along streams and estuaries. Quaternary terrace and Lowland deposits are shown in greater detail on the scale of the site area geologic map (Figure 2.5-32). Geologic cross sections in the site area indicate that the Tertiary Upland deposits are underlain by gently dipping Tertiary Coastal Plain deposits described in Section 2.5.1.2.2 (Figure 2.5-33).

The topography within 5 mi (8 km) of the site consists of gently rolling hills with elevations ranging from about sea level to nearly 130 ft (40 m) msl (Figure 2.5-4). The site is well-drained by short, ephemeral streams that form a principally dendritic drainage pattern with many streams oriented in a northwest-southeast direction (Figure 2.5-5). As shown on the site area and site topographic and geological maps, the ground surface above approximately 100 ft (30 m) msl is capped by the Upper Miocene-Pliocene Upland deposits (Figure 2.5-4, Figure 2.5-5, Figure 2.5-32, and Figure 2.5-33). These deposits occupy dissected upland areas of the Cove Point quadrangle in which the CCNPP site is located (Figure 2.5-32 and Figure 2.5-33) (Glaser, 2003a). The longest stream near the site is Johns Creek, which is approximately 3.5 mi (5.6 km) long before it drains into St. Leonard Creek (Figure 2.5-4 and Figure 2.5-34). The ephemeral stream channels near the CCNPP site are either tributary to Johns Creek or flow directly to the Chesapeake Bay. These stream channels maintain their dendritic pattern as they cut down into the underlying Choptank and St. Marys Formations (Figure 2.5-27, Figure 2.5-32 and Figure 2.5-33).

The Chesapeake Bay shoreline forms the eastern boundary of the CCNPP site and generally consists of steep cliffs with narrow beach at their base. The cliffs reach elevations of about 100 ft (30 m) msl along the eastern portion of the site's shoreline. Narrow beaches whose width depends upon tidal fluctuations generally occur at the base of the cliffs. Field observations indicate that these steep slopes fail along nearly vertical irregular surfaces. The slope failure appears to be caused by shoreline erosion along the base of the cliffs. Shoreline processes and slope failure along Chesapeake Bay are discussed in Section 2.4.9. Approximately 2500 ft (762 m) of the shoreline from the existing CCNPP Units 1 and 2 intake structure southward to the existing barge jetty is stabilized against shoreline erosion (Figure 2.5-50). The CCNPP Unit 3 will be constructed at a final grade elevation of approximately 85 ft (26 km) msl and will be set back approximately 1,000 ft (305 m) from the Chesapeake Bay shoreline.

As described in Section 2.5.1.1.1, the Chesapeake Bay was formed toward the end of the Wisconsinan glacial stage, which marked the end of the Pleistocene epoch. As the glaciers retreated, the huge volumes of melting ice fed the ancestral Susquehanna and Potomac Rivers, which eroded older Coastal Plain deposits forming a broad river valley. The rising sea level covered the Continental Shelf and reached the mouth of the Bay about 10,000 years ago. Sea level continued to rise, eventually submerging the area now known as the Susquehanna River Valley prior to sea level dropping to the current elevation. The Bay assumed its present dimensions about 3000 years ago (Section 2.4.9).

2.5.1.2.2 Site Area Geologic History

RAI 130 02.05.01-54

The site area geologic history prior to the early Cretaceous is inferred from scattered borehole data, geophysical surveys and a synthesis of published information. ~~Sparse geophysical and borehole~~These data indicate that the ~~basement~~ rock beneath the Coastal Plain sediments in the site area may be either~~consist of~~ extended or rifted exotic crystalline magmatic arc material (Glover, 1995b) or, alternatively, Triassic rift basin sediments (Benson, 1992). Although the bas~~ement~~ of the Coastal Plain section has not been penetrated directly beneath the site with drill holes, regional geologic cross sections developed from geophysical, gravity and aeromagnetic, as well as limited deep borehole stratigraphic data beyond the site area, suggest that the base of the Coastal Plain section is ~~Precambrian and Paleozoic crystalline rocks are~~ most likely ~~present~~ at a depth of about 2,600 ft (792 m) beneath the site (Section 2.5.1.2.3 and Section 2.5.1.2.4).

RAI 130 02.05.01-54

Tectonic models discussed in Section 2.5.1.1.4.3.1~~2.4~~ hypothesize that the crystalline basement was first accreted to the pre-Taconic North American margin during the Paleozoic along a suture that lies about 10 mi (16 km) west of the site (Klitgord, 1995) (Figure 2.5-17 and Figure 2.5-23). These models also suggest this basement is rifted crust that was thinned after accretion during the Mesozoic rifting of Pangea (Section 2.5.1.1.4.1.2). Therefore, the crystalline basement beneath the Coastal Plain sediments in the site area might consist of an accreted nappe-like block of Carolina-Chopawamsic magmatic arc terrane with windows of Laurentian Grenville basement cut by later phase normal faults (Figure 2.5-16 and Figure 2.5-17) (Klitgord, 1995).

RAI 130 02.05.01-54

RAI 130 02.05.01-54

As discussed in Section 2.5.1.1.2, Section 2.5.1.1.4.4.3, and Section 2.5.1.2.4, Mesozoic rift basins are exposed in the Piedmont Physiographic Province and are buried beneath Coastal Plain sediments (Figure 2.5-10). Wether or not the CCNPP Site is underlain by a Mesozoic basin (e.g., the Queen Anne Basin) preserved beneath the thick Coastal Plain section is unclear.~~The Queen Anne Basin was originally postulated by Hansen (1988) and was considered to underlie the site (Horton, 1991). However, this interpretation does not appear to be supported by most of the borehole data and current interpretations (Section 2.5.1.2.4).~~ The available data in the site area include only regional gravity and aeromagnetic data that allow multiple (often contradictory) interpretations of the location of a basin at or near the CCNPP Site beneath the Coastal Plain sediments. For example, Horton (1991) (Figure 2.5-9 and Figure 2.5-16) and Benson (1992) (Figure 2.5-10) show the CCNPP site underlain by the Mesozoic Queen Anne basin, whereas Schlische (1990) (Figure 2.5-22) and Withjack (1998) (Figure 2.5-10) do not show a Mesozoic basin beneath the site. There are no deep boreholes or seismic lines that allow for a definitive interpretation of the presence, geometry, or thickness of a Mesozoic rift basin beneath the CCNPP site. See Section 2.5.1.1.4.4.3 for further discussion regarding the Queen Anne basin.

RAI 130 02.05.01-54

During the early Cretaceous, sands, clays, sandy clays, and arkosic sands of the Arundel/Patuxent Formations (undivided) were deposited on the crystalline basement in a continental and fluviatile environment. Individual beds of sand or silt grade rapidly into sediments with different compositions or gradations, both vertically and horizontally, which suggests they were deposited in alluvial fan or deltaic environments. Clay layers containing carbonized logs, stumps and other plant remains indicate the existence of quiet-water, swamp environments between irregularly distributed stream channels. Thicker clays near the top of this unit in St Mary's County are interpreted to indicate longer periods of interfluvial quiet water deposition (Hansen, 1984).

The overlying beds of the Patapsco Formation are similar to the deposits in the Arundel/Patuxent (undivided) formations and consist chiefly of materials derived from the eroded crystalline rocks of the exposed Piedmont to the west and reworked Lower Cretaceous sediments. These sediments were deposited in deltaic and estuarine environments with relatively low relief. The Upper Cretaceous Raritan Formation appears to be missing from the site area due either to non-deposition or erosion on the northern flank of the structurally positive Norfolk Arch.

The Magothy Formation represents deposits from streams flowing from the Piedmont and depositing sediments in the coastal margins of the Upper Cretaceous sea. Subsequent uplift and tilting of the Coastal Plain sediments mark the end of continental deposition and the beginning of a marine transgression of the region. This contact is a regional unconformity marked in places by a basal layer of phosphatic clasts in the overlying Brightseat Formation.

During the Early Paleocene Epoch, the Brightseat Formation marks a marine advance in the Salisbury embayment (Ward, 2004). Uplift or sea level retreat is indicated by the burrowed contact (unconformity) of the Brightseat Formation with the overlying Aquia Formation. The marine Aquia Formation which is noted for its high glauconite content and shell beds was deposited in a shoaling marine environment indicated by a generally coarsening upward lithology (Hansen, 1996). A mix of light-colored quartz grains and greenish to blackish glauconite grains and iron staining indicated the change to a sandbank facies in the upper Aquia formation (Hansen, 1996). A marine transgression during the Late Paleocene/Early Eocene into the central portion of the Salisbury Embayment deposited the Marlboro Clay (Ward, 2004). During the Early Eocene, a moderately extensive marine transgression deposited the Potopaco Member of the Nanjemoy Formation. A subsequent transgression deposited the Woodstock Member of the Nanjemoy Formation (Ward, 2004). The most extensive marine transgression during the middle Eocene resulted in the deposition of the Piney Point Formation (Ward, 2004). The site area may have been emergent during the Oligocene as the Late Oligocene Old Church Formation indicates sea level rise and submergence to the north and south of the site area (Ward, 2004). A brief regression was followed by nearly continuous sedimentation in the Salisbury Embayment punctuated by short breaks, resulting in a series of thin, unconformity-bounded beds (Ward, 2004). A series of marine transgressions into the Salisbury Embayment during the Miocene produced the Calvert, Choptank and St. Marys Formations. Pliocene and Quaternary geologic history is discussed in Section 2.5.1.2.1.

2.5.1.2.3 Site Area Stratigraphy

RAI 130 02.05.01-54

The CCNPP site area is located on Coastal Plain sediments ranging in age from Lower Cretaceous to Recent, which, in turn, were deposited on the pre-Cretaceous basement. As discussed above in Section 2.5.1.2.2, there is uncertainty regarding whether Mesozoic rift basin deposits underlie the Coastal Plain sediments or whether the Coastal Plain sediments are deposited directly over extended crystalline basement. Figure 2.5-36 is a site-specific stratigraphic column based on correlations by Hansen (Hansen, 1996), Achmad and Hansen (Achmad, 1997) and Ward and Powars (Ward, 2004).

RAI 130 02.05.01-54

Site specific information on the stratigraphy underlying the CCNPP site is ~~limited~~constrained by the total depths of the various borings advanced by site investigators over the years.

RAI 130 02.05.01-54

Figure 2.5-35 shows the locations of the various borings at the site and identifies those completed as either water supply wells or observation wells based on the 2007 drilling program and the plot plan at that time. Many of these borings were drilled to 200 ft (61 m) in total depth; two were advanced to a total depth of 400 ft (122 m). Figure 2.5-103 includes the additional boring locations based on the 2008 drilling program. Only a few scattered borings have been advanced below the Aquia Formation (Hansen, 1986) (Figure 2.5-13). The deepest boring known to have been advanced at the site is CA-Ed 22 which was drilled to a total depth of 789 ft (240 m) and completed as a water supply well in 1968 (Hansen, 1996). This boring penetrates the full Tertiary stratigraphic section and intersects the contact between the Tertiary and the Cretaceous section at the base of the Aquia Formation.

RAI 130 02.05.01-54

The closest boring which advances to pre-Cretaceous bedrock is approximately 13 mi (21 km) south of the site at Lexington Park in St. Mary's County, (Figure 2.5-11) (Hansen, 1986). This boring cored a Jurassic diabase dike ~~in the pre-Cretaceous~~that may have intruded either Triassic rift-basin deposits or extended crystalline basement (Section 2.5.1.1.3). The few other borings that have reached basement rock ~~in~~near the site ~~area~~ are widely scattered (Figure 2.5-11) but the majority indicates that the crystalline basement ~~rock~~ beneath the site areais likely to be similar to the schists and gneisses found in the Piedmont Physiographic Province approximately 50 mi (80 km) to the west (Figure 2.5-1). Alternatively, this crystalline basement might have been accreted to the exposed Piedmont as a result of continental collision during a Paleozoic orogeny (Section 2.5.1.1.1.4 and Section 2.5.1.2.2). ~~Figure 2.5-35 shows the locations of the various borings at the site and identifies those completed as either water supply wells or observation wells. Many of these borings were drilled to 200 ft (61 m) in total depth; six were advanced to a total depth of 400 ft (122 m). Figure 2.5-36 is a site-specific stratigraphic column based on correlations by Hansen (Hansen, 1996), Achmad and Hansen (Achmad, 1997) and Ward and Powars (Ward, 2004).~~

RAI 130 02.05.01-54

~~The CCNPP site is located on Coastal Plain sediments ranging in age from Lower Cretaceous to Recent, which, in turn, were deposited on the pre-Cretaceous basement rock. The Cretaceous section shown on the site stratigraphic column is projected to the site from proximal borings which intersect the pre-Cretaceous basement (Figure 2.5-13).~~

Coastal Plain sediments were deposited in a broad basement depression known as the Salisbury Embayment extending from eastern Virginia to southern New Jersey (Figure 2.5-12) (Ward, 2004). These sediments were deposited during periods of marine transgression/ regression and exhibit lateral and vertical variation in both lithology and texture.

2.5.1.2.3.1 Lower Cretaceous Potomac Group and pre-Potomac sediments

As discussed in Section 2.5.1.1.3, Hansen and Wilson (Hansen, 1984) assign the lowermost 30 ft (9 m) of the Lexington Park well (SM-Df 84), 13 mi (21 km) south of the CCNPP site (Figure 2.5-11) (Hansen, 1986), to the Waste Gate formation. These sediments are described as gray silts and clays, interbedded with fine to medium silty fine to medium sands. Although these sediments might correlate with the Waste Gate Formation identified in a well in Crisfield, Maryland (Do-CE 88), east of the Chesapeake Bay (Figure 2.5-11), there is no direct evidence indicating whether this unit occurs beneath the CCNPP site.

The Potomac Group is comprised of a sequence of interbedded sands and silty to fine sandy clays. Because this formation was not encountered by any borings drilled at the CCNPP site, the description of these units is based on published data (Hansen, 1984) (Achmad, 1997). Regionally, the Potomac Group consists of, from oldest to youngest, the Patuxent Formation, the Arundel Formation and the Patapsco Formation. These units are considered continental in origin and are in unconformable contact with each other.

The Lower Cretaceous Patuxent Formation consists of a sequence of variegated sands and clays which form a major aquifer in the Baltimore area, approximately 50 mi (80 km) up-dip from the site, but which have not been tested in the vicinity of the site. The nearest well intercepting the Patuxent is approximately 13 mi (21 km) south of the site and here the formation contains much less sand than is found in the upper part of the Potomac Group. The Patuxent is approximately 600 to 700 ft (182 m to 213 m) thick and is overlain by the Arundel/ Patapsco formations (undivided)

In the Baltimore area, the Arundel Formation consists of clays which are brick red near the Fall Line. Further down-dip toward the southeast, the color changes to gray and this unit is difficult to separate in the subsurface from those clays present in the underlying Patuxent and overlying Patapsco formations. Consequently, the Arundel and the Patuxent are often undivided (Hansen, 1984) in the literature and referred as the Arundel/Patuxent formations (undivided). Hansen and Wilson (Hansen, 1984) describe the upper portion of the Arundel/Patuxent formations (undivided) as variegated silty clay with thin very fine sand and silt interbeds that may be as thick as 150 to 200 ft (46 to 61 m) beneath the CCNPP site (Figure 2.5-13). The Arundel Formation is not recognized in southern Maryland (Hansen, 1996).

2.5.1.2.3.2 Upper Cretaceous Formations

The Patapsco formation is the uppermost unit in the Potomac Group and consists of gray, brown and red variegated silts and clays interbedded with lenticular, cross-bedded clayey sands and minor gravels. This formation is a major aquifer near the Fall Line in the Baltimore area, but the Patapsco is untested near the CCNPP site. The thickness of the Patapsco Formation based on regional correlations is 1,000 to 1,100 ft thick beneath the CCNPP site.

The Mattaponi (?) formation described as overlying the Potomac group in Hansen and Wilson (Hansen, 1984) is no longer recognized by the Maryland Geological Survey. The section formerly assigned to the Mattaponi (?) has been included within the Patapsco Formation.

The Magothy Formation unconformably overlies the Patapsco Formation beneath the site. The Magothy is comprised chiefly of pebbly, medium coarse sand, although there are clayey portions in the upper part (Achmad, 1997). This formation is much thinner at the site than further north in Calvert County and pinches out within a few mi to the south (Achmad, 1997). The Monmouth and Matawan formations have not been differentiated from the Magothy Formation in the site area.

2.5.1.2.3.3 Tertiary Formations

The earliest Tertiary sediments beneath the site are assigned to the Lower Paleocene Brightseat Formation, a thin dark gray sandy clay identified in the deepest boring (CA-Ed 22) at the site as the Lower Confining Unit (Figure 2.5-13). The Brightseat Formation is identified in the gamma log as a higher than normal gamma response below the Aquia sand. According to Ward and Powars (Ward, 2004) the Brightseat Formation marks a marine advance in the Salisbury Embayment and occurs principally in the northeastern portion of the Embayment. This stratigraphic unit was reached by the water supply well CA-Ed 22 in 1968 (Figure 2.5-13). Achmad and Hansen (Achmad, 1997) describe the Brightseat Formation as approximately 10 ft (3 m) thick consisting mainly of very fine sand and clay with a bioturbated fabric. The absence of a bioturbated contact with the underlying beds suggests an unconformable contact.

The Aquia Formation unconformably overlies the Brightseat Formation and consists of clayey, silty, very shelly glauconitic sand (Ward, 2004). Microfossil study has placed the Aquia in the upper Paleocene. In the type section, the Aquia Formation is divided into two members, the

Piscataway Creek and the Paspotansa, but at the CCNPP site, these members are not differentiated. Achmad and Hansen (Achmed, 1997) describe the Aquia Formation as approximately 150 ft (46 m) thick. The sand becomes fine-grained in the lower 50 ft (15 m) of the formation.

The Marlboro clay is a silvery-gray to pale-red plastic clay interbedded with yellowish-gray to reddish silt occurring at the base of the Nanjemoy Formation (Ward, 2004). Achmad and Hansen (1997) describe approximately 10 ft (3 m) of clay with thin, indistinct laminae of differing colored silt. Its contact with the underlying Aquia Formation is somewhat gradational while the contact between the Marlboro and the overlying Nanjemoy appears to be sharp indicating that the Nanjemoy unconformably overlies the Marlboro. Microfossil studies indicate the presence of a mixture of very late Paleocene and very early Eocene flora. Based on geophysical logs from CA-Ed 22, the Marlboro clay appears to be approximately 15 ft (4.6 m) thick beneath the CCNPP site (Figure 2.5-13).

At the CCNPP site, the Nanjemoy Formation is divided into the Potapaco and Woodstock members between the overlying Piney Point Formation and the underlying Marlboro clay. The Nanjemoy Formation is described as olive black, very fine grained, well-sorted silty glauconitic sands (Ward, 2004). Based on electric log data, the thickness of the Nanjemoy Formation beneath the CCNPP site is approximately 180 ft (55 m). About 80 ft (24 m) of this unit was penetrated by CCNPP Unit 3 borings, B-301 and B-401 (Figure 2.5-37 and Figure 2.5-38), drilled during the subsurface investigation.

The Piney Point Formation is a thin glauconitic sand and clay unit unconformably overlying the Nanjemoy formation. According to Achmad and Hansen (Achmad, 1997), the Piney Point is approximately 20 ft (6 m) thick at the CCNPP site and extends from about the middle of Calvert County, north of the CCNPP site, toward the south to beyond the Potomac River; increasing in thickness to approximately 130 ft (40 m) at Point Lookout at the confluence of the Potomac River and Chesapeake Bay. Formerly considered late Eocene in age, the Piney Point is assigned to the middle Eocene (Achmad, 1997) (Ward, 2004). The unit has a distinctive natural gamma signature associated with the presence of glauconite and is a useful marker bed.

This distinctive natural gamma signature is present in boring B-301 at a depth of 302 ft (92 m) (205 ft (62 m) msl). This interval is described as dark greenish gray, dense clayey sand grading to very dense silty sands in their bottom 25 ft (8 m). Boring B-401 encountered the Piney Point Formation at a depth of 278 ft (85 m) (-181 ft (-55 m) msl).

According to Hansen (Hansen, 1996), the top of the Piney Point Formation occurs at an approximate elevation of -200 ft (-61 m) msl in the CCNPP site area (Figure 2.5-14). The absence of late Eocene and early Miocene sediments indicate the absence of deposition or erosion for millions of years. A structure contour map of the top of the Piney Point Formation shows an erosion surface that dips gently toward the southeast (Figure 2.5-14).

The Chesapeake Group at the CCNPP site is divided into three marine formations which are, from oldest to youngest, the Calvert Formation, the Choptank Formation and the St. Marys Formation. These units are difficult to distinguish in the subsurface due to similar sediment types and are undivided at the CCNPP site (Glaser, 2003c). Achmad and Hansen (Achmad, 1997) indicate that the Chesapeake Group is approximately 245 ft (75 m) thick beneath the CCNPP site, based on boring CA-Ed 22 data. Kidwell (Kidwell, 1997) states that the stratigraphic relations within this group are highly complex. Based on cross sections presented in Kidwell (Kidwell, 1997), the contact between the St. Marys Formation and the underlying Choptank is estimated to be approximately 22 ft (7 m) deep in boring B-301 and at 10 ft (3 m) deep in B-401.

The thickness of the Chesapeake Group (undifferentiated) is 280 ft in boring B-301 and 268 ft in B-401. The difference in these thicknesses and that in CA-Ed 22 is attributed to the geophysical log of the latter boring not continuing to the top of the boring and/or difference in the chosen top of the St. Marys Formation.

Although the formational contacts within the Chesapeake Group are difficult to impossible to identify, there are several strata which are encountered in most of the CCNPP Unit 3 investigation borings. The most persistent of these is the calcite-cemented sand shown in Figure 2.5-42 and probably is one of the units Kidwell (Kidwell, 1997) interprets as the Choptank Formation.

About 20 ft below the base of this cemented sand unit as a second, but much thinner cemented sand which is identified primarily by "N" values (the sum of the blow counts for the intervals 6 to 12 in (15 to 30 cm) and 12 to 18 in (30 to 46 cm) sample intervals in a standard SPT) higher than those immediately above and below.

The base of the Chesapeake Group (Piney Point Formation) is clearly identified in the geophysical log (Figure 2.5-37 and Figure 2.5-38) by the characteristic gamma curve response. Based on the boring log, this gamma curve response appears to be related to calcite-cemented sand.

The surficial deposits consist of two informal stratigraphic units: the Pliocene-age Upland deposits and Pleistocene to Holocene Lowland deposits. The Upland deposits consist of two units deposited in a fluviatile environment. The Upland deposits are areally more extensive in St. Mary's County than in Calvert County (Glaser, 1971). The outcrop distribution has a dendritic pattern and since it caps the higher interfluvial divides, this unit is interpreted as a highly dissected sediment sheet whose base slopes toward the southwest (Glaser, 1971) (Hansen, 1996). This erosion might have occurred due to differential uplift during the Pliocene or down cutting in response to lower base levels when sea level was lower during periods of Pleistocene glaciation.

2.5.1.2.3.4 Quarternary Formations

The Lowland deposits are considered to consist of three lithologic units. The basal unit is estimated to be 10 to 20 ft (3 to 6 m) thick and is often described as cobbly sand and gravel. This unit may represent high energy stream deposits in an alluvial environment near the base of eroding highlands to the west. The basal unit is overlain by as much as 90 ft (27 m) of bluish gray to dark brown clay that may be silty or sandy (Glaser, 1971) The uppermost of the three units consists of 10 to 30 ft (3 to 9 m) of pale gray, fairly well sorted, medium to coarse sand (Glaser, 1971). The Lowland deposits were laid down in fluvial to estuarine environments (Hansen, 1996) and are generally found along the Patuxent and Potomac River valleys and the Chesapeake Bay. These deposits occur in only a few places along the east shore of Chesapeake Bay.

Sands overlying the Chesapeake Group at the CCNPP site are mapped by Glaser (2003c) as Upland Deposits. Within the CCNPP Unit 3 power block these sands range in thickness from a feather edge in borings on the southern edge, to more than 50 ft in B-405.

Boring B-301 intersected 22 ft (7 m) of silty sand above the contact with the Chesapeake Group, while B-401 has 10 ft (3 m) of silty sand (Figure 2.5-37 and Figure 2.5-38). The sand in both borings grades into a coarser sand unit just above the contact. These sands are attributed to the Upland deposits previously mapped (Glaser, 2003c).

Terrace deposits in the CCNPP site area (Figure 2.5-32 and Figure 2.5-34) consist of interbedded light gray to gray silty sands and clay with occasional reddish brown pockets and are approximately 50 ft (15 m) thick. These units are Pliocene to Holocene in age.

Holocene deposits, mapped as Qal on the site Geologic Map, includes heterogeneous sediments underlying floodplains and beach sands composed of loose sand.

2.5.1.2.4 Site Area Structural Geology

RAI 130 02.05.01-54

The local structural geology of the CCNPP site area described in this section is based primarily on a summary of published geologic mapping (Cleaves, 1968) (Glaser, 1994) (McCartan, 1995) (Achmad, 1997) (Glaser, 2003b) (Glaser, 2003c), aeromagnetic and gravity surveys (Hansen, 1978) (Hittelman, 1994) (Milici, 1995) (Bankey, 2002), detailed lithostratigraphic profiles along Calvert Cliffs (Kidwell, 1988) (Kidwell, 1997), results of earlier investigations performed at the CCNPP site (BGE, 1968) (CEG, 2005), as well as CCNPP site reconnaissance and subsurface exploration performed for ~~this~~the CCNPP Unit 3 ~~study~~site investigation.

RAI 130 02.05.01-54

Sparse geophysical and borehole data indicate that the basement ~~likely~~ consists of exotic crystalline magmatic arc material (Hansen, 1986) (Glover, 1995b) or Triassic rift basin sedimentary rocks (Benson, 1992). Although the basement beneath the site area has not been penetrated with drill holes, regional geologic cross sections developed from geophysical, gravity and aeromagnetic, as well as limited deep borehole data from outside of the CCNPP site area, suggest that the based of the Coastal Plain section is~~Precambrian and Paleozoic crystalline rocks and, less likely, Mesozoic rift basin deposits are~~ present at a depth of approximately~~about~~ 2,500 ft (762 m) msl (Section 2.5.1.2.2).

RAI 130 02.05.01-54

Tectonic models hypothesize that the crystalline basement underlying the site was accreted to a pre-Taconic North American margin in the Paleozoic along a suture that lies about 10 mi (16 km) west of the site (Klitgord, 1995) (Figure 2.5-17 and Figure 2.5-23). The plate-scale suture is defined by a distinct north-northeast-trending magnetic anomaly that dips easterly between 35 and 45 degrees and lies about 7.5 to 9 mi (12 to 14.5 km) beneath the CCNPP site (Glover, 1995b) (Figure 2.5-17). Directly west of the suture lies the north to northeast-trending Taylorsville Basin (LeTourneau, 2003) and to the east, the postulated Queen Anne Mesozoic rift basin (Benson, 1992) (Figure 2.5-9). These ~~Mesozoic~~rift basins are delineated from geophysical data subject to alternate interpretations and a limited number of deep boreholes that penetrate the ~~crust~~Coastal Plain section located outside the Site Area,, and generally are considered approximately located where buried beneath the Coastal Plain (Jacobeen, 1972) (Hansen, 1986) (Benson, 1992) (LeTourneau, 2003). ~~Most authors interpret Mesozoic basins directly west or east of the site; however, b~~Because the available geologic information used to constrain the basin locations is sparse, some authors, but not all, depict the CCNPP site area to be underlain by a Mesozoic basin (Klitgord, 1988) (Schlische, 1990) (Horton, 1991) (Benson, 1992) (Klitgord, 1995) (Withjack, 1998) (LeTourneau, 2003) (Figure 2.5-10, Figure 2.5-12, Figure 2.5-16, and Figure 2.5-22). However, based on ~~the basis of~~ a review of existing published geologic literature, ~~site-specific data, and field reconnaissance, suggests~~ there is no known basin-related fault or geologic evidence of basin-related faulting in the basement directly beneath the CCNPP site area.

RAI 130 02.05.01-54

RAI 130 02.05.01-54

RAI 130 02.05.01-54

Recent 1:24,000-scale mapping (Glaser, 2003b) (Glaser, 2003c) for Calvert County and St. Mary's County shows the stratigraphy at the CCNPP site area consisting of nearly flat-lying Cenozoic Coastal Plain sediments that have accumulated within the west-central part of the Salisbury Embayment (Figure 2.5-32 and Figure 2.5-33). The Salisbury Embayment is defined as a regional depocenter that has undergone slow crustal and regional downwarping as a result of sediment overburden during the Early Cretaceous and much of the Tertiary. The Coastal Plain deposits within this region of the Salisbury Embayment generally strike northeast-southwest and have a gentle dip to the southeast at angles close to or less than one to two degrees (Figure 2.5-32 and Figure 2.5-33). The gentle southerly dip of the sediments result in a surface outcrop pattern in which the strata become successively younger in a southeast direction across the embayment. The gentle-dipping to flat-lying Miocene Coastal Plain deposits are exposed in the steep cliffs along the western shoreline of Chesapeake Bay and provide excellent exposures to assess the presence or absence of tectonic-related structures.

Local geologic cross sections of the site area depict unfaulted, southeast-dipping Eocene-Miocene Coastal Plain sediments in an unconformable contact with overlying Pliocene Upland deposits (Glaser, 1994) (Achmad, 1997) (Glaser, 2003b) (Glaser, 2003c) (Figure 2.5-13, Figure 2.5-32, and Figure 2.5-33). No faults or folds are depicted on these geologic cross sections. A review of an Early Site Review report (BGE, 1977), i.e. Perryman site, and a review of the Preliminary Safety Analysis Report for the Douglas Point site (Potomac Electric Power Company, 1973), located along the eastern shore of the Potomac River about 45 mi (72 km) west-southwest of the CCNPP site, also reported no faults or folds within a 5 mi (8 km) radius of the CCNPP site. The Updated Final Safety Analysis Report for the Hope Creek site, located in New Jersey along the northern shore of Delaware Bay, also was reviewed for tectonic features previously identified within 5 mi (8 km) of the CCNPP site, yet none were identified (PSEG, 2002). Review of a seismic source characterization study (URS, 2000) for a liquidefied natural gas plant at Cove Point, about 3 mi (5 km) southeast of the site, also identified no faults or folds projecting toward or underlying the CCNPP site area.

On the basis of literature review, and aerial and field reconnaissance, the only potential structural features at and within the CCNPP site area consist of a hypothetical buried northeast-trending fault (Hansen, 1986), two inferred east-facing monoclines developed within Mesozoic and Tertiary deposits along the western shore of Chesapeake Bay (McCartan, 1995), and multiple subtle folds or inflections in Miocene strata and a postulated fault directly south of the site (Kidwell, 1997) (Figure 2.5-25). The Hillville fault of Hansen and Edwards (Hansen, 1986) and inferred fold of McCartan (McCartan, 1995) and Kidwell (Kidwell, 1997) are described in Sections 2.5.1.1.4.4.4 and Section 2.5.3. As previously discussed in Section 2.5.1.1.4.4.4, none of these features are considered capable tectonic sources, as defined in RG 1.165, Appendix A. Each of these features is discussed briefly below. Only the Hillville fault has been mapped within or directly at the 5 mi (8 km) radius of the CCNPP site area (Figure 2.5-27, Figure 2.5-28, and Figure 2.5-32).

Hillville fault of Hansen and Edwards (Hansen, 1986): The 26 mile long Hillville fault approaches to within 5 mi (8 km) of the CCNPP site (Figure 2.5-32). The fault consists of a northeast-striking zone of steep southeast-dipping reverse faults that coincide with the Sussex-Currioman Bay aeromagnetic anomaly. The style and location of faulting are based on seismic reflection data collected about 9 mi (14 km) west-southwest of the site. A seismic line imaged a narrow zone of discontinuities that vertically separate basement by as much as 250 ft (76 m) (Hansen, 1978). Hansen and Edwards (Hansen, 1986) interpret this offset as part of a larger lithotectonic terrane boundary that separates basement rocks associated with Triassic rift basins on the west and low-grade metamorphic basement on the east. The Hillville fault may represent a Paleozoic suture zone that was reactivated in the Mesozoic and Early Tertiary. Based on stratigraphic correlation between boreholes within Tertiary Coastal Plain deposits, Hansen and Edwards (Hansen, 1986) speculate that the Hillville fault was last active in the Early Paleocene. There is no pre-EPRI and post-EPRI (1986) seismicity spatially associated with this feature (Figure 2.5-25) nor is there any geomorphic evidence of Quaternary deformation. The Hillville fault is not considered a capable tectonic source.

In addition, two speculative and poorly constrained east-facing monoclines along the western margin of Chesapeake Bay are mapped within the 5 mi (8 km) radius of the CCNPP site area. East-facing monoclines (McCartan, 1995): The unnamed monoclines are not depicted on any geologic maps of the area, including those by the authors, but they are shown on geologic cross sections that trend northwest-southeast across the existing site and south of the CCNPP site near the Patuxent River (McCartan, 1995) (Figure 2.5-25). East-facing monoclines are inferred beneath Chesapeake Bay at about 2 and 10 mi (3.2 to 16 km) east and southeast, respectively, from the CCNPP site. Along a northerly trench, the two monoclines delineate a continuous north-trending, east-facing monocline. As mapped in cross section and inferred in plan view, the monoclines trend approximately north along the western shore of Chesapeake Bay. The monoclines exhibit a west-side up sense of structural relief that projects into the Miocene Choptank Formation (McCartan, 1995). The overlying Late Miocene St. Marys Formation is not shown as warped. Although no published geologic data are available to substantiate the existence of the monoclines, McCartan (McCartan, 1995) believes the distinct elevation change across Chesapeake Bay and the apparent linear nature of Calvert Cliffs are tectonically controlled. CCNPP site and aerial reconnaissance, coupled with literature review, for the CCNPP Unit 3 study strongly support a non-tectonic origin for the physiographic differences across the Chesapeake Bay (Section 2.5.1.1.4.4.4). There is no pre-EPRI or post-EPRI (1986) seismicity spatially associated with this feature, nor is there geologic data to suggest that the monocline proposed by McCartan (McCartan, 1995) is a capable tectonic source.

Multiple subtle folds or inflections developed in Miocene Coastal Plain strata including a postulated fault are mapped in the cliff exposures along the west side of Chesapeake Bay. Kidwell's (Kidwell, 1997) postulated folds and fault: Kidwell (Kidwell, 1988) (Kidwell, 1997) prepared over 300 lithostratigraphic columns along a 25 mi (40 km) long stretch of Calvert Cliffs that intersect much of the CCNPP site (Figure 2.5-30). When these stratigraphic columns are compiled into a cross section, they collectively provide a 25 mi (40 km) long nearly continuous exposure of Miocene, Pliocene and Quaternary deposits. Kidwell's (Kidwell, 1997) stratigraphic analysis indicates that the Miocene Coastal Plain deposits strike northeast and dip very shallow between 1 and 2 degrees to the south-southeast, which is consistent with the findings of others (McCartan, 1995) (Glaser 2003b) (Glaser, 2003c). The regional southeast-dipping strata are disrupted occasionally by several low amplitude broad undulations developed within Miocene Coastal Plain deposits (Figure 2.5-30). The stratigraphic undulations are interpreted as monoclines and asymmetrical anticlines by Kidwell (Kidwell, 1997). In general, the undulatory stratigraphic contacts coincide with basal unconformities having wavelengths of 2.5 to 5 mi (4 to 8 km) and amplitudes of 10 to 11 ft (approximately 3 meters). Based on prominent stratigraphic truncations, the inferred warping decreases upsection into the overlying upper Miocene St. Marys Formation. Any inferred folding of the overlying Pliocene and Quaternary fluvial deposits is poorly constrained and can be readily explained by highly variable undulating unconformities.

Near Moran Landing, about 1.2 mi (1.9 km) south of the site, Kidwell (Kidwell, 1997) interprets an apparent 6 to 10 ft (2 to 3 m) elevation change in Miocene strata, and a 3 to 12 (0.9 to 3.7 m) ft elevation change in Pliocene and Quaternary (?) fluvial material (Figure 2.5-25 and Figure 2.5-30). Kidwell (Kidwell, 1997) infers the presence of a fault to explain the difference in elevation of strata across Moran Landing. The postulated fault is not shown on the Kidwell (Kidwell, 1997) section, or any published geologic map, however the inferred location is approximately 1.2 mi (1.9 m) south of the CCNPP site. The hypothesized fault is not exposed in the cliff face, but Kidwell (Kidwell, 1997) postulates the presence of a fault, and is based entirely on a change in elevation and bedding dip of Miocene stratigraphic boundaries projected across the fluvial valley of Moran Landing. Kidwell (Kidwell, 1997) postulates that the fault strikes northeast and exhibits a north-side down sense of separation across all the geologic units (Miocene through Quaternary). With regard to the apparent elevation changes for the Pliocene and Quaternary unconformities, these can be readily explained by channeling and highly irregular erosional surfaces. Field and aerial reconnaissance, coupled with interpretation of aerial photography and LiDAR data (Section 2.5.3.1 for additional information regarding the general methodology) conducted as part of this CCNPP Unit 3 study revealed no features suggestive of tectonic deformation developed in the surrounding Pliocene and Quaternary surfaces.

There is no pre-EPRI or post-EPRI study (EPRI, 1986) seismicity spatially associated with the Kidwell (Kidwell, 1997) features, the hypothetical features are not aligned or associated with gravity and magnetic anomalies, nor is there data to indicate that the features proposed by Kidwell (Kidwell, 1997) are capable tectonic sources.

The most detailed subsurface exploration of the site was performed by Dames & Moore as part of the original PSAR (BGE, 1968) for the existing CCNPP foundation and supporting structures. The PSAR study included drilling as many as 85 geotechnical boreholes, collecting downhole geophysical data, and acquiring seismic refraction data across the site. Dames and Moore (BGE, 1968) developed geologic cross sections extending from Highway 2/4 northwest of the site to Camp Conoy on the southeast which provide valuable subsurface information on the lateral continuity of Miocene Coastal Plain sediments and Pliocene Upland deposits (Figure 2.5-32 and Figure 2.5-34). Cross sections C-C' and D-D' pre-date site development and intersect the existing and proposed CCNPP site for structures trending north-northeast, parallel to the regional structural grain. These sections depict a nearly flat-lying, undeformed geologic contact between the Middle Miocene Piney Point Formation and the overlying Middle Miocene Calvert Formation at about -200 ft (-61 m) msl (Figure 2.5-41 and Figure 2.5-42).

Geologic sections developed from geotechnical borehole data collected as part of the CCNPP Unit 3 study also provide additional detailed sedimentological and structural relations for the upper approximately 400 ft (122 m) of strata directly beneath the footprint of the site. Similar to the previous cross sections prepared for the site, new geologic borehole data support the interpretation of flat-lying and unfaulted Miocene and Pliocene stratigraphy at the CCNPP site (Figure 2.5-39 and Figure 2.5-43). A cross section prepared oblique to previously mapped northeast-trending structures (i.e., Hillville fault), inferred folds (McCartan, 1995) (Kidwell, 1997), and the fault of Kidwell (Kidwell, 1997) shows nearly flat-lying Miocene and Pliocene stratigraphy directly below the CCNPP site. Multiple key stratigraphic markers provide evidence for the absence of Miocene-Pliocene faulting and folding beneath the site. Minor perturbations are present across the Miocene-Pliocene stratigraphic boundary, as well as other Miocene-related boundaries, however these minor elevation changes are most likely related to the irregular nature of the fluvial unconformities and are not tectonic-related.

Numerous investigations of the Calvert Cliffs coastline over many decades by government researchers, stratigraphers, and by consultants for Baltimore Gas and Electric, as well as investigations for the CCNPP Unit 3, have reported no visible signs of tectonic deformation within the exposed Miocene deposits near the site, with the only exception being that of Kidwell (Kidwell, 1997) (Figure 2.5-44). Collectively, the majority of published and unpublished geologic cross sections compiled for much of the site area and site, coupled with regional sections (Achmad, 1997) (Glaser, 2003b) (Glaser, 2003c) and site and aerial reconnaissance, indicate the absence of Pliocene and younger faulting and folding. A review and interpretation of aerial photography, digital elevation models, and LiDAR data of the CCNPP site area, coupled with aerial reconnaissance, identified few discontinuous north to northeast-striking lineaments. None of these lineaments were interpreted as fault-related, nor coincident with the Hillville fault or the other previously inferred Miocene-Pliocene structures mapped by

McCartan (McCartan, 1995) and Kidwell (Kidwell, 1997) (Section 2.5.3). A review of regional geologic sections and interpretation of LiDAR data suggest that the features postulated by Kidwell (Kidwell, 1997), if present, are not moderate or prominent structures, and do not deform Pliocene and Quaternary strata. In summary, on the basis of regional and site geologic and geomorphic data, there are no known faults within the site area, with the exception of the poorly constrained Hillville fault that lies along the northwestern perimeter of the 5 mi (8 km) radius of the site (Hansen, 1986).

2.5.1.2.5 Site Area Geologic Hazard Evaluation

No geologic hazards have been identified within the CCNPP site area. No geologic units at the site are subject to dissolution. No deformation zones were encountered in the exploration or excavation for CCNPP Units 1 and 2 and none have been encountered in the site investigation for CCNPP Unit 3. Because the CCNPP Unit 3 plant site is located at an elevation of approximately 85 ft (26 m) msl and approximately 1,000 ft (305 m) from the Chesapeake Bay shoreline, it is unlikely that shoreline erosion or flooding will impact the CCNPP site.

2.5.1.2.6 Site Engineering Geology Evaluation

2.5.1.2.6.1 Engineering Soil Properties and Behavior of Foundation Materials

Engineering soil properties, including index properties, static and dynamic strength, and compressibility are discussed in Section 2.5.4. Variability and distribution of properties for the foundation bearing soils will be evaluated and mapped as the excavation is completed.

Settlement monitoring will based on analyses performed for the final design.

2.5.1.2.6.2 Zones of Alteration, Weathering, and Structural Weakness

No unusual weathering profiles have been encountered during the site investigation. No dissolution is expected to affect foundations. Any noted desiccation, weathering zones, joints or fractures will be mapped during excavation and evaluated.

2.5.1.2.6.3 Deformational Zones

No deformation zones were encountered in the exploration or excavation for CCNPP Units 1 and 2 and none have been encountered in the site investigation for CCNPP Unit 3. Excavation mapping is required during construction and any noted deformational zones will be evaluated. No capable tectonic sources as defined by Regulatory Guide 1.165 (NRC, 1997) exist in the CCNPP site region.

2.5.1.2.6.4 Prior Earthquake Effects

Outcrops are rare within the CCNPP site area. Studies of the CCNPP Unit 1 and 2 excavation, available outcrops, and extensive exposures along the western shore of Chesapeake Bay have not indicated any evidence for earthquake activity that affected the Miocene deposits. Potential liquefaction features were investigated as part of the CCNPP Unit 3 site investigation, which included a review of existing literature, discussion with researchers familiar with the local Quaternary geology, aerial and field reconnaissance, and review of site vicinity aerial photography (multiple vantages within a 5 mile radius of the site). During the field reconnaissance along the Potomac and Patuxent Rivers, and where outcrops of Quaternary deposits were available, exposures were evaluated for liquefaction-related deformation features. Quaternary fluvial deposits inset into Calvert Cliffs and partially exposed along the west side of Chesapeake Bay were evaluated for liquefaction-related features. No liquefaction features were identified. Several small tributaries intersecting the site were also inspected;

RAI 72
02.05.02-28

RAI 72 02.05.02-28

however, no suspicious features were identified in the limited exposures available for review. The aerial reconnaissance consisted of traverses across the Potomac, Patuxend and Rappahannock Rivers where Quaternary fluvial terraces (e.g., potentially liquefiable deposits) were inspected for features that could be related to earthquake-induced liquefaction. A similar aerial reconnaissance of the Demarva Peninsula was performed. There is no evidence of earthquake-induced liquefaction in the State of Maryland (Crone, 2000) (Wheeler, 2005).

2.5.1.2.6.5 Effects of Human Activities

No mining operations, excessive extraction or injection of ground water or impoundment of water has occurred within the site area that can affect geologic conditions.

2.5.1.2.6.6 Site Ground Water Conditions

A detailed discussion of ground water conditions is provided in Section 2.4.12.

2.5.1.3 References

This section is added as a supplement to the U.S. EPR FSAR.

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Table 2.5-1 {Definitions of Classes Used in the Compilation of Quaternary Faults, Liquefaction Features, and Deformation in the Central and Eastern United States}

Class Category	Definition
Class A	Geologic evidence demonstrates the existence of a Quaternary fault of tectonic origin, whether the fault is exposed for mapping or inferred from liquefaction to other deformational features.
Class B	Geologic evidence demonstrates the existence of a fault or suggests Quaternary deformation, but either (1) the fault might not extend deeply enough to be a potential source of significant earthquakes, or (2) the currently available geologic evidence is too strong to confidently assign the feature to Class C but not strong enough to assign it to Class A.
Class C	Geologic evidence is insufficient to demonstrate (1) the existence of tectonic fault, or (2) Quaternary slip or deformation associated with the feature.
Class D	Geologic evidence demonstrates that the feature is not a tectonic fault or feature; this category includes features such as demonstrated joints or joint zones, landslides, erosional or fluvial scarps, or landforms resembling fault scarps, but of demonstrable non-tectonic origin.

Figure 2.5-1 {Map of Physiographic Province}

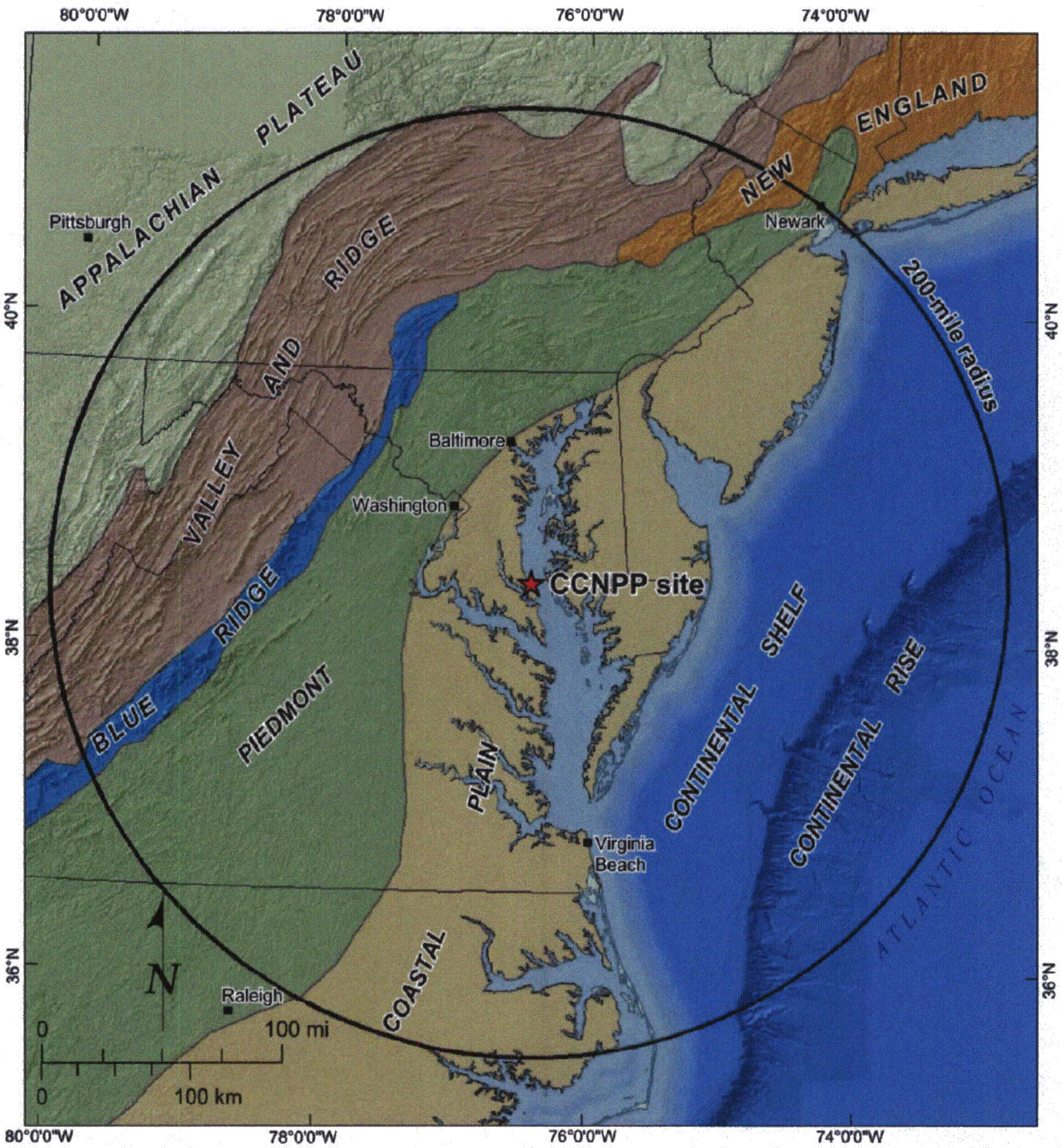


Figure 2.5-2 {Site Vicinity Topographic Map 25-Mile (40-Km) Radius}

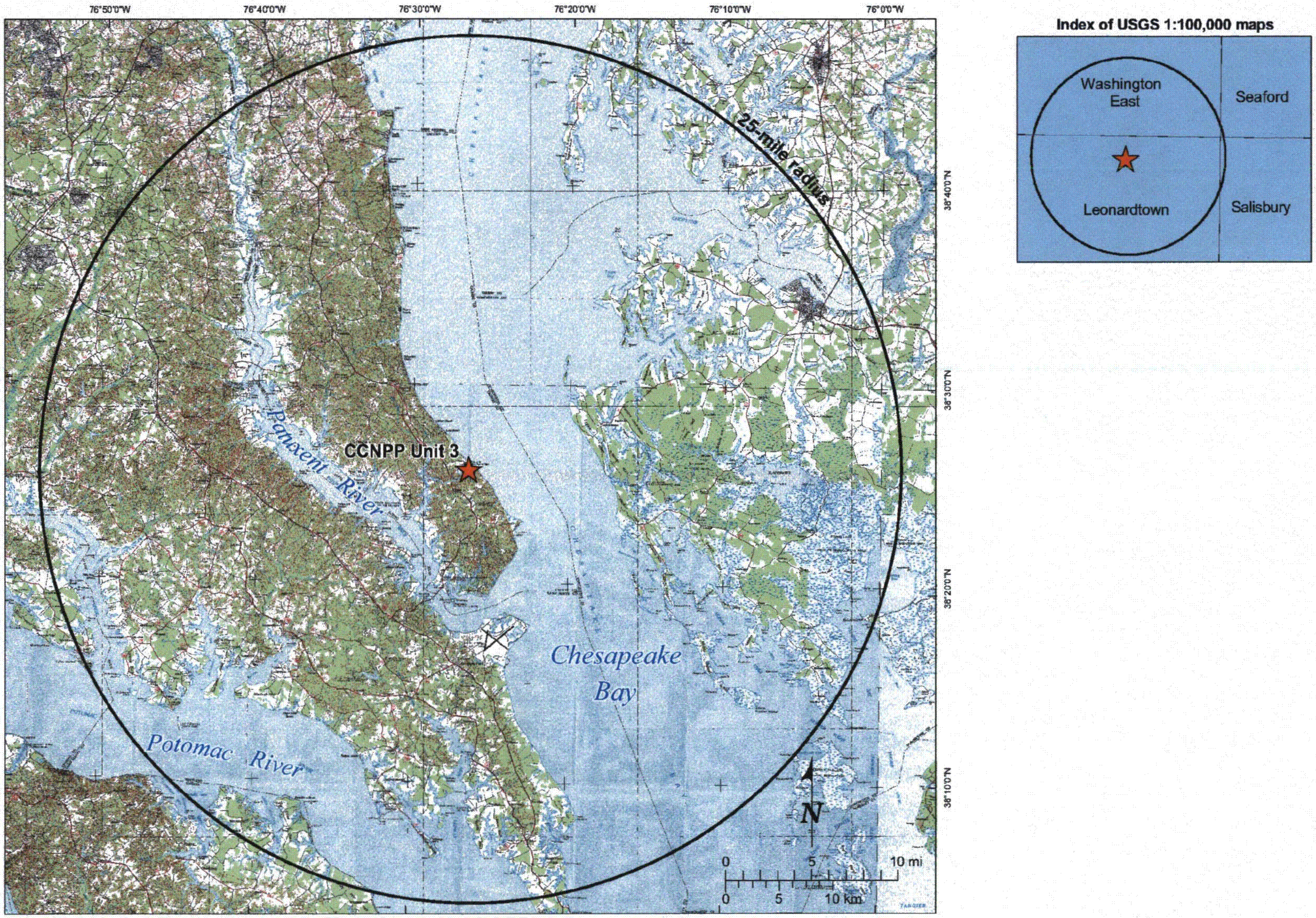
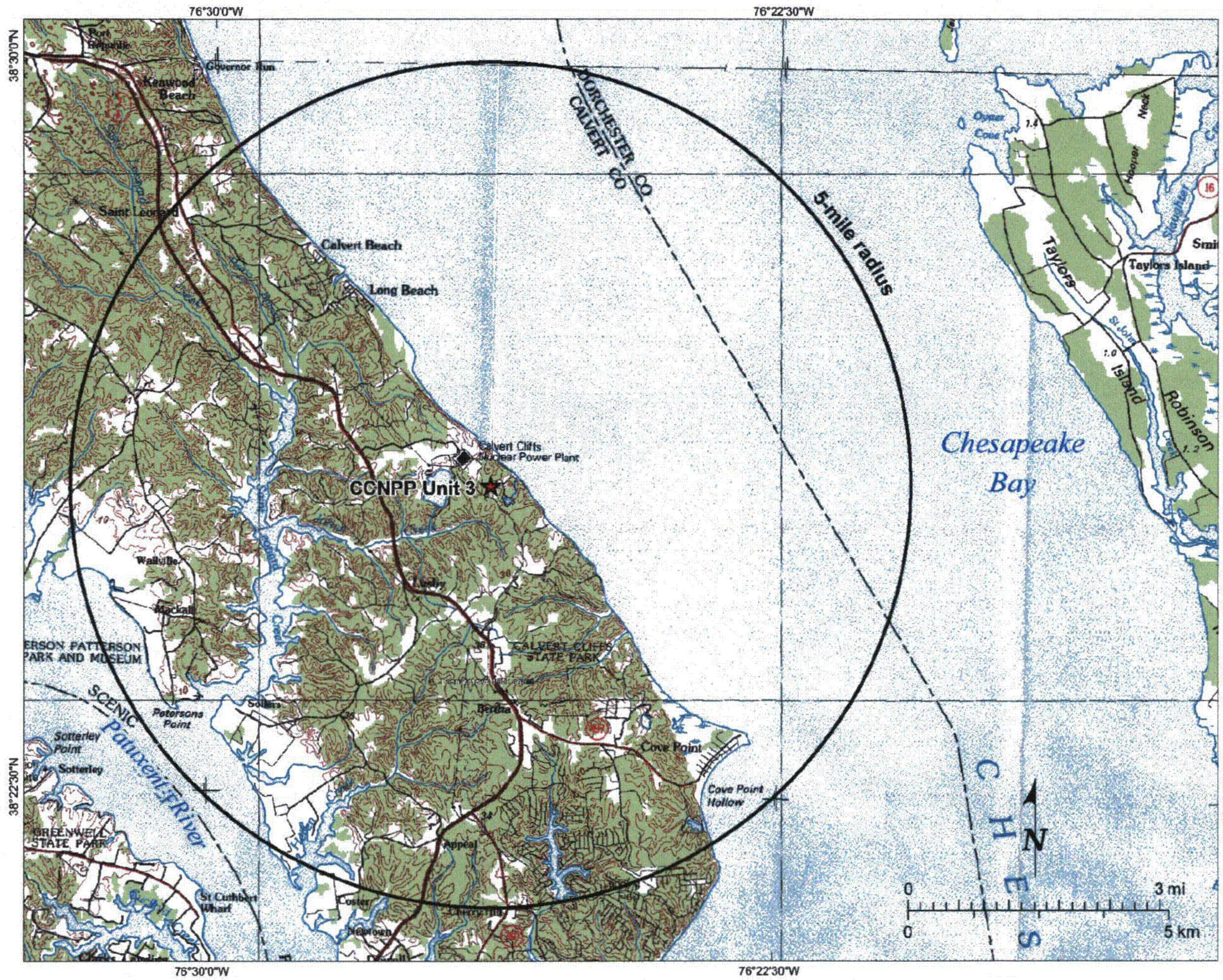


Figure 2.5-3 {Site Area Topographic Map 5-Mile (8-Km) Radius}



Base map: Leonardtown 30' x 60' U.S. Geological Survey Topographic Map

Figure 2.5-4 {Site Topographic Map 0.6-Mile (1-Km) Radius}

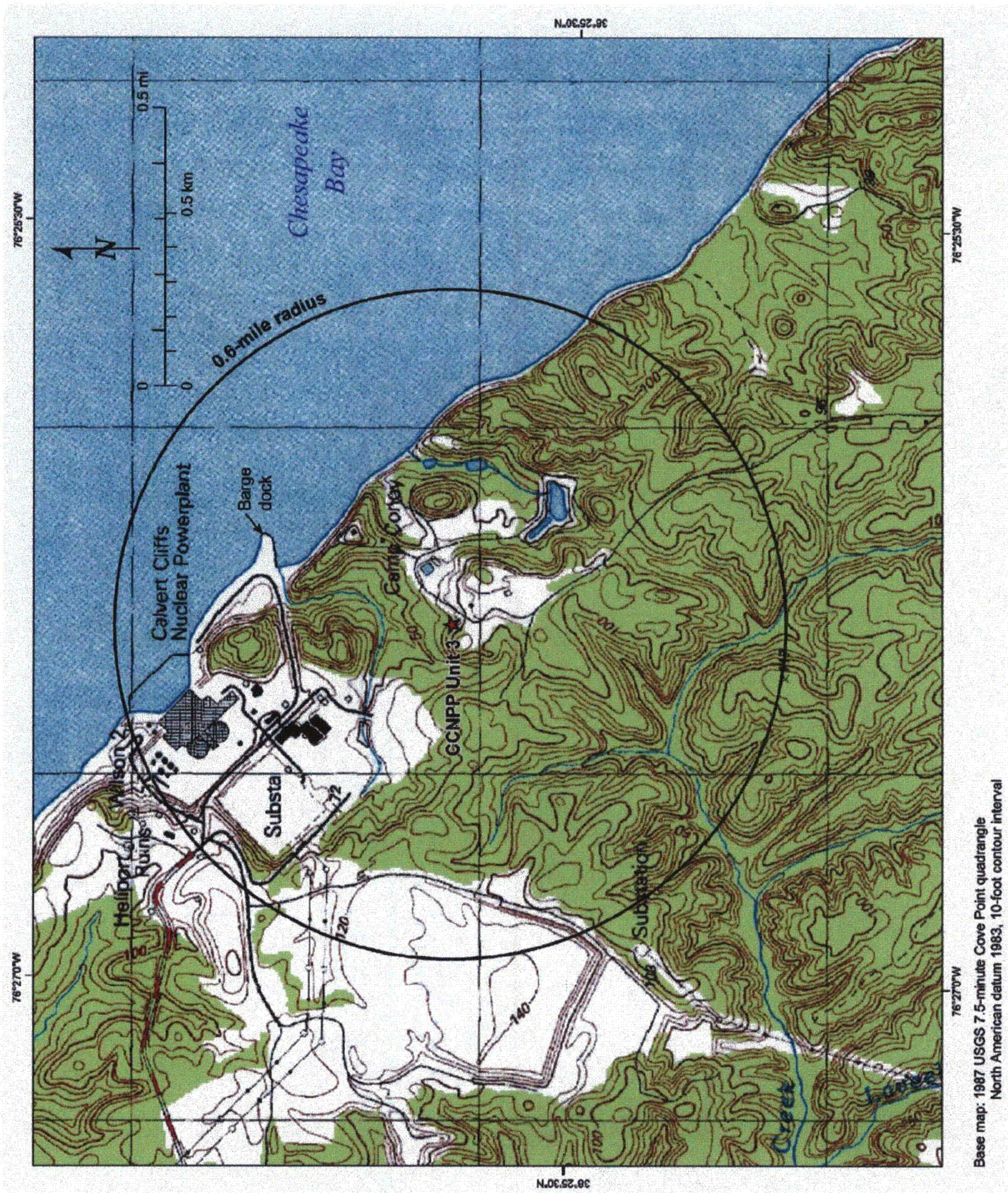


Figure 2.5-5 {Regional Geologic Map 200-mile (320-km) Radius}

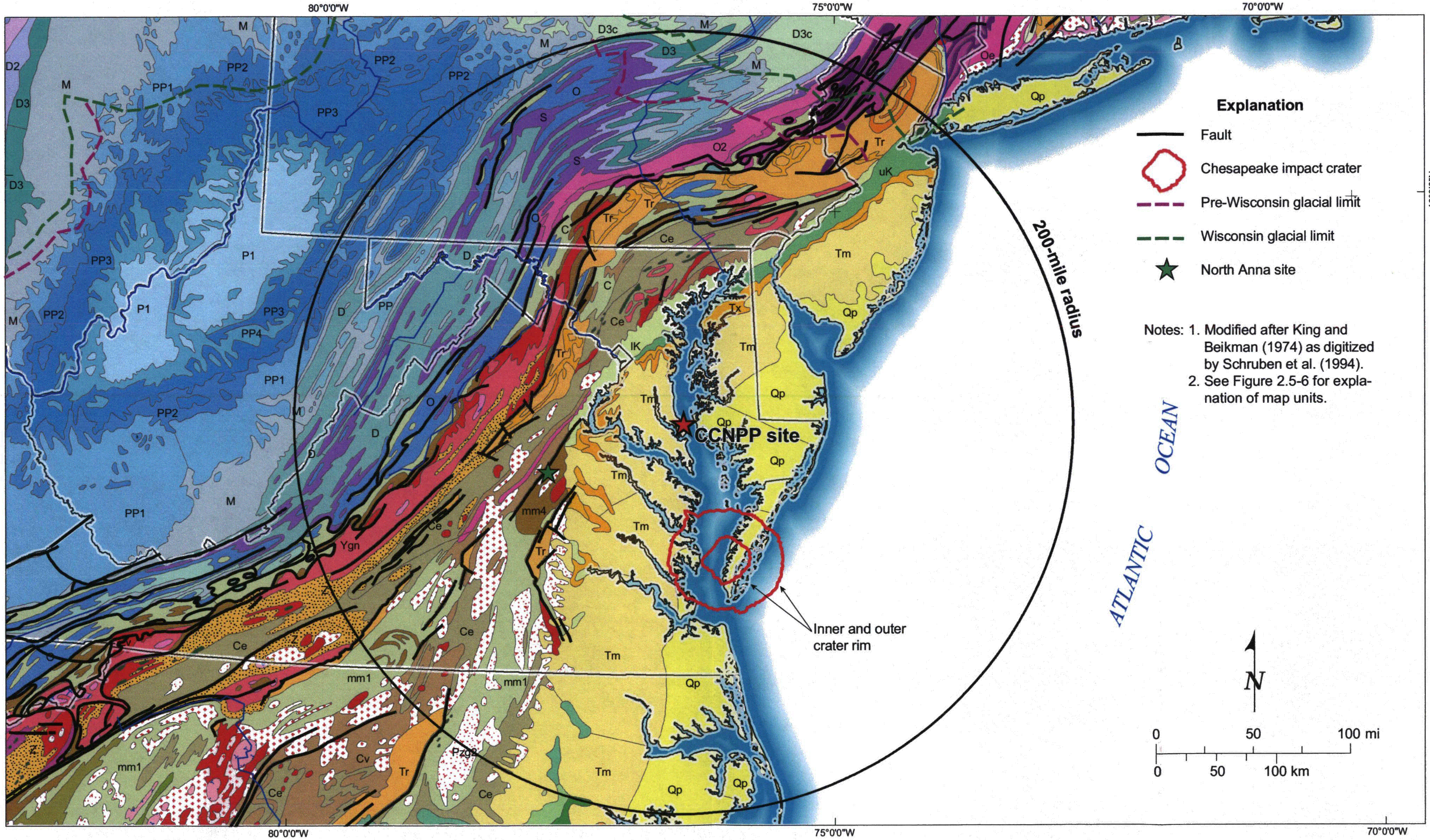


Figure 2.5-6 {Regional Geologic Map 200-mile (320-km) Radius Explanation}

Stratified Sequence

Cenozoic	Quaternary (Q)	Qp	Pleistocene		
Cenozoic	Tertiary (T)	Tm	Miocene		
		Te	Eocene		
		Tx	Paleocene		
Mesozoic	Cretaceous (K)	uK	Upper Cretaceous	uK4	Navarro Group
				uK1	Woodbine and Tuscalossa Group
		lK	Lower Cretaceous	lK3	Washita Group
	Triassic (Tr)			Tr	Triassic
Paleozoic	Permian (P)			P1	Wolfcampian Series
	Pennsylvanian (PP)	Upper		PP4	Virgilian Series
				PP3	Missourian Series
		Middle		PP2	Des Moinesian Series
		Lower		PP1	Atokan and Morrowan Series
	Mississipian (M)	M	Mississippian		
	Devonian (D)	D3	Upper Devonian	DS	Devonian and Silurian
		D2	Middle Devonian		
		D1	Lower Devonian		
	Silurian (S)	S	Silurian		
	Ordovician (O)	O2	Middle Ordovician-Mohawkian		
		O1	Lower Ordovician-Canadian	OC	Lower Ordovician and Cambrian-carbonate rocks
				lPz	Lower Paleozoic
	Cambrian (C)	C	Cambrian		
Precambrian	Precambrian	Z		Z	Z sedimentary rocks
		Y			

Continental Deposits

- D3c — Upper Devonian continental
- Cq — Basal Lower Cambrian clastic rocks

Eugeosynclinal Deposits

- De — Devonian eugeosynclinal
- Oe — Ordovician eugeosynclinal
- Ce — Cambrian eugeosynclinal

Volcanic Rocks

- Trv — Mafic Lava interbedded in Triassic Newark Group
- Ov — Ordovician volcanic rocks
- Cv — Cambrian volcanics
- Zv — Z volcanic rocks

Plutonic and Intrusive Rocks

- Tri — Triassic mafic intrusives
- um — Ultramafic rocks
- Pzg3 — Upper Paleozoic granitic rocks
- Pzmi — Paleozoic mafic intrusives
- Pzg2 — Middle Paleozoic granitic rocks
- Pzg1 — Lower Paleozoic granitic rocks
- Zg — Z granitic rocks
- Ya — Anorthosite

Metamorphic Rocks

- cat — Cataclastic rocks
- Metamorphic Complexes:
  - mm1 — Felsic paragneis and schist
  - mm2 — Mafic paragneiss (= hornblende, amphibolite)
  - mm3 — Migmatite
  - mm4 — Felsic orthogneiss (= granite gneiss)
- Ym — Paragneiss and schist
- Ygn — Orthogneiss

Figure 2.5-7 {Physiographic Map of Maryland}

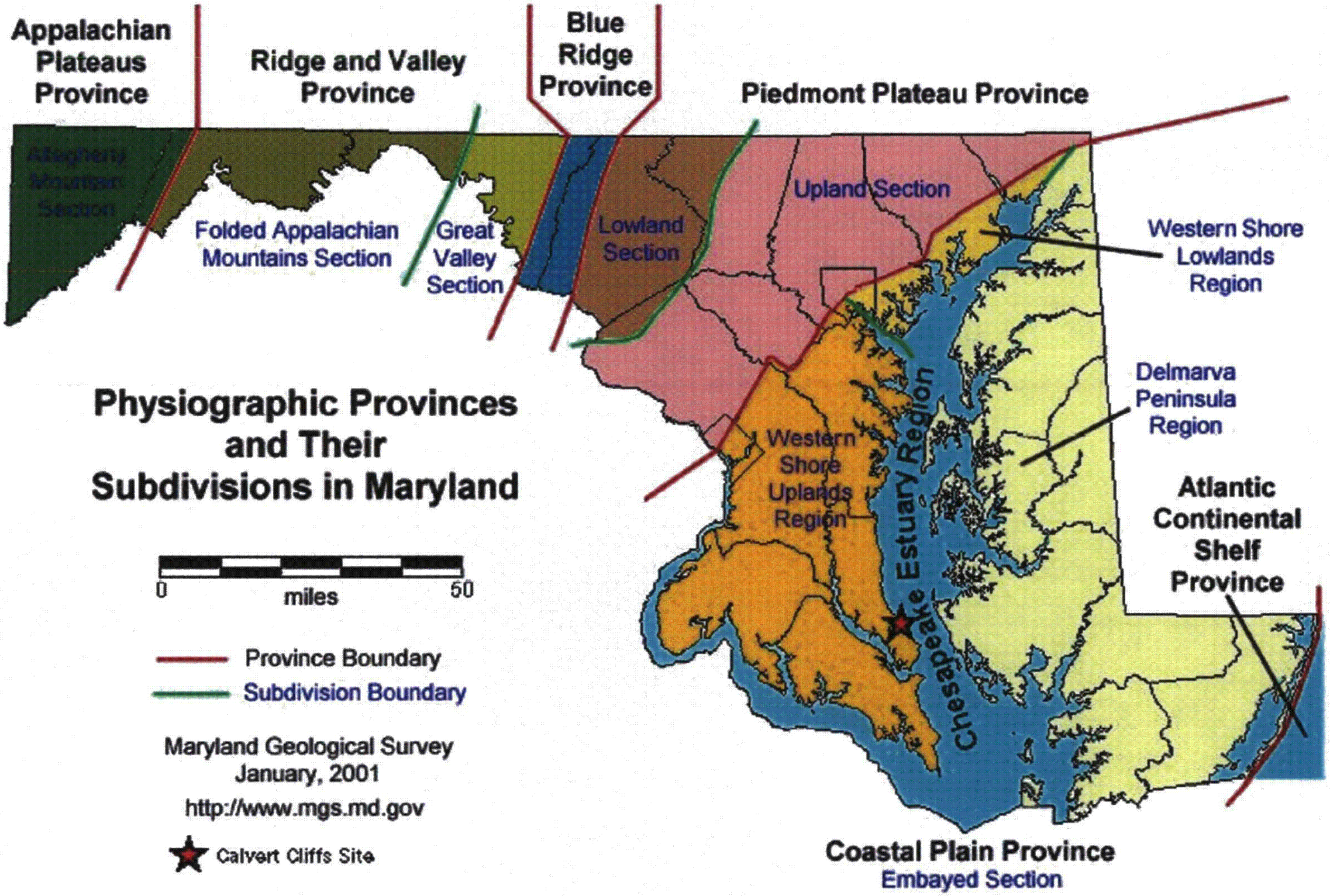


Figure 2.5-8 {Evolution of the Applachian Orogen}

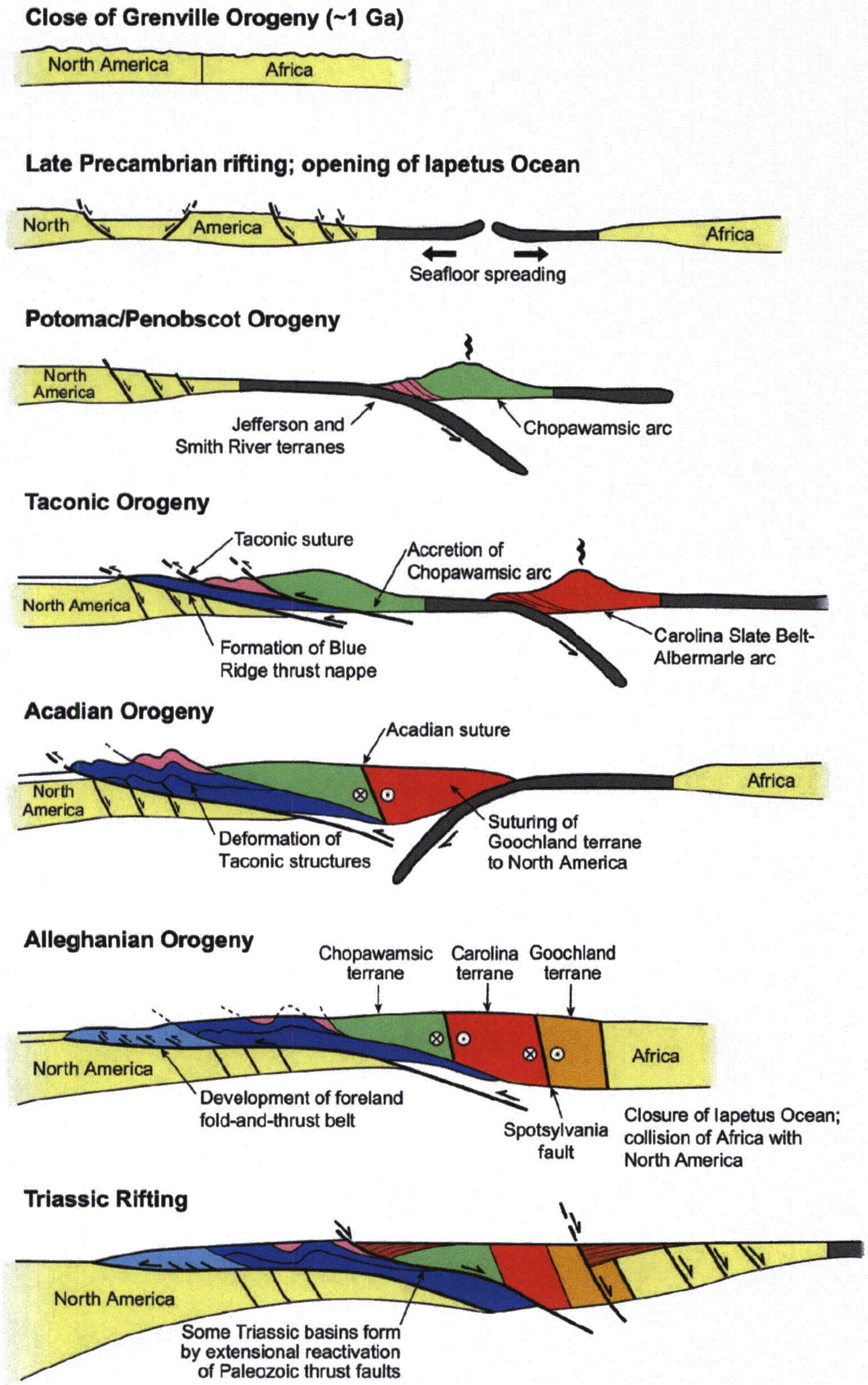


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Figure 2.5-9 {General Technostatigraphic Terrane Map 200-mile (320 km) Radius (modified from Horton 1991)}

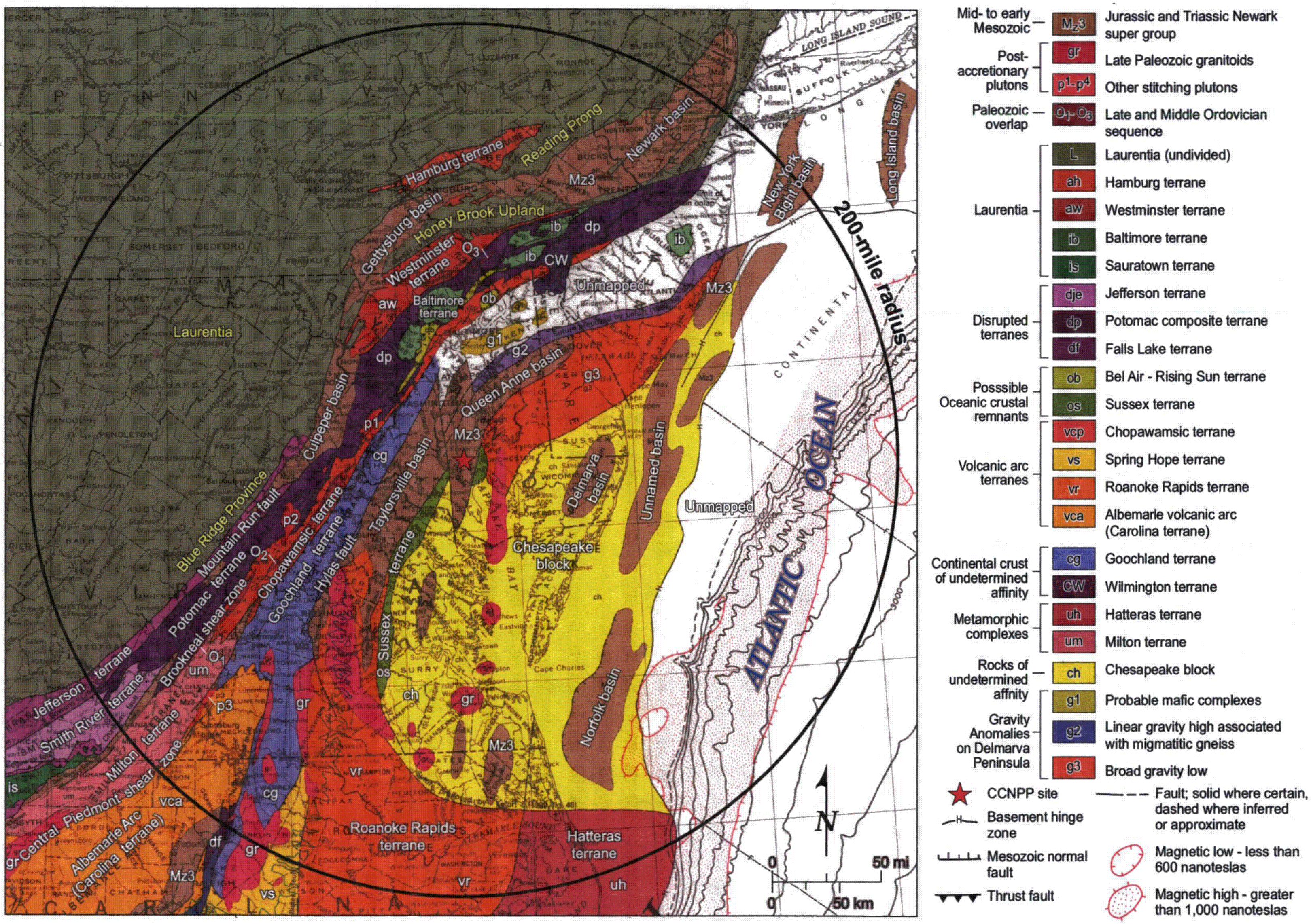

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Figure 2.5-10 {Map of Mesozoic Basins}

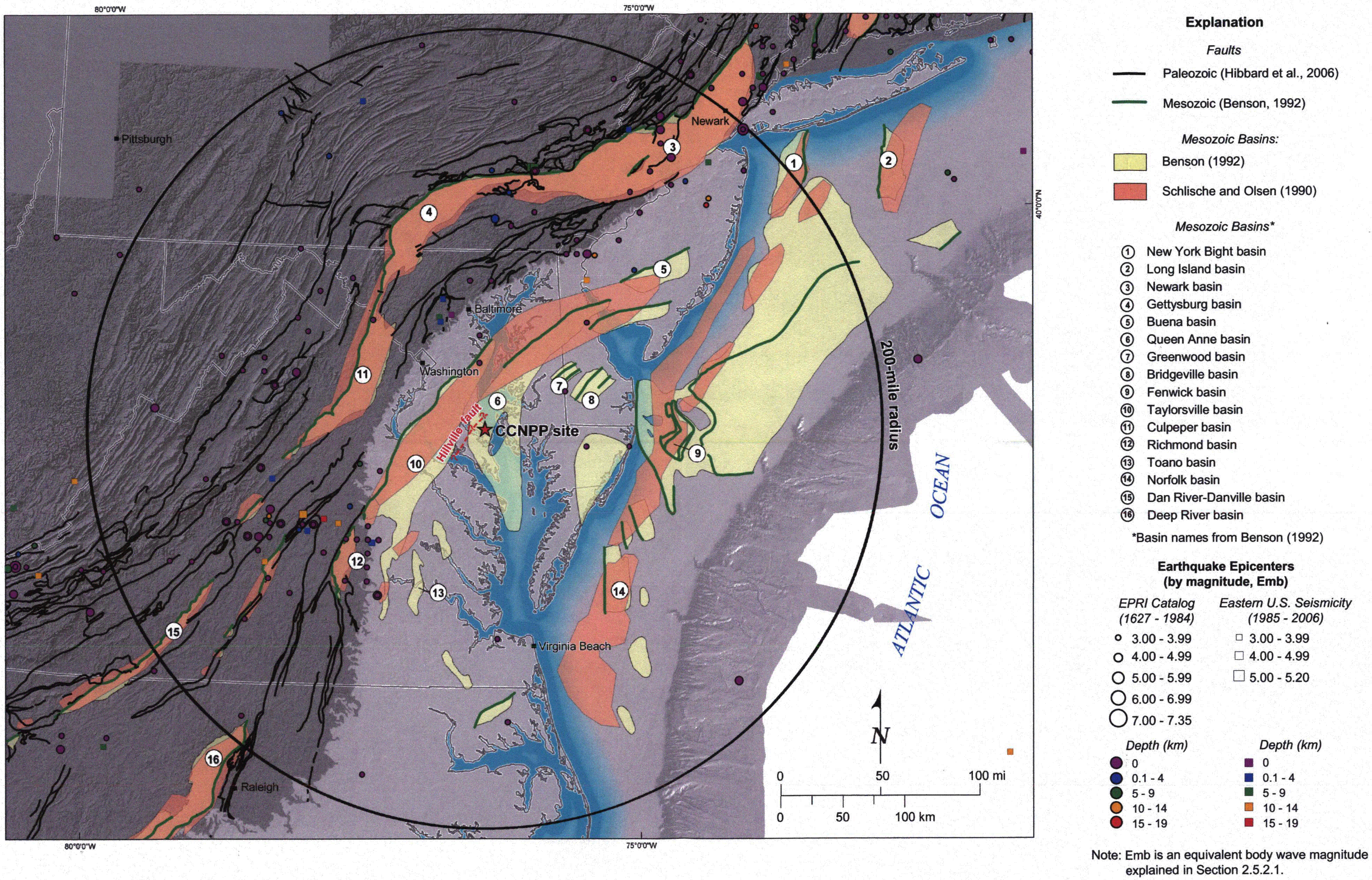

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Figure 2.5-11 {Lithologies of Basement Rocks from Coastal Plain Wells}

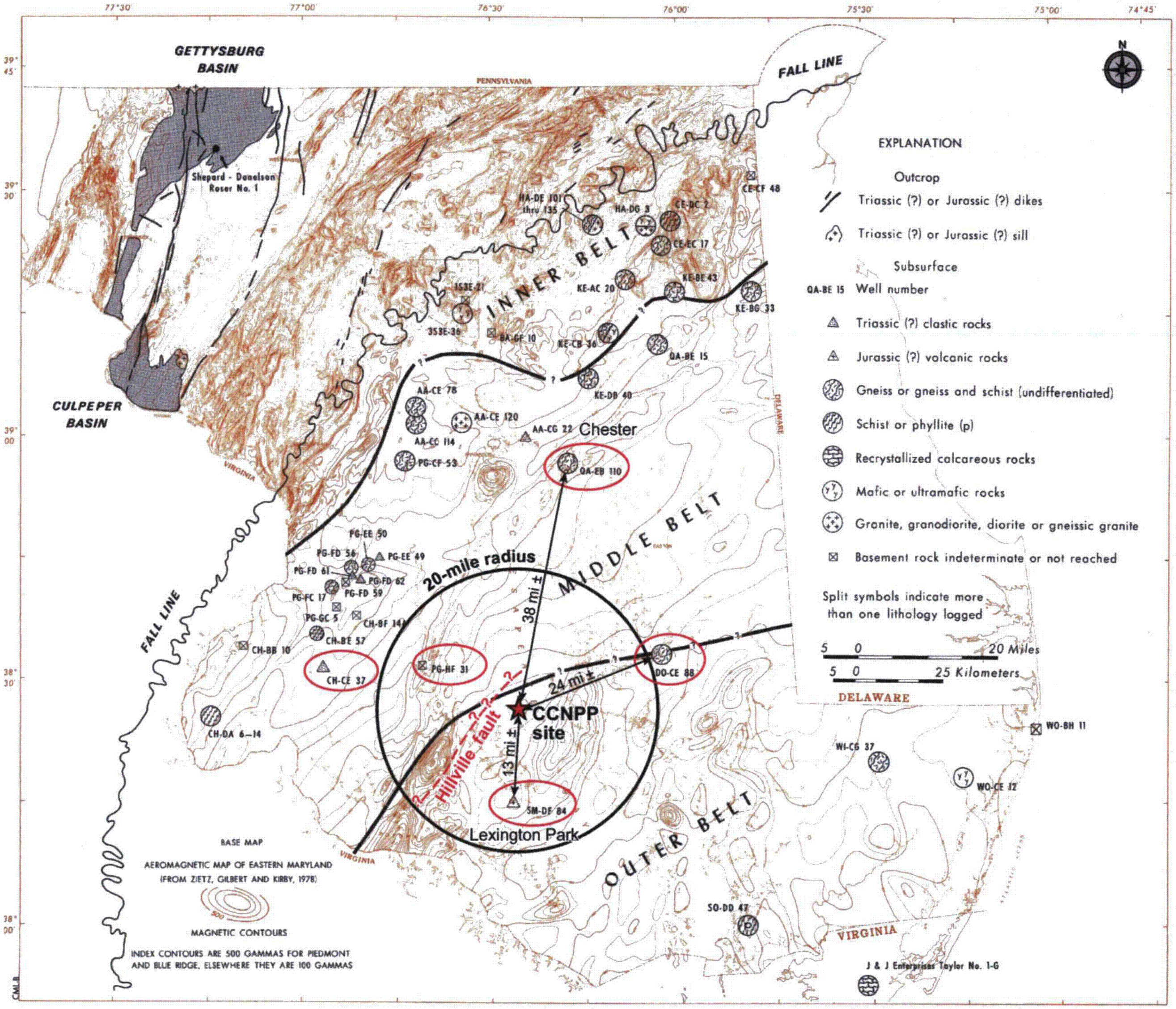


Figure 2.5-12 {Tectonic Features of the Mid-Atlantic Passive Margin}

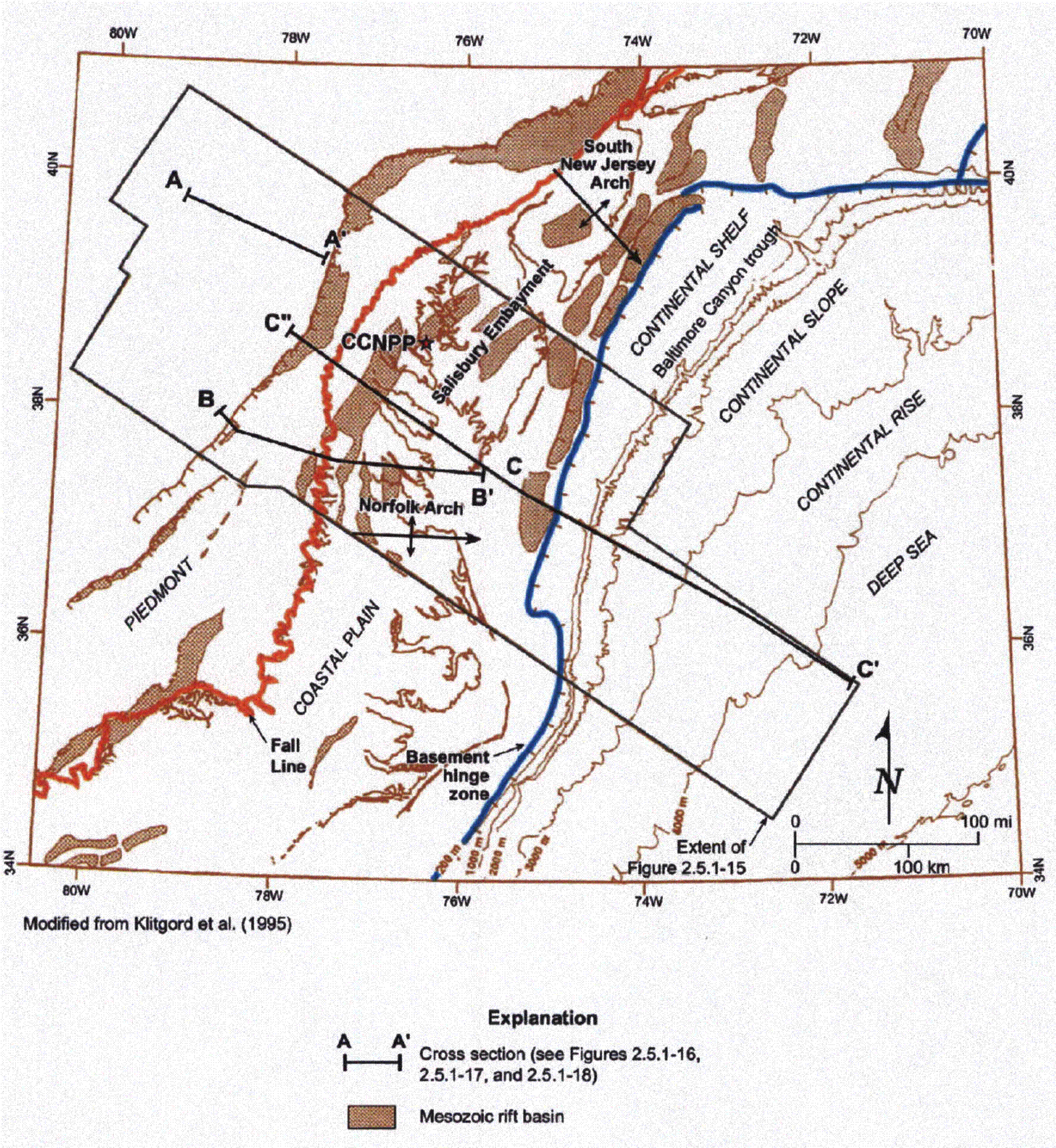
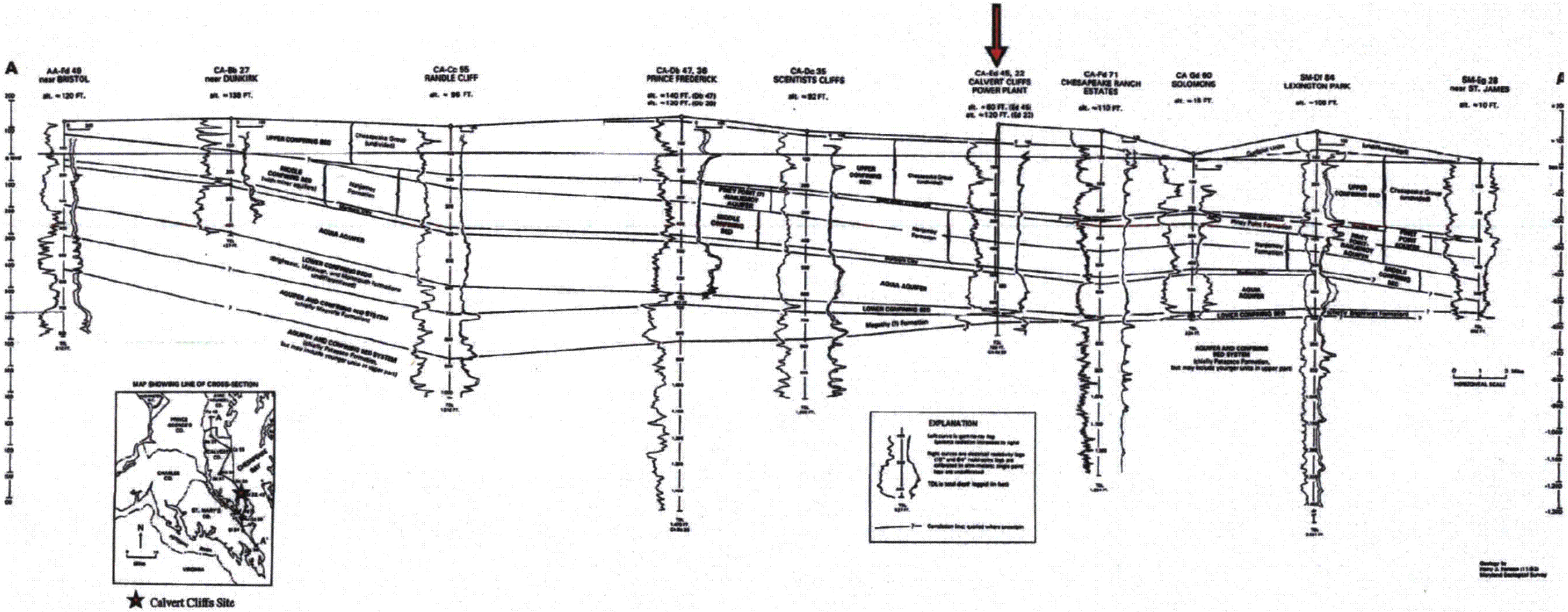
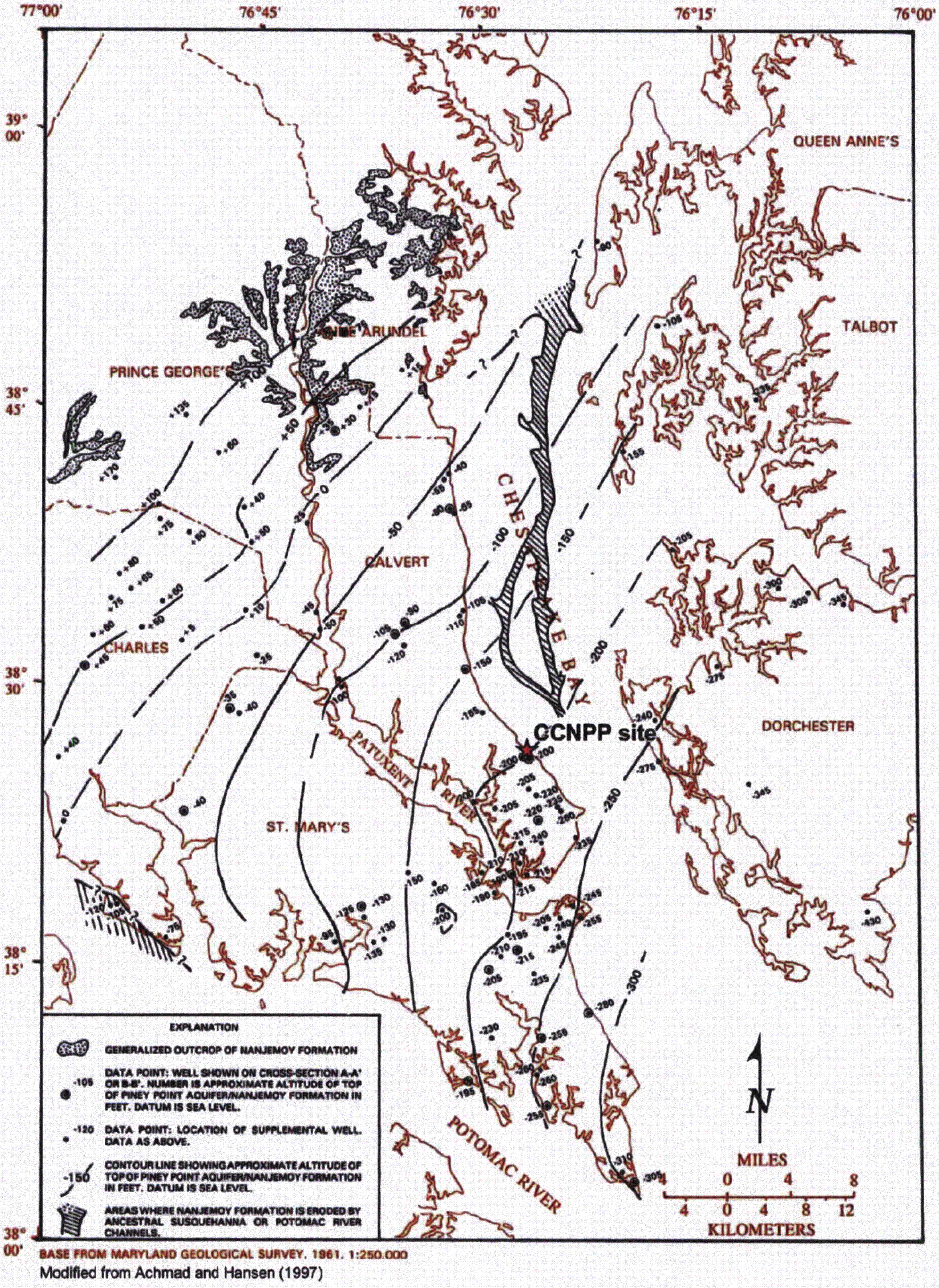


Figure 2.5-13 {Stratigraphic Cross-Section Through Anne Arundel, Calvert and St. Mary's Counties}



THE CCNPP SITE IS REPRESENTED BY THE WELLS CA-Ed 45 AND 22.

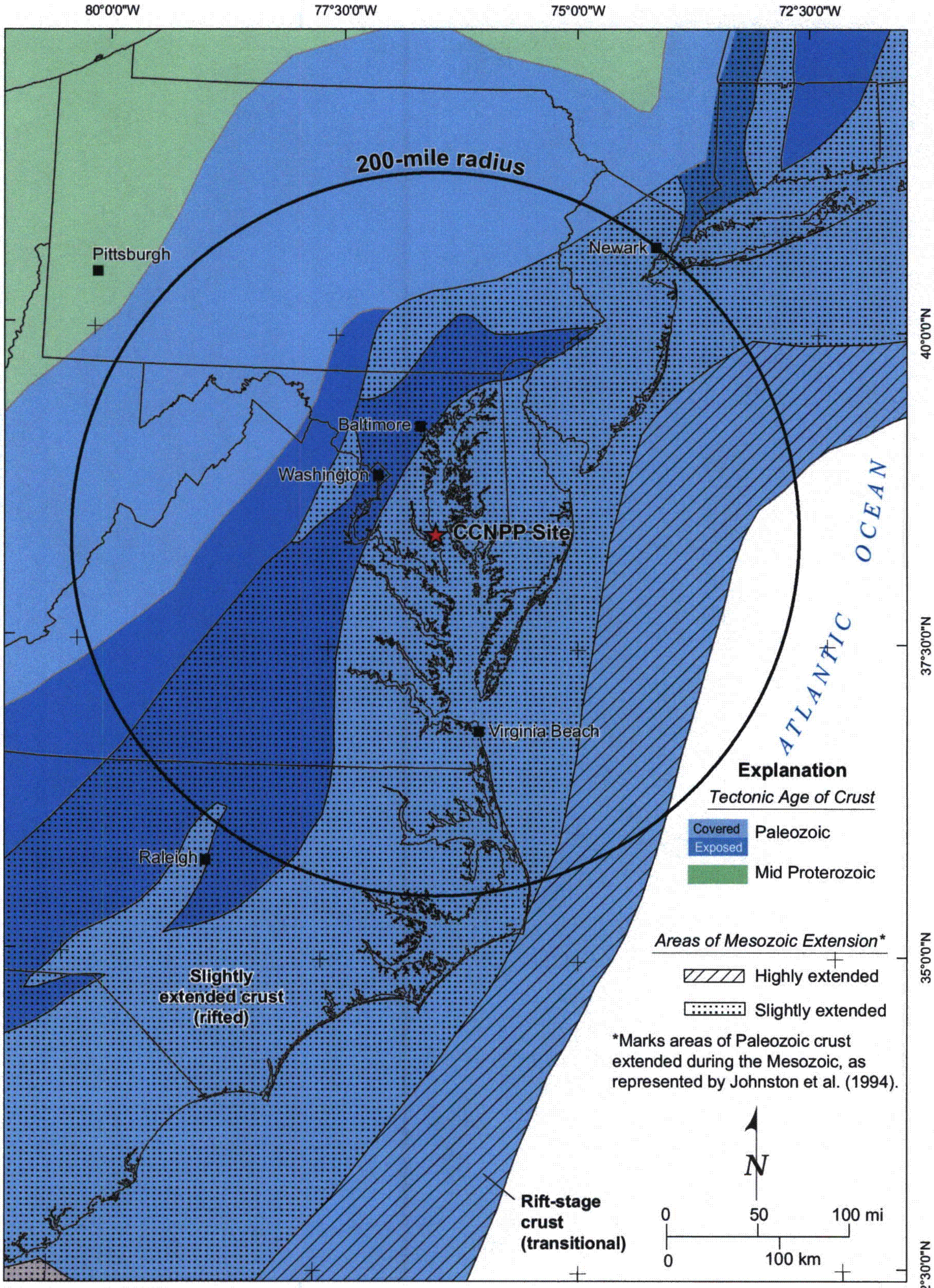
Figure 2.5-14 {Structure-Contour Map of the Top of the Piney Point-Nanjemoy Aquifer}



Modified from Achmad and Hansen (1997)

RAI 71
02.05.01-29

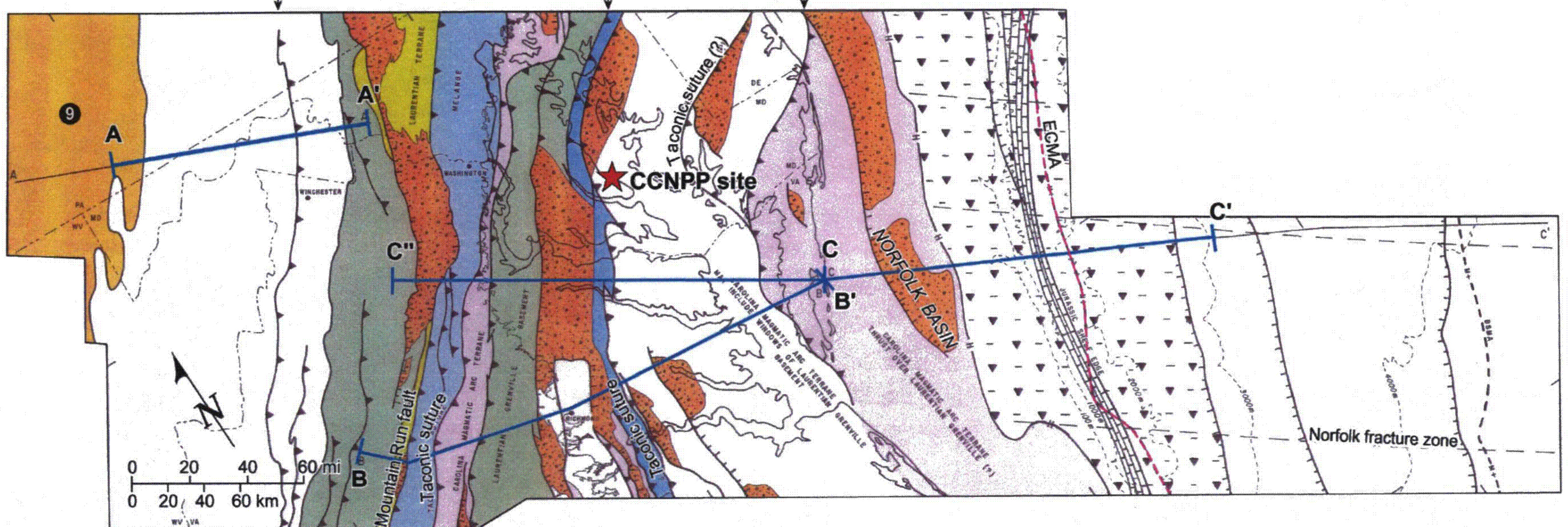
Figure 2.5-15 {Tectonic Age of Crust~~Crusted Ages~~}



Johnston et al. (1994)

RAI 130
02.05.01-40

Figure 2.5-16 {Regional Strip Maps Showing Tectonostratigraphic Divisions and Regional Cross-Section Lines}

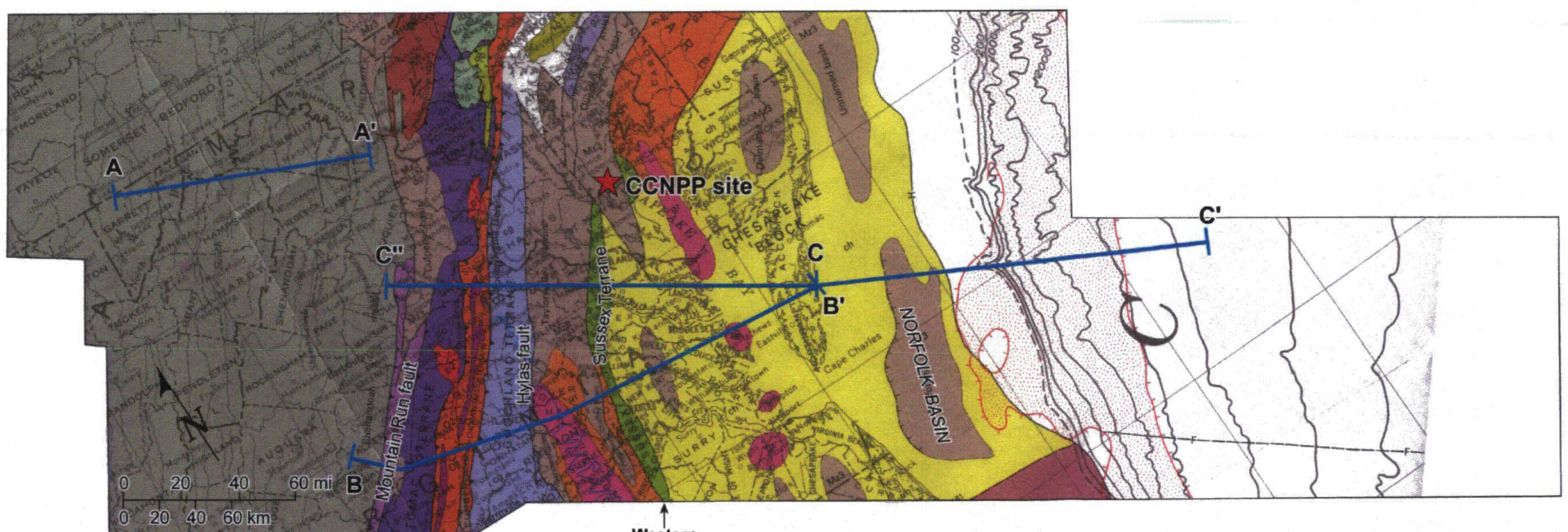


Glover and Klitgord (1995)

Explanation
(Glover and Klitgord, 1995)

Symbols
- Bathymetric isobath
- Normal fault
- Basement hinge zone
- C C' Cross section (see Figures 2.5-17, 18, and 19)
- Magnetic lineament

Map Units
- Cenozoic
- Early Mesozoic basin
- Cretaceous - stratified rock
- Permian - Upper Mississippian stratified rock
- Lower Jurassic - Upper Triassic stratified rock
- Jurassic volcanic wedge
- Cambrian - stratified rock
- Cambrian - intrusive rock



Horton et al. (1991)

Explanation
(Horton et al., 1991)

Symbols
- Basement hinge zone
- Mesozoic normal fault
- Thrust fault
- C C' Cross-section line (see Figures 2.5-17, 18, and 19)
- Fault; solid where certain, dashed where inferred or approximate
- Magnetic low - less than 600 nanoteslas
- Magnetic high - greater than 1,000 nanoteslas

Map Units
- M_z3 Jurassic and Triassic Newark super group
- gr Late Paleozoic granitoids
- p^1, p^2 Other stitching plutons
- O_2 O_1 Late and Middle Ordovician sequences interpreted as Taconian successor basin deposits
- L Laurentia (undivided)
- aw Westminster terrane
- ib Baltimore terrane
- dje Jefferson terrane
- dp Potomac composite terrane
- ob Bel Air - Rising Sun terrane
- os Sussex terrane
- vcp Chopawamsic terrane
- vr Roanoke Rapids terrane
- cg Goochland terrane
- ch Chesapeake block

Gravity Anomalies on Delmarva Peninsula
- g1 Probable mafic complexes
- g2 Linear gravity high associated with migmatitic gneiss
- g3 Broad gravity low

Figure 2.5-17 {Crustal-Scale Cross Section Through the Appalachian Orogen and Coastal Plain}

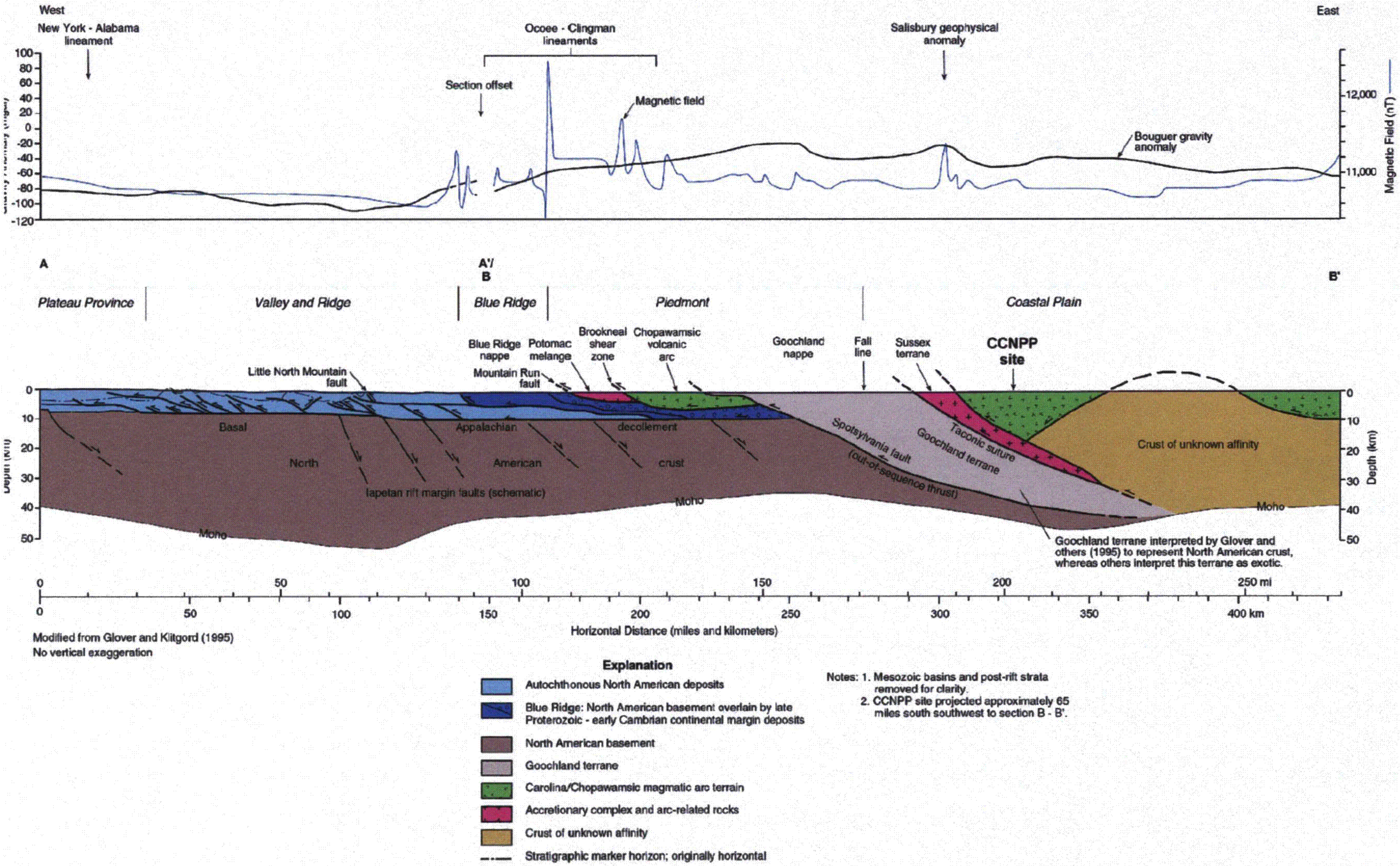
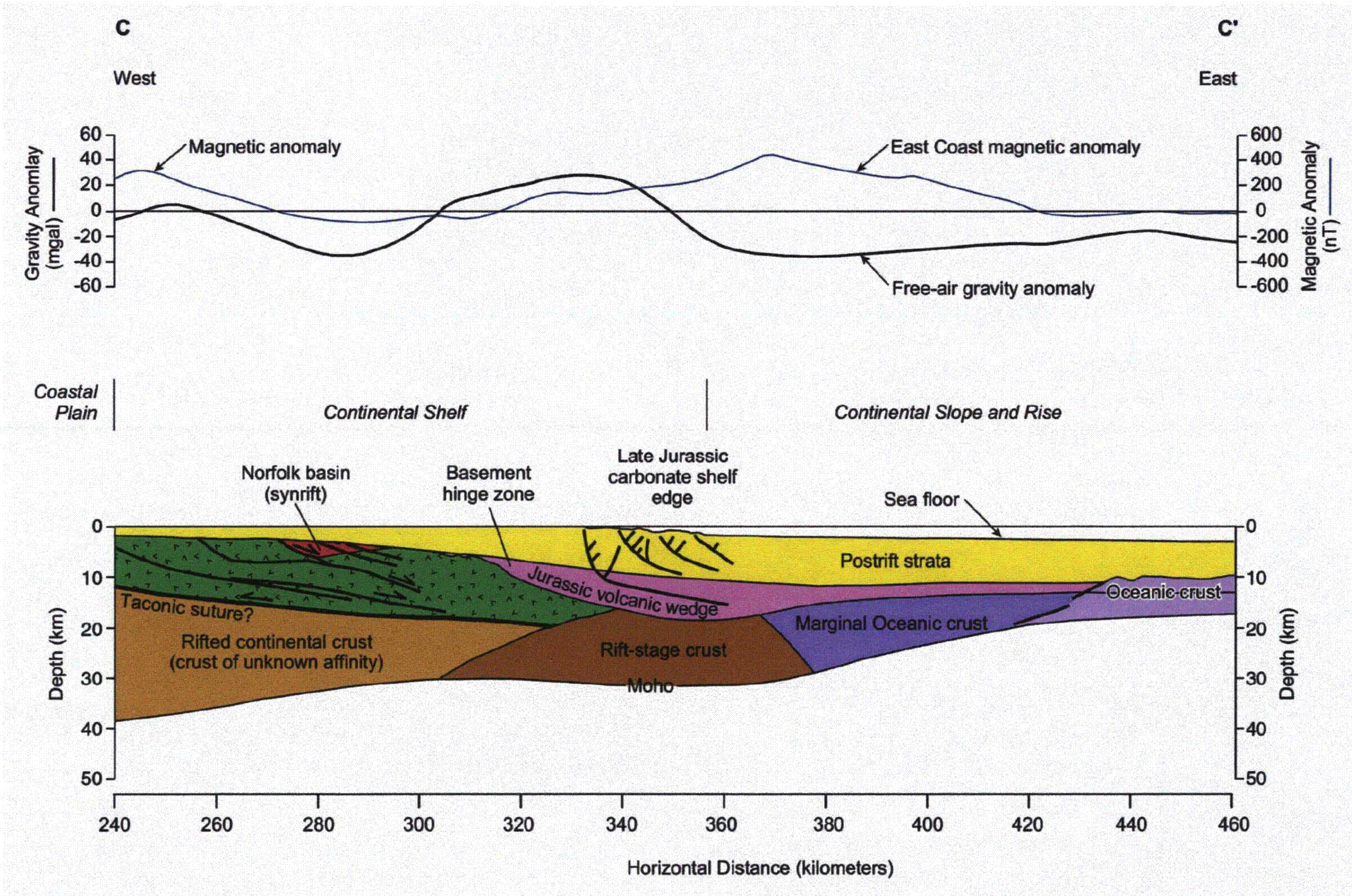
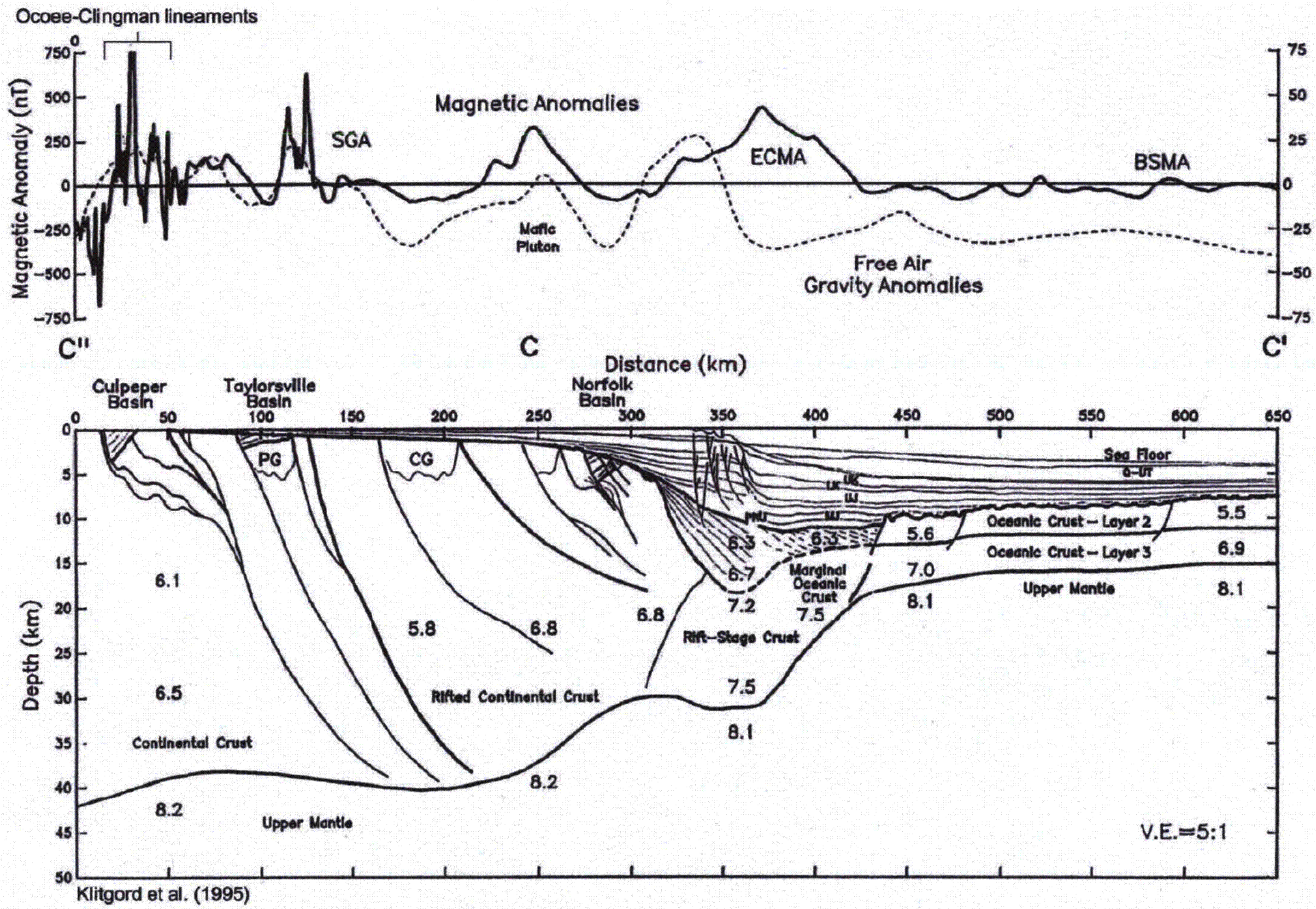


Figure 2.5-18 {Crustal-Scale Cross Section Across the Mid-Atlantic Continental Shelf, Slope and Rise}



Modified from Glover and Klitgord (1995)

Figure 2.5-19 {Crustal-Scale Cross Section of the Mid-Atlantic Passive Margin}



Cross section along line C" - C - C' displaying selected crustal fractures. Surface features along segment C" - C are taken directly from the geologic map panel. Subsurface features have been projected northward onto the profile from cross section B - B'. Magnetic and gravity anomaly profiles along the section and selected refraction velocity values (in km/sec) are shown. Major sub-horizontal crustal boundaries are indicated by heavy lines. Sedimentary strata are indicated by the light lines above the upper heavy line. SGA - Salisbury geophysical anomaly; ECMA = East Coast magnetic anomaly; BSMA = Blake Spur magnetic anomaly; PG = Petersburg Granite; CG = Chesapeake Granite. See Figure 2.5.1-15 for section location. C- C' is the same as Figure 2.5.1-17, but represents an alternative interpretation.

Figure 2.5-20 {Regional Magnetic Anomaly Map}

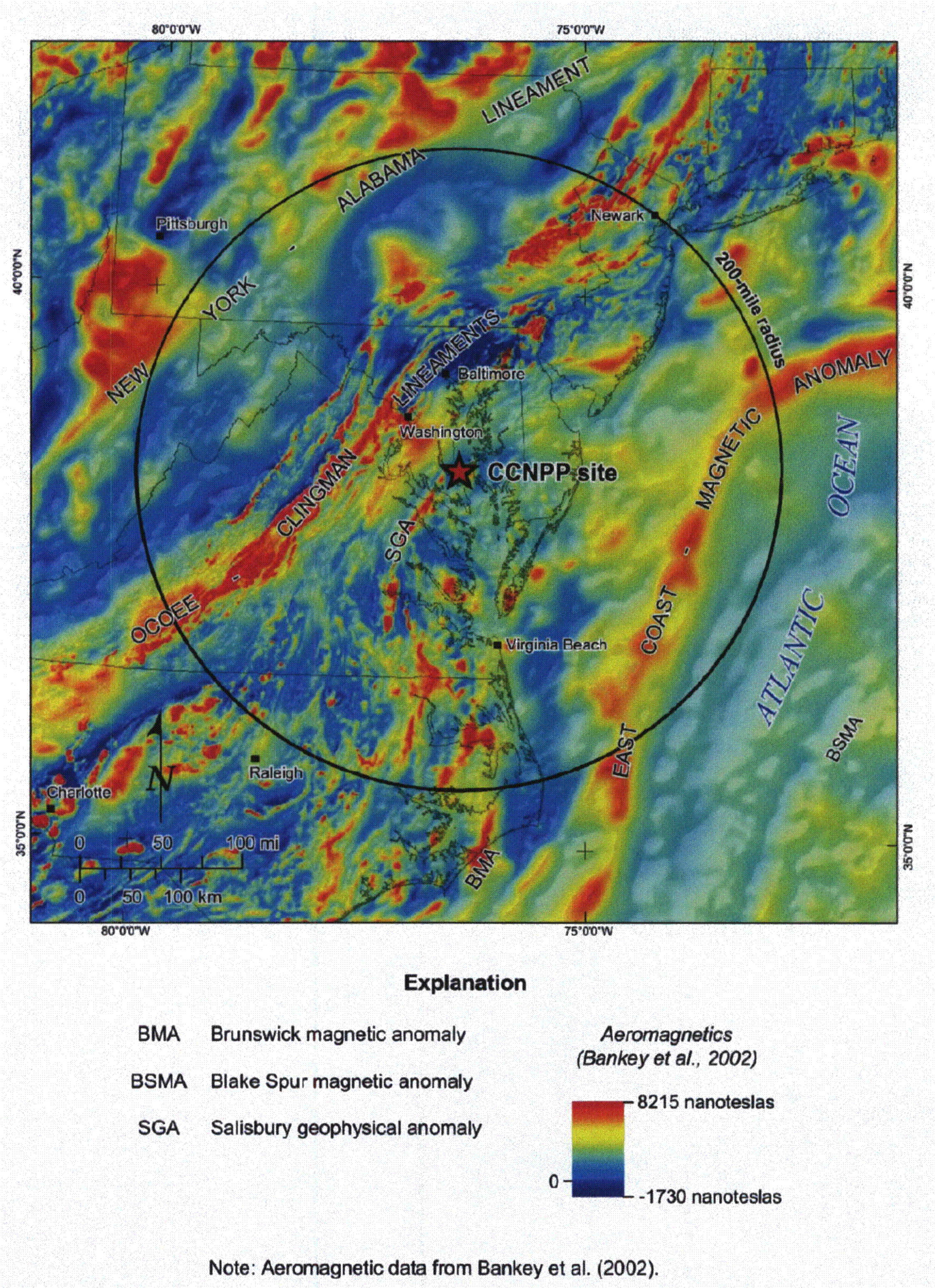
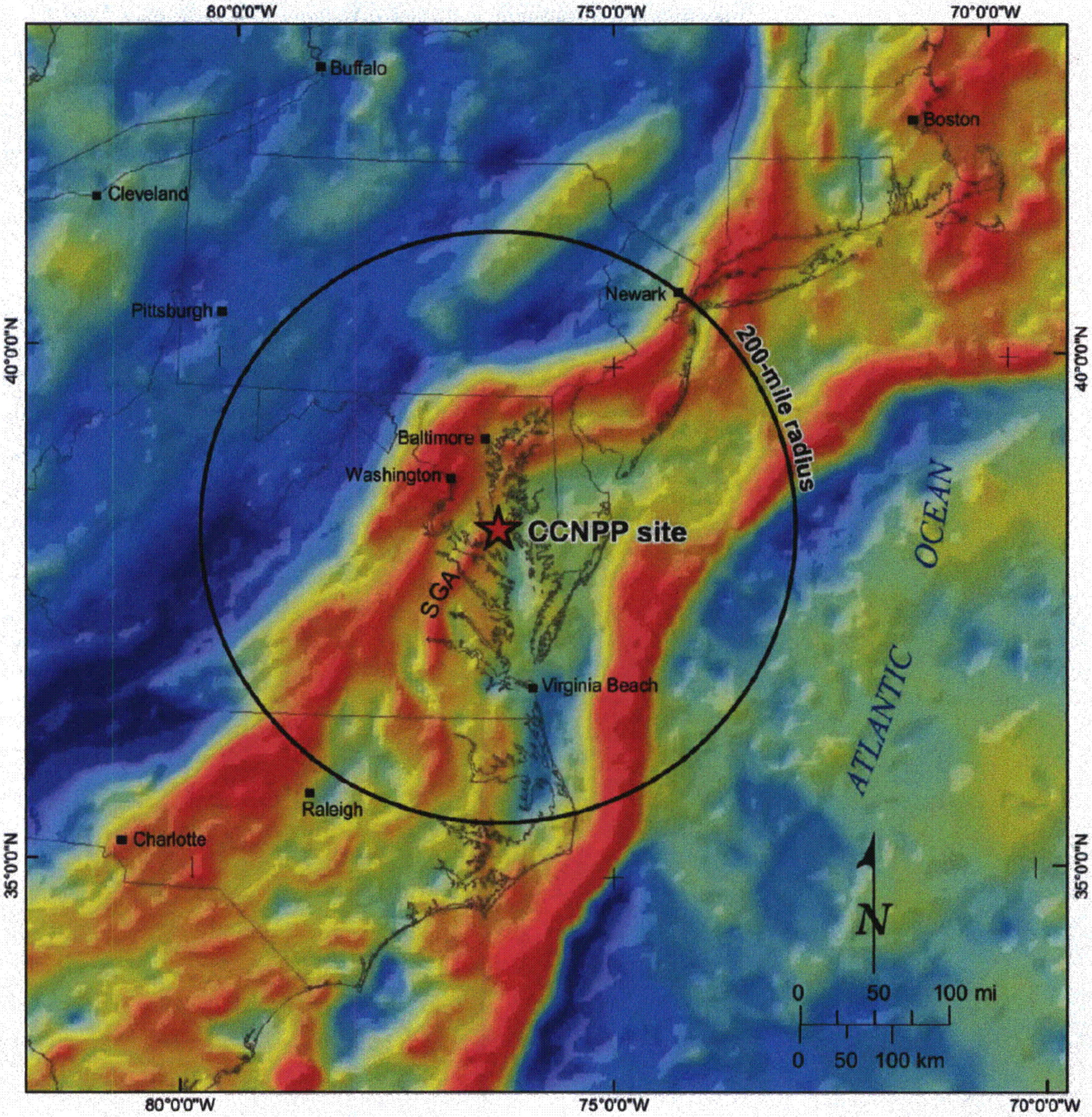


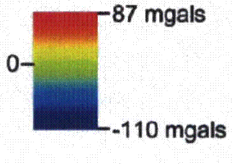
Figure 2.5-21 {Regional Gravity Anomaly Map}



Explanation

SGA Salisbury geophysical anomaly

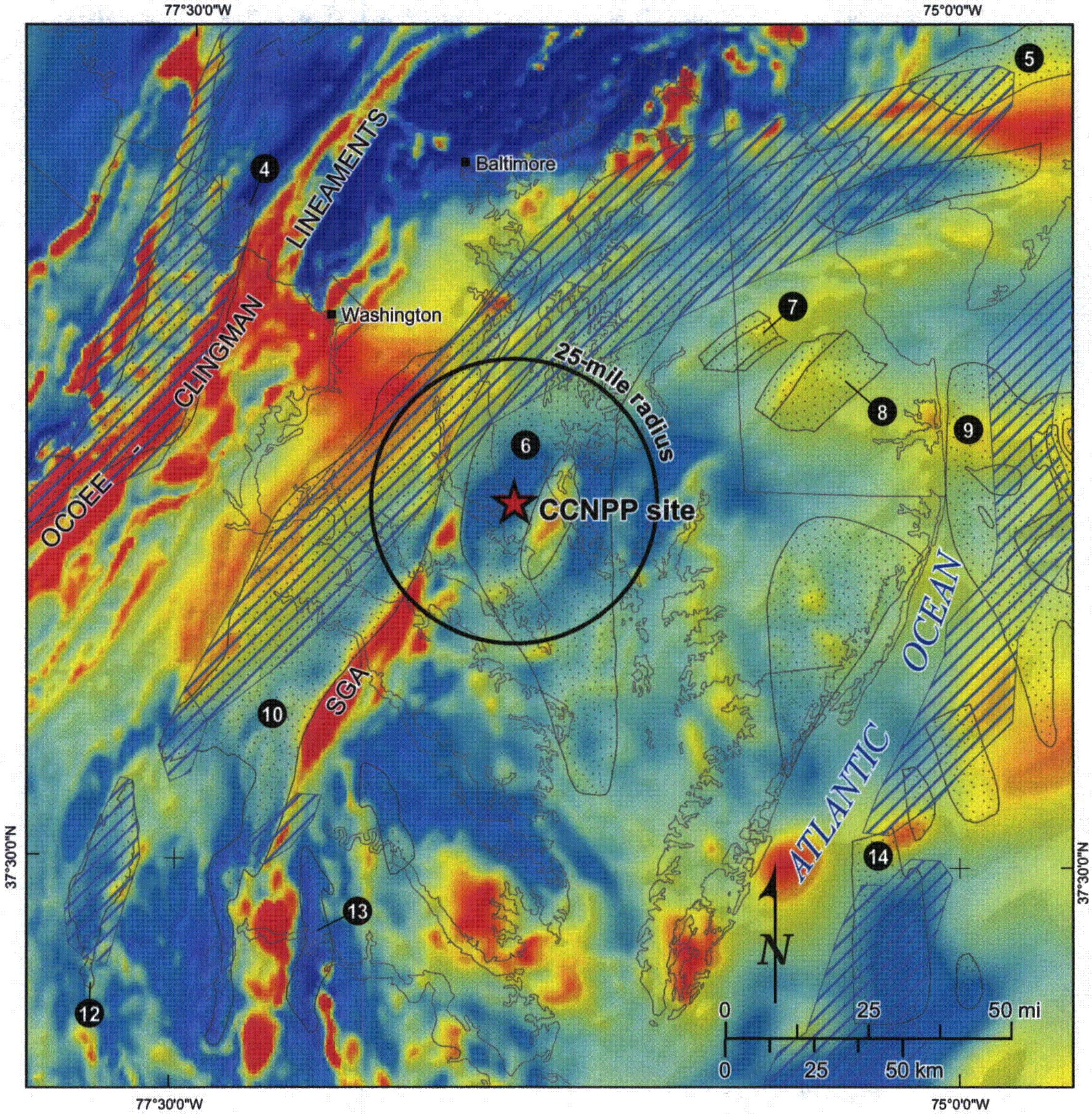
Gravity Anomaly



Notes: 1. Gravity data from Hittelman et al. (1994).
2. Gravity measurements over land are Bouger gravity anomalies.
3. Gravity measurements over water are free-air anomalies.

Figure 2.5-22 {Chesapeake Bay Region Magnetic Anomalies with Mesozoic Basins}

RAI 71
02.05.01-11



Explanation

Mesozoic basin, Schlische and Olsen (1990)

Mesozoic basin, Benson (1992)

SGA Salisbury geophysical anomaly

14 Mesozoic basin names listed on Figure 2.5-10 (Benson 1992)

Aeromagnetics

2116 nanoteslas

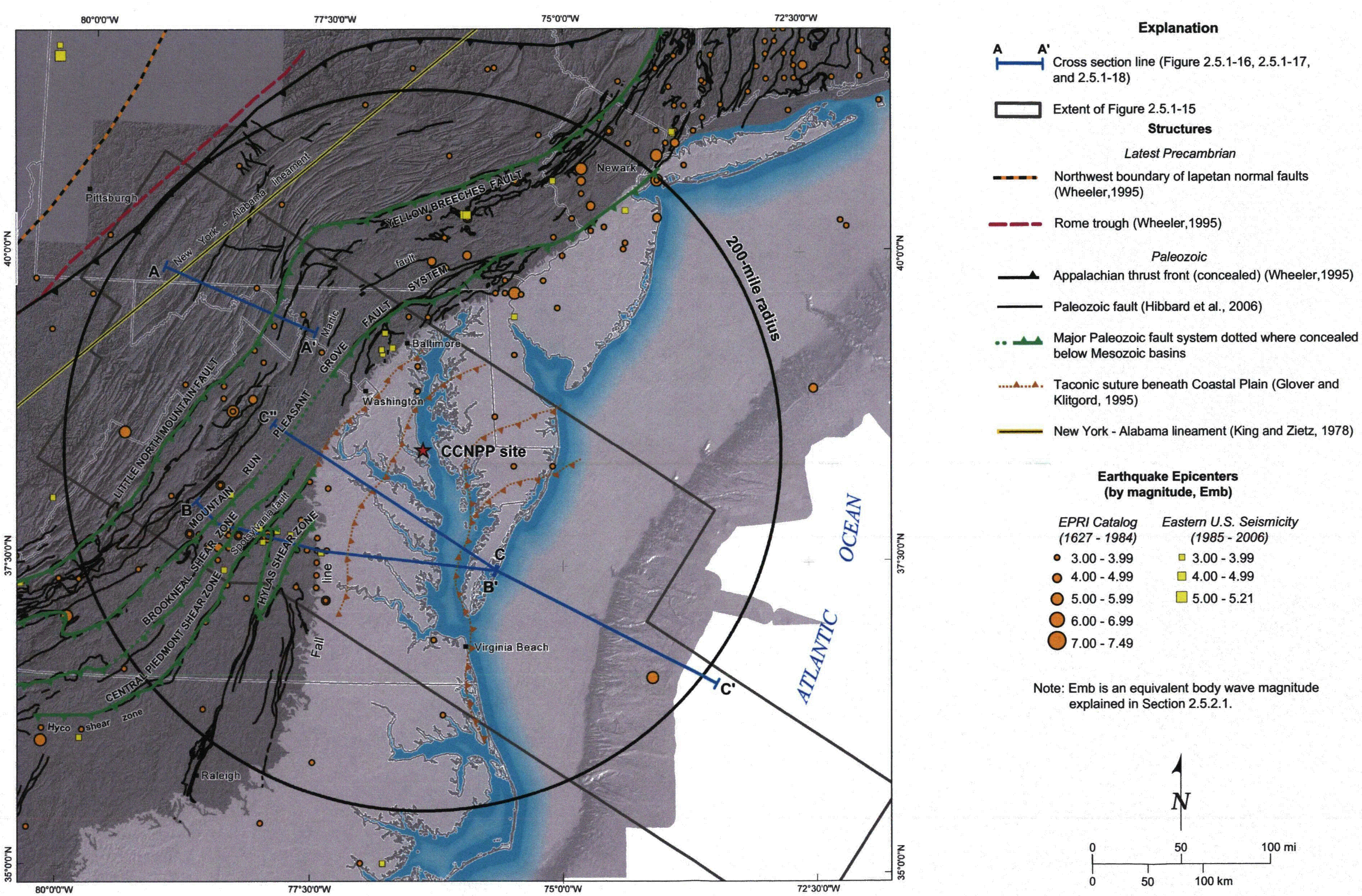
0

-819 nanoteslas

Note: Aeromagnetic data from Bankey et al. (2002).

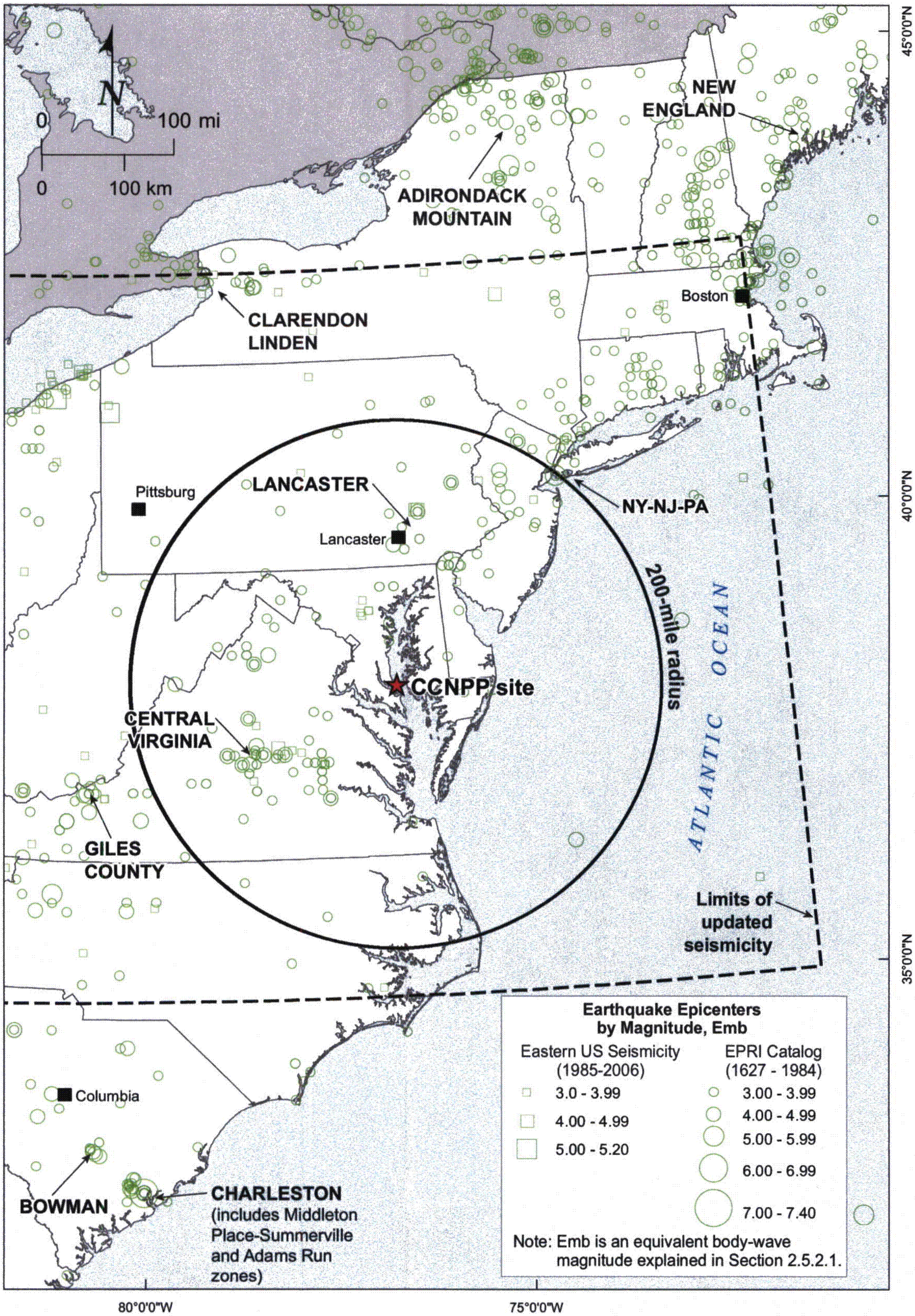
RAI 130
02.05.01-40

Figure 2.5-23 {Late Proterozoic and Paleozoic Tectonic Features}



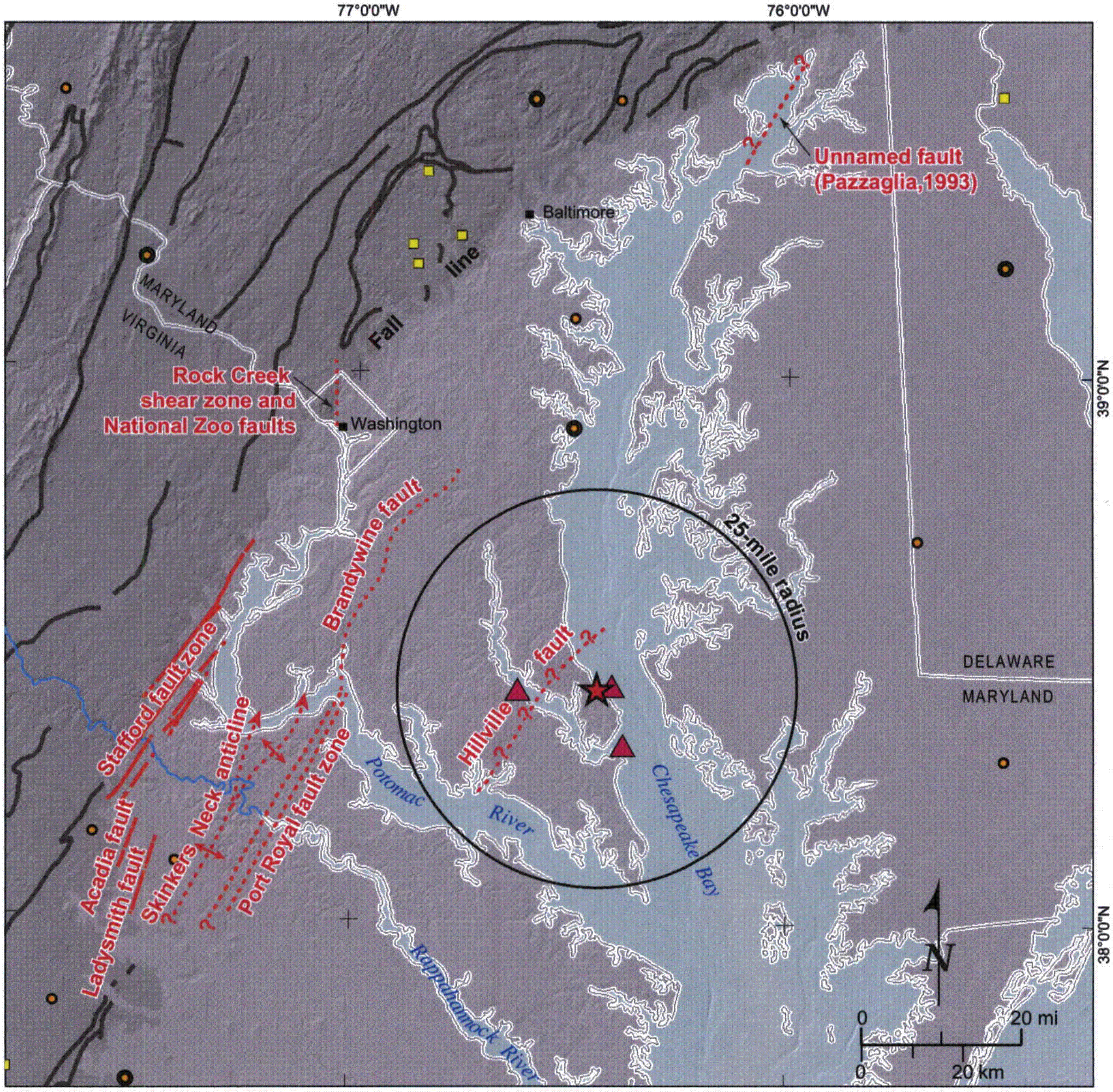
RAI 71
02.05.01-1

Figure 2.5-24 {Seismic Zones and Seismicity in CEUS}



RAI 71
02.05.01-21

Figure 2.5-25 {Map of Tertiary Tectonic Features}



Explanation

CCNPP site

McCartan et al. (1995) features (inferred monoclines)

Tertiary fault; dashed where uncertain; dotted where buried

Pre-Tertiary fault (Hibbard et al., 2006)

Anticline

Physiographic Provinces

Coastal Plain

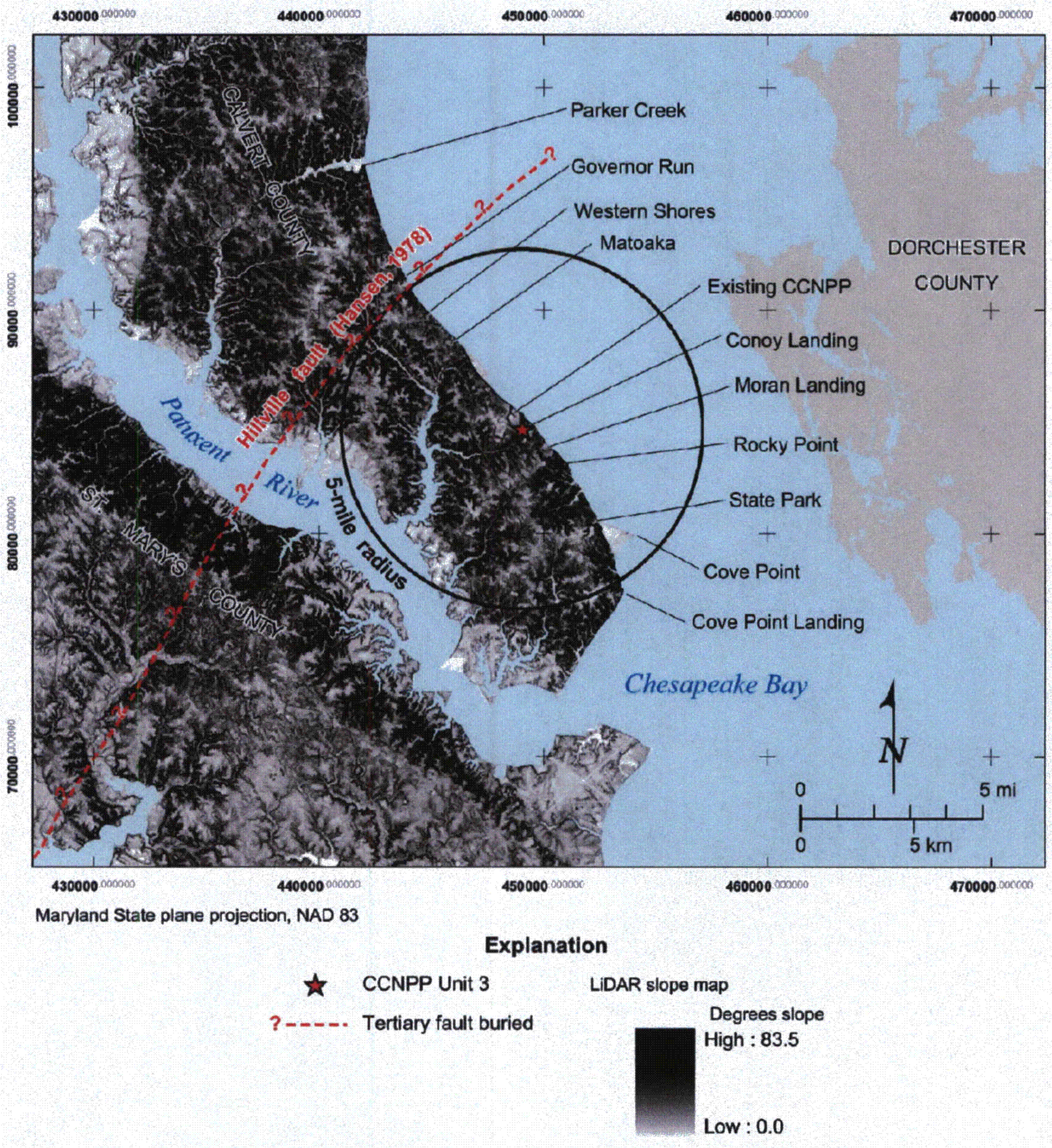
Piedmont

Earthquake Epicenters (by magnitude, Emb)

EPRI catalog (1627-1984)	*Eastern U.S. Seismicity (1985-2006)*
3.00 - 3.99	3.00 - 3.99
4.00 - 4.99	4.00 - 4.99
5.00 - 5.99	5.00 - 5.21
6.00 - 6.99	
7.00 - 7.40	

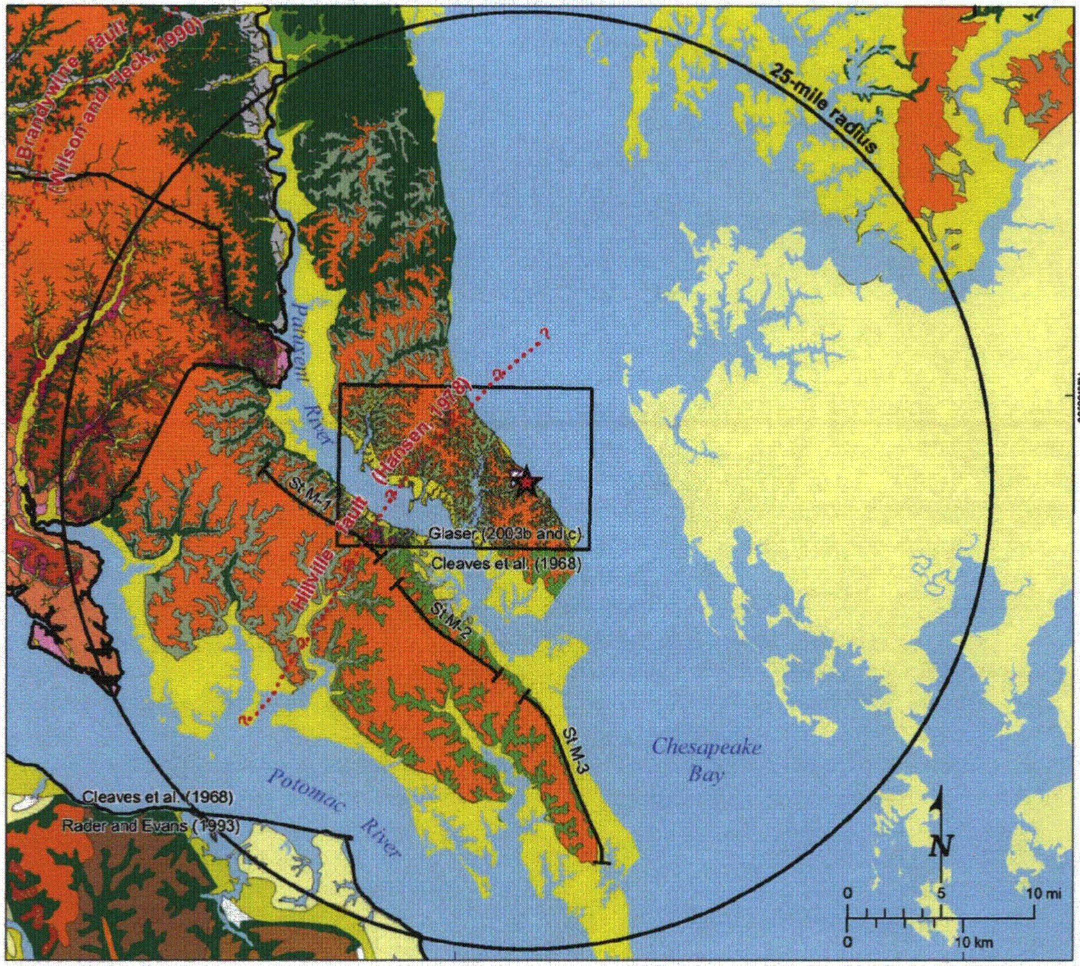
Note: Emb is an equivalent body-wave magnitude explained in Section 2.5.2.1.

Figure 2.5-26 {LiDAR Data for Calvert and St. Mary's Counties}

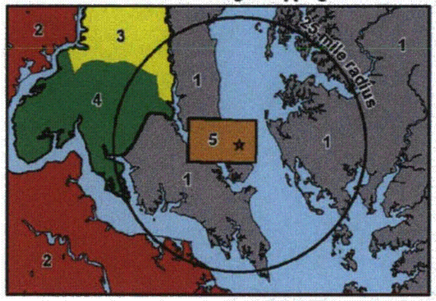


Note: LiDAR data for Calvert and St. Mary's County has an ~2-meter resolution. Text leadlines refer to reference locations from Kidwell (1997).

Figure 2.5-27 {Site Vicinity Geologic Map 25-Mile (40-Km) Radius}



Index of Geologic Mapping



1. State Map of Maryland (Cleaves et al., 1968) digitized by Dicken et al. (2005)
2. State Map of Virginia (Rader and Evans, 1993) digitized by Dicken et al. (2005)
3. Geologic map of Prince George's County (Glaser, 2003a)
4. Geologic map of Charles County (McCartan, 1989a)
5. Geologic maps of Cove Point and Broomers Island 7.5-minute quadrangles (Glaser, 2003b and c)

Explanation

CCNPP Unit 3

Source map boundary

Tertiary fault (buried)

St M-1 Seismic line (St M-1 through ST M-3) (Hansen, 1978)

For description of map units see Figure 2.5.1-26b