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2.5 Geology, Seismology, and Geotechnical Engineering

This section presents information on the geological, seismological, and geotechnical characteristics of the VCS site and the region surrounding the site. The data and analyses in this section document Exelon's evaluation of the suitability of the site. Section 2.5 provides sufficient information to support evaluations of the site-specific ground motion response spectra and provides information to permit adequate engineering solutions to geologic conditions and seismic effects at the proposed site.

References in this section to "Unit 1" and/or "Unit 2" are based on an assumed dual unit plant arrangement, with Unit 1 located in the western half of the designated ESP Power Block Area and Unit 2 located in the eastern half of the designated ESP Power Block Area (Reference Figures 1.2-1 and 1.2-2). In cases where more than two units would be built (e.g., mPower reactor design), the units would be divided between the eastern and western halves of the power block area.

Section 2.5 is divided into five subsections that generally follow the organization of RG 1.206.

- 2.5.1 — Basic Geologic and Seismic Information
- 2.5.2 — Vibratory Ground Motion
- 2.5.3 — Surface Faulting
- 2.5.4 — Stability of Subsurface Materials and Foundations
- 2.5.5 — Stability of Slopes

The VCS site is located within the Texas Gulf Coastal Plains physiographic province about 120 miles (193 km) southwest of Houston and about 13 miles (21 km) south of the city of Victoria, the county seat of Victoria County. The site topography consists of gently-to-moderately rolling hills covered by grassland. The local terrain is covered with shallow residual soils overlying the Pleistocene Beaumont Formation, consisting of sands and clays deposited in a fluvial-deltaic environment during the Pleistocene Sangamon interglacial stage.

The geological and seismological information presented in this section was developed from a review of published geologic literature, interviews with experts in the geology and seismotectonics of the site region, aerial photo analysis, and geologic fieldwork performed as part of the ESP application (including a site investigation of this greenfield location and two geologic field reconnaissances). A list of the references used to compile the geological, seismological, and geotechnical information presented in the following sections is provided at the end of each major subsection within Section 2.5.

The review of regional and site geologic, seismic, and geophysical information and an evaluation of the updated earthquake catalog confirmed the use of appropriate seismic sources in the probabilistic seismic hazard analysis (PSHA) as well as the need to include updated New Madrid seismic source zones to reflect current information on source geometries, maximum earthquake magnitudes, and recurrence parameters. Borings at the site provided geologic and geotechnical data to characterize material properties of the soil.

Subsection 2.5.1.1 describes the geologic and structural setting of the region within a 200-mile (322-km) radius of the site. Subsection 2.5.1.2 describes the geology and structural setting of the site vicinity (25-mile or 40-km radius), site area (5-mile or 8-km radius), and site (0.6-mile or 1-km radius). The geological and seismological information was developed in accordance with the guidance presented in RG 1.206, Subsection 2.5.1, Basic Geologic and Seismic Information, and RG 1.208, *A Performance-Based Approach to Define the Site-Specific Earthquake Ground Motion*, and is intended to satisfy the requirements of 10 CFR 100.23(c). The geological and seismological information presented in this subsection is used as a basis for evaluating the detailed geologic, seismic, and man-made hazards at the site.

Subsection 2.5.2 describes the methodology used to develop the ground motion response spectra (GMRS) for the VCS site. RG 1.208 further requires that the geological, seismological, and geophysical database be updated and any new data evaluated to determine whether revisions are required to the 1986 seismic source model developed by the Electric Power Research Institute (EPRI) in conjunction with the Seismicity Owners Group (SOG). Subsection 2.5.2 provides an update of the geological, seismological, and geophysical database for the VCS site, focusing on whether any data published since 1986 indicates a significant change to the 1986 seismic source model, referred to hereafter as the 1986 EPRI-SOG study (Reference 2.5.1-1).

Bechtel Power Corporation, supported by William Lettis & Associates, Inc. and Risk Engineering, Inc., conducted an assessment of ground motion at the VCS site using the guidance provided in RG 1.208. The starting point for this site assessment is the EPRI-SOG PSHA evaluation (Reference 2.5.1-1). RG 1.208 incorporates developments in ground motion estimation models, updated models for earthquake sources, methods for determining site response, and new methods for defining a site-specific, performance-based earthquake ground motion that satisfy the requirements of 10 CFR 100.23 and lead to the establishment of the safe shutdown earthquake (SSE) ground motion. The purpose of Subsection 2.5.2 is to develop the site-specific GMRS characterized by horizontal and vertical response spectra determined as free-field motions at the GMRS horizon using performance-based procedures. The GMRS represents the first part in the development of an SSE for the site as a characterization of the regional and local seismic hazard under Regulatory Position 5.4 of RG 1.208. In the case of the VCS site, the GMRS incorporates site-specific horizontal ground motion amplification factors using estimates of subsurface properties. The GMRS is used to determine the adequacy of the certified seismic design response spectra for

the plant parameter envelope. The certified seismic design response spectra are the SSE for the site, the vibratory ground motion for which certain structures, systems and components are designed to remain functional, pursuant to Appendix S of 10 CFR 50.

Subsection 2.5.3 documents an evaluation of the potential for tectonic and non-tectonic surface deformation at the VCS site. The data was developed as a result of literature and data reviews, interpretations of aerial and satellite imagery, field and aerial reconnaissance, and discussions with current researchers, and an analysis of seismicity with respect to geologic structures. This data indicates that there are no Quaternary faults or capable tectonic sources within 25 miles (40 km) of the site.

Subsection 2.5.4 describes the site subsurface investigation, which consisted of borings, cone penetration tests, test pits, geophysical logging (including P-S suspension logging and seismic cone penetration tests), groundwater observations and monitoring, and laboratory testing of soil samples. Site-specific subsurface conditions are described, and design geotechnical engineering parameters are derived. The power block structures will be founded on natural soil strata and/or structural fill. Foundation bearing capacities are calculated, and foundation settlements are estimated. The potential for site soils to liquefy in a seismic event is additionally evaluated, and adequate factors of safety are calculated.

Subsection 2.5.5 describes the design of nonsafety-related earth dams and slopes for the cooling basin, which occupies the bulk of the VCS site. Case histories of similar earth dam structures are presented. Slope stability analyses for various design conditions (end-of-construction stage, steady-state seepage conditions, rapid drawdown conditions, and seismic conditions) are described, and adequate factors of safety against slope failure of site earth dams are calculated. Additionally, analyses for through- and under-dam seepage are described, and adequate factors-of-safety against piping failure of site earth dams are calculated. Static and dynamic settlements are estimated.

2.5.1 Basic Geologic and Seismic Information

The geological and seismological information presented in this section was developed from a review of published geologic literature, interpretation of aerial photography, a site subsurface investigation, and an aerial reconnaissance conducted for preparation of the VCS ESP application.

This subsection demonstrates compliance, in part, with the requirements of 10 CFR 100.23(c).

The geological and seismological characteristics of the VCS region, site vicinity, site area, and site are described in Subsection 2.5.1. The geologic and tectonic characteristics of the site region and site vicinity are described in Subsection 2.5.1.1 and are shown in Figures 2.5.1-1, 2.5.1-2a, and 2.5.1-3. The geologic and tectonic characteristics of the VCS area and site are described in

Subsection 2.5.1.2 and shown in Figures 2.5.1-4 and 2.5.1-5, respectively. The geological and seismological information was developed in accordance with RG 1.206 and RG 1.208.

2.5.1.1 Regional Geology

This subsection provides information on the physiography, geologic history, stratigraphy, structures, and tectonic setting within the 200-mile (322-km) radius of VCS. The nomenclature used in this subsection is consistent with terms used by the Texas Bureau of Economic Geology.

The regional geologic map (Figure 2.5.1-2a) with explanation (Figure 2.5.1-2b) (References 2.5.1-4, 2.5.1-5, 2.5.1-6, and 2.5.1-7) contains information on the geology, stratigraphy, structures, and tectonic setting of the region surrounding VCS. Summaries of these aspects of regional geology are presented in the following subsections to provide the framework for evaluation of the geologic and seismologic hazards.

The VCS lies within the Coastal Prairies subprovince of the Gulf Coastal Plains physiographic province as shown in Figures 2.5.1-1 and 2.5.1-3. The subprovince is composed of young, unconsolidated deltaic sands, silts, and clays sloping to the southeast that are incised by meandering streams discharging into the Gulf of Mexico. The ground surface elevation in the subprovince ranges from sea level to 300 feet (91 meters). The geologic and tectonic setting of the region is the product of a complex one-billion-year history of continental collisions and rifting followed by deposition of sediments in the newly formed Gulf of Mexico basin. Site regional stratigraphy consists of undifferentiated Precambrian basement rock overlain by up to 21,000 feet (4 miles or 6.4 km) of Mesozoic sedimentary rock units, which are in turn overlain by up to 20,000 feet (3.8 miles or 6.1 km) of Cenozoic well to poorly lithified sediments.

2.5.1.1.1 Regional Physiography and Geomorphology

The VCS lies within the Coastal Prairies subprovince of the Gulf Coastal Plains physiographic province as shown in Figures 2.5.1-1 and 2.5.1-3. The region within a 200-mile (322-km) radius of the site encompasses portions of five physiographic provinces from the North American platform south into the Gulf of Mexico: (1) the Edwards Plateau, (2) the Central Texas or Llano Uplift, (3) three subprovinces of the Gulf Coastal Plains province (Blackland Prairies, Interior Coastal Plains, and Coastal Prairies), (4) the Texas-Louisiana Shelf, and (5) the Texas-Louisiana Slope. Each of these physiographic provinces is described briefly in the following subsections.

2.5.1.1.1.1 Edwards Plateau Physiographic Province

The Edwards Plateau is a hilly area clearly demarcated by the Balcones Escarpment to the east and south, but it grades into the Chihuahuan Desert to the west and the Llano Uplift and Great Plains to the north. The Balcones Escarpment traces a series of en echelon normal faults that follow the

Ouachita tectonic front along the southern margin of the North American platform. A proposed dextral bend connects the Ouachita front into the Grenville tectonic front along the entire length of the Eastern seaboard of North America. The rocks of the Edwards Plateau consist mainly of upthrust limestones and dolomites of Upper Cretaceous age, in which caverns and sinkholes are common. Hard and soft strata have created stair-step topography. Streams have eroded the surface into ravines as deep as 1800 feet (550 meters). Elevations range from 450 to 3000 feet (137 to 914 meters) in the principal part of the province. Higher elevations occur in the Stockton Plateau, which is the western portion of the Edwards Plateau (Reference 2.5.1-8). Based on limited well drilling data in Gillespie County, north and west of the VCS site, the horizontal Cretaceous beds of the Edwards Plateau overlie an uneven floor of Carboniferous Paleozoic rocks (Reference 2.5.1-9).

2.5.1.1.1.2 Central Texas or Llano Uplift

The Central Texas Uplift (also known as the Llano Uplift) is a roughly circular uplifted dome exposing a central core of polydeformed metamorphic and igneous rock emplaced over the continental platform along at least two shear zones. The Mesoproterozoic metamorphic rocks found in the uplift were involved in a Grenville-age orogenic event (ca. 1.3 to 1.0 billion years before present (Giga-annum or “Ga”) along the southern margin of the North American craton. The exposed rocks include the Precambrian Town Mountain Granite plutons (1.12 Ga to 1.0 Ga) (Reference 2.5.1-10) at the center of a basin with a rolling floor studded with rounded granite hills 400 to 600 feet (122 to 183 meters) high. The Town Mountain Granite is surrounded by a ring of meta-igneous and meta-sedimentary rocks, including serpentinite, the Packsaddle Schist, Lost Creek Gneiss, and the Valley Spring Gneiss (Reference 2.5.1-11). A major ductile shear zone separates the southwestern most Coal Creek arc terrane from the Packsaddle basinal sedimentary and volcanic rocks deposited along the southern Laurentian margin. To the north, the Valley Spring gneiss, consisting of plutonic and supracrustal rocks, was thrust northeastward beneath Packsaddle units along a mylonitic shear zone. Together, these imbricate stacked thrust units represent the oldest rocks in Texas. The metamorphic units are intruded by younger granite plutons (Reference 2.5.1-12). Around the granitic rocks of the central basin are two overlying formations eroded to concentric rims: the inner is a rim of resistant lower Paleozoic formations. Outside the Paleozoic rim is a second ridge formed of limestones like those of the Edwards Plateau.

2.5.1.1.1.3 Gulf Coastal Plains Physiographic Province

The Gulf Coastal Plains physiographic province, as shown in Figure 2.5.1-3 (Reference 2.5.1-8), extends southeast and east from the edge of the Edwards Plateau for 200 to 300 miles (322 to 483 km) to the shore of the Gulf of Mexico. The Gulf Coastal Plains have been divided into three subprovinces from north to south: The Blackland Prairies, the Interior Coastal Plains, and the Coastal Prairies.

The northernmost of the Gulf Coastal Plains subprovince is the Blackland Prairies, separated from the Edwards Plateau carbonates on the north and west by the normal fault system of the Balcones escarpment. The Balcones escarpment was most recently active in the Miocene epoch when down-to-the-south displacement of over 1200 feet (366 meters), resulting from the weight of sediment deposition, offset the Coastal Plains sediments. The Blackland Prairies subprovince consists of upper Cretaceous chalk and marls. Soils consist of fertile, deep, black clay. The surface is gently rolling and agriculturally developed (Reference 2.5.1-8). Ground surface elevations range from 450 to 1000 feet (137 to 305 meters) North American Vertical Datum of 1988 (NAVD 88).

South of the Blackland Prairies lies the Interior Coastal Plains subprovince. It begins at or near the contact between Quaternary and Tertiary sediments and extends to the northwest 75 to 150 miles (121 to 241 km) as shown in Figures 2.5.1-1, 2.5.1-2a, and 2.5.1-3. The sediments are uncemented red and brown sands and clays that are relatively resistant to erosion. At least two down-to-the-coast normal fault systems parallel the coast. Ground surface elevations range from 300 to 800 feet (91 to 244 meters) NAVD 88. Several thousand feet of unconsolidated Cenozoic-age sands and clays underlie the surficial sediments. Subsection 2.5.1.1.3.4 contains a detailed description of these sediments.

The Coastal Prairies subprovince is located south of the Interior Coastal Plains subprovince. The Coastal Prairies subprovince is approximately 50 to 75 miles (80 to 121 km) in width and stretches to the Gulf of Mexico. The land surface has an almost negligible slope to the southeast (Figure 2.5.1-6). The sediments are composed of young (Pleistocene and Holocene) unconsolidated deltaic sands, silts, and clays incised by meandering streams that discharge into the Gulf of Mexico. Approximately 21,000 feet (3.8 miles or 6.1 km) of unconsolidated Cenozoic sediments underlie the surface of this subprovince. Subsection 2.5.1.1.3 contains a description of Coastal Plains stratigraphy. The ground surface elevation ranges from sea level to approximately 300 feet (91 meters) in the subprovince, and from 25 to 85 feet (7.6 to 26 meters) NAVD 88 at the VCS.

The four periods of continental glaciation that occurred during the Pleistocene resulted in rising and falling sea levels along the Gulf of Mexico and in worldwide (eustatic) changes in sea level. Rivers draining the continental interior flowed across the Gulf Coastal Plains and built deltas as they discharged into the Gulf. The most recent of these glacial advances, the Wisconsinan glacial stage of the late Pleistocene, lowered sea levels, and the coalescing deltas of rivers draining the continental interior during the eustatic low stand of the sea resulted in the deposition of the Beaumont Formation, which forms the present surface of the Coastal Prairies subprovince. Post-Beaumont erosion and deposition has created terraces within incised channels. These sediments comprise the undifferentiated Deweyville terrace deposits.

A rise in sea level beginning approximately 18,000 years ago initiated the geomorphic process of longshore-drift of sands and deposition of those sands as barrier islands and lagoons. Narrow

lagoons separate the barrier islands from the mainland. Previous high sea level stands can be identified by a series of late Pleistocene ridges (former barrier bars) on the north side of the lagoons. Smaller rivers have discharged sediments into the lagoons, nearly filling them.

2.5.1.1.1.4 Texas-Louisiana Shelf

The continental shelf off the Texas Gulf Coast is termed the Texas-Louisiana Shelf (Reference 2.5.1-13). It has experienced a net progradation during the Tertiary and Quaternary periods. Clastic materials derived from the uplands to the north and west has spread across the shelf as the seas transgressed for more than 66 million years. This depositional pattern has been present in the Gulf of Mexico since the Jurassic, and the shelf has prograded approximately 186 miles (300 km) in that time. The offshore Texas-Louisiana Shelf is a broad, nearly featureless plain. Thin Holocene sediments cover a late Pleistocene fluvial plain. Entrenched stream channels are common, and are filled by Holocene sediments. Carbonate banks occur in places, including true algal-reefs off Galveston, Texas (Reference 2.5.1-13).

2.5.1.1.1.5 Texas-Louisiana Slope

The continental slope, known as the Texas-Louisiana Slope off the shore of Texas, covers 46,332 square miles (120,000 km^2) of knoll-and-basin seafloor. The average gradient is less than 1 degree but slopes greater than 20 degrees occur near knolls and basins. The extreme change in relief is the result of salt diapirs that have moved upward from the deeper Jurassic-age beds. Because of rapid sedimentation, growth faults are common and tend to accentuate the shelf-edge break. (Reference 2.5.1-13)

2.5.1.1.2 **Regional Geologic History**

The geologic and tectonic setting of the VCS site region is the product of a complex history of continental collisions and rifting, which spanned a period of more than one billion years. Major tectonic events in the site region include three compressional deformational events (orogenies) and at least two major extensional events. Direct evidence for most of these events is largely buried beneath the coastal plain sediments in the site region. Continental rifting in the Jurassic followed by deposition of sediments in the newly formed Gulf of Mexico basin shaped the current south Texas terrain.

Figure 2.5.1-7 is a geologic time scale that provides a framework for this subsection.

2.5.1.1.2.1 Grenville Orogeny

The earliest of the orogenies recorded in the rocks of the region is the Grenville orogeny of Middle to Late Precambrian (Proterozoic) time, approximately 1.3 to 1.0 Ga, as a result of continent-to-continent impact along the edges of Laurentia, the ancestral North American craton.

Some reconstructions show only the ancestral African continental mass impacting on the eastern edge of Laurentia, but recent evidence (References 2.5.1-14 and 2.5.1-15) indicates that Amazonia, the ancestral South American continent, was cojoined with Laurentia along the southern coastal area, Australia and Antarctica were positioned against the proto-North American west coast, and a proto-north central African craton and numerous volcanic arcs were somehow caught between these colliding masses to form one of the oldest known supercontinents, Rodinia (Reference 2.5.1-16).

Rocks of Grenville age are poorly exposed in Texas, occurring only in the Llano Uplift in central Texas, in the Franklin Mountains, and the West Texas Uplifts in west Texas. Based on a few oil and gas borings, Renfro et al. (Reference 2.5.1-17) indicate that large areas of the state have granitic basement related to the Llano Uplift. The basement beneath Victoria County is believed to be meta-igneous continental crust material (Reference 2.5.1-18), but no data is available to confirm that the thick pile of Cenozoic sediments overlie Precambrian basement because this contact is estimated to be approximately 41,000 feet (7.8 miles or 12.5 km) below the surface, far deeper than normal petroleum exploration wells.

2.5.1.1.2.2 Late Proterozoic Laurentian Rifting

Following the Grenville orogeny, late Proterozoic crustal extension and rifting occurred around 700 million years before present (Mega-annum before present, or Ma), causing the separation of three continents: proto-Laurasia, the Congo craton, and proto-Gondwana (minus the Congo craton and Antarctica). These three continents rotated into positions close to the poles, creating a short-lived supercontinent, Panotia during a time of intense glaciation. Within another 160 million years, Panotia separated into four continents: Laurentia, Baltica, Siberia and Gondwana. Laurasia (comprised of Laurentia, Baltica, Siberia, Kazakhstania, and the North China and East China cratons.) moved away from Gondwanaland (comprised of Antarctica, South America, Africa, Madagascar, Australia-New Guinea, New Zealand, Arabia, and the Indian subcontinent), creating the ancestral African continent and the proto-Atlantic (Iapetus) Ocean. In the region of the Gulf of Mexico, shallow seas transgressed across the wide area between Alabama and Texas as the land subsided, possibly over a period of more than 200 million years. Thomas suggests that this period included early extension, pervasive rifting, and late-stage rifting. Failed rifts (aulacogens) formed graben systems such as the Reelfoot Rift in eastern Arkansas, and the Rome Trough (Figure 2.5.1-8), located northeast of the Gulf of Mexico. These and other failed rifts indicate that late crustal extension was pervasive along the rift margin. (References 2.5.1-19 and 2.5.1-20)

From the Cambrian to the Early Mississippian periods, the region between the southern Appalachians and the Marathon Embayment in west Texas was covered by shallow seas whose deposits record periods of transgression and regression. The early Paleozoic continental margin was well inland from the present VCS site. According to Thomas, (References 2.5.1-19 and 2.5.1-20), the VCS area received sediments of the pre-orogeny Ouachita facies—shale, chert, micrite, and

sandstone. These sediments have an unconfirmed total thickness of approximately 4000 feet (1219 meters).

2.5.1.1.2.3 Ouachita Orogeny

From the Middle Mississippian to the Permian periods, the tectonics in the southern edge of Laurentia changed from the spreading (extensional) phase to a closing (compressional) phase. This is equivalent to the Alleghenian orogeny along eastern Laurentia and the Ouachitan orogeny along southern Laurentia when the ancestral supercontinent, Gondwana, collided with Laurentia, resulting in the closing of the proto-Atlantic (Iapetus) Ocean (Reference 2.5.1-19).

Rates of deposition were high in the pre-orogenic Ouachita trough, extending from Mississippi to Mexico (Reference 2.5.1-21). As the Ouachita orogenic belt developed, thrusting of the sediments in the fore-arc basin toward the north and northwest formed the Ouachita orogen in North America. There is evidence that this tectonic event began in the Devonian because isotopic ages from metamorphic clasts in the Haymond boulder beds south of the Marathon region indicate Devonian deformation and metamorphism (References 2.5.1-21 and 2.5.1-22). These boulders must have originated from a source south of the Ouachita trough, as Devonian deformation is not known to have affected rocks from the Laurentia (proto-North America) side of the proto-Atlantic (Iapetus) Ocean.

After late Paleozoic (Late Pennsylvanian-Permian) thrusting created the Ouachita Mountains, the closing of the proto-Atlantic (Iapetus) Ocean and the assembly of Pangea was complete.

2.5.1.1.2.4 Mesozoic Rifting (Opening of the Gulf of Mexico and the Atlantic)

Although a basin (the proto-Gulf of Mexico) appears to have developed before the Cretaceous period, the sedimentary record indicates that there was no connection to the Atlantic Ocean until the Early Cretaceous period. The basis for this interpretation is that no Triassic marine sediments have been documented in the region of the present Gulf of Mexico. The earliest indication of marine deposition in the present Gulf of Mexico region is extensive salt deposits of the Middle Jurassic, about the time the initial breakup of Pangea began. These salt deposits were derived from the evaporation of Pacific Ocean seawater that covered the Mexico-Central American region. Salvador (Reference 2.5.1-23) states that the salt was deposited over continental or transitional crust and that the thick salt suggests that subsidence kept pace with salt deposition.

The initial rifting and crustal extension probably began in the Late Triassic and continued into the early Late Jurassic periods. By the Late Jurassic, the emplacement of new oceanic crust in the Gulf of Mexico basin had ended and the new crust cooled and subsided. The distribution and nature of the early Late Jurassic sedimentary sequences indicate that there still was no connection between the proto-Gulf of Mexico and the Atlantic Ocean because the Florida and Yucatan platforms were above sea level during the Jurassic and probably connected to each other (Reference 2.5.1-23). On the

basis of regional stratigraphic information, Salvador (References 2.5.1-23 and 2.5.1-24) has assumed that a connection between the Gulf of Mexico and the Atlantic was not established until the late Kimmeridgian (middle Late Jurassic).

By the Early Cretaceous period, the Gulf of Mexico basin was tectonically stable. The Florida platform had been covered by a transgressing sea connected to the Atlantic Ocean, and the deposition of sediments from the North American continent had begun in the northern part of the basin brought to the Gulf by the ancestral Mississippi River (Reference 2.5.1-23).

2.5.1.1.2.5 Laramide Orogeny

The lithology of deposits along the western and northwestern flanks of the Gulf of Mexico basin indicates that the Laramide orogeny began in the Late Cretaceous period (Reference 2.5.1-23). In addition, volcanism was occurring in the Balcones fault zone in central and south Texas and offshore Louisiana (Reference 2.5.1-23). The principal effects that the Laramide orogeny had on the Gulf of Mexico were providing a western source of clastic sediments (the Rocky Mountains) and reducing the connection between the Pacific Ocean and the Gulf of Mexico (Reference 2.5.1-23). Subsidence continued in the central part of the Gulf of Mexico basin while there was uplift in the Mississippi Embayment.

2.5.1.1.2.6 Cenozoic History

Early Cenozoic (Paleocene and early Eocene) rocks and geologic structures, represented by the Chihuahua tectonic belt and the Sierra Madre Oriental thrust belt, record the final thrust faulting and folding of the Laramide orogeny in the region. Along with the uplift of the Rocky Mountains came an influx of clastic sediments originating from the new terrestrial source. Subsidence of the Gulf of Mexico basin previously due to cooling of newly emplaced oceanic crust now was primarily due to loading of the crust by prograding wedges of sediments. Marine deposits record sequences of transgression (sea level rise) and regression (sea level fall) throughout the Cenozoic, but especially during the Pleistocene when continental glaciations held huge volumes of water as icecaps, which, when melted during the interglacial periods, caused sea levels to rise world-wide. These repeated sea level changes along with natural basin subsidence deposited discontinuous beds of sand, silt, clay, and gravel under a fluvial-deltaic to shallow-marine environment. Rapid burial of the fluvio-deltaic sediments restricted expulsion of pore water and caused the development of over-pressured zones in the subsurface that act as detachment planes for faults (Reference 2.5.1-25). In addition to the loading of the crust, the weight of these rapidly accumulating sediments contributed to the development of salt diapirs in the Jurassic salt, growth faults in the sediments themselves, and shale diapirs in the lower Cenozoic over-pressured shales, respectively (Reference 2.5.1-23). The location of the depositional centers changed over time. The thickest accumulations during the Paleocene and early Eocene were in northeast Texas, northeast Louisiana, southeast Arkansas, and western Mississippi. During the late Eocene and early Oligocene, the

depocenter moved to south Texas, then towards southwest Louisiana. The Pliocene and Pleistocene depocenters occurred west of the Mississippi Delta then south of south Texas (Reference 2.5.1-23). While volcanic materials are found in the Cenozoic sediments of the Gulf Coastal Plains, these are derived from areas in Mexico and other western areas, not within the Coastal Plains province itself.

The continental glacial events that took place during the Pleistocene modified the sedimentary patterns that had previously characterized the Cenozoic of the northern Gulf of Mexico region. Sea level dropped by as much as 500 feet (152 meters) during glacial periods, exposing large areas of the northern shelf. The rise of sea level submerged the shelf, shifting the areas of active deposition landward and causing the reworking of sediments deposited during the previous glacial period (References 2.5.1-23 and 2.5.1-13).

2.5.1.1.3 **Regional Stratigraphy**

This subsection contains information on the regional stratigraphy within the Coastal Plains physiographic province. Figure 2.5.1-9 contains a regional cross section illustrating the regional stratigraphy.

2.5.1.1.3.1 Basement Rock

Because the Mesozoic and Cenozoic sections below the Coastal Plains physiographic province is thick (approximately 41,000 feet, equivalent to 7.8 miles or 12.5 km) and the oil and gas industry considers the sediments below the Triassic to be barren, petroleum geologists have drilled only a limited number of borings through the full Mesozoic and Cenozoic sections. Except for northern Florida, southern Georgia, and southeastern Alabama, few wells within the basin have penetrated pre-Pennsylvanian rocks (Reference 2.5.1-23). As a result, there is sparse data on the pre-Cenozoic sediments overlying basement rock.

The history of investigations of the Gulf of Mexico basin contains contradictory views on the origin and crustal type present beneath the marine and non-marine sediments that are known to be present. As late as 1967, some geologists favored the concept that the basin formed at the end of the Precambrian period and existed then as it is today. Recent interpretation of geophysical (seismic) survey data suggests that the crust beneath the Jurassic sediments is continental (Reference 2.5.1-23).

2.5.1.1.3.2 Paleozoic Stratigraphy

Little is known about the Paleozoic strata that are basinward from the structural rim of the Gulf of Mexico basin. Unmetamorphosed Paleozoic rocks are exposed in the southern Appalachian Mountains, the Ouachita Mountains, the Llano Uplift, and the Marathon Uplift of west Texas, plus in two small areas on the eastern edge of the Sierra Madre Oriental in Mexico.

A thickness of 45,000 to 52,000 feet (8.5 to 10 miles or 14 to 16 km) of Paleozoic rocks are exposed in the Ouachita Mountains of eastern Oklahoma and western Arkansas, with approximately 75 percent of this sequence being Late Mississippian through Middle Pennsylvanian in age. The Paleozoic section is thinner in the Marathon Uplift area, with a total of about 18,000 feet (3.4 miles or 5.5 km), of which about 66 percent are Upper Mississippian to Middle Pennsylvanian sandstones and shales with some interbedded limestones. In the Llano Uplift, the Paleozoic section is thinner, with only 3500 feet (1.1 km) of sandstones, shale, and limestone.

As recently as 1991, no wells in southern Texas (References 2.5.1-24 and 2.5.1-26) had been drilled deep enough to identify Paleozoic or older rocks. At that time, only two wells that penetrated Ouachita facies rocks had been drilled on the structural rim of the Gulf of Mexico basin. The scarcity of wells penetrating beneath the Jurassic in the onshore portions of south Texas makes it difficult to interpret the paucity of stratigraphic data and identify the crustal material, or to interpret the geophysical (seismic) data.

The absence of late Paleozoic and early Mesozoic marine sediments beneath the Coastal Plains (Reference 2.5.1-24) surrounding the Gulf of Mexico suggests that this was a positive, stable area until the Middle Jurassic. Rifting that accompanied the opening of the Atlantic and the Gulf of Mexico created depositional basins for Middle Jurassic salt deposits, as noted in the description of Mesozoic rifting in Subsection 2.5.1.1.2.4 above.

2.5.1.1.3.3 Mesozoic Stratigraphy

Geologic and geophysical evidence indicates that the site of the present Gulf of Mexico was part of Pangea, a supercontinent landmass at the beginning of the Mesozoic (Reference 2.5.1-23). The development of the Gulf of Mexico basin occurred in the Mesozoic with the breakup of Pangea and the opening of the Atlantic Ocean. Sediments from North America began to fill the newly formed basin. The text that follows describes the opening of the basin and the deposition of approximately 21,000 feet (4 miles or 6.4 km) of Mesozoic sediments in the Victoria County region. Figure 2.5.1-10 is a generalized Mesozoic stratigraphic column (Reference 2.5.1-27).

2.5.1.1.3.3.1 Triassic Stratigraphy

The Triassic was a period of tectonic activity comprising rifting in the Gulf of Mexico basin and breakup of Pangea. Red beds of Triassic-Jurassic ages are found in the Mesozoic rift basins; however, there are no outcrops of Triassic stratigraphic units within the Victoria County region (Reference 2.5.1-24). Red beds have been encountered in petroleum exploration wells in the Eagle Mills Formation in northeastern Texas. This formation is predominantly composed of red, greenish, or mottled shales and siltstones, which are similar to strata present in the Newark Basin and other Triassic grabens of the Appalachian region.

2.5.1.1.3.3.2 Jurassic Stratigraphy

The Gulf of Mexico basin did not appear as a structural feature until the Middle Jurassic period (Reference 2.5.1-23). Diabase dikes and sills associated with the rifting are present in the Gulf of Mexico basin and have been dated from Lower to Middle Jurassic by isotopic methods (References 2.5.1-28 and 2.5.1-29). Stratigraphic evidence within the region indicates that at approximately 200 Ma (Figure 2.5.1-10), the Middle Jurassic Louann Salt formed the basal Jurassic unit. A limited number of wells and seismic investigations are the source of data on the Louann Salt. The seawater that was evaporated to form the “mother” salt beds originated in the Pacific Ocean and entered the shallow Gulf of Mexico depression across the Mexico platform (Reference 2.5.1-24).

The Louann Salt is composed mainly of coarsely crystalline halite, with anhydrite as the chief additional mineral, but making up at most 10 percent of the rock (Reference 2.5.1-24). The Louann Salt is inferred to be present in the Victoria County region due to the presence of salt domes in the subsurface, as shown in Figures 2.5.1-11 and 2.5.1-12. The thickness of the Louann Salt bed varies with location. The original thickness may have ranged from more than 3300 feet (1 km) in east Texas, north Louisiana, and Mississippi salt basins to as much as 13,000 feet (2.5 miles or 4 km) in the Texas-Louisiana Gulf of Mexico slope area (Reference 2.5.1-24).

Over most of the Gulf of Mexico basin, migration of the salt has formed diapirs. Borehole and geophysical data indicates that the base of the salt shows little deformation and unconformably overlies the underlying rocks (Reference 2.5.1-24).

Following the deposition of the Middle Jurassic Louann Salt, the region was covered by a marine transgression (References 2.5.1-23 and 2.5.1-24) that was continuous through the Upper Jurassic. The Middle Jurassic Norphlet Formation represents the basal coarse clastic stratum in the Gulf of Mexico basin, extending from the Florida Panhandle to northeastern Mexico. Norphlet Formation sediments resulted from erosion of the Appalachian highlands to the north and east. The Norphlet Formation is mainly composed of sandstones and conglomeritic sandstones with a thickness less than 100 to 200 feet (30 to 61 meters) in the site region (Reference 2.5.1-24).

The Upper Jurassic sediments generally form a transgressive and conformable sequence with each successive unit pinching out further landward. This sequence has been interpreted as coastal onlap due to eustatic sea level rise. This information is based on stratigraphic data from petroleum wells. The Upper Jurassic section does not crop out in the United States part of the Gulf of Mexico basin (Reference 2.5.1-24). However, the Upper Jurassic sediments outcrop in Mexico.

In the Gulf of Mexico basin the Upper Jurassic is predominately marine, with non-marine fluvial and deltaic clastic sediments present in the northern and northwestern basin margins (References 2.5.1-23 and 2.5.1-24). The ancestral Mississippi River appears to have contributed clastics to the Gulf beginning late in the Jurassic – for perhaps 150 million years.

The Upper Jurassic sediments do not show evidence of large-scale tectonics (Reference 2.5.1-23). The strata are relatively uniform in lithology and do not abruptly change in thickness. The characteristics of Upper Jurassic strata were affected by (a) syndepositional movement of regional tensional normal fault zones, (b) syndepositional flow of the underlying Middle Jurassic salt, (c) fluctuations of sea level, and (d) preexisting topography (Reference 2.5.1-24).

The Smackover Formation, dated from the lower (Oxfordian) stage of the Upper Jurassic, conformably overlies the Middle Jurassic Norphlet Formation and is composed mainly of carbonates and calcareous shales. The lower unit is a dark-colored carbonate mudstone deposited in a low energy environment while the upper unit, which is a reservoir rock for oil and gas fields, is an öolitic carbonate deposited in a high energy, shallow marine environment. The thickness of the Smackover Formation might reach as much as 1600 feet (488 meters) in the VCS region (Reference 2.5.1-24).

In the United States, the Kimmeridgian units of the Gulf Coastal Plains are composed of clastics, carbonates, and evaporites (Reference 2.5.1-23). The term “Buckner” has been applied to the lower evaporite section and the overlying, predominately clastic section has been called “Haynesville.” The Gilmer limestone identifies the limestone equivalent of the Haynesville Formation. Salvador (Reference 2.5.1-24) uses the “Haynesville” terminology to describe the entire section between the Oxfordian stage Smackover Formation and the base of the younger Tithonian stage Bossier Formation. The Buckner and Gilmer are defined as stratigraphic members of the Haynesville Formation away from the San Marcos Arch region. The description of these units in the Victoria County region follows the grouping proposed by Salvador (Reference 2.5.1-24).

The Buckner member is characterized by white, pink, or gray massive or nodular anhydrite in thick, massive beds with thinner interbeds of dolomite, argillaceous limestone, anhydritic limestone, and anhydritic or dolomitic mudstone. Downdip, the Buckner evaporites appear to grade into öolitic limestone similar to that in the older Smackover Formation. In the Victoria County region, the upper part of the Haynesville Formation is the carbonate Gilmer member. The Haynesville Formation may be more than 1600 feet (488 meters) thick in the Victoria County region (Reference 2.5.1-24).

As described previously, there is no evidence that the proto-Gulf of Mexico had a connection to the Atlantic Ocean until after the deposition of the Buckner evaporites during the Kimmeridgian stage of the Upper Jurassic. Throughout the region, the Buckner represents low energy, hypersaline coastal lagoons overlying the high energy, shallow water marine upper portions of the Smackover Formation (Reference 2.5.1-24). In the eastern part of the basin, the upper Haynesville Formation is composed of terrigenous clastic sediments, but in the Victoria County region, the influx of clastic sediments was not as strong, and the Gilmer limestone was deposited in high energy, shallow marine conditions.

Along the northern part of the basin, the upper stage of the Upper Jurassic, the Tithonian, consists of a thick wedge of coarse clastic sediments, whereas on the western and southwestern flanks the

Tithonian section is thinner and composed of fine clastic sediments (shales, calcareous shales) with non-clastic sediments (argillaceous limestones) (References 2.5.1-23 and 2.5.1-24). Sediments within the northern (United States) part of the Gulf of Mexico basin comprise the Cotton Valley Group, which is further divided into the lower Bossier Formation, a predominately marine shale unit, and the overlying non-marine clastics of the Schuler Formation. In the VCS region, the Cotton Valley Group is shaly, but it becomes increasingly sandy toward the North. Deposition of the Schuler Formation continued into the Lower Cretaceous.

The offshore Bossier Formation is composed of dark gray to black marine shales and calcareous shales with occasional thin beds of fine-grained sandstone. The Bossier Formation is time transgressive, and in the deeper portions of the Gulf of Mexico basin, the Bossier Formation is time-equivalent to the upper part of the Smackover and Haynesville formations and most of the Cotton Valley Group. The Schuler Formation is composed of a variety of clastics, including mudstones, shales, siltstones, sandstones, and conglomerates. The coarse clastics give way basinward to finer grained sediments that are shades of red and maroon while the coarser clastics shoreward are greenish gray to white. The maximum thickness of the Tithonian sediments in the south Texas region is between 1600 to 2300 feet (488 to 701 meters). This stratigraphic data indicates that the developing Gulf of Mexico was connected to the Atlantic Ocean during this last stage of the Upper Jurassic period. (Reference 2.5.1-24)

2.5.1.1.3.3.3 Cretaceous Stratigraphy

The thickness of the Lower Cretaceous section varies from 8000 to 11,000 feet (2.4 to 3.4 km) along the northern shelf, thinning to less than 6000 feet (1.8 km) toward the central part of the basin. In the Victoria County region, the Lower Cretaceous section is represented by the carbonates and shales of the Trinity Group, comprised of the Hosston, Sligo, Pearsall, Glen Rose, and Stuart City formations. The estimated total thickness of the Trinity Group in the site region is about 3200 feet (1 km) (Reference 2.5.1-30).

The Hosston Formation unconformably overlies the Cotton Valley strata in the Victoria County region. To the east, where Cretaceous streams drained the southern Appalachians and the continental interior, the Hosston Formation consists of fine-to-coarse clastics. The Hosston Formation also contains a large amount of chert in the Victoria County region. Across the northern shelf, the Hosston interfingers with the argillaceous reef limestones of the Sligo Formation. (Reference 2.5.1-30)

The Pearsall Formation conformably overlies the Sligo Formation. The lower member of the Pearsall Formation, the Pine Island Shale, consists of dark shale interbedded with gray, thin limestones. The James Limestone member (also known as the Cow Creek in the site region) overlies the Pine Island Shale from south Texas to Florida (Reference 2.5.1-30). The James Limestone is a dense,

non-porous gray limestone interbedded with shale. Overlying the James Limestone is the Bexar Shale member, forming the uppermost member of the Pearsall Formation.

Conformably overlying the Pearsall Formation are the limestones of the Glen Rose Formation. In south Texas, these rocks are gray argillaceous dolomite with anhydrite layers. Downdip toward the basin, the Glen Rose shelf carbonates interfingers with reef limestones containing rudist bivalves, corals, mollusks, and other shallow water bank fauna. Where mappable, this lithology has been named the Stuart City Formation.

Unconformably overlying the Glen Rose limestones are the shelf limestones of the Fredericksburg Group. In the VCS region, this group includes the Edwards Formation, which is subdivided into the transgressive West Nueces Limestone member, the overlying McKnight Limestone member, and the McKnight Evaporite member. Downdip, the Fredericksburg Group interfingers with the Stuart City Formation, which in turn grades into the basinal micrites and shales of the Atascosa Formation (Reference 2.5.1-30).

The deposition of Fredericksburg Group strata came to a halt as the land gradually rose and the shoreline regressed at the beginning of Washita Group deposition. The Washita Group includes (from older to younger): the Upper Cretaceous Georgetown Formation, the Del Rio Formation, and the Buda Limestone Formation. The Georgetown limestones grade basinward into the upper reef facies of the Stuart City Formation (Reference 2.5.1-30).

During the Lower Cretaceous, the ancestral Gulf of Mexico was connected with the Atlantic and Pacific oceans as well as with the Western Interior Seaway (Reference 2.5.1-33). The surrounding platforms were stable and were covered with shallow seas. Clastic sediments were deposited along the northern margin by major streams that drained the southern Appalachian Mountains. Long-shore currents flowed westward, carrying these clastics away from the deltas. The maximum extent of carbonate reef deposition took place during the end of the Lower Cretaceous. (References 2.5.1-23 and 2.5.1-30)

In the Gulf of Mexico basin, the Upper Cretaceous was a period of generally high sea levels that supported the deposition of a continuous cover of Upper Cretaceous sediments. These sediments thicken downdip (basinward) to a shelf edge that is postulated to follow the reef of the Lower Cretaceous period (Reference 2.5.1-23). In the northern shelf areas, clastic sediments were derived from the southern Appalachian Mountains as they had been in the Lower Cretaceous. However, toward the end of the period, clastics from a western source, possibly from the nascent Laramide Uplift, make a more pronounced contribution. The thickest Upper Cretaceous sections are found in the major embayments, with as much as 5500 feet (1676 meters) in the Rio Grande Embayment (Figure 2.5.1-13).

The middle of the Upper Cretaceous occurs within the Washita Group. The base of the "Gulfian Series" occurs at the physical break in the mid-Cenomanian, the lowermost stage in the Upper Cretaceous. The magnitude of this unconformity represents a profound change in the depositional regime over most of the Gulf of Mexico basin. The widespread distribution of this unconformity here and in many other regions of the world suggests a major eustatic lowering of the sea level. (Reference 2.5.1-23)

The Upper Cretaceous Washita Group includes the marly, fossiliferous limestones of the Georgetown Formation, the gypsiferous clays and fossiliferous shales of the Del Rio Formation, and the dense, porcelaneous Buda Limestone Formation that unconformably overlies the Fredericksburg Group (Reference 2.5.1-31).

The Woodbine Formation is the first Upper Cretaceous unit deposited above the unconformity (Reference 2.5.1-23). In general, this unit is a fluvio-deltaic to marginal marine sequence that is highly variable and complicated. The Woodbine Formation thins south of the type locality in Woodbine, Texas, and the sand content decreases. The outcrop area consists of black non-calcareous shales that are the upper part of the Woodbine Formation (Reference 2.5.1-32). The shale thins southward and pinches out on the San Marcos Arch, and therefore may not be present beneath the VCS site.

The Eagle Ford Group is the fine-grained phase of terrestrial deposition that began with the coarse-grained deposits of the Woodbine Formation in north Texas. By the time the Eagle Ford Group reaches the VCS/San Marcos Arch region (see Figure 2.5.1-14), the strata are thin and consist of fissile, calcareous, carbonaceous black shale with interbeds of dark limestone. The Woodbine/Eagle Ford strata becomes thinner or are locally absent over the San Marcos Arch, and may reach a thickness of about 225 feet (69 meters) in south Texas (Reference 2.5.1-32).

The Austin Group disconformably overlies the Eagle Ford Group and reflects a change in depositional environment in many areas from clastic sediments to limestone. In the VCS region, the Austin Group (undifferentiated) consists of a lower chalk and a limestone that transition to calcareous shale and overlying limestone. The undifferentiated group of strata is estimated to be 555 feet (170 meters) thick toward the Rio Grande Embayment, thinning over the San Marcos Arch.

The Anacacho Limestone of the Taylor Group disconformably overlies the Austin Group. The Anacacho Limestone is mainly a clayey, bioclastic limestone interbedded with clay and marl and can be as much as 800 feet (244 meters) thick. Downdip, the Anacacho Limestone grades into shelf mudstones of the Upson Formation. The Upson Formation consists of as much as 500 feet (152 meters) of fossiliferous dark to greenish gray clay. Outcrops of the Upson Formation are limited, in part due to erosion, but it is commonly recognized in the subsurface where it is conformably overlain by the San Miguel Formation.

The San Miguel Formation consists of as much as 400 feet (122 meters) (in outcrop) of fossiliferous sands and sandy limestones interbedded with gray clays; as much as 1150 feet (351 meters) have been identified in the subsurface. The San Miguel Formation probably was deposited in a wave-dominated deltaic system that was transitional from the underlying Upson Formation, a shallow water shelf deposit and transitional to the overlying Olmos Formation. All of these strata are truncated against the San Marcos Arch (Figure 2.5.1-12). In some areas of south Texas, there was uplift and erosion so that the San Miguel and Upson formations are missing and the Escondido Formation (uppermost Cretaceous) disconformably overlies the Anacacho Limestone directly. This uplift may be related to the early stages of the Laramide orogeny, which probably contributed to deposition of the clastics in the deltaic complex of the San Miguel Formation (Reference 2.5.1-32).

The Laramide orogeny continued to create a source for clastics in the Upper Cretaceous. In the area of the San Marcos Arch, the claystone, chalky marls, and sandy strata of the east Texas Navarro Group give way to coarser clastics of the Olmos Formation. In outcrop, the Olmos Formation is mainly non-marine, irregularly bedded clays, shales, and sandstones, accompanied by seams of coal, lignite, fire clay, and carbonaceous shales (Reference 2.5.1-32). The Olmos Formation ranges between 400 to 500 feet (122 to 152 meters) in outcrop, but thickens to more than 1300 feet (396 meters) in the Rio Grande Embayment subsurface.

The Olmos Formation is transitional to the Escondido Formation, the uppermost Cretaceous unit in south Texas. The lower three-quarters of the formation consist of bioturbated mudstones interbedded with sandstones and coquina beds. Some sandstone contains ripple marks or cross-bedding that indicates a shallow-water depositional environment. The upper quarter consists of inner-shelf deposits of sandy mudstone, siltstone, and impure limestone. The Escondido Formation is about 800 feet (244 meters) in outcrop in Texas, thickening gulfward to approximately 1300 feet (396 meters) (Reference 2.5.1-32).

Along the northern Gulf of Mexico coast, Tertiary sediments lie disconformably on Upper Cretaceous strata. Basal Paleocene units contain reworked Cretaceous fossils along with those from the Paleocene in most areas. However, in the Rio Grande Embayment and the Brazos River sequence in Central Texas, deposition may have been continuous from the Cretaceous through the Paleocene. This is largely due to the continued uplift in the Rocky Mountains resulting from the Laramide orogeny. The connection between the Gulf of Mexico and the Western Interior Seaway was probably closed during the Upper Cretaceous (Reference 2.5.1-33).

2.5.1.1.3.4 Cenozoic Stratigraphy

Early Cenozoic rocks and geologic structures record the final tectonic activity of the Laramide orogeny west of the region. The uplift of the Rocky Mountains caused an influx of clastic sediments originating from the new terrestrial source to the northwest. Large volumes of clastic sediments

began to accumulate in an offlapping regressive (sea level fall) depositional style that continued throughout the Cenozoic in Texas (Reference 2.5.1-34). Subsidence of the basin was due mainly to sedimentary loading of the crust by prograding wedges of these clastic sediments and, during the Pleistocene, the variation of ice volumes that had a eustatic effect on sea levels. These rapidly accumulating sediments contributed to the development of growth faults, as well as the development of salt and shale diapirs in the Jurassic salt and lower Cenozoic over-pressured shales, respectively (Reference 2.5.1-23).

The considerable heterogeneity of the Cenozoic sediments, the discontinuity of the beds, and a general absence of index fossils and diagnostic electric log signatures in the subsurface make correlation of the lithologic units problematic. Since 1903, at least seven stratigraphic classifications have been proposed (Reference 2.5.1-25). Figure 2.5.1-15 contains a generalized Cenozoic stratigraphic column. Thicknesses shown in Figure 2.5.1-15 are estimates taken from the Texas Bureau of Economic Geology cross sections. Groundwater is obtained from the Coastal Lowlands aquifer system, which has a lower bound in the Catahoula confining system at depths of 5000 to 6000 feet (1.5 to 1.8 km). Various authors estimate that there are approximately 20,000 feet (3.8 miles or 6.1 km) of Cenozoic sediments beneath the site region (References 2.5.1-35 and 2.5.1-25).

2.5.1.1.3.4.1 Paleocene Stratigraphy

The Midway Group contains the basal Cenozoic sediments along the Gulf Coastal Plains and overlies Upper Cretaceous strata in disconformable contact. The Kinkaid Formation, the basal unit of the Midway Group, is mostly composed of limestone, calcareous sand, and sandstone and is usually less than 100 feet (30 meters) thick. The overlying Wills Point Formation, mainly a dark-gray to black, micaceous clay, is present throughout the area and represents the bulk of the Midway Group. The thickness of the Wills Point Formation is more than 2500 feet (762 meters) in southern Texas (Reference 2.5.1-35).

Major Cenozoic deltaic sediment sources on the Texas Gulf Coast are shown in Figure 2.5.1-16. The Lower Wilcox Group reflects the first Cenozoic episode of major deltaic offlap with the influx of clastic sediment into the west and central Gulf of Mexico basin along the Houston and Central Mississippi-Holly Springs delta complexes (Reference 2.5.1-34). The considerable sediment loading by the delta complexes was responsible for the first of many episodes of growth faulting and salt mobilization during the Cenozoic (Reference 2.5.1-24). These were the first systems to contribute to growth faulting and salt mobilization (Reference 2.5.1-34) through application of lithostatic pressure developed by the large volume of sediments deposited in the delta. In south Texas, a strandline-shelf system developed that generally covered the same area as the Cretaceous limestone platforms. The sands were reworked from the deltaic lobes and were transported southwestward along the coast. The sands grade into shelf mud toward the Gulf of Mexico.

The Wilcox Group is undifferentiated in Texas where it lacks regionally mappable units (Reference 2.5.1-35). In the Mississippi Embayment, it has a maximum thickness of 1200 feet (366 meters), and it becomes thousands of feet thick toward the Gulf. It is estimated to be 2000 feet (610 meters) thick in the Victoria County region. The Wilcox Group typically consists of sandy-clayey deposits; lignite is common and is characterized by coarser sandy, deltaic, and non-marine sediments.

2.5.1.1.3.4.2 Eocene Stratigraphy

In the Lower to Middle Eocene, the sediment distribution systems established in the Paleocene generally continued (Reference 2.5.1-34). Sediments eroded from the Laramide uplift prograded into the western Gulf of Mexico basin as much as 20 miles (32 km). A broad alluvial coastal plain derived from fluvial systems flowing from the continent extended along the central and south Texas coast. The sediments constitute the Claiborne Group (Figure 2.5.1-15 and Reference 2.5.1-24), a classic example of strata produced by alternating transgressive and regressive marine-non-marine depositional cycles.

The Carrizo Sand is the basal Claiborne unit in Texas. Deposited unconformably on the clays of the Wilcox Formation, the Carrizo Sand varies in thickness from 100 to 1200 feet (30 to 366 meters) in Texas and is estimated to be approximately 800 feet (244 meters) thick in the Victoria County region (Reference 2.5.1-35).

The Reklaw Formation conformably overlies the Carrizo Sand. It is a transgressive (rising sea level) marine clay unit composed largely of dark shales and sands. The lower part of the Reklaw Formation is glauconitic and partly non-marine and reaches a thickness of approximately 50 feet (15 meters) in the Victoria County region (Reference 2.5.1-35).

In the Middle Eocene, the Queen City Sand and the Sparta Sand were deposited in two marine-non-marine depositional cycles. During the depositional period, a wave-dominated barrier island complex extended from the present coast of Mexico northward to the Houston Embayment. The outer shelf, slope, and deep basin of the Gulf of Mexico remained sediment starved (Reference 2.5.1-34).

The Queen City Sand is predominately light-gray to grayish-brown very fine to medium quartz sand interbedded with dark carbonaceous shale, silt, and lignite. In the region of the VCS, the Queen City Sand thickness is estimated to be approximately 50 feet (15 meters) thick (Reference 2.5.1-35).

The Weches Formation overlies the Queen City Sand. It is predominately glauconitic and has been mined as an iron ore where leaching of the glauconite has concentrated the iron. The sands and clays are highly cross-bedded and interbedded with dark-gray to black glauconitic clay and shale (Reference 2.5.1-35). Its thickness within the region has been estimated at 60 feet (18 meters)

(Reference 2.5.1-36). The Weches Formation is highly fossiliferous, which distinguishes it from the units above and below. It is present beneath the Victoria County region.

The Sparta Sand overlies the Weches Formation and is composed of very fine to medium unconsolidated quartz sand interbedded with sandy clay and clay. Lignite is common. The Sparta Sand is about 100 feet (30 meters) thick at the outcrops to more than 1000 feet (305 meters) thick in the subsurface near the Mississippi Embayment axis. The thickness of the Sparta Sand in the region of the VCS is estimated at about 100 feet (30 meters) (Reference 2.5.1-35).

The Cook Mountain Formation overlies the Sparta Sand in what may be a gradational contact (Reference 2.5.1-35). The formation in south Texas is more than 900 feet thick and has a large proportion of sandy clay and sand containing disseminated glauconite. Interbedded clays are bluish gray to black and become the dominant lithologic type as the formation thickens downdip. In the Victoria County region, the Cook Mountain Formation is estimated to be 300 to 400 feet (91 to 122 meters) thick.

The Yegua Formation is composed of massive laminated and cross-bedded, fine- to medium-grained sand. Sandy clay and clay, thin lignite beds and glauconitic sands are present in some places. The maximum thickness of the Yegua Formation is more than 1800 feet (550 meters) beneath the Texas Gulf Coast (Reference 2.5.1-35). Beneath the region of Victoria County, the formation thickness is estimated to be 800 to 1000 feet (244 to 305 meters). Volcanic ash in the Yegua sediments reflects the uplift and crustal heating in the Mexican Cordillera and western Gulf of Mexico. Areas in the western Gulf basin were uplifted, and the area of active deposition of ash extended eastward (Reference 2.5.1-34). The Gulf of Mexico margin prograded as much as 15 miles (24 km) during this time.

Jackson Group sediments conformably overlie the Yegua Formation deposits of the underlying Claiborne Group. The Jackson Group was deposited during the Upper Eocene and stratigraphic equivalents are present throughout the Gulf Coastal Plains. The presence of volcanic ash and coarser volcanic-derived clastics in the Jackson Group reflect volcanic activity in Mexico and the southwestern United States. This group and its stratigraphic equivalents were deposited during the last major marine transgression that covered the Coastal Plains and occupied the Mississippi Embayment (Reference 2.5.1-33 and Figure 2.5.1-14).

The lowermost unit of the Jackson Group is the Caddell Formation. The lower part of the Caddell Formation is typically a marine deposit and is composed of gray calcareous sandstone and greenish calcareous clays, which may contain some glauconite. The upper part exhibits fewer marine characteristics and locally contains lignite and thin chocolate shales and interbedded sands. The Caddell Formation ranges from 30 to 300 feet (9 to 91 meters) in thickness in the VCS region (Reference 2.5.1-35).

The Wellborn Sandstone Formation, which overlies the Caddell Formation, is a massive, gray sandstone with interbedded marine clay units. The middle part is a highly fossiliferous marine facies consisting of sandy, marly clays. The upper part is a massive gray to white clayey sandstone. The Wellborn Sandstone is between 100 and 300 feet (30 and 91 meters) thick in the region of the VCS (Reference 2.5.1-35).

The Manning Clay Formation consists of carbonaceous, dark brown clay alternating with two beds of gray sandstone and overlies the Wellborn Sandstone. The clay is essentially non-marine, but some marine shale beds are present locally. The thickness of the Manning Clay is between 250 and 350 feet (76 and 107 meters) in the VCS region (Reference 2.5.1-35).

The Whitsett Formation, the uppermost unit in the Jackson Group, is mainly non-marine cross-bedded sandstone interbedded with tuffaceous shale and fine sandy tuff. The sands are generally fine to medium grained but may be very coarse and conglomeratic in places. The Whitsett Formation is about 135 feet (41 meters) thick in the region of the VCS (Reference 2.5.1-35).

2.5.1.1.3.4.3 Oligocene Stratigraphy

During the Oligocene, multiple fluvial systems developed resulting in an influx of sediments from Mexico and the southwestern United States. The Norias Delta system on the Rio Grande axis merged with the Norma Delta to the south. A third system developed along the Houston Embayment while the fourth developed along the central Mississippi margin (Figure 2.5.1-16). The eastern Gulf of Mexico continued to receive a minimum of clastic sediments during the Oligocene. Clastic sediments comprise the Vicksburg Group and the overlying Catahoula Sandstone (tuff) Formation (Reference 2.5.1-34).

The Vicksburg Group is composed of a variety of marine sediments, varying from sandstones and clays to marl and limestone (Reference 2.5.1-35). The Frio Clay (not to be confused with the Frio Formation of the overlying Catahoula Formation) is probably time-equivalent with the Vicksburg Group (Reference 2.5.1-35). The Frio Clay facies typically is composed of massive dark, greenish-gray, red, and blue gypsiferous clay interbedded with sandy clay, sand, and sandstone. The thickness of the Vicksburg Group in the region of the VCS ranges from 400 feet (122 meters) (Reference 2.5.1-37) to 1000 feet (305 meters) (Reference 2.5.1-38).

A major influx of clastic sediments was deposited in the region during the Upper Oligocene; it is generally designated at the surface as the Catahoula Formation (Reference 2.5.1-33). The Catahoula Formation consists of three units that are identified only in the subsurface. These are, from oldest to youngest: the Frio Formation (not the Frio Clay facies of the Vicksburg), the Anahuac Formation, and the upper part of the Catahoula Tuff. The Catahoula Formation consists of approximately 60 percent volcanic material, mainly airborne ash from Mexican volcanoes, and 30 percent sand (Reference 2.5.1-25). The Catahoula Sandstone Formation ranges from 200 to

800 feet (61 to 244 meters) in thickness and thickens downdip to thousands of feet where an accretionary wedge of fossiliferous marine clay, called the Anahuac Formation, occurs in the upper section. The thickness of the entire Catahoula/Frio/Anahuac in the region of the VCS is estimated to be approximately 800 feet (244 meters) (Reference 2.5.1-25). The age of the Catahoula Sandstone is uncertain, but the U.S. Geological Survey (USGS) designates the unit as both Oligocene and Miocene (Reference 2.5.1-25).

2.5.1.1.3.4.4 Miocene Stratigraphy

The Miocene was a period of relative paleogeographic stability in the Gulf of Mexico basin (Reference 2.5.1-34). The increased presence of Cretaceous limestone clastics in the Miocene fluvial deposits indicates uplift of the Edwards Plateau north of the Gulf of Mexico basin and adjacent areas. During the Middle Miocene epoch, the Corsair Delta developed in the region of the Colorado River, an area that had not previously been a depocenter. The Corsair Delta apron formed a sandy depositional element on the Gulf of Mexico floor (Reference 2.5.1-34). The Oakville Sandstone member is a sandy facies in the lower part of the Fleming Formation that unconformably overlies the Catahoula Formation in the site region (References 2.5.1-25 and 2.5.1-35). The Oakville Sandstone is composed of non-marine, irregularly bedded clastics consisting of coarse sands and interbedded clay. The thickness of the Oakville Sandstone ranges from about 200 to about 500 feet (about 61 to about 152 meters) regionally (Reference 2.5.1-35).

The Fleming Formation unconformably overlies the Oakville Sandstone. The Fleming Formation and the Oakville Sandstone are similar lithologically, but the Oakville Sandstone is much coarser grained. In the Victoria County region, the Fleming Formation is largely variegated yellow, green-red, pink-blue, and purplish gray or greenish gray clay (Reference 2.5.1-35). The strata may be calcareous and contain thin chalky limestone as well as cross-bedded sands. The Fleming Formation is about 200 feet (61 meters) thick in outcrop and can be thousands of feet thick in the subsurface. The estimated thickness of the Fleming Formation is 1500 feet (457 meters) in the Victoria County region (Reference 2.5.1-25).

2.5.1.1.3.4.5 Pliocene Stratigraphy

By the early Pliocene, sediment supply and accumulation had shifted eastward to the Mississippi depositional axis. Sediments were carried east and west from the Mississippi deltas, forming gulf shore deposits that were deposited as thin veneers, compared with previous deposits (Reference 2.5.1-34).

In the late Pliocene, the three central Gulf fluvial systems, the Red River, Central Mississippi River, and Eastern Mississippi River formed a composite delta system in which the Red River continued to dominate the sediment supply (Reference 2.5.1-34). The fluvial systems of the western and northwestern Gulf were dormant.

The initial influence of continental glaciation on drainage patterns and sediment load and discharge to the Gulf was recorded in the late Pliocene, including initiation of the first phase of the Mississippi fan system (Reference 2.5.1-34).

The Goliad Sand Formation was the dominant unit deposited during the Pliocene epoch along the Texas Gulf Coast. The Goliad Sand overlies the Fleming Formation and ranges from a basal strata of coarse undivided sediments to indurated units, whitish to pinkish gray in color, and ranging in grain size from very fine to coarse. The cement is typically calcium carbonate (caliche). Clay interbeds are grayish and may be calcareous. The bedding suggests that the Goliad Sand was deposited as river-bottom sediments (Reference 2.5.1-35). The estimated thickness of the Goliad Sand ranges from 1000 to 1700 feet (305 to 518 meters) in the Victoria County region (Reference 2.5.1-39).

2.5.1.1.3.4.6 Pleistocene Stratigraphy

The Pleistocene depositional record shows pulses of sandy outwash during glacial retreats and cyclic sea level changes. The inland Mississippi and Red River fluvial systems were separate inland but coalesced on the depositional coastal plain. During the Pleistocene, the Rio Grande and Colorado/Brazos deltas formed minor depositional centers on the Texas Gulf Coast (Reference 2.5.1-34). Sediments in the Victoria County area are attributed to the Guadalupe and Colorado/Brazos fluvial systems (Figure 2.5.1-17).

The extensive Pleistocene alluvial-deltaic plains of the Texas Gulf Coast represent a succession of valley fill deposits designated as the Willis Sand (oldest), the Lissie, and the Beaumont (youngest) formations (Reference 2.5.1-40).

The Willis Sand Formation overlies and interfingers with the Goliad Sand. It is locally extensive in the region, but occurs over a small geographic area (Reference 2.5.1-39). The Willis Sand Formation is a sequence of unfossiliferous sand and gravelly sands with subordinate clays. It is estimated to range from 400 to 500 feet (122 to 152 meters) in the Victoria County region.

Where the Willis Sand is absent, the overlying Lissie Formation is in unconformable contact with the underlying Goliad Sand. The Lissie Formation (also differentiated as the Bentley and Montgomery formations in Subsection 2.4.12) crops out in a 30-mile wide band parallel to and about 20 miles inland along the Texas Gulf Coast. The sediments are partially flood plain deposits and partially deltaic sands, silts, and mud. The sediments are described as reddish, orange, and gray fine-to-coarse grained, cross-bedded sands with the base of the formation often indicated by caliche (Reference 2.5.1-25). In the Victoria County region, the Lissie Formation is estimated to range from 500 to 700 feet (152 to 210 meters) in thickness (Reference 2.5.1-39).

The near-surface sediments in the Victoria County region belong to the Beaumont Formation. From the Louisiana/Texas border to the Rio Grande, the Beaumont Formation is recognized as a series of

multiple, cross-cutting and/or superimposed incised stream channel fills and over-bank deposits formed during glacio-eustatic cycles (Reference 2.5.1-40). The Beaumont Formation is composed of poorly bedded, marly, reddish-brown clay interbedded with lenses of sand (Reference 2.5.1-25); its thickness varies up to 500 feet (152 meters) (Reference 2.5.1-41).

2.5.1.1.3.4.7 Holocene Stratigraphy

The Holocene surficial alluvial systems exhibit the largest outcrop area of all the units in the Texas Gulf Coast. The Brazos, Trinity, Nueces, and Rio Grande alluvial basins consist of terrace gravels, buried sand deposits, and point bar deposits with grain sizes ranging from clay to gravel. The flat-lying floodplain deposits typically consist of sand and gravel in the lower part and silt and clay in the upper part (Reference 2.5.1-39). In the Gulf Coastal Plains, the “Deweyville” terraces extend from Texas to east of the Mississippi River. They consist of three main erosional face terraces. These terraces are filled with alluvial deposits from nearby Pleistocene glacial streams such as the Sabine, Trinity, and Colorado. The “Deweyville” terrace deposits most likely formed as the result of variability in stream meander geometry of the ice age discharge streams. This may be a reflection of the influence of bank stabilizing mud (Reference 2.5.1-42).

2.5.1.1.4 **Regional Tectonic Setting**

In the late 1980s, EPRI-SOG developed a seismic source model for the central and eastern United States (CEUS), including the region surrounding the VCS site. This source model is the basis for the seismic hazard characterization of the VCS site and is referred to here as the EPRI-SOG model. The CEUS is a stable continental region that is characterized by low rates of crustal deformation and no active plate boundary conditions. As such, the EPRI-SOG source model was based on interpretations of the seismic potential of tectonic features within the CEUS and historical seismicity rates. Six independent earth science teams (ESTs) developed tectonic interpretations (Reference 2.5.1-1), and these interpretations represented the general state of knowledge incorporating the range of uncertainties within the geoscience community as of 1986 with respect to the seismic potential of tectonic features within the CEUS. The EPRI-SOG source model (Reference 2.5.1-1) for the site region is thoroughly described in the 1986 reports and is summarized in Subsection 2.5.2.2.

Since 1986, various investigators have completed additional geological, seismological, and geophysical studies in the CEUS and in the VCS site region. The purpose of this subsection is to summarize the current state of knowledge of the tectonic setting and tectonic structures in the site region and to summarize information acquired since 1986 that is relevant to the assessment of seismic sources.

A global review of earthquakes demonstrated that nearly 70 percent of stable continental region earthquakes with a moment magnitude (**M**) ≥6 occurred in areas of crust extended during the

Mesozoic and Cenozoic eras (Reference 2.5.1-43). Additional data indicates an association between Late Proterozoic rifts and modern seismicity in eastern North America (References 2.5.1-44, 2.5.1-43, and 2.5.1-45). Regional gravity and magnetic data indicates that continental crust extended during the Proterozoic and Mesozoic eras underlies at least part of the 200-mile (322-km) site region (References 2.5.1-46 and 2.5.1-47) (Figures 2.5.1-18 and 2.5.1-19). As described in this subsection, however, there is no positive evidence for late Cenozoic seismogenic activity of any tectonic feature or structure in the site region (References 2.5.1-48 and 2.5.1-49). Although research during the last two decades has modified our understanding of the tectonic evolution and processes in the Gulf Coastal Plains and the Gulf of Mexico basin, no new structures or features have been identified in the site region since 1986 that show clear evidence of seismogenic potential greater than what was recognized and incorporated in the EPRI-SOG seismic source model (Reference 2.5.1-1).

The regional tectonic setting of the VCS site is presented in the following subsection. This subsection includes descriptions of: (a) plate tectonic history of the Gulf of Mexico and the Gulf Coastal Plains, (b) origin and orientation of regional tectonic stresses, and (c) principal regional tectonic structures.

2.5.1.1.4.1 Tectonic History of the Site Region

2.5.1.1.4.1.1 Overview

As described in Subsection 2.5.1.1.1, the VCS site lies within the Gulf Coastal Plains physiographic province (Figures 2.5.1-1 and 2.5.1-3), which extends from Mexico on the west and southwest to Florida on the east (Reference 2.5.1-50). The Gulf Coastal Plains developed as part of the geologic and physiographic evolution of the Gulf of Mexico basin (Reference 2.5.1-51), an ocean basin that opened in the Triassic/Jurassic due to extensional rifting of Pangea along the trend of the Ouachita orogenic belt (Reference 2.5.1-21). The Ouachita orogenic belt is a Paleozoic thrust belt that formed during the closing of an ocean basin along the southeastern margin of ancestral North America (i.e., Laurentia) and extends for more than 2100 miles (3400 km) between western Texas and Mississippi (References 2.5.1-52 and 2.5.1-21). Rocks of the Ouachita belt are exposed in the Ouachita Mountains of eastern Oklahoma and western Arkansas as well as in the Llano Uplift of central Texas and the Marathon Uplift of the Big Bend area of west Texas. Between these widely separated exposures, the Ouachita orogenic belt extends continuously beneath Mesozoic and Tertiary marine sediments that fringe the northern margin of the Gulf Coastal Plains (References 2.5.1-53, 2.5.1-52, and 2.5.1-21). The tectonic events associated with the formation of the Ouachita belt have significantly influenced the structure of the crust in the site region and are summarized in the following subsections.

2.5.1.1.4.1.2 Precambrian to Paleozoic Plate Tectonic History

Because outcrops of Precambrian rocks are limited in Texas, geologic evidence recording the tectonic history of the Precambrian is sparse. The oldest known rocks in Texas are from the Early to

Middle Proterozoic Sierra-Grande Chaves tectonic cycle (approximately 1.4 Ga), a composite and incomplete Wilson cycle (i.e., an incomplete episode of continental rifting that forms an ocean basin and the subsequent closure of that basin through collisional orogenesis) (Reference 2.5.1-54). Rocks associated with the Sierra-Grande Chaves tectonic cycle have been positively identified within the subsurface in northwestern Texas where they are buried beneath younger sediments and may also be represented by the Coal Creek rocks of the Llano Uplift (Subsection 2.5.1.1.1.2) (Reference 2.5.1-14). Much of the information on these rocks comes from exposures of equivalent lithologies and formations in New Mexico and Oklahoma (Reference 2.5.1-54).

The oldest subaerially exposed rocks in Texas are those related to the Middle Proterozoic Grenville tectonic cycle, locally referred to as the Llano cycle (Reference 2.5.1-54). Rocks from this cycle have limited surficial exposures and crop out primarily in the Llano Uplift and to a more limited extent in westernmost Texas (References 2.5.1-55, 2.5.1-54, and 2.5.1-56) (Figures 2.5.1-2a, 2.5.1-3, and 2.5.1-12). These rocks are interpreted as the remains of the arc-continent and continent-continent collision that occurred along southern Laurentia, the ancestral North American continent (Reference 2.5.1-57).

In late Precambrian to Cambrian time, Laurentia broke up along a series of north-northeast-trending rifts as a result of seafloor spreading and the development of an ocean basin that contained the Iapetus Ocean (References 2.5.1-58 and 2.5.1-21). The rift system along the eastern margin of Laurentia was approximately coincident with the present Appalachian mountain belt (Reference 2.5.1-58). The locus of rifting stepped abruptly westward near the latitude of Alabama and Mississippi along a west-northwest-trending transform fault zone (Reference 2.5.1-19). This transform fault zone terminated to the west at the northern end of a north-northwest-trending rift system that was located approximately between the Sabine Uplift and the Fort Worth Basin (See Reference 2.5.1-59, Figure 1) (References 2.5.1-59 and 2.5.1-62). The transform fault system formed a jog in the ancestral continental margin that roughly parallels the northern rim of the present Gulf Coastal Plains (Reference 2.5.1-21). Two failed rift basins, or aulacogens, developed along Laurentia's southern margin (Reference 2.5.1-60). One of the aulacogens, the Reelfoot Rift, trends northeast-southwest and underlies the Mississippi Embayment (Figure 2.5.1-8). The other failed rift basin, the southern Oklahoma aulacogen, trends west-northwest to east-southeast along the border between southern Oklahoma and northeastern Texas (References 2.5.1-60 and 2.5.1-21).

The rifted southern margin of Laurentia in early Paleozoic (Cambrian) time included structures typical of passive margins, such as grabens and fault-bounded basins open to the ocean (References 2.5.1-59 and 2.5.1-21). As active rifting ceased, shallow-water shelf environments emerged along the developing passive margin and are recorded in sequences of early to middle Paleozoic marine clastic and carbonate deposits (Reference 2.5.1-21). Progressive subsidence of the Laurentian passive margin and the Iapetus basin in Late Ordovician through Early Mississippian

time is indicated by deep-water siliceous chert and shale overlying the older clastic and carbonate rocks (Reference 2.5.1-21).

The onset of collisional tectonics and closing of the Iapetus Ocean during the Carboniferous (Mississippian and Pennsylvanian time) are marked in the stratigraphic record of the Ouachita orogenic belt by the abrupt appearance of coarse clastic deposits overlying older, deep-water deposits. This east-west trending Ouachita belt formed roughly coincident with the Appalachian orogenic belt, which trends generally northeast-southwest along the length of the eastern United States and Canada (Reference 2.5.1-53) (Figure 2.5.1-8). Hatcher et al. (Reference 2.5.1-58) present a comprehensive synthesis of the Appalachian-Ouachita orogen and Viele and Thomas (Reference 2.5.1-21) present a detailed summary of the Ouachita belt development within this larger orogeny. Viele and Thomas (Reference 2.5.1-21) invoke progressive closure of the Iapetus Ocean along a subduction zone that dipped southward beneath an island arc located off the southern coast of Laurentia. Convergence between Laurentia and the island arc across the subduction zone is interpreted to be oblique, which resulted in diachronous, westward-propagating closure of the Iapetus Ocean (Reference 2.5.1-59). As the ocean basin was consumed, rocks of the accretionary prism above the subduction zone were thrust northward over deposits of the Laurentian passive margin. The passive margin strata subsequently were caught in the deformation and thrust northward as the island arc overrode the continental margin, forming the ancestral Ouachita Mountains. Mickus and Keller (Reference 2.5.1-61) developed a lithospheric-scale model of the collisional orogen east of the VCS site from a synthesis of seismic, borehole, and gravity data. The Mickus and Keller model reflected the known tectonic evolution of the area, but many new relationships between major structural blocks were identified. Specifically, the stable midcontinental Ozark uplift region has an average crustal thickness of about 27 miles (43 km). The Arkoma basin to the south shows approximately 9 miles (15 km) of deep-water sediments, possibly thickened by folding and thrusting, beneath the allochthonous Ouachita Mountains. The deep water sediments overlie oceanic or highly extended continental (transitional) crust. Further south, the Sabine block is probably a Grenville age accreted terrane and arc overlying Paleozoic rhyolites and tuff. Growth of the Ouachita Mountains loaded the Laurentian continental crust, forming a series of broad arches and basins in the foreland region north and west of the orogenic belt (References 2.5.1-62 and 2.5.1-63). Examples of these foreland structures that are relatively close to the VCS site include the Kerr Basin, Llano Uplift, and Fort Worth Basin (Reference 2.5.1-64) (Figures 2.5.1-11 and 2.5.1-12).

2.5.1.1.4.1.3 Mesozoic and Cenozoic Plate Tectonic History

The development of the Ouachita orogenic belt in late Paleozoic time marked the end of a full Wilson cycle, defined as the cycle of the opening and closing of an ocean basin through continental rifting and collision. According to Mickus and Keller (Reference 2.5.1-61), the Triassic grabens in southern Arkansas are of limited extent and depth and have limited sedimentary cover, indicating that extension in this area was small. Further south, near the Texas-Louisiana coastline, is an area of

thinned continental crust with an underlying region of anomalous low-density mantle, possibly the result of mantle upwelling and intrusion into the lower crust during rifting. Finally, the successful opening of the Gulf of Mexico formed the Sigsbee deep, with clear oceanic crust. The opening of the modern Gulf of Mexico represents the commencement of another Wilson cycle. The opening of the Gulf of Mexico began in the Late Triassic and persisted into the Late Jurassic. The opening of the Gulf was accommodated by the extension of pre-rift continental crust and the formation of new oceanic crust (References 2.5.1-23 and 2.5.1-65).

The opening of the Gulf of Mexico has been represented as occurring in two stages. The initial stage of rifting during the Late Triassic and Early Jurassic occurred along the trend of the Ouachita belt (References 2.5.1-23 and 2.5.1-65). Detailed modeling of gravity data suggests that the locus of rifting and crustal extension occurred south of the main Ouachita collisional suture, approximately beneath the present continental shelf and rise in the offshore region of the Gulf of Mexico (Reference 2.5.1-62). This extension was thought to have occurred along preexisting zones of crustal weaknesses and sutures resulting from the earlier Precambrian rifting and Late Paleozoic Ouachita orogeny (References 2.5.1-66, 2.5.1-62 and 2.5.1-47). The majority of the extension and rifting occurred during the second stage of the opening in the Middle Jurassic and was characterized by the formation of widespread rift basins that were in filled with non-marine sediments and volcanics (Reference 2.5.1-23).

These two stages of rifting affected large portions of the existing crust and created distinct variations in crustal thickness due to the variable amount of thinning experienced by the crust. Researchers commonly classify the crust within the Gulf of Mexico region as one of four types: (1) continental, (2) thick transitional, (3) thin transitional crust, or (4) oceanic (References 2.5.1-66, 2.5.1-67, and 2.5.1-47) (Figures 2.5.1-18 and 2.5.1-19). The region classified as continental crust was largely unaffected by the rifting and extension, but some fault-bounded basins associated with the rifting have been identified. Current crustal thickness estimates for the continental crust range between 22 to 25 miles (35 to 40 km) (Reference 2.5.1-47). In contrast, the thick transitional crust underwent moderate thinning, with post-extension crustal thicknesses ranging between 13 to 22 miles (20 to 35 km). This variable thinning of the thick transitional crust occurred along trends perpendicular to the margins of the Gulf of Mexico (Reference 2.5.1-47) and is proposed by some to have influenced the location of the gulf-perpendicular basement highs and lows that form the alternating arches and embayments of the Gulf coastal margin (e.g., Sabine Uplift, Houston Embayment, San Marcos Arch, and Rio Grande Embayment) (Reference 2.5.1-68) (Figures 2.5.1-11 and 2.5.1-12).

The thin transitional crust, with post-extension crustal thicknesses of 5–9 miles (8–15 km), underwent considerably more thinning than the thick transitional crust (References 2.5.1-113b and 2.5.1-47). Sawyer et al. (Reference 2.5.1-47) hypothesized that the contrast in thinning is due to the thin transitional crust having originally been weaker due to locally elevated crustal temperatures. In contrast to the thick transitional crust, the major crustal thickness variations in the thin transitional

crust are parallel to the Gulf margin (Figures 2.5.1-18 and 2.5.1-19) (Reference 2.5.1-47). Throughout the later of the two periods of rifting, significant accumulations of non-marine clastic sediment, volcanic rocks, and salt were deposited in fault-bounded basins (References 2.5.1-23, 2.5.1-65, and 2.5.1-47). In particular, thick deposits of Jurassic salt accumulated in long-lived basins along the developing rift margin.

Formation of the oceanic crust in the Gulf of Mexico occurred in the final stage of rifting and extension from the Middle to Late Jurassic (References 2.5.1-69, 2.5.1-70, 2.5.1-67, 2.5.1-23, and 2.5.1-47). The extent of oceanic crust within the Gulf of Mexico basin is limited (Figures 2.5.1-18 and 2.5.1-19) compared to the accumulative relative plate motion, reflecting the observation that over 50 percent of the relative plate motion during rifting was accommodated by crustal extension rather than by the creation of new oceanic crust (Reference 2.5.1-71). The thickness of the oceanic crust is difficult to determine due to the thick sections of overlying sediments and salt in many regions, but considerable structural variation within the oceanic crust is apparent from seismic evidence and is thought to be related to variations in the initiation, duration, and rate of spreading (Reference 2.5.1-47). The precise boundaries between the four crustal classifications within the Gulf of Mexico, and in particular the boundary between the thin-transitional continental crust and the oceanic crust, are debated within the literature (References 2.5.1-72, 2.5.1-71, 2.5.1-75, 2.5.1-73, 2.5.1-46, and 2.5.1-47). Much of the uncertainty in the boundary locations is due to the difficulty of imaging the crust through the thick sedimentary sequences and salt and the difference of opinion among experts with respect to the kinematics of the opening of the Gulf of Mexico (Reference 2.5.1-72).

The advent of plate tectonic theory provided the framework for the development of many alternative kinematic models describing the formation of the Gulf of Mexico (References 2.5.1-75, 2.5.1-23, 2.5.1-76 and references therein). However, there is a general consensus among the kinematic models proposed in the past 20–30 years that the opening of the Gulf of Mexico was dominated by the southward translation of what is referred to as the Yucatan block from the proto-Texas coast to its present position (Figures 2.5.1-20 and 2.5.1-21). The main difference between these models with respect to their implications for tectonic structures within the Texas region is their treatment of the southward translation of the Yucatan block. Hall and Najmuddin (Reference 2.5.1-75) classify Gulf kinematic models into three groups (Figure 2.5.1-20): (a) models where Yucatan rotates counterclockwise as an independent block (References 2.5.1-29, 2.5.1-71, 2.5.1-73, 2.5.1-77, 2.5.1-72, 2.5.1-79, 2.5.1-80, and 2.5.1-81) (Figure 2.5.1-20a), (b) models where Yucatan is part of the diverging South American-African plate and rotates about the pole of rotation describing the opening of the North Atlantic (References 2.5.1-69, 2.5.1-82, and 2.5.1-83) (Figure 2.5.1-20b), and (c) models where Yucatan rotates clockwise as an independent block (e.g., Reference 2.5.1-84) (Figure 2.5.1-20c).

In recent years a consensus has developed around models that include a counterclockwise rotation of the Yucatan block (Reference 2.5.1-80) (Figure 2.5.1-20a), but there are still relatively minor

differences between these models with respect to the original position of the block and the large-scale structures that accommodated the block's translation to its current position (References 2.5.1-85, 2.5.1-86, 2.5.1-79, and 2.5.1-87). James Pindell's model (References 2.5.1-77, 2.5.1-78, 2.5.1-88, 2.5.1-79, 2.5.1-87, and 2.5.1-80) is one of the more widely cited models. The main features of the Pindell model, with respect to the opening of the Gulf of Mexico, are as follows (Figure 2.5.1-21):

1. In the Lower Jurassic, before the formation of oceanic crust in the proto-central Atlantic, the Yucatan block was oriented so that it partially overlapped the modern-day coast of Texas (Figure 2.5.1-21). Further east, the proto-Gulf of Mexico was filled with the Florida block, a section of continental crust partially comprised of the southern tip of Florida (Figure 2.5.1-21).

2. By Bathonian time (Middle Jurassic, approximately 169–164 Ma), rifting from continental extension in the Gulf of Mexico caused the Yucatan block to rotate counterclockwise about an Euler pole (a pole of rotation on a sphere commonly used to describe the motion of lithospheric plates) located within the central Atlantic. Also during this time, the Florida block began to move southeastwards out of the Gulf of Mexico along the Bahamas fracture zone.

3. By the early Oxfordian (Upper Jurassic, approximately 158 Ma) the opening of the Gulf of Mexico was dominated by the creation of oceanic crust and further counterclockwise motion of the Yucatan block about a different Euler pole located between Florida and the Yucatan block (Figure 2.5.1-21). The rotation of Yucatan was accommodated on the west side of the Gulf of Mexico by motion along the East Mexican transform (Figure 2.5.1-21). This kinematic regime persisted until the formation of oceanic crust ceased and the Yucatan block was in its present position. By this time the Florida block had moved out of the Gulf of Mexico, emplacing southern Florida into its present position.

4. By the early Cretaceous (approximately 130 Ma) spreading in the Gulf of Mexico had ceased, and the Yucatan block was in its modern relative position.

A potentially important implication of this and similar kinematic models (References 2.5.1-85, 2.5.1-86, 2.5.1-79, and 2.5.1-87) is the existence of large-scale, basement involved tectonic structures (e.g., the Bahamas fracture zone and East Mexican transform) (Figure 2.5.1-21). However, there is no evidence of Quaternary activity along any of these features (References 2.5.1-48, 2.5.1-49, 2.5.1-89, and 2.5.1-90), and there is no spatial correlation between seismicity and these structures (see description in Subsection 2.5.2.3) (Figure 2.5.1-21 and 2.5.1-22).

After the relatively rapid phase of continental extension and rifting associated with the opening of the Gulf of Mexico ended, a long period of tectonic quiescence ensued during which the newly passive margin subsided and thick deposits of Late Jurassic and Cretaceous marine sediments accumulated

(References 2.5.1-66 and 2.5.1-23). Enormous volumes of sediment were deposited along the northern and northwestern margins of the ancestral Gulf of Mexico by streams draining the interior of North America, causing flexural loading of the crust and progressive southward migration of the shoreline toward the axis of the basin (Reference 2.5.1-33). The long-term migration of the shoreline is marked by bands of offlapping marine strata in the Gulf Coastal Plains that become progressively younger to the south (Figures 2.5.1-2a, 2.5.1-9, and 2.5.1-17). During the period of relative quiescence within the Gulf of Mexico region the early Tertiary Laramide orogeny was occurring along the paleo-west coast of North America. Researchers have suggested that compressional stresses generated by subduction and collision during the orogeny were transmitted to the Gulf of Mexico region and influenced the formation of the San Marcos Arch, Sabine Uplift, and intervening basins (Figure 2.5.1-12) (References 2.5.1-91,and 2.5.1-92). Deformation and thinning of the Cretaceous deposits constrain the timing of this deformation (See description in Subsection 2.5.1.1.4.3.3).

The VCS site region is located within the northwestern progradational margin of the Gulf of Mexico basin, which extends generally from the eastern edge of the Cordilleran compressional deformation near the border of Mexico and Texas eastward to the western most part of Florida and into the southwestern portion of Alabama (Reference 2.5.1-51). The northwestern progradational margin is subdivided into the interior zone and coastal zone, with the interior zone being the more landward of the two zones (Figure 2.5.1-12). The interior zone is primarily associated with broad, relatively shallow Mesozoic embayments that locally host salt diapir provinces overlying Paleozoic basement. According to Ewing (Reference 2.5.1-51), the principal structures of the interior zone are Mesozoic-age normal faults associated with opening of the Gulf of Mexico. South of the interior zone is the coastal zone, which is characterized by a very thick (6 to 9 miles, or 10 to 15 km) section of Late Mesozoic to Cenozoic strata that buries highly-extended Paleozoic crust and Mesozoic oceanic crust (Reference 2.5.1-51). The boundary between the interior and coastal zones lies along a trend of Lower Cretaceous reefs within the Gulf Coastal Plains section (Figure 2.5.1-12). The location of this reef trend is interpreted as a hinge zone reflecting the transition between thick and thin transitional crust and the greater net subsidence of the thin transitional crust due to sedimentary loading in the basin to the south (References 2.5.1-51 and 2.5.1-47).

Strata of the Gulf Coastal Plains have been deformed by the diapiric rise of salt bodies, and by growth faults, which have formed in response to sedimentary compaction, subsurface migration of salt, and down-to-the-basin slumping of the Gulf Coastal Plains section (References 2.5.1-51, 2.5.1-23, and 2.5.1-93), Stratigraphic relationships in the Gulf Coastal Plains section document salt-related deformation and growth fault activity beginning in Late Mesozoic time and continuing to recent time. Growth faults occur almost exclusively in the coastal zone (Reference 2.5.1-51) of the Gulf Coastal Plains, within the thickest section of Mesozoic to Cenozoic strata. Growth faults terminate against or sole into bodies of salt or detachment horizons within the Gulf Coastal Plains section (Reference 2.5.1-94). These structures do not penetrate the crystalline basement rocks

beneath the Gulf Coastal Plains strata, and thus are characterized as “Class B” structures by the U.S. Geological Survey (Reference 2.5.1-95); i.e., faults that “might not extend deeply enough to be a potential source of significant earthquakes” (References 2.5.1-48 and 2.5.1-49).

Broad epeirogenic uplift is interpreted to have occurred in west-central Texas and New Mexico during the Miocene, coeval with development of the Rio Grande Rift and Basin and Range province to the west (see summary description in Reference 2.5.1-51). This uplift resulted in widespread erosion of Paleocene and Upper Cretaceous strata in central Texas, and was accompanied by down-to-the-southeast flexure of the interior zone along a northeast-trending hinge line. The relatively uplifted area northwest of the hingeline is known as the Edwards Plateau, and is characterized, in general, by horizontally bedded rocks of the Cretaceous Edwards limestone. The northeast-trending Balcones and Luling fault zones are spatially associated with the hingeline and geomorphic transition from the Edwards Plateau to the interior zone of the Gulf Coastal Plains (Reference 2.5.1-96). The Balcones and Luling fault zones (Figures 2.5.1-11 and 2.5.1-12) are interpreted to extend through the Gulf Coastal Plains strata and either terminate against the upper surface of the Paleozoic basement (Reference 2.5.1-97) or continue downward into the Ouachita rocks (Reference 2.5.1-83). Major activity on the Balcones and Luling faults is interpreted to have occurred in Late Oligocene or Early Miocene time (Reference 2.5.1-98) and possibly was driven by crustal flexure and tilting along the hingeline associated with sedimentary loading of the Gulf of Mexico basin.

The long-term southward migration of the Gulf shoreline has been overprinted in late Cenozoic time with relatively minor marine regressions and transgressions associated with sea level changes during glacial and interglacial periods. Within the site vicinity, some of these glacial cycles are recorded in the deposition of the Beaumont and Lissie formations, the major Pleistocene formations (see Subsection 2.5.1.2) (Figures 2.5.1-17 and 2.5.1-23). Both formations were deposited during interglacial transgressions as facies of alluvial fan-delta systems.

2.5.1.1.4.2 Tectonic Stress

As part of the 1986 EPRI-SOG study (Reference 2.5.1-1), participating earth science teams (ESTs) evaluated the intra-plate stress of the CEUS and concluded that the tectonic stress in the CEUS is primarily characterized by northeast-southwest-directed horizontal compression. In general, the ESTs 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.

The ESTs focused on evaluating the state of stress in the mid-continent and Atlantic seaboard regions, for which stress indicator data was relatively abundant. Fewer stress indicator data was available for the Gulf of Mexico, Gulf Coastal Plains and Western Great Plains, and thus these areas received less scrutiny in the EPRI-SOG study (Reference 2.5.1-1). Notably, the Dames & Moore, Law, and Bechtel ESTs observed that the orientation of maximum horizontal compression in the Gulf Coastal Plains and west Texas may be perturbed from the regional northeast-southwest orientation that characterizes much of the CEUS.

Since 1986, an international effort to collate and evaluate stress indicator data culminated in publication of a World stress map in 1989 (References 2.5.1-99 and 2.5.1-100) that is periodically updated (Reference 2.5.1-101). Plate-scale trends in the orientations of principal stresses were assessed qualitatively based on analysis of the highest quality stress indicators (Reference 2.5.1-102), and previous delineations of regional stress provinces were refined. Statistical analyses of stress indicators confirm that the trajectory of the maximum compressive principal stress is uniform across broad continental regions at a high level of confidence (Reference 2.5.1-103). In particular, the northeast-southwest orientation of principal stress in the CEUS inferred by the EPRI ESTs is statistically robust and is consistent with the theoretical orientation of compressive forces acting on the North American plate from the Mid-Atlantic Ridge (Reference 2.5.1-99).

According to the continental United States stress map of Zoback and Zoback (Reference 2.5.1-100), the VCS site is located within the Gulf Coast stress province, which generally coincides with the belts of growth faults in the coastal regions of Texas, Louisiana, Mississippi, Alabama, and northwestern Florida (Figure 2.5.1-24). The Gulf Coast stress province is characterized by north-south-directed tensile stress (Reference 2.5.1-100) and is spatially associated with down-to-the-Gulf extension and slumping of the Gulf Coastal Plains stratigraphic section. Because these strata are deforming above sub-horizontal detachment faults and/or large bodies of Jurassic salt, gravitational tensile stress driving growth faulting is confined to the sedimentary section, and thus decoupled mechanically from the state of stress in the underlying crystalline basement.

The state of stress in the crystalline basement underlying the Gulf Coastal Plains strata is very poorly constrained by data (Reference 2.5.1-104) and may be affected by flexural loading of the lithosphere due to rapid and voluminous sedimentation in the Gulf of Mexico during the Pleistocene. Detailed numerical modeling of flexural deformation associated with sedimentary loading in the Gulf by Nunn (Reference 2.5.1-105) suggests that large, elastic bending stresses may be present in the crust and systematically vary from north-south tension in the Gulf Coastal Plains, to north-south compression in an approximately 63-mile-wide (100-km-wide) zone in the northern offshore region directly adjacent to the coast, to north-south tension at distances of greater than 63 miles (100 km) from the coast.

North of the Gulf Coastal Plains stress province, the generalized continental United States stress map of Zoback and Zoback (Reference 2.5.1-99) shows a northwest-trending boundary between two major crustal stress provinces in central Texas (Figure 2.5.1-24):

- The mid-plate stress province, which includes the CEUS and is characterized by northeast-southwest horizontal compression.
- The southern Great Plains stress province, which is characterized by northeast to north-northeast horizontal tension.

Zoback and Zoback (Reference 2.5.1-99) interpret the southern Great Plains stress province to be a transition between tensile stress and active extension in the Basin and Range to the west, and compressive stress in the tectonically stable mid-continent to the east. The boundary between the mid-plate and southern Great Plains stress provinces is shown as approximately located by Zoback and Zoback (Reference 2.5.1-99) (Figure 2.5.1-24), which reflects the paucity of stress indicator data to precisely constrain the location of the boundary. Zoback and Zoback (Reference 2.5.1-99) observed that the southern Great Plains province "generally coincides with the major topographic gradient (about 325 feet/211 miles or 99 meters/225 km) separating the thermally elevated western Cordillera from the mid-continent area." If this correlation is applicable in Texas, then the boundary between the mid-plate and southern Great Plains stress provinces probably is located near the eastern base of the mountains in west Texas, west of the VCS site.

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. Numerical models of lithospheric stress (References 2.5.1-104, 2.5.1-106, and 2.5.1-107) have generally considered the contribution to total tectonic stress in the lithosphere from three classes of forces:

- Horizontal stresses driven by buoyancy forces caused by lateral variations in lithospheric density. Richardson and Reding (Reference 2.5.1-104) emphasized 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 (Reference 2.5.1-108). The force is an integrated effect over oceanic lithosphere ranging in age from about 100 Ma to the present (Reference 2.5.1-109). The ridge-push force is transmitted as stress to the interior of continents by the elastic strength of the lithosphere.
- Shear tractions and compressive stresses applied across major plate boundaries (strike-slip faults and subduction zones) and transferred as elastic stress to plate interiors.

- Shear tractions acting on the base of the lithosphere from relative flow of the underlying asthenospheric mantle.

Richardson and Reding (Reference 2.5.1-104) concluded that the observed northeast-southwest trend of principal stress in the mid-plate stress province of the CEUS dominantly reflects ridge-push body forces associated with the Mid-Atlantic Ridge. They estimated the magnitude of these forces to be about 2 to 3 x 10^{12} N/m (i.e., the total vertically integrated force acting on a column of lithosphere 3.28 feet [1 meter] wide, which corresponds to average equivalent stresses of about 40 to 60 MPa distributed across a 30-mile-thick [48-km-thick] elastic plate).

The tensile stress regime in the southern Great Plains stress province is interpreted by Humphreys and Coblentz (Reference 2.5.1-107) to arise from positive buoyancy forces associated with the high potential energy of the elevated Cordilleran topography to the west. Essentially, the tensile stress in the western Cordillera, and in the southern Great Plains along its southeastern flank, is an on-land version of the ridge-push buoyancy force. The magnitude of the positive buoyancy force and resulting tensile stress decays eastward in the southern Great Plains stress province, coincident with the eastward decrease in topography and potential energy from the southern Rocky Mountains to the interior of the continent as noted by Zoback and Zoback (Reference 2.5.1-99).

Richardson and Reding (Reference 2.5.1-104) found that the fit of the model stress trajectories to data was improved by adding compressive stress (about 5 to 10 MPa) acting on the San Andreas fault and Caribbean plate boundary structures. The fit of the model stresses to data further indicates that shear stresses acting on these plate boundary structures must also be in the range of 5 to 10 MPa. Humphreys and Coblentz (Reference 2.5.1-107) also found that the fit of numerical stress models for the North American plate was improved by imposing large compressive stresses on the San Andreas fault and Caribbean plate boundary structures.

Richardson and Reding (Reference 2.5.1-104) noted that the general northeast-southwest orientation of principal stress in the CEUS also could be reproduced in numerical models that assume horizontal shear tractions acting on the base of the North American plate. Richardson and Reding (Reference 2.5.1-104) did not favor this as a significant contributor to total stress in the mid-continent region because their model would require an order-of-magnitude increase in the horizontal compressive stress from the eastern seaboard to the Great Plains. Using numerical models, Humphreys and Coblentz (Reference 2.5.1-107) also evaluated the contribution of shear tractions on the base of the North American lithosphere to intra-continental stress and concluded that:

- There is a viscous drag or resisting force acting on the cratonic root of North America as it moves relative to the asthenospheric mantle and that this drag supports part of the ridge-push force acting from the east and creates a stress shadow for the western United States.

- Shear tractions on the base of North America from convection of the underlying asthenospheric mantle are a minor contribution to stress in the mid-continental lithosphere. Humphreys and Coblentz (Reference 2.5.1-107) concluded that the dominant control on the northeast-southwest orientation of the maximum compressive principal stress in the CEUS is ridge-push force from the Atlantic basin.

To summarize, research on the state of stress in the continental United States since publication of the EPRI-SOG study (Reference 2.5.1-1) has confirmed observations that stress in the CEUS is characterized by relatively uniform northeast-southwest compression, and that this regional trend may be perturbed in the vicinity of the VCS site due to the influence of buoyancy forces in the uplifted Cordillera to the west and the flexure of the crust due to sedimentary loading of the Gulf of Mexico. Very little new data has been reported since the EPRI-SOG study (Reference 2.5.1-1) to better determine the orientations and relative magnitudes of the principal stresses in the VCS site region. Given that the current interpretation of the orientation of principal stress is similar to that adopted in the EPRI-SOG study (Reference 2.5.1-1), 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 Gulf Coastal Plains since development of the EPRI-SOG seismic source characterization (Reference 2.5.1-1) and there are no significant implications for existing characterizations of potential activity of tectonic structures.

2.5.1.1.4.3 Principal Tectonic Structures

The following subsections contain descriptions of specific tectonic features (Figures 2.5.1-11 and 2.5.1-12) and their evidence for activity published since the EPRI-SOG study. In summary, no new information has been published since the EPRI-SOG study on any tectonic feature within the site region that would cause a significant change in the seismic source characterizations used in the EPRI-SOG model for the region surrounding the VCS site (Reference 2.5.1-1).

Principal tectonic structures within the 200-mile (322-km) site region can be divided 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 to early Tertiary structures are related to major plate tectonic events and are mapped regionally on the basis of geological and/or geophysical data. Late Proterozoic structures include normal faults active during rifting and formation of the Iapetus Ocean passive margin. Paleozoic

structures include thrust and reverse faults active during the Ouachita orogeny. Mesozoic structures include normal faults and other structures active during formation of the Gulf of Mexico.

Tertiary and Quaternary structures within the site region are related to the tectonic environment of the Gulf of Mexico passive margin. This passive margin environment is characterized by southwest-northeast-oriented horizontal principal compressive stress (see description in Subsection 2.5.1.1.4.2), large-scale basinward slumping of the Gulf Coastal Plains section toward the basin, and vertical crustal motions. The vertical crustal motions are associated with flexural loading of the Gulf Coastal Plains and offshore sedimentary basins, and erosion and exhumation of the Great Plains (Reference 2.5.1-51).

2.5.1.1.4.3.1 Late Proterozoic Tectonic Structures

No significant Late Proterozoic structures are mapped within the 200-mile (322-km) radius of the site. The only exposures of Proterozoic rocks in the site region are in the erosional window through Mesozoic strata across the axis of the Llano Uplift (Figures 2.5.1-11, 2.5.1-12, and 2.5.1-14) (References 2.5.1-55 and 2.5.1-51). The Proterozoic rocks in the Llano Uplift show evidence for multiple phases of penetrative ductile deformation that predate late Proterozoic rifting of Laurentia and formation of the Iapetan margin (References 2.5.1-57 and 2.5.1-15). Normal faults and fault-bounded basins associated with Late Proterozoic to Early Paleozoic rifting of Laurentia are inferred from geophysical surveys to lie beneath overthrust rocks of the Late Paleozoic Ouachita orogenic belt and Mesozoic and Tertiary Gulf Coastal Plains strata (References 2.5.1-51 and 2.5.1-21), but these structures are not exposed in central Texas, and are not well documented in available geologic literature.

2.5.1.1.4.3.2 Paleozoic Tectonic Structures

The major Paleozoic tectonic structures in the 200-mile (322-km) radius of the site are associated with the Late Paleozoic Ouachita orogeny. These structures can be divided into two main groups: (1) structures of the Ouachita orogenic belt; and (2) basins and arches developed in the foreland of the Ouachita orogenic belt.

As described in Subsection 2.5.1.1.4.1, the Ouachita belt in central Texas is buried entirely by Mesozoic and Tertiary strata of the Gulf Coastal Plains; therefore, faults, folds, and other structures that developed during the Late Paleozoic Ouachita orogeny are not exposed at the surface (Reference 2.5.1-21). Based on analysis of borehole and other subsurface data from the Gulf Coastal Plains, the Ouachita belt in central Texas is divided into a 20- to 40-mile-wide (32- to 64-km-wide) frontal zone, consisting primarily of rocks of the Paleozoic passive margin sequence that were transported northward and westward along low-angle thrust faults, and an interior metamorphic belt consisting of intensely deformed fragments of accreted granitic basement overlain by sandstone and marble, all subjected to weak to low-grade regional metamorphism

(References 2.5.1-110, 2.5.1-59, and 2.5.1-21). The southern boundary of the interior metamorphic belt has not been penetrated by drill holes, but is assumed to be down structural dip to the south beneath the Coastal Plains strata (Reference 2.5.1-110). The total minimum width of the Ouachita belt in the subsurface of east Texas is about 50 miles (80 km) (Reference 2.5.1-21).

Like the better-exposed and better-studied Appalachian orogenic belt, researchers have interpreted the Ouachita orogenic belt to be underlain by a major décollement that dips basinward (south) and separates the allochthonous Ouachita rocks from the autochthonous crust of the Laurentian margin (Reference 2.5.1-52). The autochthonous rocks below the décollement probably range from a full thickness of ancestral North American continental crust beneath the northwestern part of the Ouachita belt, to transitional crust and oceanic crust farther to the south and southeast. The upper surface of the Ouachita rocks beneath the basal Mesozoic unconformity is a low-relief erosion surface that dips 1 degree or less toward the Gulf of Mexico (Reference 2.5.1-52).

The boundary between the frontal belt and interior metamorphic belt was called the “Luling front” by Flawn et al. (Reference 2.5.1-111) and interpreted as an overthrust fault (i.e., the “Luling thrust”) (Reference 2.5.1-21). Subsequent work has established that the “Luling front” probably is not a single fault, but rather a zone of distributed thrust deformation that is up to several miles wide and locally difficult to define with precision in the subsurface (Reference 2.5.1-112). Culotta et al. (Reference 2.5.1-97) interpreted a deep seismic reflection profile along the north-northwest-trending San Marcos Arch to image the Luling thrust as a folded, south-dipping structural contact between deformed autochthonous rocks of the frontal zone and accreted rocks of the interior zone. It is important to note that the “Luling front” or “Luling thrust” is a structure of the buried Ouachita belt inferred from analysis of subsurface data. This structure is distinct from the Luling fault zone, which is a Tertiary fault mapped at the surface in the northern Gulf Coastal Plains in central Texas (Subsection 2.5.1.1.4.3.4.3).

The Kerr Basin and Fort Worth Basin, located to the southwest and northeast of the Llano Uplift, respectively (Figures 2.5.1-11 and 2.5.1-12), are late Paleozoic marine basins that developed in the foreland of the Ouachita orogenic belt. These foreland basins are buried by Cretaceous and younger strata and are interpreted from subsurface data gathered during oil and gas exploration (References 2.5.1-22 and 2.5.1-64). The basins primarily formed by flexural loading of the crust as the Ouachita orogen developed structural and topographic relief. Geophysical data from other parts of the Ouachita foreland indicates that these basins typically subsided along down-to-the-south normal faults, which in some cases were overthrust by the frontal zone thrust sheets during the latter stages of the Ouachita orogeny (References 2.5.1-22 and 2.5.1-21). Although comparable late Paleozoic foreland basin faults may be present beneath the Coastal Plains section in the site region, they are not extensively documented in the geologic literature (e.g., Reference 2.5.1-113a).

2.5.1.1.4.3.3 Mesozoic Tectonic Structures

Major Mesozoic structural features in the VCS site region include:

- Faults that accommodated renewed crustal rifting in the Triassic Structures associated with seafloor spreading in the young Gulf of Mexico
- Jurassic basins that formed in the early stages of the opening of the Gulf of Mexico Structures related to the movement of Jurassic salt deposits
- Large basement-involved uplifts and arches that are hypothesized to have developed coeval with the Late Cretaceous-Early Tertiary Laramide orogeny to the west.

The initial stages of rifting related to breakup of the post-Appalachian/Ouachita orogeny supercontinent of Pangea occurred in the Late Triassic and accommodated relatively little of the overall extension and thinning that formed the modern Gulf of Mexico (Reference 2.5.1-47). This stage of rifting was characterized by the formation of grabens and half-grabens filled with non-marine sediments commonly referred to as “red beds” and rift-related volcanics (References 2.5.1-23, 2.5.1-65, and 2.5.1-47). These basins ring the modern day Gulf of Mexico, but are primarily concentrated along the western Gulf in Mexico and the north to northeastern Gulf from Texas to northern Florida (References 2.5.1-23 and 2.5.1-65).

The closest known red beds to the VCS site occur within the East Texas Basin (Figure 2.5.1-12), suggesting that this basin initially formed as part of the breakup of Pangea (Reference 2.5.1-23). After the red bed deposition, the East Texas Basin accumulated thick deposits of salt in the late Middle Jurassic, followed by a large influx of clastic deposits during the Late Jurassic and Early Cretaceous (Reference 2.5.1-51). Stratigraphic relations document that the Jurassic salt deposits along the northeast-trending axis of the basin were mobilized beginning in Late Jurassic time to form numerous diapirs by Early Cretaceous time; these structures now comprise the East Texas diapir province (Reference 2.5.1-51). Presently the East Texas Basin is bounded on the west and north by the Mexia-Talco fault system, on the east by the Late Cretaceous to early Tertiary Sabine Uplift, and on the southeast by a south-facing homocline (Reference 2.5.1-51).

As described in Subsection 2.5.1.1.4.1.3, the bulk of the rifting associated with the opening of the Gulf of Mexico occurred in Middle to Late Jurassic and was accommodated almost equally by extension of continental crust and, at a later stage, by seafloor spreading (Reference 2.5.1-71). Extension occurred both as thinning within the thin transitional crust and, to a lesser degree, within the thick transitional crust (Reference 2.5.1-47) (Figures 2.5.1-18 and 2.5.1-19). Basement block-bounding faults formed during the extensional episode have been identified within both the thick and thin transitional crust based on analyses of gravity, magnetic and seismic data (References 2.5.1-113a, 2.5.1-113b, 2.5.1-113c, 2.5.1-91, 2.5.1-113d, and 2.5.1-147). The precise

locations of these faults and details of their geometry are difficult to determine given the thick accumulations of overlying sedimentary rocks. More importantly, no seismicity within the site region has been attributed to movement on this type of basement fault (References 2.5.1-114, 2.5.1-115, and 2.5.1-116).

As described in Subsection 2.5.1.1.4.1.3, there is no consensus within the technical community regarding the kinematics of Middle and Late Jurassic oceanic crust formation within the Gulf of Mexico (Figures 2.5.1-18 and 2.5.1-19). The lack of consensus is partially due to the thick accumulations of sedimentary rocks and salt overlying the crust (References 2.5.1-67 and 2.5.1-47) that make plate reconstructions based on magnetic anomalies particularly difficult and uncertain. Individual variations not withstanding, all of the kinematic models for the opening of the Gulf of Mexico predict that large transform faults accommodated lateral variations in spreading rate and translation of large crustal blocks (References 2.5.1-69, 2.5.1-66, 2.5.1-82, 2.5.1-71, 2.5.1-84, 2.5.1-83, 2.5.1-67, 2.5.1-77, 2.5.1-78, 2.5.1-79, 2.5.1-80, and 2.5.1-81). Given the uncertainty associated with such plate reconstruction models and the lack of a consensus model, the presence of transform faults within the site region is unconfirmed. Because the potential location of such faults is limited to the oceanic crust and/or paleo-block boundaries, the closest approach of these structures to the VCS site is approximately 50 miles (80 km) (Figures 2.5.1-12, 2.5.1-18, and 2.5.1-19). No seismicity within the site region has been attributed to this type of structure (References 2.5.1-114, 2.5.1-115, and 2.5.1-116).

Mesozoic fault systems of the Gulf of Mexico region are interpreted as related to bodies of Jurassic salt at depth. These fault systems include the Mexia-Talco, Milano, Charlotte-Jourdanton, Karnes, and Mt. Enterprise-Elkhart Graben (MEEG) fault systems (Figures 2.5.1-11 and 2.5.1-12). In general, these fault systems lie updip of and sole into salt pinchouts or welds, and motion on the faults is related to the salt migration that ultimately caused the formation of the welds and pinchouts (References 2.5.1-94 and 2.5.1-117).

The Mexia-Talco fault system, which bounds the western and northern margins of the East Texas Basin, is mapped continuously from the northeastern flank of the San Marcos Arch in central Texas to the Arkansas border (Reference 2.5.1-51) (Figures 2.5.1-11 and 2.5.1-12). In detail, the fault system is divided into three segments: the Talco fault zone in northeastern Texas, the Mexia fault zone in north-central Texas, and the Milano fault zone in central Texas. The Mexia-Talco fault system is characterized by a series of asymmetric grabens ranging from 5 to 8 miles (8 to 12 km) in width that are linked by left-stepping, down-to-basin (i.e., down-to-the-south) normal faults. Upper Jurassic and Lower Cretaceous strata systematically thicken within the grabens, indicating that movement began in the Jurassic. Stratigraphic relations also demonstrate that movement continued through Mesozoic and into Paleocene to Eocene time (Reference 2.5.1-51). Major movement on the Mexia-Talco fault system primarily occurred in Late Oligocene or Early Miocene time (References 2.5.1-28 and 2.5.1-118).

Seismic reflection and borehole data indicates that the Mexia-Talco fault system is located directly updip of the pinchout of Jurassic salt in the subsurface of the East Texas Basin and that individual graben segments typically develop where salt pinchout parallels strike (References 2.5.1-51 and 2.5.1-120). A structural cross section across the Mexia fault zone by Locklin (Reference 2.5.1-119) reproduced in Ewing (Reference 2.5.1-51) shows the fault zone terminating downward at a depth of about 9000 feet to 10,000 feet (about 2.7 to 3 km) at the unconformity between the Jurassic Louann Salt at the base of the Mesozoic Gulf Coastal Plains section and the top of the Paleozoic Ouachita rocks. These relationships strongly suggest that activity of the Mexia-Talco fault system is related to movement of salt and does not involve the underlying crystalline basement (Reference 2.5.1-120).

The Charlotte-Jourdanton fault system lies along the northeastern margin of the Rio Grande Embayment (Figures 2.5.1-11 and 2.5.1-12). The fault system is interpreted by some as the southwestern continuation of the Mexia-Talco fault system (Reference 2.5.1-28) with the Karnes fault zone acting as the structural link to the Mexia-Talco fault system (Reference 2.5.1-51). Collectively, these basin-bounding structures are referred to as the peripheral graben system (Reference 2.5.1-51). Like the Mexia-Talco fault system, the Charlotte-Jourdanton fault system is comprised of a series of en-echelon, graben-forming normal faults. Stratigraphic growth relations across faults of the Charlotte-Jourdanton fault system indicate that movement began in the Jurassic and continued into the early Tertiary. The youngest documented rocks displaced by the Charlotte-Jourdanton fault system are undifferentiated strata of Paleocene-Eocene age (Reference 2.5.1-28).

The MEEG fault system is a zone of normal faults that obliquely crosses the southeastern margin of the East Texas Basin and extends eastward to the western flank of the Sabine Uplift (References 2.5.1-51 and 2.5.1-120). The MEEG fault system strikes east-northeast and extends for a total distance of about 90 miles (145 km) from south of Carthage to the Trinity River near Palestine, Texas (Figures 2.5.1-11 and 2.5.1-25). At its closest approach, the MEEG is located over 200 miles (322 km) northeast of the VCS site. Like the Mexia-Talco fault system, the MEEG is characterized by a structurally complex series of grabens that are interpreted to root in Jurassic Louann Salt, and which were primarily active in Late Jurassic-Early Cretaceous time with lesser activity through the Eocene (References 2.5.1-81 and 2.5.1-120). Postulated evidence for late Quaternary activity of the MEEG fault system is described in Subsection 2.5.1.1.4.3.5.1.

The Gulf Coastal Plains is partly characterized by a series of Mesozoic, Gulf-perpendicular, large-scale arches and basins including the Rio Grande Embayment, San Marcos Arch, Houston Embayment, East Texas Basin, and Sabine Uplift (Figures 2.5.1-11 and 2.5.1-12). The presence of these features is apparent in the relief on the base of Mesozoic sediments, with greater depths associated with basins and shallower depths associated with arches (Figure 2.5.1-26). The San Marcos Arch, which lies between the Rio Grande Embayment and East Texas Basin, is a northwest-trending, southeast-plunging antiform with an axial trace greater than 250 miles (402 km)

long that crosses the northeast-southwest structural trend of the Ouachita belt in the northwestern part of the VCS site region (Reference 2.5.1-97). The Llano Uplift is the northern reach of the San Marcos Arch (Reference 2.5.1-96). The Sabine Uplift is the general term for the conglomeration of smaller north-, northeast-, and northwest-trending, doubly plunging anticlines that extend between the coast and the Sabine Uplift (Reference 2.5.1-91). The more complicated shape of the Sabine Uplift relative to the San Marcos Arch is attributed to distortion by local flexures and salt structures (Reference 2.5.1-91).

Both the San Marcos Arch and Sabine Uplift are defined by broad, open folding of Paleozoic structures and strata, as well as Jurassic subcrop trends. Cretaceous units overlying the arches and intervening basins have gentle 0.2 degree to 1 degree dips, and the structural relief between Lower Cretaceous rocks on the arches and in the basins is on the order of 0.6 mile (1 km) (Reference 2.5.1-91). Ewing (Reference 2.5.1-51) characterized the San Marcos Arch as a "...broad area of lesser (Mesozoic) subsidence between the Rio Grande embayment [sic] and East Texas Basin..." The onset of deformation for both arches occurred in Late Cretaceous time (References 2.5.1-51 and 2.5.1-91); subsequent growth of the arches is indicated by thinning of Late Cretaceous marine strata across the axes of the structures (References 2.5.1-51 and 2.5.1-91).

The formation of the series of arches and basins along the Texas Gulf Coastal Plains was likely caused by the combination of Mesozoic rifting of Pangea and the Late Cretaceous Laramide orogeny. As described in Subsection 2.5.1.1.4.1.3 and above, thinning and rifting related to extension within the modern thick-transitional crust may have caused gulf-perpendicular trends in basement thickness that in turn allowed for variable subsidence of crustal blocks and the creation of variable amounts of accommodation space for sedimentation observed in the modern day arches and embayments (Reference 2.5.1-47). Several researchers (References 2.5.1-51, 2.5.1-91, and 2.5.1-92) have proposed that the arches are genetically related to east-west compressive stresses during the Late Cretaceous to early Tertiary Laramide orogeny that occurred further to the west and were likely related to subduction processes.

2.5.1.1.4.3.4 Tertiary Tectonic Structures

The Gulf Coastal Plains was tectonically quiescent throughout most of the Tertiary. Regional deformation during the Tertiary is primarily characterized by slow sedimentary loading near the coast and down-to-the-south flexure of the lithosphere, resulting in progressive southward migration of the Gulf shoreline. Sedimentary loading of deeply buried Jurassic salt, combined with migration of the shelf margin toward the basin, compaction of the Gulf Coastal Plains strata, and gravitational slumping toward the Gulf basin contributed to the development of diapir provinces and systems of growth faults that accommodate down-to-the-basin subsidence. Although stratigraphic relations indicate that salt migration and growth faulting began in Cretaceous time (Reference 2.5.1-51), the

evolution of these structures in the Tertiary has significantly affected patterns of deposition and geomorphic development of the Gulf Coastal Plains.

2.5.1.1.4.3.4.1 Tertiary Salt Structures

Mobilization of Jurassic salt deposits in subbasins that formed during Mesozoic time led to the development of distinct diapir provinces in the Gulf Coastal Plains region. Major concentrations of salt diapirs in the site region include the East Texas, Rio Grande, and Houston diapir provinces (Figures 2.5.1-11 and 2.5.1-12). Ewing (Reference 2.5.1-51) noted that individual diapir provinces are characterized by distinctive spatial clustering of salt bodies, as well as distinct patterns and styles of salt movement. Major zones of diapiric salt movement in the offshore Gulf of Mexico region include the Northwest Slope and Perdido provinces south of the site and the much larger Texas-Louisiana Slope province to the southeast of the site. Although initial movement of salt began in Mesozoic time, deformation continued locally on structures in Tertiary time (Reference 2.5.1-33). Within the VCS site vicinity there are no known salt domes, as reported within published geologic literature (References 2.5.1-31 and 2.5.1-28). The absence of salt domes in this region is consistent with the noted scarcity of salt deposits within the region of the San Marcos Arch (Reference 2.5.1-122). Although regional in scale Figure 2.5.1-12 demonstrates the absence of salt and suggests that there are no salt structures within the site vicinity and site area. As part of the site-specific investigation, additional effort was placed on determining whether or not there was evidence of salt diapirism within the site vicinity. This effort included an extensive literature review for any documentation of salt structures within the greater site vicinity, analysis of aerial photos and LiDAR-derived topography for any surficial evidence of shallow salt domes (e.g., circular depressions or uplifts), analysis of Geomap subsurface structural contour maps, and analysis of seismic reflection data.

The literature review indicates that the closest mapped salt dome is over 50 miles (80 km) from the site. The Geomap data show no evidence for salt diapirism within the site vicinity above the deepest horizon mapped by Geomap. This deepest horizon varies between approximately 4000 and 10,000 feet (1.2 to 3 km) depth within the site vicinity. The Geomap data are considered to be a good resource for identifying salt diapirs and other salt structures because the mapping is based on structural horizons defined within well logs from petroleum exploration or production wells. Salt is impermeable to fluid flow and thus commonly acts as a hydrocarbon trap, so salt structures are common targets of petroleum production and exploration. It is expected that any salt structure would have numerous wells surrounding it and would be well defined within the Geomap data. This pattern of wells surrounding salt structures is observed in Geomap data from other regions of the Gulf of Mexico.

Analysis of aerial photos and the LiDAR-derived topography also reveal no evidence of anomalous geomorphic features that may be related to shallow salt structures.

As described in Subsection 2.5.1.2.4.2, seismic reflection data were licensed and analyzed to investigate the geologic structure within the site area. This investigation focused on growth faults, but all geologic structures, including salt diapirs, were investigated and mapped within the cross sections developed from the seismic reflection data (see Figures 2.5.1-45 through 2.5.1-48). The seismic reflection data used to develop these cross sections extend to over 18,000 feet (5.5 km) depth and are well suited to identify salt diapirs within this depth range. The presence of salt diapirs would be apparent as a diapiric shape with incoherent reflectors that would consistently truncate continuous and discontinuous reflectors at its edges. No such relationships were observed within the reflection data, and thus it is concluded that there is no evidence of salt diapirs within the seismic reflection data.

In summary, a diverse set of data were used to evaluate the potential existence of salt diapirs within the site vicinity. No evidence of salt diapirism is apparent or is visible in these data, thus supporting the regional inference that salt diapirs are unlikely within the site and the site vicinity.

2.5.1.1.4.3.4.2 Tertiary Growth Faults

Syndepositional growth faults are generally parallel to the trend of the Gulf coastline and are clustered in distinct spatial groups by age and structural style (References 2.5.1-94 and 2.5.1-28) (Figures 2.5.1-11, 2.5.1-12, and 2.5.1-14). The locations of individual growth fault zones are thought to be related to positions of the clastic Gulf shelf margin and progressively southward sediment loading of the basin in Late Cretaceous, Tertiary and Quaternary time (References 2.5.1-51, 2.5.1-125, and 2.5.1-126). From north to south, the major growth fault systems within the site region include the Wilcox fault zone, the Yegua fault zone, the Vicksburg fault zone, and the Frio fault zone (Figures 2.5.1-11 and 2.5.1-12). The next major growth fault zone to the south in the offshore region is the Corsair or Brazos fault zone. The VCS site lies between the generalized traces of the Vicksburg and Frio fault zones as shown by many researchers (References 2.5.1-127, 2.5.1-124, 2.5.1-128, and 2.5.1-129). However, more detailed mapping shows that the site lies between two major Vicksburg growth faults (References 2.5.1-37 and 2.5.1-235), and thus the site is within the Vicksburg fault zone proper.

The common characteristic of all growth faults is that they sole into or terminate against low-angle detachment horizons within the Gulf Coastal Plains section. These detachments are variously bodies of Jurassic salt and/or shale horizons (References 2.5.1-94 and 2.5.1-117), but shale horizons dominate in south Texas (Reference 2.5.1-125). Growth faults do not extend through the Gulf Coastal Plains section into the basement. Characteristics of the major growth fault zones in the site region are summarized in greater detail, as follows.

The Wilcox fault zone is Paleocene-Eocene in age and related to the shelf-margin progradation marked by the deposition of the deltaic Wilcox Group strata in south Texas and Louisiana

(Reference 2.5.1-130). Interpretation of a deep seismic reflection profile along the San Marcos Arch suggests that the Wilcox fault zone is localized along the buried edge of a Cretaceous reef system (Reference 2.5.1-59), which marks the boundary between the Interior Zone and Coastal Zone of the Gulf Coastal Plains (Reference 2.5.1-51) (Figure 2.5.1-12). The Wilcox fault zone consists of about 5 to 10 closely spaced, moderately to steeply dipping regional faults that terminate against or are rooted in a detachment in highly pressurized Cretaceous strata at depth (Reference 2.5.1-51). Wilcox faults that cross the Houston diapir province northeast of the site are localized above pre-existing salt pillows and are deformed by salt diapirs (References 2.5.1-94 and 2.5.1-130).

The Yegua fault zone is associated with middle to late Eocene southward progradation of the clastic shelf margin (Reference 2.5.1-130), and is best expressed in the Houston Embayment east-northeast of the San Marcos Arch (Figures 2.5.1-11 and 2.5.1-12). The Yegua fault zone is characterized by a series of fault-bounded blocks that are rotating domino-style against a low-angle detachment or detachments at depth, which may in part be inherited from structures of the older Wilcox system.

The Vicksburg fault zone, also historically referred to as the Sam Fordyce-Vanderbilt fault zone, extends from northeastern Mexico along the Gulf Coastal Plains and through Houston becoming less well defined as it continues to the east (References 2.5.1-51, 2.5.1-28, and 2.5.1-129). The Vicksburg fault zone formed at the shelf margin during an Oligocene deltaic progradation in response to rapid sedimentation (References 2.5.1-37 and 2.5.1-51). The depositional environment of deposits contemporaneous to faulting appear to have been unaffected by the fault activity suggesting that both sides of the fault system were subsiding during deposition with the basinward, downthrown side subsiding at a higher rate (Reference 2.5.1-129). The contemporaneous nature of faulting and sedimentation is widely reflected in the large number of structural closures observed in Vicksburg hydrocarbon fields (Reference 2.5.1-129). Sedimentation rates apparently matched the differential subsidence as there is no evidence of an escarpment across the fault zone (Reference 2.5.1-129). The majority of the slip along Vicksburg growth faults occurred in the Oligocene and Early Miocene, and faulting had largely ceased after the deposition of the upper Frio Formation (References 2.5.1-37, 2.5.1-131, and 2.5.1-129). However, some faults have either remained active at a much lower rate or have been re-activated as evident in the faults that have extended above the Frio and have minor topographic expressions within Pleistocene units (References 2.5.1-132 and 2.5.1-133) (see description in Subsection 2.5.1.2.4.2).

The Vicksburg fault zone is characterized predominantly by down-to-the-basin, steeply dipping (40 degrees to 70 degrees) normal faults that become listric at depth and commonly terminate against or sole into bodies of salt, shale, and detachment horizons within the Texas Gulf Coastal Plains section (References 2.5.1-137, 2.5.1-94, 2.5.1-134, and 2.5.1-28). Secondary to these main faults are numerous antithetic up-to-the-basin normal faults (References 2.5.1-51 and 2.5.1-129). In contrast to the domino-style faulting of the hanging wall observed in the Yegua fault zone, the

Vicksburg faults are characterized by “escalator-style glide faults,” in which the hanging wall moves down as a relatively intact block and is continuously buried by large bodies of syntectonic sediment (Reference 2.5.1-51). Thickening of the sedimentary section and offsets observed across the Vicksburg fault zone are highly variable with the greatest amount of thickening (approximately 10 times) and largest offset (approximately 5000 feet or 1524 meters) occurring in the thick sedimentary sections of the Houston and Rio Grande Embayments (References 2.5.1-51, 2.5.1-28, and 2.5.1-129). Stratigraphic thickening and fault offset within the region of the San Marcos Arch, and thus the site vicinity (Figures 2.5.1-11 and 2.5.1-12), is not as pronounced (Reference 2.5.1-37). Also, within the region of the San Marcos Arch there is relatively little salt, so many of the growth faults are associated with shale and not salt structures (References 2.5.1-135, 2.5.1-125, 2.5.1-37, 2.5.1-128). Also associated with Vicksburg faulting is the formation of rollover anticlines that, in many cases, are productive hydrocarbon reservoirs (References 2.5.1-131, 2.5.1-28, and 2.5.1-129). The amount of down bending observed in these rollovers decreases upsection, reflecting the decrease in fault activity with time (Reference 2.5.1-129).

The Frio fault zone developed in response to Late Oligocene shelf progradation in Texas and Louisiana (Reference 2.5.1-130). The Frio fault zone is about 38 miles (61 km) wide and characterized by moderately dipping sinuous normal faults, spaced at 3- to 6-mile (4.8- to 9.7-km) intervals and rooted in a deep detachment system (Reference 2.5.1-130). The hanging walls of the major south-dipping normal faults contain roll-over anticlines (Reference 2.5.1-136) and some antithetic, north-dipping normal faults (Reference 2.5.1-51). Variations in structural style along the Frio growth fault trend are attributed to the relative influence of salt tectonism and associated structures (salt domes, salt-cored anticlines and salt-withdrawal features), shale tectonism (e.g., shale diapirs and ridges), and the depositional environment of the Frio-aged strata involved in the deformation (Reference 2.5.1-130).

The Corsair fault zone south of the site formed in response to middle Miocene shelf progradation. Like the Vicksburg fault zone, the Corsair fault zone is an escalator-style glide-fault system (Reference 2.5.1-51).

2.5.1.1.4.3.4.3 Tertiary Basement-Involved Faults

The Balcones and Luling fault zones strike northeast-southwest, subparallel to the trend of the buried Ouachita orogenic belt, and are exposed on the San Marcos Arch southeast of the Llano Uplift (Figures 2.5.1-11 and 2.5.1-12). The Balcones fault zone is dominated by down-to-southeast normal faults with maximum displacements up to 1625 feet (495 meters), and the Luling fault zone is dominated by down-to-the-northwest normal faults with a cumulative throw of 1000 to 2000 feet (305 to 610 meters) (Reference 2.5.1-51). Together, the Balcones and Luling fault zones form a 31-mile-wide (50-km-wide) graben system (Reference 2.5.1-51). Displacements on the faults

diminish to the northeast and southwest with distance from the axis of the San Marcos Arch (Reference 2.5.1-51).

Initial movement on the Balcones and Luling fault zones may have occurred in the Mesozoic, because Late Cretaceous volcanic rocks of the Balcones igneous province generally are exposed along the trend of the fault zones, and in some cases volcanic centers are aligned along the faults (Reference 2.5.1-51). Collins (Reference 2.5.1-137) stated that most of the displacement on the Balcones fault zone occurred in Late Oligocene and Early Miocene, however, he did not provide a basis for this assessment.

The downdip geometry of the Luling fault zone was imaged in a deep seismic reflection profile acquired by the Consortium for Continental Reflection Profiling (COCORP) along the axis of the San Marcos Arch. Culotta et al (Reference 2.5.1-97) interpreted the COCORP data to show the Luling fault terminating at a depth of 0.6 to 1.25 miles (1 to 2 km) against the unconformity between Cretaceous limestone and underlying Paleozoic rocks of the Ouachita orogenic belt. The Ouachita rocks beneath the Luling fault are associated with an antiformal pattern of reflectors that Culotta et al. (Reference 2.5.1-97) interpreted as an antiformal structural duplex of Ouachita-age thrust sheets.

Culotta et al. (Reference 2.5.1-97) proposed that the Tertiary Luling fault may represent localized reactivation of south-dipping Ouachita structures in response to flexure along the subsiding Gulf margin. The authors speculated that the location and magnitude of flexure, and thus the Balcones and Luling fault zones, may be controlled in part by pre-existing structures in the Ouachita orogenic belt. Ewing (Reference 2.5.1-138) suggested that extension represented by these faults may be the shallow expression of down-to-the-basin motion on reactivated south-dipping thrust faults in the Ouachita belt, which may have acted as glide planes. Alternatively, the graben formed by the Balcones and Luling fault zones may be a “keystone graben” formed along the Early Miocene hingeline that accommodated sedimentary loading and flexure of the lithosphere (Reference 2.5.1-138).

The Balcones fault zone is associated with the southeast-facing Balcones Escarpment, a prominent geomorphic feature in central Texas (Reference 2.5.1-139). Rocks exposed on the up-thrown side of the fault zone are dominantly Lower Cretaceous carbonates, which are relatively resistant to erosion, whereas strata on the downthrown side are non-resistant Upper Cretaceous chalk and mud rocks (Reference 2.5.1-140). The Balcones Escarpment is a fault-line scarp produced by differential erosion of these units.

2.5.1.1.4.3.5 Quaternary Tectonic Structures

The VCS site region is part of a tectonically stable continental margin. No capable tectonic faults were identified within the subject site region during the 1986 EPRI-SOG study (Reference 2.5.1-1), and subsequent studies have confirmed this conclusion (References 2.5.1-141, 2.5.1-142, 2.5.1-48,

2.5.1-143, 2.5.1-144, 2.5.1-145, 2.5.1-49, and 2.5.1-90). The only geologic features within the site region noted by researchers since publication of the EPRI-SOG study (Reference 2.5.1-1) with potentially arguable tectonic activity in the Quaternary are the Balcones fault zone and the Gulf Coast growth faults. The MEEG fault system, also a geologic feature with potentially arguable Quaternary tectonic activity (e.g., Reference 2.5.1-51), is located beyond the northeastern extent of the site region. However, the available evidence reviewed in Subsection 2.5.1.1.4.3.5.1 through 2.5.1.1.4.3.5.3 suggests that none of these features are capable tectonic structures.

Due to the relatively low levels of background seismicity (see Subsection 2.5.2.1) and the lack of capable of tectonic sources within the site region, tectonic structures outside of the site region may be important in determining the GMRS at the VCS site. The following subsections contain descriptions of three of the closest and most significant tectonic structures with documented Quaternary activity (the Meers fault, the Rio Grande Rift, and the New Madrid Seismic Zone) as background information for the seismic hazard assessment in Subsection 2.5.2.

2.5.1.1.4.3.5.1 Mt. Enterprise-Elkhart Graben System

It is widely accepted that the most recent activity along the MEEG fault system was likely Eocene or younger in age (References 2.5.1-51, 2.5.1-146, 2.5.1-120, and 2.5.1-28). However, one publication (Reference 2.5.1-147) that predates the EPRI-SOG study (Reference 2.5.1-1) presents several lines of evidence that suggest Quaternary motion and active creep along the MEEG:

- Three faults at the western end of the MEEG fault system in the Trinity River Valley near Palestine, Texas, displace late Quaternary (37,000-year-old) deposits overlying Eocene Claiborne strata (Reference 2.5.1-147). Maximum normal displacement of the Eocene strata on the faults at this site is 46.5 inches (118 centimeters), with maximum offset of the overlying Quaternary gravels of 26 inches (66 centimeters). Based on an estimated age of 37,000 years for the late Quaternary gravels (Reference 2.5.1-147) the implied average, late Quaternary separation rate across the fault is about 0.0008 inches/year (0.02 millimeters per year).

- Geodetic leveling data shows a relative displacement of about 5 inches (130 millimeters) across the MEEG fault system between 1920 and the mid-1950s, with a down-to-the-south displacement (Reference 2.5.1-147). If this motion is due to slip on normal faults of the MEEG, then the average vertical separation rate is 0.17 inch per year (4.3 millimeters per year), assuming a window of 30 years between leveling surveys.

- Historical and instrumentally located seismicity is reported as spatially associated with the MEEG, including: the 1891 Rusk earthquake (maximum intensity MMI VI; magnitude [unspecified scale] 4.0 and location estimated from felt effects), four earthquakes in 1957

(maximum intensity III to V; magnitudes [unspecified scale 3.0 to 4.7, and locations estimated from felt effects), and the 1981 Center (mb 3.0) and Jacksonville (mb 3.2) earthquakes (References 2.5.1-147, 2.5.1-114, and 2.5.1-116).

As described in Subsection 2.5.1.1.4.3.3, seismic reflection data suggests that the MEEG is rooted in the Jurassic Louann Salt at maximum depths of 2.9 to 3.75 miles (4.5 to 6 km) (References 2.5.1-120 and 2.5.1-148). This suggests that observed Late Quaternary displacement and contemporary creep across the MEEG is driven by movement of salt at depth, indicating that the fault does not accommodate tectonic deformation and thus is not an independent source of moderate to large earthquakes. Presumably, this was the evaluation of the EPRI-SOG ESTs, which had access to the pre-1986 literature on the MEEG and did not specifically characterize it as a Quaternary tectonic fault and potentially capable structure. However, Ewing (Reference 2.5.1-51) briefly comments in a post-EPRI publication that, “surface strata are displaced and seismicity suggests continuing deformation” on the MEEG.

Based on a review of post-EPRI scientific literature, no new data has been published to support an interpretation that the MEEG is a capable tectonic structure. Recent reviews of suspected Quaternary tectonic features in the CEUS by Crone and Wheeler (Reference 2.5.1-48) and Wheeler (Reference 2.5.1-49) did not identify or discuss the MEEG as a potential tectonic fault. The documented association of the MEEG with Jurassic salt deposits, and the high rate of active creep measured by geodetic methods support the interpretation that Quaternary activity of the MEEG is related to salt migration at depth. The 5 inches (130 millimeters) of displacement observed across the fault zone in approximately 30 years is highly anomalous for a fault located in a stable continental block and is similar to fault slip rates of about 0.16 to 0.2 inches per year (4 to 5 millimeters per year) characteristic of faults associated with active plate boundaries. There is broad consensus within the informed geoscience community that the Gulf Coastal Plains is part of stable North America and not part of an active plate boundary, so the high geodetic deformation rates, if accurate, are most simply explained by movement of salt at depth and do not reflect whole-crustal strain. In conclusion, there is no new information bearing on the Quaternary activity of the MEEG fault system requiring a revision of the EPRI seismic source characterization of the Gulf Coastal Plains region.

2.5.1.1.4.3.5.2 Balcones Fault Zone

As described in Subsection 2.5.1.1.4.3.4.3, the Balcones fault and Luling fault zones comprise an approximately northeast-southwest-trending graben system located approximately 110 miles (176 km) north of the site. Collins (Reference 2.5.1-137) interpreted the most significant displacements on the Balcones fault to have occurred in Late Oligocene-Early Miocene time. In a publication postdating the EPRI-SOG study (Reference 2.5.1-1), Collins et al. (Reference 2.5.1-149) reported that downward tapering, wedge-shaped fractures filled with weathered colluvium have been observed along individual faults of the Balcones zone. Collins et al. (Reference 2.5.1-149)

speculated that the fractures may have formed during surface-rupturing events on the associated faults, and were subsequently filled with colluvial material. Based on the degree of weathering and soil profile development in the colluvium, Collins et al. (Reference 2.5.1-149) hypothesized that the deposits are Pleistocene in age. If the wedges of colluvium do fill fractures that formed during surface-rupturing events on the Balcones fault zone, then the faults could have generated moderate to large earthquakes during the Quaternary. Collins et al. (Reference 2.5.1-149) also noted, however, that strands of the Balcones fault zone are overlain by unfaulted Quaternary terrace deposits, and that these relations suggest the fissure-fill deposits probably are not related to co-seismic faulting.

Based on a review of literature postdating the EPRI-SOG study (References 2.5.1-98, 2.5.1-48, 2.5.1-144, and 2.5.1-49), including a later publication by Collins (Reference 2.5.1-98), there is no new data or research that documents Quaternary activity of the Balcones fault zone (Reference 2.5.1-1). The colluvial relations described by Collins et al. (Reference 2.5.1-149) are equivocal evidence for late Cenozoic activity at best, and the stratigraphic relationships of unfaulted Quaternary terrace deposits overlying the Balcones fault zone are positive evidence for no Quaternary activity. Collins' current opinion is that there is no evidence to support the interpretation of the Balcones fault zone as a capable fault (Reference 2.5.1-270).In conclusion, there is no post-EPRI information on the Balcones fault zone that requires a revision of the EPRI seismic source characterization of the Coastal Plains region.

2.5.1.1.4.3.5.3 Quaternary Growth Faults

Evidence for Quaternary activity in the form of surface deformation has been documented on some growth faults in the Texas Coastal Plains (References 2.5.1-149a, 2.5.1-149b, 2.5.1-134, 2.5.1-149c, 2.5.1-149d, 2.5.1-149e, 2.5.1-149f, 2.5.1-133, and 2.5.1-149g). Most of this deformation has been attributed to the extraction of fluids and gas from underground reservoirs resulting in compaction of sediments on the down-thrown side of growth faults and thus motion along those faults. The consensus among the scientific community is that motion on these growth faults is incapable of producing damaging earthquakes. As noted by Wheeler (Reference 2.5.1-95):

> "The Gulf-margin normal faults in Texas are assigned as Class B structures because [of] their low seismicity and because they may be decoupled from underlying crust, making it unclear if they can generate significant seismic ruptures that could cause damaging ground motion."

The definition of a Class B structure, per USGS criteria (Reference 2.5.1-49), is as follows:

> "Class B: Geologic evidence demonstrates the existence of 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."

This definition is in contrast to that of Class A faults, which are defined as tectonic faults with Quaternary slip, and Class C faults are defined as having no evidence of being tectonic faults or having Quaternary slip (Reference 2.5.1-49).

The assessment of the USGS (Reference 2.5.1-95) that the Gulf of Mexico growth faults are not capable sources and that they do not extend into the crystalline basement is consistent with the results of the EPRI-SOG study (Reference 2.5.1-1) and numerous studies published since the EPRI-SOG study (References 2.5.1-141, 2.5.1-142, 2.5.1-48, 2.5.1-143, 2.5.1-144, 2.5.1-150, 2.5.1-145, 2.5.1-49, and 2.5.1-90) that have not considered the Texas Coastal Plains growth faults as seismogenic sources. The implication of these assessments is that the growth faults and the weak sedimentary material through which the faults cut are incapable of storing the elastic strain energy required to generate damaging earthquakes when subsidence of the downthrown side of the fault occurs. Instead, this subsidence drives aseismic slip on the growth fault.

In summary, no new information has been published since the EPRI-SOG study (Reference 2.5.1-1) that would require updating the characterization of growth faults in the Coastal Plain as capable faults. The potential for a contribution to seismic hazard at VCS from the growth faults is adequately captured by the EPRI-SOG model (Reference 2.5.1-151) as modified to reflect new information published since 1986 on background seismicity in the Gulf of Mexico (see description in Subsections 2.5.2.2 and 2.5.2.6.2).

2.5.1.1.4.3.5.4 Meers Fault

The Meers fault is the southern boundary of the Frontal Wichita fault system in southern Oklahoma and is over 400 miles (644 km) from the VCS site (Figure 2.5.1-22). The history of the Meers fault, like the majority of the Frontal Wichita fault system, largely reflects the history of rifting and orogenesis in southern Oklahoma. The Meers fault may have originally formed as a rift-bounding normal fault during the formation of the southern Oklahoma aulacogen (Reference 2.5.1-152). During the Permian, the Meers fault accommodated some shortening associated with closing of the Atlantic ocean, and the Ouachita orogeny that led to the formation of the Wichita Uplift (References 2.5.1-153a, 2.5.1-152, 2.5.1-154, 2.5.1-155, and 2.5.1-156). Slip on the Meers fault during this time was characterized by up-to-the-north motion on a southward dipping fault with an unknown component of left-lateral slip. Ultimately approximately 7.5 miles (12 km) of vertical offset is thought to have occurred across the Frontal Wichita fault system, and roughly 1.2 miles (1.9 km) was accommodated on the Meers fault (References 2.5.1-152, 2.5.1-154, and 2.5.1-156).

Since formation of the Wichita Uplift, the Meers fault has been reactivated at least twice: during the Late Permian and late Holocene. During the known reactivations, the sense of vertical slip on the Meers fault reversed from north-down to south-down. The change in slip during the Permian reactivation was determined from observations of sedimentary material derived from the north,

up-thrown side of the fault occurring in deposits on the south, down-thrown side of the fault References 2.5.1-153a and 2.5.1-156). The second known reactivation of the Meers fault began sometime in the Quaternary with the most recent slip in the late Holocene (References 2.5.1-157, 2.5.1-152, 2.5.1-158, 2.5.1-156, and 2.5.1-159).

The modern state of knowledge regarding the Quaternary activity of the Meers fault is primarily based on the result of four sets of studies: the studies of Ramelli and others (References 2.5.1-156 and 2.5.1-160), the studies of Madole (References 2.5.1-161 and 2.5.1-158), the study of Crone and Luza (2.5.1-157), and the studies of Swan and others (References 2.5.1-162 and 2.5.1-159). Other investigations of the Meers fault have been conducted (References 2.5.1-163, 2.5.1-164, and 2.5.1-156), but these studies do not significantly add to the modern state of knowledge of the Meers fault as a potential seismic source. A summary of the results of each of the four studies relevant to the seismic source characterization is presented in Table 2.5.1-2 and briefly reviewed below.

The most detailed and comprehensive study of the Meers fault to date was conducted by Swan and others through funding from NRC. The study results were summarized in an NRC conference proceedings (Reference 2.5.1-162) and fully reported in a draft report to NRC (Reference 2.5.1-159). Other studies were useful in constraining the length of the Holocene rupture of the Meers fault (References 2.5.1-156 and 2.5.1-160) and in providing initial estimates of the timing of Holocene and earlier events on the Meers fault (References 2.5.1-157, 2.5.1-161, and 2.5.1-158), but the studies of Swan and others (References 2.5.1-162 and 2.5.1-159) are the primary resource for constraining the timing and number of Quaternary events on the fault due to the detailed trenching and radiocarbon dating conducted in those studies. For this reason, the following briefly summarizes the results of the Swan et al. (Reference 2.5.1-159) study.

The Swan et al. study (Reference 2.5.1-159) of the Meers fault consisted of numerous trenches, soil pits, hand auger samples, surveys of offset features, and over 30 calibrated radiocarbon dates from four sites along the Meers fault: the valley site, the northwest ponded alluvium site, the southeast ponded alluvium site, and the Canyon Creek site.

At the valley site, Swan et al. (Reference 2.5.1-159) excavated one trench and four soil pits and observed stratigraphic relationships that supported the occurrence of two Holocene surface-rupturing events. Swan et al. (Reference 2.5.1-159) reported calibrated radiocarbon ages from key stratigraphic horizons within these excavations that can be used to help constrain the timing of the events. An age of 2918 years before present (BP) was determined from the youngest unit faulted in the oldest event, and two ages of 1942 and 1610 years BP were determined on alluvium from the scarp of the oldest event. These two ages were interpreted as minimum ages for the oldest event and maximum ages for the youngest event. Four ages of 1296, 1296, 777, and 777 years BP from colluvium and alluvium post dating the youngest event were interpreted as constraining the minimum age of that event. At the site Swan et al. (Reference 2.5.1-159) also measured a stratigraphic

separation of 12 ± 2 feet (3.6 ± 0.6 meters) associated with the fault. Lateral offset at the site was not as well constrained, but Swan et al. (Reference 2.5.1-159) estimated an approximate left-lateral offset of 30 ± 7 feet (9 ± 2 meters).

At the northwest ponded alluvium site Swan et al. (Reference 2.5.1-159) excavated seven trenches that document two surface-rupturing events on the Meers fault. As with the Canyon Creek site, Swan et al. (Reference 2.5.1-159) reported calibrated radiocarbon ages from key stratigraphic horizons within these excavations that can be used to help constrain the timing of the events. An age of 1484 years BP from faulted colluvium and two ages from unfaulted alluvium (1238 and 1265 years BP) were interpreted as constraining the youngest event, and an age of 1912 years BP from faulted colluvium was interpreted as a minimum age for the oldest event. A buried channel within the trench also allowed Swan et al. (Reference 2.5.1-159) to measure fault offset across the channel thalweg. Their best estimates of lateral and vertical offset are 10 ± 3.3 feet (3.1 ± 1.0 meters) of left-lateral offset and 7.9 ± 1 feet (2.4 ± 0.4 meters) of vertical offset for the combined two events.

At the southeast ponded alluvium site Swan et al. (Reference 2.5.1-159) excavated nine trenches that document two surface-rupturing events on the fault. Swan et al. (Reference 2.5.1-159) collected an extensive set of radiocarbon dates at the site that constrained the timing of the two events. Ages of 3397 and 2039 years BP from faulted alluvium and colluvium were interpreted as bounding the age of the oldest event. Younger faulted colluvium with a date of 1669 years was interpreted as a maximum age for the youngest event. Swan et al. (Reference 2.5.1-159) also estimated the minimum age of the youngest event as between 1336–648 years BP based several ages from unfaulted deposits post-dating the event. The southeast ponded alluvium site excavations also exposed channel thalwegs that Swan et al. (Reference 2.5.1-159) use to estimate fault displacement. Their reported best estimates of lateral and vertical offset from the thalwegs are 11 ± 3.3 feet (3.4 ± 1.0 meters) of left-lateral offset and 8.9 ± 3.3 feet (2.7 ± 1.0 meters) of vertical offset for the upper thalweg and 12 ± 3.3 feet (3.7 ± 1.0 meters) of left-lateral offset and 8.9 ± 2 feet (2.7 ± 0.7 meters) of vertical offset for the lower thalweg. Swan et al. (Reference 2.5.1-159) also noted that minor topographic ridge crests at the site are offset further than these thalwegs, suggesting that additional Quaternary events besides the two Holocene events are required to generate the observed ridge crest offsets.

At the Canyon Creek site Swan et al. (Reference 2.5.1-159) used terrace surveys, nine test pits, and three hand-auger boreholes to estimate the elapsed time between the two Holocene events and any previous Quaternary events. Based on the similarity in offset in the bedrock contact between the Holocene Browns Creek alluvium (17 ± 5.3 feet or 5.2 ± 1.6 meters) and the Pleistocene Porter Hill alluvium (17 ± 3.9 feet or 5.1 ± 1.2 meters), Swan et al. (Reference 2.5.1-159) concluded that no events occurred on the Meers fault since deposition of the Porter Hill alluvium except for the two Holocene events. Swan at el. (Reference 2.5.1-159) also correlated the soil development of the Porter Hill alluvium to a soil at a distant site that overlies a 560 thousand years before present

(Kilo-annum before present, or ka) ash deposit to infer that the Porter Hill alluvium was deposited around 500 to 200 ka years and to estimate the minimum time since the last pre-Holocene event on the Meers fault.

Full characterizations of the seismic potential of the Meers fault (i.e., magnitudes and recurrence rates) (References 2.5.1-157, 2.5.1-152, 2.5.1-158, 2.5.1-156, and 2.5.1-159) were not known until after the EPRI-SOG source characterizations had been completed (Reference 2.5.1-1). For that reason, the EPRI-SOG model does not include a characterization of the Meers fault that reflects the current state of knowledge. Therefore, the Quaternary activity of the Meers fault should be accounted for in the VCS site analysis to determine if the Meers fault contributes significantly to the seismic hazard at the VCS site. The seismic source characterization used in the screening study for the VCS is presented in Subsection 2.5.2.4.4.

2.5.1.1.4.3.5.5 Rio Grande Rift

The Rio Grande Rift (RGR) is a north-south-trending continental rift system that is recognized to extend from central Colorado through New Mexico, Texas, and into northern Mexico (References 2.5.1-165, 2.5.1-166, 2.5.1-167, 2.5.1-168, 2.5.1-168a, and 2.5.1-169). Research post-dating the EPRI-SOG study has documented previously unrecognized late Quaternary fault activity within the RGR (References 2.5.1-170, 2.5.1-171, 2.5.1-172, 2.5.1-173, 2.5.1-174, 2.5.1-175, 2.5.1-176, and 2.5.1-177). These studies indicate that the RGR is a zone of distinct and elevated tectonic activity relative to other regions at a similar distance from the VCS site. Based on these observations, the tectonic features of the RGR are relevant to VCS, despite the greater than 400 mile distance between the RGR and the site because the faults of the RGR are some of the closest capable tectonic features (Figure 2.5.1-22).

The RGR is commonly thought to have developed in two main stages. The first stage, from approximately 30 Ma to 20 Ma, involved basaltic volcanism and low-angle normal faulting. The second stage, from approximately 10 Ma to 3 Ma, involved more expansive basaltic volcanism and high-angle normal faulting that cut across and overprinted the earlier faulting (Reference 2.5.1-180). The precise cause of the rifting during these two phases of activity is debated, but the rifting is generally attributed to a combination of elevated lithospheric temperatures and east-west tensional stress caused by plate interactions in western North America. The elevated lithospheric temperatures and east-west tensional stress led to thinning of the lithosphere and associated faulting and volcanism (References 2.5.1-178, 2.5.1-179, and 2.5.1-180). Numerous faults within the RGR have been active during the Quaternary (References 2.5.1-170, 2.5.1-171, 2.5.1-172, 2.5.1-173, 2.5.1-174, 2.5.1-175, 2.5.1-176, and 2.5.1-177).

Presently the RGR is characterized by north-trending grabens centered on a broad topographic high, elevated heat flow, and a tensile stress regime. (References 2.5.1-181, 2.5.1-167, 2.5.1-168, and

2.5.1-180). The east-west extent of the RGR surficial expression (e.g., faults and elevated topography) occupies a narrower region than the lithospheric structure of the RGR (region of tensile stress, thinned crust, elevated mantle, gravity anomaly) (References 2.5.1-107, 2.5.1-182, 2.5.1-168, 2.5.1-183, and 2.5.1-101). This observation suggests that the state of stress and the processes driving the Quaternary seismic activity observed within the RGR also extend beyond the region of the surficial expression of the rift (e.g., Reference 2.5.1-184).

An example of this phenomena is the April 14, 1995, Alpine earthquake in west Texas, described in Subsection 2.5.2.4.3, that occurred significantly eastward of the nearest mapped RGR fault (Figure 2.5.2-10). The focal mechanism for this event is characterized by normal faulting with the minimum principal stress oriented north-northeast and the maximum horizontal stress oriented east-west (Reference 2.5.1-185). This event and others with similar focal mechanisms have been interpreted as reflecting the interaction of the topographically high RGR with relatively stable and low-lying Great Plains further east (References 2.5.1-186, 2.5.1-187, and 2.5.1-188). Essentially, the RGR region is characterized by large gradients in gravitational potential energy caused by a combination of excess topography and lateral variations in lithospheric density. These potential energy gradients create a tensile stress regime at the eastern edge of the RGR with the maximum horizontal compressive stress generally oriented east-west. These tensile stresses partially drive deformation within and well eastward of the physiographic RGR (References 2.5.1-189 and 2.5.1-107) as evident with the 1995 Alpine earthquake.

Quaternary faulting within the RGR has been reported in numerous studies that are well summarized and documented in the USGS Quaternary Fault and Fold Database of the United States (Reference 2.5.1-176). Summaries of individual faults are not presented here due to the large number of structures. However, some of these faults have been studied in enough detail to generate complete seismic source characterizations, and these faults are included in the 2002 USGS National Seismic Hazard Maps (Reference 2.5.1-144).

Because the current state of knowledge regarding the seismic potential of the RGR has evolved significantly since the EPRI-SOG study (Reference 2.5.1-1) and because the RGR is one of the closest capable tectonic sources to the VCS site, the RGR is included in a screening study for significant seismic sources. This source characterization and the screening study are presented in Subsection 2.5.2.2.

2.5.1.1.4.3.5.6 New Madrid Seismic Zone

The New Madrid Seismic Zone (NMSZ) extends from southeastern Missouri to southwestern Tennessee and is located over 500 miles (800 km) northeast of the VCS site (Figure 2.5.1-22). The NMSZ lies within the Reelfoot Rift and is defined by post-Eocene to Quaternary faulting imposed on previous older seismic activity. Quaternary activity of the NMSZ was recognized and accounted for by

the six ESTs in the EPRI-SOG study (Reference 2.5.1-1). Since the EPRI-SOG study, however, significant new research has been conducted that suggests the existing EPRI-SOG source models do not adequately characterize the seismic hazard posed by the NMSZ. The NMSZ is relevant to the VCS site because this new research requires revisions to the existing EPRI-SOG source models. In addition, the relatively low levels of seismic hazard at the VCS site may result in the NMSZ being a significant contributor to seismic hazard at the site. The source characterization of the NMSZ used for the VCS ESP application is presented in Subsection 2.5.2.4.4.

The NMSZ is approximately 125 miles (201 km) long and 25 miles (40 km) wide. Research conducted since the EPRI-SOG study has identified three distinct fault segments embedded within the seismic zone, consisting of a southern northeast-trending dextral slip fault, a middle northwest-trending reverse fault, and a northern northeast-trending dextral strike-slip fault (Reference 2.5.1-90). In the current east-northeast to west-southwest directed regional stress field, Precambrian and Late Cretaceous extensional structures of the Reelfoot Rift appear to have been reactivated as right-lateral strike-slip and reverse faults.

The NMSZ produced a series of historical, large-magnitude earthquakes between December 1811 and February 1812 (Reference 2.5.1-190). The December 16, 1811 earthquake is associated with strike-slip fault displacement along the southern part of the NMSZ. Johnston (Reference 2.5.1-191) estimated a moment magnitude of $\mathbf{M}_w$ 8.1 ± 0.31 for the December 16, 1811 event. However, Hough et al. (Reference 2.5.1-190) reevaluated the isoseismal data for the region and concluded that the December 16 event had a magnitude of M_w 7.2 to 7.3. Bakun and Hopper Reference 2.5.1-192) similarly concluded this event had a magnitude of M_w 7.2.

The February 7, 1812 New Madrid earthquake is associated with reverse fault displacement along the middle part of the NMSZ (Reference 2.5.1-193). This earthquake most likely occurred along the northwest-striking Reelfoot fault that extends approximately 43 miles (69 km) from northwestern Tennessee to southeastern Missouri. The Reelfoot fault is a northeast-dipping reverse fault. The Reelfoot fault does not extend updip to the earth's surface, but a topographic scarp has developed above the buried tip of the fault as a result of fault-propagation folding (References 2.5.1-194, 2.5.1-195, and 2.5.1-196). Johnston (Reference 2.5.1-191) estimated a magnitude of M_w 8.0 ± 0.33 for the February 7, 1812, event. However, Hough et al. (Reference 2.5.1-190) reevaluated the isoseismal data for the region and concluded that the February 7 event had a magnitude of M_w 7.4 to 7.5. More recently, Bakun and Hopper (Reference 2.5.1-192) estimated a similar magnitude of M_w 7.4.

The January 23, 1812 earthquake is associated with strike-slip fault displacement on the East Prairie fault along the northern part of the NMSZ. Johnston (Reference 2.5.1-191) estimates a magnitude of M_w 7.8 ± 0.33 for the January 23, 1812, event. Hough et al. (Reference 2.5.1-190), however, reevaluated the isoseismal data for the region and concluded that the January 23 event had a

magnitude of M_w 7.1. More recently, Bakun and Hopper ((Reference 2.5.1-192) estimated a similar magnitude of M_w 7.1. The upper-bound M_{max} values used in the EPRI-SOG studies (References 2.5.1-1 and 2.5.1-2) for the NMSZ range from mb 7.2 to 7.9, generally consistent with the revised magnitudes for the three events reviewed here.

Because there is very little surface expression of faults within the NMSZ, earthquake recurrence estimates are based largely on dates of paleoliquefaction and offset geological features. The most recent summaries of paleoseismologic data (References 2.5.1-197, 2.5.1-198, and 2.5.1-199) suggests a mean recurrence time of 500 years, which was used in the 2002 USGS model (Reference 2.5.1-144). This recurrence interval is half of the 1000-year recurrence interval used in the 1996 USGS hazard model (Reference 2.5.1-143), and an order of magnitude less than the seismicity-based recurrence estimates used in the 1986 and 1989 EPRI-SOG studies (References 2.5.1-1 and 2.5.1-2, respectively).

2.5.1.1.5 Regional Gravity and Magnetic Data

The primary sources of magnetic data reviewed for this application are those of Bankey et al. (References 2.5.1-200 and 2.5.1-201) and Keller (Reference 2.5.1-202). The data from Bankey et al. (References 2.5.1-200 and 2.5.1-201) within the VCS site area is shown in Figures 2.5.1-19 and 2.5.1-27. The primary sources of gravity data reviewed for this ESP application are: the compilation of on-land Bouguer gravity anomalies and offshore free-air gravity anomalies published by the Geological Society of America (GSA) (Reference 2.5.1-203), which are available through the National Geophysical Data Center (Reference 2.5.1-204); the on-land Bouguer gravity anomalies of Lyons et al. (Reference 2.5.1-205); the on-land Bouguer gravity anomalies of Keller (Reference 2.5.1-202); and the offshore free-air gravity anomalies of Sandwell and Smith (Reference 2.5.1-206), which are available from the Scripps Institution of Oceanography (Reference 2.5.1-207). The gravity data presented in the GSA compilation are shown on Figures 2.5.1-18, 2.5.1-26, 2.5.1-28, and 2.5.1-30. With the exception of the Sandwell and Smith (References 2.5.1-206 and 2.5.1-207) free-air gravity anomaly data, each of the magnetic and gravity anomaly datasets are primarily composed of compiled data from other sources that have been in some cases reprocessed. Details of the compilations are given within the reference for each dataset. The free-air gravity anomaly of Sandwell and Smith (References 2.5.1-206 and 2.5.1-207) is an original dataset derived from satellite measurements.

The grid spacing of the potential field datasets is generally less than 6 miles (10 km), making the data useful in identifying and assessing gravity and magnetic anomalies with wavelengths on the order of tens of miles or greater. The majority of these features within the site region are associated with three major tectonic events described in Subsection 2.5.1.1.4.1:

- Late Precambrian to Cambrian rifting that led to the break up of Laurentia and the opening of the Iapetus Ocean basin (References 2.5.1-58 and 2.5.1-21).
- The Paleozoic Ouachita orogeny that marked the close of the Iapetus Ocean basin (Reference 2.5.1-21).
- Mesozoic rifting that formed the present-day Gulf of Mexico (References 2.5.1-23 and 2.5.1-93).

Both rifting episodes and the Ouachita orogeny have contributed to creating a complicated modern-day basement structure beneath the Gulf Coastal Plains and Gulf of Mexico. The basement structure is expressed in the gravity and magnetic anomaly data as long-wavelength features subparallel to the modern coastline (Figures 2.5.1-18 and 2.5.1-19). As described in Subsections 2.5.1.1.5.1 and 2.5.1.1.5.2, there is some uncertainty in the interpretations of which basement features are causing any particular gravity and magnetic anomaly (References 2.5.1-72, 2.5.1-64, 2.5.1-94, 2.5.1-84, 2.5.1-75, 2.5.1-62, 2.5.1-74, 2.5.1-208, 2.5.1-113c, and 2.5.1-210). Part of this uncertainty is due to the difficulty in collecting seismic data within the Gulf of Mexico and the proximal regions of the Gulf Coastal Plains, where thick deposits of sediments and salt make it challenging to accurately image basement structure (Reference 2.5.1-47).

The expression of these three tectonic events within the site region as gravity and magnetic anomaly features was recognized at the time of the 1986 EPRI study (References 2.5.1-211, 2.5.1-84, 2.5.1-212, 2.5.1-74, 2.5.1-213, 2.5.1-214, and 2.5.1-214a). Higher quality gravity and magnetic anomaly datasets postdating the 1986 EPRI-SOG study have generally refined the interpretation and identification of features related to these main tectonic events, rather than imaging new basement faults that were unidentified before the 1986 EPRI-SOG study (References 2.5.1-215, 2.5.1-72, 2.5.1-75, 2.5.1-62, 2.5.1-208, 2.5.1-113c, and 2.5.1-210).

2.5.1.1.5.1 Gravity Data

Gravity anomaly data encompassing the site region is shown in Figure 2.5.1-28. The data is a compilation of on-land Bouguer gravity anomalies and offshore free-air gravity anomalies published by GSA (Reference 2.5.1-203) available through the National Geophysical Data Center (Reference 2.5.1-204). A profile of the gravity field along profiles B' to B through VCS and perpendicular to the regional structural trend is shown in Figure 2.5.1-29.

The longest wavelength variations in the on-land Bouguer gravity anomalies correlate to the thickness of the Mesozoic sediments deposited after the opening of the Gulf of Mexico. Figure 2.5.1-26 shows this correlation with the gravity anomaly data overlain by contours of basement depth (Reference 2.5.1-24), where basement is defined as the base of Mesozoic sediments (Reference 2.5.1-47). As is apparent in Figure 2.5.1-26, long-wavelength gravity lows

correlate with regions of deep basement and thick sedimentary sequences, such as in the Houston Embayment and Rio Grande Embayment, while long-wavelength gravity highs correlate to regions of shallow basement and thin sedimentary sequences, such as the San Marcos Arch and Sabine Uplift. As described in Subsection 2.5.1.1.4.1.3, these arches and embayments were probably formed due to a combination of variable crustal thinning during Jurassic rifting and east-west compression related to the Laramide orogeny (References 2.5.1-91 and 2.5.1-47). The correlation between increasing sediment thickness and declining gravity values is due to the lower density of the Mesozoic sediments relative to the pre-Mesozoic basement. Within the offshore region, the free-air gravity anomaly correlates less with basement depth (Figure 2.5.1-26) than with bathymetry (Figure 2.5.1-30) due to the fact that free-air gravity anomalies are not corrected for variable water depths.

In addition to these long-wavelength variations, nine individual features within the gravity data, referred to as gravity features A through I, are described below and shown in Figure 2.5.1-18.

Gravity Feature A

Gravity feature A refers to a prominent gravity high to the northwest of Austin and north of San Antonio. In Figure 2.5.1-18, the feature appears as a roughly 75-mile (121-km)-wide circular region of high gravity. In Figure 2.5.1-29, the feature appears as an approximate 25 milligal (mgal) gravity high on the northwest end of the profile. This feature correlates to the Llano Uplift, described in Subsections 2.5.1.1.1.2 and 2.5.1.1.4.3.1 The high gravity anomaly of this feature is likely due to the relatively high density of the Proterozoic crystalline rocks comprising the core of the uplift (References 2.5.1-57, 2.5.1-15, and 2.5.1-11).

Gravity Feature B

Gravity feature B refers to a distinct arcuate gravity low adjacent and to the south-southeast of gravity feature A that passes through San Antonio, Austin, and Waco (Figure 2.5.1-18). In Figure 2.5.1-29, gravity feature B is apparent as an approximately 50-mile (80-km)-wide gravity low adjacent to gravity feature A. The negative anomaly of gravity feature B has been attributed to the presence of low-density sediments within buried foreland basins of the Ouachita system that, in some cases, have been overridden by thrusts sheets of the Ouachita orogeny (References 2.5.1-113c and 2.5.1-214).

Gravity Feature C

Gravity feature C refers to a prominent gravity high directly south and southeast of gravity feature B (Figure 2.5.1-18). In Figure 2.5.1-29, gravity feature C appears as an approximate 25-mile (40-km) wide gravity high adjacent to the gravity low of gravity feature B. The high of gravity feature C has been attributed to a variety of sources (Reference 2.5.1-113c), but gravity modeling studies have suggested that the dominant signal is due to a major transition in basement structure from

unextended continental crust to thick transitional crust as described in Subsection 2.5.1.1.4.1.3 (Figure 2.5.1-18) (References 2.5.1-64, 2.5.1-113c, and 2.5.1-214). Locally the anomaly may be enhanced by the intrusion of igneous rocks associated with Mesozoic rifting (Reference 2.5.1-214).

Gravity Feature D

Gravity feature D refers to the broad regional increase in gravity extending south and east from gravity feature C to approximately 47 miles (76 km) seaward of the coastline (see Figure 2.5.1-18). In Figure 2.5.1-29, this feature is apparent as an approximate 60 mgal increase in gravity over 120 miles (193 km). This feature is commonly interpreted as representing regional scale thinning of the continental crust and is apparent throughout much of the Gulf Coastal Plains (References 2.5.1-62, 2.5.1-113c, 2.5.1-214, and 2.5.1-61). The thinned crust has been hypothesized to be a preserved feature of the Paleozoic continental margin, the signature of the transition between thick transitional and thin transitional crust formed during Mesozoic rifting, or a combination of the two (Reference 2.5.1-113c).

Gravity Feature E

Gravity feature E refers to the short wavelength gravity lows overprinting the regional gravity increase of gravity feature D (see Figure 2.5.1-18). In Figure 2.5.1-29, the variations are apparent as approximate 5 to 10 mgal oscillations in gravity superimposed on the regional increase. The exact cause of each variation is not well documented, but many of them are interpreted to be associated with horst and graben structures formed during Mesozoic rifting that preferentially thinned some regions and created local deposition centers for low-density sediments (Reference 2.5.1-214). Rifting-related volcanism may also contribute to some of the positive gravity anomalies (Reference 2.5.1-214).

Gravity Feature F

Gravity feature F refers to the prominent gravity high offshore and subparallel to the coastline (Figure 2.5.1-18). In Figure 2.5.1-29, the feature is apparent as an approximately 10 mgal increase in gravity at the apex of the regional gravity increase of Feature D. Gravity feature F is interpreted as an outer marginal basement high (Figure 2.5.1-18) related to a variations in crustal thickness within the thin transitional crust, with more extended crust gulfward of the high (Reference 2.5.1-84).

Gravity Feature G

Gravity feature G refers to the broad regional decrease in gravity directly south of gravity feature F (Figure 2.5.1-18). In Figure 2.5.1-29, the feature is apparent as an approximate 90 mgal decrease in gravity. Gravity feature G reflects the effect of the increasing water depth with distance from the coast in the Gulf of Mexico (Reference 2.5.1-216) on the free-air gravity anomaly (Figure 2.5.1-30). Bouguer gravity anomaly maps for the Gulf of Mexico that correct for the increasing water depth show a continuation of the regional gravity increase of gravity feature D. This increase in the Bouguer

gravity anomaly most likely indicates the continued thinning of the crust and positive relief on the Mohorovicic discontinuity (i.e., the contact between the crust and mantle lithosphere) postulated to cause gravity feature D (References 2.5.1-75, 2.5.1-74, 2.5.1-217, and 2.5.1-208).

Gravity Feature H

Gravity feature H refers to the short-wavelength gravity anomalies overprinted on the regional gravity decrease of Feature G (Figure 2.5.1-18). In Figure 2.5.1-29, this feature is apparent as an approximately 10 mgal increase in gravity. The exact source of each of these anomalies is not well documented, but they are likely due to a combination of variations in bathymetry, crustal thickness, and crustal composition created during Mesozoic extension and rifting. (References 2.5.1-215, 2.5.1-84, 2.5.1-74, and 2.5.1-214)

Gravity Feature I

Gravity feature I refers to the abrupt increase in gravity anomaly in the southeastern portion of Figure 2.5.1-18. A spatially correlated bathymetric high (Bryant and Bryant, 1990) likely causes a portion of the gravity increase. However, a postulated mafic basement high caused by a Late Jurassic mantle plume (i.e., “hot spot”) may also contribute to the amplitude of gravity feature I (Reference 2.5.1-72).

2.5.1.1.5.2 Magnetic Data

Magnetic data encompassing the site region is shown in Figure 2.5.1-27. The data is from aeromagnetic surveys reported by Bankey et al. (References 2.5.1-200 and 2.5.1-201). A profile of the magnetic field along profiles B’ to B through the VCS site and perpendicular to the regional structural trend is shown in Figure 2.5.1-29. Six major features present within the magnetic data, referred to as magnetic features A through F, are described in detail below and shown in Figure 2.5.1-19.

Magnetic Feature A

Magnetic feature A refers to the irregular pattern of distinct magnetic highs and lows in the region of the Llano Uplift (Figures 2.5.1-11 and 2.5.1-19). In Figure 2.5.1-29, this feature is apparent as approximately 600 nanotesla variations in the magnetic anomaly at the northwest end of the profile. Magnetic feature A is likely due to variations in susceptibility within the Proterozoic igneous intrusions comprising the core of the Llano Uplift (References 2.5.1-57 and 2.5.1-45). Magnetic feature A is spatially correlated with gravity feature A.

Magnetic Feature B

Magnetic feature B refers to a subtle, arcuate magnetic low adjacent and to the south-southeast of magnetic feature A (Figure 2.5.1-19). In Figure 2.5.1-29, this feature is apparent as an approximate

40 nanotesla decrease in magnetic anomaly. Magnetic feature B is spatially associated with the same foreland basins of the Ouachita system as gravity feature B and is proposed to indicate the presence of deep sedimentary basins (References 2.5.1-113a and 2.5.1-214).

Magnetic Feature C

Magnetic feature C refers to a subtle magnetic high directly south and southeast of magnetic feature B trending subparallel to magnetic feature B (Figure 2.5.1-19). In Figure 2.5.1-29, magnetic feature C is seen as an approximate 25-mile (40-km)-wide magnetic high of approximately 40 nanoteslas. This feature is spatially associated with gravity feature C. The more subdued nature of magnetic feature C relative to gravity feature C has been suggested to indicate that the source of the magnetic feature is at considerable depth (Reference 2.5.1-113c). As with gravity feature C, magnetic feature C is thought to represent a major transition in basement structure from unextended continental crust to thick transitional crust as described in Subsection 2.5.1.1.4.1.3 (Figure 2.5.1-29) (References 2.5.1-64, 2.5.1-113c, and 2.5.1-214). Locally the anomaly may be enhanced by the intrusion of igneous rocks associated with Mesozoic rifting (Reference 2.5.1-214).

Magnetic Feature D

Magnetic feature D refers to the prominent magnetic high inland of the coastline passing through Houston (Figure 2.5.1-19). In Figure 2.5.1-29, this feature is apparent as an approximately 100-mile (161-km)-wide magnetic high with an overall increase in magnetic anomaly of approximately 500 nanoteslas. This feature is referred to as the “Houston magnetic anomaly” and is hypothesized to reflect the presence of a mafic dike complex injected into the thin transitional crust during Mesozoic rifting, the presence of ultramafic rocks emplaced during the Ouachita orogeny, or a combination of both. (Reference 2.5.1-84)

Magnetic Feature E

Magnetic feature E refers to the increase in magnetic anomaly approximately 100 miles (161 km) gulf-ward of the coastline (Figure 2.5.1-19). In Figure 2.5.1-29, this feature is apparent as an approximately 100 nanotesla increase in magnetic anomaly. This feature has been attributed to a significant contrast in magnetism within the thin transitional crust (Reference 2.5.1-84) and is likely due to variations in the volume of rift-related intrusions.

Magnetic Feature F

Magnetic feature F refers to the distinct magnetic highs located in the southeast of Figure 2.5.1-19. This feature is spatially associated with gravity feature I. Magnetic feature F potentially reflects the presence of mafic rocks intruded into the crust during passage of the lithosphere over a Late Jurassic mantle plume (Reference 2.5.1-72).

2.5.1.2 Site Area Geology

2.5.1.2.1 Site Area Physiography and Geomorphology

The VCS site covers an area of approximately 11,500 acres (46.5 km^2) and is located in Victoria County, Texas, approximately 13 miles (21 km) southwest of the city of Victoria and 25 miles (40 km) northwest of Matagorda Bay. The site area is located within the Coastal Prairies subprovince of the Gulf Coastal Plains physiographic province. It is bordered by the Guadalupe River, Linn Lake, and Victoria Canal to the east and U.S. Highway 77 and Kuy Creek to the west (Figures 2.5.1-31 and 2.5.1-32).

The site vicinity geologic map (Figure 2.5.1-23) shows that the terrain consists largely of the Beaumont Formation. The Beaumont Formation sediments are predominantly a sequence of sands and clays deposited by Pleistocene streams in a deltaic environment that existed during the last eustatic sea level high as described in Subsection 2.5.1.2.3. The sands are indicative of channel lag deposits, while the clays represent deposition as natural levees and overbank deposits.

The site generally has approximately 15 feet (4.6 meters) of natural relief from its northern to southern boundary and approximately 65 feet (20 meters) of relief west to east. The northern portion of the site is at an elevation of approximately 80 feet (24 meters) NAVD 88, whereas the southeastern section is at an elevation of approximately 65 feet (20 meters) NAVD 88. The Guadalupe River flows east of the site's eastern boundary at an elevation of approximately 15 feet (4.6 meters) NAVD 88. The site map shows that the proposed approximately 4900-acre (19.8-km^2) cooling basin is the predominant feature in the site area.

Figure 2.5.1-4 shows the VCS site area geologic map. Most of the surficial sediments at the site consist of Beaumont Formation flood and ridge deposits. Examination of the site geologic map (Figure 2.5.1-5) shows that the VCS site lies exclusively on top of Beaumont Formation flood deposits consisting of silt and clay with a few isolated areas lying northeast and southwest of the reactor power block building areas consisting of construction fill at their surface.

Unnamed streams that discharge into Dry Kuy Creek drain from the northern portion of the site. Some of these are fed by stock ponds, which often are wet enough that a little drainage is present year-round.

2.5.1.2.2 Site Area Geologic History

Major tectonic events in the region surrounding the site include three compressional orogenies (Grenville, Ouachita, and Laramide) and a minimum of two major extensional events (late Proterozoic Laurentian rifting and Mesozoic rifting). Subsection 2.5.1.1.2 contains a detailed description of each of these events. The basement rock beneath the site is believed to be continental

crustal material from the Grenville orogeny overlain by more than 41,000 feet (7.8 miles or 12.5 km) of sedimentary section.

Regional subsidence of the Gulf Coast basin occurred simultaneously with inland uplift of the Rocky Mountain Cordillera throughout the Tertiary. This newly uplifted terrestrial source provided a great influx of sediment into the Gulf of Mexico basin and, for the first time, subsidence of the basin was primarily due to loading of the crust by prograding wedges of clastic sediment instead of cooling of new oceanic crust. Subsection 2.5.1.1.2.6 contains descriptions of these events and their regional effects. A result of this sediment influx was the migration of the Gulf Coast depocenter toward the south to its current location approximately 70 miles (114 km) southeast of Matagorda Bay. As a result, the VCS site is located on a thick wedge of clastic sediments with the formations that comprise it thickening toward the Gulf.

Four periods of glaciation followed by interglacial climatic episodes occurred during the Pleistocene that affected the geology of the site due to falling and rising sea levels along the Gulf Coast. During glacial periods, sea levels were lower causing the processes of valley cutting and widespread erosion to dominate the landscape. Interglacial periods were marked by higher sea levels along the Gulf Coast, which lead to the deposition of coalescing alluvial and deltaic plains by ancestral river systems. The Willis, Lissie, and Beaumont formations and the undifferentiated Deweyville terrace sediments were deposited following these interglacial episodes (Figures 2.5.1-4 and 2.5.1-23). As stated in Subsection 2.5.1.2.1, the VCS site is underlain by the Beaumont Formation (Figures 2.5.1-4 and 2.5.1-23). The Beaumont Formation, which was deposited during a short interglacial in the early Wisconsinan glacial stage of the late Pleistocene, was encountered in all VCS site characterization boreholes.

2.5.1.2.3 Site Area Stratigraphy

The VCS site is located on Cenozoic Coastal Plains sediments estimated to be approximately 20,000 feet (61 km) thick, which, in turn, rest on older sediments estimated to be approximately 21,000 feet (6.4 km) thick. The basement on which the sediments rest is believed to be continental crust (Reference 2.5.1-23). Figure 2.5.1-15 shows the strata encountered during the VCS subsurface investigations and the deeper underlying strata as described in the literature. Figure 2.5.1-15 also shows a correlation between geologic/geotechnical and hydrogeologic units described in detail in Subsection 2.4.12. These strata are described below as they occur from the ground surface to depth beneath the VCS site. Most borings drilled for the VCS ESP penetrate the Beaumont Formation. The two deeper borings (B-2174 and B-2274) may have penetrated into the underlying Lissie Formation, but this contact is very difficult to determine in the subsurface due to the similarity of the two units. The subunits identified by letters are informal, site-specific units. Boring logs are included in References 2.5.4-1, 2.5.4-2 and 2.5.4-3, which are provided in Part 5 of the ESP application.

Figure 2.5.1-4 shows the site area geologic map. Figure 2.5.1-23 shows the site vicinity geologic map.

The Texas Coastal Plains sediments are part of a thick sequence of sediments deposited in a subsiding basin. The surficial deposits at the site consist of the Pleistocene Beaumont Formation and a veneer of soil. The Beaumont Formation is a sequence of sand and clay deposited by ancestral Guadalupe River streams as a delta discharging into a sea that was at a higher elevation during that part of the Pleistocene than the present sea level. The Beaumont Formation is reported to be about 400 feet (122 meters) thick beneath the site; however, the exact thickness is unknown because the contact between the Beaumont Formation and the underlying Lissie Formation is difficult to determine due to the similar lithology of both formations. The Lissie Formation and the Beaumont Formation are the two dominant subdivisions of the Pleistocene deltaic system.

The older Lissie Formation crops out in the site vicinity as levee deposits, distributary sands, and flood basin mud with a combined thickness of roughly 200 feet (61 meters) (Reference 2.5.1-218). The formation was deposited in low energy depositional environments, resulting in clay-rich surfaces (Reference 2.5.1-219). The subaerially exposed Lissie surface is morphologically subdued and has a relatively uniform seaward dip of 4.4 to 6.6 feet per mile (0.8 to 1.3 meters per km) (Reference 2.5.1-219). The primary morphological features observed on the surface are rounded shallow depressions and pimple mounds (Reference 2.5.1-218). The distinct gradient of the Lissie Formation surface allows it to be easily distinguished from stratigraphically higher and chronologically younger units like the Beaumont Formation (Reference 2.5.1-219). The age of the Lissie Formation has been bracketed by seismic reflectors linked with faunal succession data that constrain the approximate age of the formation to between 1.4 Ma and 400 ka (Reference 2.5.1-219). Based on the projection of the subsurface reflectors, polarity characteristics, and other surface data, the surface of the Lissie Formation is estimated to be 700 ka (Reference 2.5.1-219).

In contrast to the Lissie Formation, the Beaumont Formation within the site vicinity is very heterogeneous and composed of multiple noncontiguous soil types deposited within transgressive, aggradational, and progradational environments (References 2.5.1-40 and 2.5.1-219). The Beaumont Formation is 100 to 200 feet (30 to 61 meters) thick and is composed of sand-rich zones, clay-rich zones, and barrier island beach deposits (Reference 2.5.1-218). Within the VCS site vicinity, the Beaumont Formation east of the Guadalupe River (Figures 2.5.1-6 and 2.5.1-23) is relatively homogeneous with a subtle surface morphology characterized by dry lakebeds and dunes (Reference 2.5.1-218). Within the site area west of the Guadalupe River the Beaumont Formation is characterized by coalescing low-gradient alluvial fans, inset fluvial terraces, incised river paleochannels, point bars, natural levees, backswamp deposits, and relict barrier islands/dunes (Figures 2.5.1-4 and 2.5.1-31) (References 2.5.1-219 and 2.5.1-218). The overall gradient of the Beaumont Formation (1.4 to 4.8 feet per mile or 0.3 to 0.9 meters per km) is less than that of the

Lissie Formation due to differences in sea level relative to the depositional zones at the time of deposition for the two units (References 2.5.1-240 and 2.5.1-219).

Barnes (Reference 2.5.1-218) described uncertainty in the age span of the Beaumont Formation. Part of this uncertainty exists due to discrepancies in nomenclature used to describe Pleistocene deposits with the Texas Coastal Plains (Reference 2.5.1-132). For example, Beaumont and Lissie are commonly used inconsistently as stratigraphic and morphologic descriptors. In subsurface investigations, Beaumont is commonly used to describe all Pleistocene deposits, and in surface geomorphology studies the Pleistocene is commonly divided into the older Lissie Formation and the younger Beaumont Formation, as described here (Reference 2.5.1-219).

When the Beaumont and Lissie formations are differentiated as distinct Pleistocene formations, there is uncertainty in the oldest bounding age of the Beaumont Formation due to the diversity of deposits within the formation and the scarcity of data constraining the age of the deposits (Reference 2.5.1-40). The majority of studies have estimated the Beaumont Formation to have deposits from 150 ka to 100 ka based on an association of the unit with the last interglacial highstand (References 2.5.1-218 and 2.5.1-132). However, this range of ages implies a significant depositional time gap between the Lissie and Beaumont Formations (Reference 2.5.1-219). In the Colorado River valley, approximately 50 miles (80 km) east of the VCS site, Blum and Price (Reference 2.5.1-230) and, more recently, Blum and Aslan (Reference 2.5.1-40) have accounted for this depositional time gap by documenting the presence of older paleosol deposits, up to approximately 350 ka, within the Beaumont Formation. This relatively new work suggests that deposition occurred throughout the Late Pleistocene and not just during the 100,000 to 150,000-year time span previously estimated.

Within the VCS site vicinity, disagreement among researchers as to the location of the contact between the Lissie and Beaumont Formations and has resulted in the site being mapped on both the Lissie and Beaumont formations by different authors (Figures 2.5.1-17 and 2.5.1-23) (References 2.5.1-219 and 2.5.1-218). Barnes (Reference 2.5.1-218) mapped the contact between the two units downstream of the intersection of the Guadalupe and San Antonio rivers immediately south of site area (Figures 2.5.1-6 and 2.5.1-17). This mapping has the Lissie Formation as the surficial unit at the site. In contrast, alternative mapping of the site vicinity by Winker (Reference 2.5.1-219) shows the Lissie and Beaumont contact approximately 25 miles (40 km) upstream of the confluence. Winker (Reference 2.5.1-219) based this contact on the projection of known stratigraphic horizons identified in seismic reflection data, the presence of geomorphic features characteristic of the Beaumont Formation (e.g., relict meanders), and the distinct contrast in surface gradient between the Lissie (4.4 to 6.6 feet per mile or 0.84 to 1.22 meters per km) and Beaumont Formations (1.5 to 4.8 feet per mile or 0.35 to 0.9 meters per km).

Despite the discrepancy between the two sets of maps, both Barnes (Reference 2.5.1-218) and Winker (Reference 2.5.1-219) describe the morphology of Lissie and Beaumont formations in similar

terms (subdued and relatively featureless Lissie surface, abundant relict meanders in Beaumont deposits, distinct surface gradients for each formation). The similarity in descriptions yet difference in mapping suggests that the mapping discrepancy may be due to the scale and/or detail of original mapping. Detailed original mapping of the site area for the VCS ESP application is based on field observations, topographic analysis, aerial photo interpretation, and published soil surveys identified the presence of fluvial terraces, river paleochannels, point bars, natural levees, and backswamp deposits throughout the site area and a relatively subdued surface gradient (approximately 3 feet/mile). These characteristics of deposits within the site area are consistent with the descriptions of the Beaumont Formation as provided by Barnes (Reference 2.5.1-218) and Winker (Reference 2.5.1-219) and suggest that the site is within the Beaumont Formation, as mapped by Winker (Reference 2.5.1-219), and not the Lissie, as mapped by Barnes (Reference 2.5.1-218).

As part of the VCS site characterization program, subsurface information was collected from over 230 geotechnical borings and cone penetrometer tests (CPTs). Of the 230 subsurface field testing locations, 186 are borings. Of the 186 borings drilled, 93 are located within the western portion of the power block area, also referred to as the Unit 1 area (2100- and 3100-series borings) and 93 are located within the eastern portion of the power block area, also referred to as the Unit 2 area, (2200- and 3200-series borings) (see the site boring plan on the Figure 2.5.1-33). Subsection 2.5.4 contains a more detailed description of the comprehensive geotechnical investigation employed to characterize the site subsurface.

Of the 186 borings drilled as part of the investigations for the VCS, 4 were advanced to the depth of 620 feet (189 meters) below ground surface (bgs). The remaining 182 borings ranged from 100 to 400 feet (30 to 122 meters) bgs with an average of approximately 190 feet (57 meters) bgs. This subsurface investigation obtained detailed information about the near-surface geologic structure and composition of sediments underlying the site as well as provided information regarding subsurface materials at greater depths. The two deep borings (B-2174 and B-2274) may have penetrated the entire Beaumont Formation beneath the site at about 400 feet (122 meters) bgs. Information gathered from the regional investigation coupled with information gathered in deeper borings that were drilled as part of the VCS subsurface investigations strongly indicate that the stratigraphy found under VCS is as described in the stratigraphic column presented in Figure 2.5.1-15 (References 2.5.1-261 and 2.5.1-267).

Figure 2.5.1-33 shows the location and orientation of two cross sections at the site beneath the power block area, developed from borings, CPTs, and from correlations interpreted from a suite of geophysical boring logs. Detailed boring logs are located in the geotechnical data report in References 2.5.4-1, 2.5.4-2 and 2.5.4-3, which are provided in Part 5 of the ESP application. Cross sections interpreted from site borings are shown in Figures 2.5.1-34 and 2.5.1-35. The cross sections presented in Appendix 2.5.1-A are interpreted from a suite of geophysical logs.

Cross section A-A' (Appendix 2.5.1-A) extends through the power block area and includes the two deep borings B-2174 and B-2274. Correlations based on gross lithology are tentative as the sediments were deposited by distributary streams in the ancient delta. These streams were similar to the distributary streams in the modern Mississippi Delta that flow away from the main channel and do not return to it.

Each log on the cross sections has a total of seven curves that are recorded in two passes into the boring, which commonly is filled with a clay-based drilling fluid or mud. Beginning on the left side of the log are two natural gamma logs that record the natural gamma radiation in the sediments. One gamma log is part of a suite of logs called the electric log suite; the other is collected as part of the caliper log that measures the boring diameter. The two gamma logs are virtually identical. The gamma curve recorded with the caliper log is used to verify the alignment of the caliper log on the paper. Gamma logs are used to identify lithology, with gamma counts of sands generally lower (moving to the left side) and clays generally higher (moving to the right) because clays adsorb uranium particles more readily than sand.

The other curve on the left side of the paper is the spontaneous potential or SP curve, which can be used to identify lithology, although it is not as sensitive to changes in lithology as the natural gamma curves.

On the right side of the sheet are three resistivity curves that record the resistivity of the formation at various depths away from the boring to see the effects of the fresh-water drilling fluid at different levels. These three curves track similarly but not identically and are used to identify lithology with sandy units moving the curve to the right, and clays shown by a curve moving to the left. The resistivity curves and the gamma curves move in opposite directions.

Finally, the fourth curve on the right side is the caliper log that measures the diameter of the boring. The key at the top of each log identifies each of the curves. A more detailed description of the down-hole geophysical logging is available in the geotechnical data report in References 2.5.4-1, 2.5.4-2, and 2.5.4-3 provided in ESP Application Part 5.

Similarities in the log curves from boring to boring are highlighted by the correlation lines connecting points of similarity in each log. The thickness of the sands and clays vary from boring to boring. This reflects the environment of deposition and is emphasized by the changes in the sand body located at an approximate depth of 100 feet (30 meters). In boring B-2174, the sand is present from 96 to 102 feet (29 to 31 meters). Toward the northeast, in boring B-2176, the sand has thickened to 94 to 106 feet (29 to 32 meters) and contains a thin clay from 101 to 104 feet (31 to 32 meters). One thousand feet (305 meters) further to the northeast in boring B-2274, the overall sandy interval has thickened to 94 to 110 feet (29 to 34 meters) and contains a clay stratum from 98 to 106 feet (29 to 32 meters). The clay is absent toward the northeast in B-2276, and the sand is a slightly thicker 92 to 112 feet (28

to 34 meters). Similar geometry changes can be traced across the logs of B-2302 on the west and B-2307 on the east.

Cross sections B-B' and C-C' of Appendix 2.5.1-A are the north-south cross sections through the Unit 1 and Unit 2 power block area locations, respectively.

Detailed descriptions of the lithology are found in the geotechnical data report in References 2.5.4-1, 2.5.4-2 and 2.5.4-3, which are provided in Part 5 of the ESP application. In general, the sands are light gray, well-sorted (poorly graded - SP) fine-grained units. Occasionally gravels are reported in borings in the cooling basin investigation and boring B-2324, which has more than 30 feet (9 meters) of gravel in the top of the boring. Boring B-2324 is located on the flood plain of the Guadalupe River.

Caliche (calcium carbonate) occurs in several depths (Figures 2.5.1-34 and 2.5.1-35) and since these layers represent a series of paleosoils, they are interpreted as time-stratigraphic markers across the site. In particular, the drilling in the power block area identified a caliche interval consistently at approximately 100 feet (30 meters) bgs.

The sands appear to be distributary channel sands based on interpretation of the electric log character and the Beaumont Formation dispositional environment. The clays are overbank and flood plain sediments.

2.5.1.2.4 **Site Area Structural Geology**

2.5.1.2.4.1 Basement Structure Beneath the Site Area

The site area is located within the coastal zone of the Gulf of Mexico basin (Reference 2.5.1-51) and is underlain by approximately 11 to 12 km of Mesozoic and Cenozoic strata above the crystalline basement (Reference 2.5.1-65). The basement below the VCS site is interpreted to be "thin transitional crust" between the tectonically thickened crust of the Paleozoic Ouachita orogenic belt to the northwest, which was not significantly affected by Mesozoic rifting, and Mesozoic oceanic crust beneath the deep Gulf of Mexico basin to the southeast (Reference 2.5.1-47). Although scientific literature published since the EPRI-SOG study (Reference 2.5.1-1) has improved the understanding of the crustal-scale structure along the buried rifted margin of the Gulf of Mexico, there is no new data that indicate the presence of previously unknown discrete basement faults or basement structures in the subsurface beneath the site area.

2.5.1.2.4.2 Growth Faults

The term "growth fault" refers to a geologic structure across which displacement occurs contemporaneously with sediment deposition, resulting in the thickening of sediment on the downthrown side of the fault (References 2.5.1-125 and 2.5.1-28). Within the Texas Gulf Coastal Plains, the term "growth fault" more specifically refers to the collection of normal faults that formed

contemporaneously with rapid sediment deposition during the Late Mesozoic and Cenozoic (References 2.5.1-51 and 2.5.1-95) (Figures 2.5.1-11 and 2.5.1-12) (see Subsection 2.5.1.1.4.3.4.2). The growth faults of the Texas Coastal Plains originally formed due to the gulfward creep and subsidence of these sediments (e.g., Reference 2.5.1-126), largely driven by compaction and dewatering of the sediment, salt and shale migration, large-scale slumping of the coastal plain due to lateral gradients in gravitational forces, and differential compaction caused by abrupt facies changes (References 2.5.1-51, 2.5.1-128, 2.5.1-134, 2.5.1-149c, 2.5.1-222, 2.5.1-23, 2.5.1-93, 2.5.1-223, and 2.5.1-126). These processes led to the formation of steeply dipping, listric growth faults that root into regional, bedding-parallel growth fault detachment surfaces at depth (References 2.5.1-224 and 2.5.1-117).

As described in Subsection 2.5.1.1.4.3.4.2, the VCS site is within the Vicksburg fault zone, a zone of growth faulting that extends from northeastern Mexico along the Gulf Coastal Plains and through the Houston area (Reference 2.5.1-51). In general, Vicksburg growth faults dip moderately to steeply towards the gulf (40 to 70 degrees) (Reference 2.5.1-28) and terminate against or sole into bodies of salt, shale, and detachment horizons within the Texas Gulf Coastal Plains section (References 2.5.1-94 and 2.5.1-134). However, the Vicksburg growth faults most proximal to the site overlay the San Marcos Arch, a region with relatively little salt, so many of the growth faults are associated with shale ridges, massifs, or diapirs (References 2.5.1-135, 2.5.1-125, and 2.5.1-128). Thickening of the sedimentary section and offsets observed across the Vicksburg fault zone are highly variable, with the greatest amount of thickening (approximately ten times) and largest offset (approximately 5000 feet or 1534 meters) occurring in the thick sedimentary sections of the Houston and Rio Grande embayments (References 2.5.1-51, 2.5.1-28, and 2.5.1-129). Section expansion and fault offset within the region of the San Marcos Arch, and thus within the site vicinity (Figures 2.5.1-11 and 2.5.1-12), is not as pronounced (Reference 2.5.1-37).

The processes thought to be responsible for the development of systems of growth faults like the Vicksburg fault zone are tied to major pulses of sedimentation at the paleo-continental shelf of the Gulf of Mexico (References 2.5.1-51 and 2.5.1-95). As the locus of deposition has stepped gulfward with time, so has the locus of growth fault formation and activity. Hence, the processes that led to the development of the Vicksburg growth faults within the site vicinity have ceased or are occurring at such low rates that late Cenozoic movement or surface deformation has not been attributed to these processes. However, other mechanisms have been reported as reactivating existing growth faults and causing deformation of the ground surface (References 2.5.1-132, 2.5.1-134, 2.5.1-149, 2.5.1-149d, 2.5.1-226, 2.5.1-222, 2.5.1-149e, 2.5.1-149f, 2.5.1-133, 2.5.1-149g, and 2.5.1-229).

The primary mechanism associated with modern growth fault movement and related surface deformation is the withdrawal of hydrocarbon and groundwater resources from the subsurface. Because growth faults commonly act as fluid traps due to the different lithologies found on opposite sides of the fault and their propensity to form rollover anticlines, fluids are commonly trapped within

the downthrown side of growth faults. Sedimentary compaction from the decrease in pore fluid pressure associated with fluid extraction causes differential subsidence and slip along the growth fault. Such activity has been extensively documented around the greater Houston-Galveston area due to the large amounts of groundwater and hydrocarbon withdrawal and the obvious effects on growth fault activity on infrastructure and buildings (i.e., warped roads and sidewalks, damaged houses) (References 2.5.1-134, 2.5.1-149c, 2.5.1-149d, 2.5.1-149e, 2.5.1-149f, 2.5.1-133, and 2.5.1-149g). In general, surface deformation from reactivated growth fault movement due to fluid withdrawal has been observed over much of the Gulf Coastal Plains (References 2.5.1-132, 2.5.1-134, 2.5.1-231, 2.5.1-236, 2.5.1-222, 2.5.1-133, and 2.5.1-229).

Surficial evidence of growth faults causing deformation of Quaternary deposits may be very subtle (References 2.5.1-132, 2.5.1-134, 2.5.1-149c, 2.5.1-149d, 2.5.1-226, 2.5.1-222, 2.5.1-149e, 2.5.1-149f, 2.5.1-133, 2.5.1-149g, and 2.5.1-229). The typical surface expression is broad, low-amplitude warping of the ground surface (e.g., several feet of relief occurring over several hundreds of feet perpendicular to the trend of the fault). The magnitude and wavelength of the warping varies greatly between individual growth faults and has been proposed by some to be related to the rate and age of movement, with younger and more rapidly moving growth faults having the most robust surface expression (Reference 2.5.1-132).

Historically, this type of surface deformation has been identified by using aerial photographs to delineate continuous lineations of broad warping (References 2.5.1-134, 2.5.1-149c, 2.5.1-149d, 2.5.1-149f, 2.5.1-149g, 2.5.1-229, and 2.5.1-232). Recent studies of growth faults also have used high-resolution (on the order of a centimeter) topographic data developed using light detection and ranging (LiDAR) techniques. This type of topographic data has been particularly useful in identifying the low-amplitude and long-wavelength surface deformation associated with some growth faults (see Subsection 2.5.1.2.4.2.1.4 for further description of LiDAR data) (Reference 2.5.1-149e).

The general consensus of the scientific community is that the growth faults of the Gulf Coastal Plains move aseismically and are not capable of generating strong vibratory ground motion (e.g., Reference 2.5.1-150). This interpretation is supported by: (1) the NRC in its classification of growth faults as non-capable and non-tectonic faults in RG 1.208, (2) the USGS in its classification of Gulf Coastal Plains growth faults as “Class B” structures that are not thought to be sources with the potential for generating significant earthquakes (References 2.5.1-95 and 2.5.1-49), and (3) many other studies that have concluded growth faults are not seismogenic sources, including the EPRI-SOG study (References 2.5.1-141, 2.5.1-142, 2.5.1-1, 2.5.1-143; e.g., 2.5.1-144 and 2.5.1-145). In contrast to capable faults that are driven by tectonic forces occurring within competent basement rocks, growth faults of the Gulf Coastal Plains only occur within the unconsolidated Mesozoic and younger sediments overlying the Gulf basement and move in response to gravitational forces, not plate tectonic stresses within the basement. These gravitational forces cause growth-fault-bounded blocks of sediment to slip gulfward along the faults, thus reducing the

gravitational potential energy of the entire sedimentary packages. This slip is observed as aseismic creep rather than rapid seismogenic slip (References 2.5.1-134, 2.5.1-149c, 2.5.1-149d, 2.5.1-149f, 2.5.1-149g, 2.5.1-229, and 2.5.1-232) and most likely reflects the inability of the unconsolidated sediments and growth faults to support sufficient elastic strain energy to cause earthquakes. As such, growth faults do not pose a ground motion hazard to the VCS site and only need to be considered with respect to their potential for permanent ground deformation (see Subsection 2.5.3).

The VCS site lies within the Vicksburg fault zone. The majority of slip along Vicksburg growth faults occurred in the Oligocene and Early Miocene and, based on stratigraphic thicknesses observed at depth, faulting had largely ceased after the deposition of the upper Frio Formation (References 2.5.1-37, 2.5.1-31, and 2.5.1-129). In general Vicksburg growth faults do not deform sediments stratigraphically higher than the Frio Formation (Reference 2.5.1-51), although some faults do extend above the Frio and have minor topographic expression within Pleistocene units (References 2.5.1-132 and 2.5.1-133). Given the large number of Vicksburg faults identified at depth within the site vicinity (Reference 2.5.1-123), the potential for surface deformation from reactivated Vicksburg growth faults was evaluated.

The majority of the growth faults identified within the site vicinity have been identified by the petroleum industry due to their importance as hydrocarbon traps (References 2.5.1-224, 2.5.1-233, 2.5.1-127, 2.5.1-234, and 2.5.1-235). Some of the most productive deposits within the site vicinity have traditionally been within the Frio Formation (References 2.5.1-236, 2.5.1-130, 2.5.1-128, 2.5.1-122, 2.5.1-237, and 2.5.1-234) (see description in Subsection 2.5.1.2.6), which is present at depths of approximately 2000 to 8000 feet (610 to 2438 meters) in the site vicinity and approximately 4000 feet (1219 meters) beneath the site (References 2.5.1-238, 2.5.1-239, and 2.5.1-123) (see description in Subsection 2.5.1.2.3). Currently there is considerable production and exploration for hydrocarbon plays within the site vicinity (see description in Subsection 2.5.1.2.5), generating large commercial interest in identifying subsurface structure, including growth faults. The commercial interest in growth faults has lead to a scarcity of publicly available information on the locations of growth faults within the site vicinity relative to amount of proprietary information held by the petroleum industry, and the information that is publicly available focuses on the location of growth faults within the productive Frio Formation (References 2.5.1-236, 2.5.1-233, 2.5.1-127, 2.5.1-240, 2.5.1-238, 2.5.1-239, 2.5.1-241, 2.5.1-234, and 2.5.1-235). Of the readily available public resources, only Galloway et al. (Reference 2.5.1-239) and Dodge and Posey (Reference 2.5.1-238) report any growth faults within the site area.

Given the importance of growth faults with respect to the potential for surface deformation at the VCS site and the lack of publicly available information, a multidisciplinary investigation to identify growth faults within the site vicinity and site area was undertaken. This effort consisted of:

- Compiling publicly available information regarding growth faults.

- Licensing commercial subsurface mapping that identifies growth faults within the Frio Formation.
- Analyzing aerial photographs and high-resolution, LiDAR-derived elevation data for characteristic expressions of growth faults deforming the ground surface.
- Licensing and reprocessing commercially available seismic reflection data to characterize the structure of growth faults with the potential to impact the site.
- Performing field reconnaissance to validate the remotely sensed data (e.g., aerial photos, LiDAR, seismic reflection) and document additional evidence for the presence or absence of growth faulting in the field.

The goal of this effort was to develop a general understanding of growth fault structures within the site vicinity, develop a detailed understanding of the location, depth, geometry, and activity of growth faults within the site area, and to evaluate the potential for growth faults to cause permanent ground deformation at the site.

2.5.1.2.4.2.1 Growth Fault Data Sources

Information on growth faults within the site vicinity was compiled from five classes of sources. Each of these sources is described in detail below.

2.5.1.2.4.2.1.1 Published Materials

A wide variety of published literature was reviewed to compile a list of growth faults that have been identified within the site vicinity (References 2.5.1-236, 2.5.1-224, 2.5.1-242, 2.5.1-243, 2.5.1-125, 2.5.1-127, 2.5.1-240, 2.5.1-37, 2.5.1-238, 2.5.1-243, 2.5.1-244, 2.5.1-256, 2.5.1-246, 2.5.1-128, 2.5.1-122, 2.5.1-239, 2.5.1-131, 2.5.1-241, 2.5.1-134, 2.5.1-17, 2.5.1-247, 2.5.1-235, 2.5.1-133, 2.5.1-229, 2.5.1-232, 2.5.1-248). Many of these studies either present generalized growth fault locations mapped at a scale too small to adequately characterize faults within the site vicinity, or did not map any growth faults within the site vicinity. Of the reviewed publications, only the regional cross sections of Dodge and Posey (Reference 2.5.1-238) and Galloway et al. (Reference 2.5.1-239) reported growth faults within the site vicinity that were mapped at a scale sufficient for use in this study (Figure 2.5.1-36).

Dodge and Posey (Reference 2.5.1-238) published a series of structural cross sections across the Texas Gulf Coastal Plains. The purpose of these sections was to illustrate the gross regional Tertiary stratigraphy of sandstone and shale units, but the cross sections also identified major growth faults. The cross sections were constructed through the compilation of proprietary and publicly available data. Within the cross sections, numerous growth faults are identified by offsets in stratigraphic markers, but no stratigraphic data is presented above the upper Frio Formation, preventing offsets

above this horizon from being determined. In Dodge and Posey (Reference 2.5.1-238), cross section 14-14' is located within several miles of the VCS site (Figure 2.5.1-36). Along this cross section, nine faults are identified within the greater site vicinity. These faults are designated in this text with the acronym DP and a unique number (e.g., DP-3) (Table 2.5.1-1). Each of these faults was projected to the ground surface using the dip of the fault as presented in the original Dodge and Posey cross section (Reference 2.5.1-238). Characteristics of each fault (depth of projection, dip direction, presence of surface expression) are presented in Table 2.5.1-1, and the surface projections are shown in Figure 2.5.1-36.

Galloway et al. (Reference 2.5.1-239) also published a series of cross sections through the Tertiary section of the Texas Gulf Coastal Plains. The purpose of these cross sections was to locate regional unconformities, condensed sections, and significant stratigraphic units. Galloway et al. (1994) state that part of the motivation for developing the cross sections was to update the work of Dodge and Posey (Reference 2.5.1-238). While developing the cross sections, Galloway et al. (Reference 2.5.1-239) also identified and interpreted the location of growth faults. The cross sections were primarily constructed using sequence stratigraphic techniques to correlate units between well logs. The data provided within the cross sections does not document whether fault offsets exist above the Frio-Lower Miocene boundary. Cross section B-B' in Galloway et al. (Reference 2.5.1-239) approaches to within several miles of the VCS site (Figure 2.5.1-36). The 1994 Galloway study identifies nineteen faults within the site vicinity along the B-B' cross section that are referred to in this text with the acronym G and a unique number (e.g., G-2) (Figure 2.5.1-36) (Table 2.5.1-1). Each of these faults was projected to the ground surface using the dip of the fault as presented in the original Galloway et al., cross section (Reference 2.5.1-239). Characteristics of each fault (depth of projection, dip direction, presence of surface expression) are presented in Table 2.5.1-1, and the surface projections are shown in Figure 2.5.1-36.

2.5.1.2.4.2.1.2 Geomap Company Structural Maps

The Geomap Company develops proprietary, commercially available structure contour maps of key stratigraphic horizons in the Gulf Coastal Plains derived from analysis primarily of well logs. The maps also show interpreted intersections between growth faults and these horizons. This study for the VCS ESP application used Geomap maps to identify growth faults. These maps provide the most comprehensive known and available compilation of growth faults at depth within the site vicinity. The Geomap Company maps that were used are the Upper Texas Gulf Coast map 327 and Middle Texas Gulf Coast maps 260, 262, and 263 (Reference 2.5.1-123). Each map consists of a set of two contoured stratigraphic horizons: an upper A horizon, and a lower B horizon. The precise stratigraphic marker mapped for each horizon (horizon A or B) varies within and between the individual map sets. The stratigraphic marker for the A horizon of the maps used here varies between the top of the Wilcox Formation and the top of the Frio Formation, and the B horizon varies between the lower Wilcox Formation and the middle Frio Formation (e.g., for a single A horizon map, one

region of the map may map the top of the Wilcox, and another region of the map may map the lower Frio).

The Geomap mapping of growth faults in the A and B horizons provides a discrete trace of the growth faults in three dimensions at the two different horizon depths and allows for the growth faults to be projected updip to the ground surface. Figure 2.5.1-36 shows the location of the surface projection of growth faults within the site vicinity based on the Geomap data (Reference 2.5.1-123) and Table 2.5.1-1 lists characteristics of each fault (depth of projection, dip direction, and presence of surface expression). These faults are referred to with the prefix GM and a letter designation (e.g., GM-F). The surface projections were determined by: (1) using the fault traces at both horizons to calculate the fault dip along the growth faults, and (2) using the calculated dip to project the position of the fault to sea level assuming the calculated fault dip is constant updip of the highest fault trace. All faults within the Geomap data with sufficient traces within two horizons were projected to the surface for this study. Faults mapped only within one horizon on the Geomap maps, or were not mapped in enough detail to allow projection, were not projected.

It should be noted that the faults were projected to sea level and not the ground surface. Within the site area the elevation of the ground surface ranges between several feet to tens of feet (Figures 2.5.1-31 and 2.5.1-32), and given the typical dips of growth faults from the Geomap data (on the order of 40 to 60 degrees), not projecting the faults to the ground surface contributes tens of feet of error at most to the projected position of the faults. The potential error introduced by this simplification is significantly less than the uncertainty in the projection locations (on the order of several miles) that comes from: (1) projecting the faults from the three-dimensional traces in the Geomap maps and (2) assuming the faults have a constant dip between the A horizon and the surface.

2.5.1.2.4.2.1.3 Aerial Photographs

Analysis of aerial photography has been traditionally used as a method of identifying lineaments and geomorphic features potentially related to growth fault activity (References 2.5.1-134, 2.5.1-149g, and 2.5.1-232). Stereo-paired aerial photographs from the U.S. Department of Agriculture Farm Service Agency were used for the VCS ESP application to identify potential lineations caused by surface deformation related to growth faulting. The photographs used in this effort were 1:20,000 scale photographs taken by the Farm Service Agency. Photographs of areas within approximately 10 miles (16 km) of the site were analyzed from the 1966 Victoria County flight lines, the 1969 Refugio County flight lines, and the 1964 Calhoun County flight lines. Lineaments identified within the site area are shown in Figure 2.5.1-37 and described in Subsection 2.5.1.2.4.2.3.

2.5.1.2.4.2.1.4 LiDAR-Derived Topography

As described in the introduction to Subsection 2.5.1.2.4.2.1.2, growth faults that cause surface deformation tend to have a characteristic broad and low-amplitude monoclinal surface expression. Recent studies have shown that LiDAR-derived topography is useful in identifying these features (Reference 2.5.1-149e). In 2007, the Texas Natural Resources Information System, in conjunction with the Federal Emergency Management Agency, completed the collection and processing of LiDAR topographic data for all Texas coastal counties. The data is publicly available through the Texas Natural Resources Information System, and data for Victoria, Calhoun, and Refugio Counties were used to identify growth faults with surficial expression. These three counties comprise the majority of the VCS site vicinity. LiDAR data from Goliad County, comprising the northwest portion of the site vicinity, were not collected by the Texas Natural Resources Information System and thus were not available for this effort.

The LiDAR data was collected with a spacing of 4.6 feet (1.4 meters). The resultant elevation model developed by the Texas Natural Resources Information System has a horizontal and vertical accuracy of 2.4 feet (0.73 meter) and 1.2 feet (0.37 meter), respectively (References 2.5.1-249 and 2.5.1-250). Typically the vertical resolution of LiDAR data is significantly finer than the vertical accuracy (i.e., relative differences in elevation less than 1.2 feet (0.37 meter) can be detected, but the absolute position of the surfaces may be incorrect by up to 1.2 feet). Field inspection of the LiDAR data along gently inclined paved roads within the VCS site region suggests that the resolution is on the order of centimeters. The LiDAR-derived topography was used to generate shaded relief images of the site vicinity with illumination from the northwest (Figures 2.5.1-38, 2.5.1-39, 2.5.1-40, 2.5.1-41, 2.5.1-42, and 2.5.1-43). These images then were used to identify lineaments that were classified as either potentially related to growth fault induced deformation or other geomorphic processes (see description in Subsection 2.5.1.2.4.2.2) (Figures 2.5.1-42 and 2.5.1-44).

2.5.1.2.4.2.1.5 Seismic Reflection Data

The subsurface Geomap data provides a detailed compilation of growth faults at depth within the site vicinity, and the combination of aerial photographs and LiDAR data provide a complimentary set of resources to identify and map topographic features that are potentially related to growth faults. However, assessing the relationship between lineaments observed at the surface to growth faults at depth, and documenting the presence or absence of shallow growth faults without surface expression, is difficult without more detailed subsurface information over the depth range between the deep growth faults and the surface. Exelon licensed four proprietary seismic reflection lines within the greater site area to help elucidate the structure of growth faults within the site area. This proprietary seismic reflection information is provided in Part 6 of the ESP application.

Exelon licensed an array of four two-dimensional reflection profiles from Seismic Exchange, Inc. (SEI), a geophysical data broker based in Houston, Texas that primarily serves the oil and gas

industry. Each profile was acquired by one of four different exploration companies in the late 1970s and early 1980s for petroleum exploration. Subsequently, the legal rights to license the data for use came under the control of SEI. The license agreement between Exelon and SEI restricts Exelon from releasing details of the profiles into the public domain (e.g., precise line locations, shot point locations, raw or processed seismic data). Therefore, only derivative products of the profiles (i.e., interpreted line drawings) are presented in this subsection.

The four profiles licensed from SEI were chosen based on the quality and coverage of available data that would best elucidate the structure of growth faults within the site area. Particular attention was given to obtaining and processing profiles that would image any growth faults in the shallow subsurface (1000 to 2000 feet or 305 to 610 meters deep) near the site (Figure 2.5.1-41). The four profiles included:

- Two “dip lines” oriented northwest-southeast (lines TGS and GDI), which cross the subsurface trend of the growth faults identified by Geomap (Reference 2.5.1-123) at a high angle and thus are best oriented to image the fault geometry. Line GDI extends through the proposed VCS site, and line TGS parallels line GDI to the northeast. GDI is the closest line to the site with a closest approach of 275 feet (84 meters) perpendicular and to the west of the center of the Unit 1 containment. The closest approach of TGS to the site is 6514 feet (1985 meters) perpendicular and to the east of the centerpoint of the eastern portion of the power block area.

- One north-south oriented oblique line (line GSI) that crosses the growth faults identified by Geomap (Reference 2.5.1-123) at an oblique angle. Line GSI is west of the site with a closest approach of 4250 feet (1295 meters) perpendicular to the centerpoint of the western portion of the power block area.

- One “strike line” oriented northeast-southwest (Line PLJ) that is oriented parallel to the strike of the growth faults identified by Geomap (Reference 2.5.1-123) and roughly perpendicular to the dip lines. Line PLJ crosses the other three survey lines south of the site and provides a tie among the lines for correlating geologic structures and stratigraphic marker horizons.

Figure 2.5.1-41 shows the area that envelopes the extent of the four profiles. Survey lines GDI, GSI, and PLJ all have a 6-second record length (corresponding to a maximum imaging depth of approximately 18,000 feet or 5486 meters), and line TGS has a 10-second record length (corresponding to a maximum imaging depth of about 30,000 feet or 9144 meters). The approximate line lengths are 10 miles (16 km), 14 miles (23 km), 14 miles (23 km), and 20 miles (32 km) for survey lines PLJ, GDI, GSI, and TGS, respectively.

Given that the four seismic profiles were originally acquired by different firms and at different times, the acquisition methods, data parameters, and processing of the final sections varied among survey lines. For the VCS ESP application, the raw data for each profile was reprocessed by Excel Geophysical Services, Inc. (EGS) of Greenwood Village, Colorado. SEI provided the raw seismic reflection data in a standard format, along with scanned copies of observers' notes and survey notes to EGS for reprocessing. EGS performed industry standard processing of the data using ProMAX software (version 2003.12.1; distributed by Halliburton Company) and Green Mountain Geophysical (GMG) software (version 33; distributed by ION Geophysical Corporation) to convert the raw seismic reflection data, consisting of arrival times of reflected energy, to 2-D cross sections that display subsurface reflectors in both time and depth. Two time sections and a depth section were developed for each of the four seismic reflection profiles. EGS provided detailed ProMAX processing flows that document the processing steps and parameters used for each of the profiles.

The first stage of the processing sequence was integrating information about line geometry from the observer's notes and survey notes using the GMG software to input and perform interactive checks of data acquisition geometries. The seismic reflection waveform data then was imported into the ProMAX software and deconvolved using spectral whitening. Interactive consistency checks of geometry information were performed in ProMAX by visually inspecting each shot gather from all the profiles to detect undocumented source location shifts and to calculate appropriate corrections when such shifts were found. The GMG software was used to pick first arrival times and calculate refraction statics and refractor velocities. The refractor depths estimated for lines PLJ and TGS corresponded to the shot hole depths, so the uphole shot times were used to calculate the weathering zone statics for these lines. First break picks from the GDI and PLJ lines produced more variable refraction velocity estimates than the other two lines, so elevations statics were used for the lines GDI and PLJ.

The normal-moveout corrections derived from initial statics and velocity models were applied to obtain supergather stacks for interactive velocity analyses (IVA) using the ProMAX software. Surface-consistent residual statics were calculated after each iteration of IVA and updated stacking velocities obtained from IVA were used to start new iterations of IVA until velocities converged to consistent values. Two independent analysts within EGS checked the IVA estimates of stacking velocities to ensure that consistent estimated velocities were obtained from this analysis.

The final velocity models were used along with final residual statics in the ProMAX software to calculate final stacked time-domain sections for each profile. Filtering and frequency-distance (FX) deconvolution were used to improve signal-to-noise ratio for the profiles. Kirchoff time and depth migrations were used with the final stacked time-domain sections in the ProMAX software to obtain time and depth migrations for each profile. Smoothed interval velocity models derived from the stacking velocities were used to develop the depth migrations. The end-product of the EGS processing included a series of reflection profiles for each line that consisted of three profiles in the time domain (unfiltered data, migrated data, and final stacked data) and one depth migration. The

depth migration of the lines was truncated at approximately 4.5 seconds in an attempt to prevent uncertainty in the deeper velocity structure from distorting the geometry of structures and their relationships to stratigraphic units.

These profiles were obtained from EGS in standard SEG-Y format and were analyzed and interpreted using Seismic Micro-Technology, Inc.'s Kingdom Suite software (version 8.2) to display and interpret the data. All of the profiles for each line were analyzed and interpreted to check for consistency, but final interpretations were made using the final stacked time-domain profiles and the depth migrated profiles (see Subsection 2.5.1.2.4.2.3.1 for details).

2.5.1.2.4.2.2 Site Vicinity Growth Faults

Growth faults in the site vicinity were identified from publications indicating the presence of growth faults in the subsurface (References 2.5.1-238 and 2.5.1-239), the subsurface mapping of the Geomap Company (Reference 2.5.1-123), and the 2007 and 2008 LiDAR data from the Texas Natural Resources Information Systems (TNRIS) (References 2.5.1-249 and 2.5.1-250, respectively), all of which are described in Subsection 2.5.1.2.4.2.1. The compilation of growth faults identified within the subsurface data is shown in Figure 2.5.1-36 and Table 2.5.1-1. The compilation figure shows estimated positions of the surface projections of the faults developed as described in Subsection 2.5.1.2.4.2.1. It is important to emphasize that the surface projections were derived from data at depths between approximately 2000 and 7000 feet (610 to 2134 meters) (Table 2.5.1-1), and have estimated surface location uncertainties on the order of several miles.

The projections of the growth faults identified within the Geomap data (Reference 2.5.1-123) in the site vicinity strike northeast-southwest, subparallel to the Gulf of Mexico coastline. The faults generally are spaced about 2 to 20 miles (3.2 to 32 km) apart, exhibit braided and branching patterns, and cluster in three main groups based on their geographic position relative to the site (Figure 2.5.1-36):

- A group of several faults in the southeast part of the site vicinity that pass through and skirt the northern shore of San Antonio Bay (faults GM-AE, GM-AD, GM-AH, GM-AG).
- A group of over 20 individual faults in the central part of the site vicinity, bounded by fault GM-A on the southeast and fault GM-T to the northwest. The three longest faults in this group, with continuous mapped traces of 40 miles (64 km) in length or more, include faults GM-T, GM-D, and GM-L. Fault GM-D is the closest growth fault to the site.
- A group of five faults in the northwest part of the site vicinity, bounded by fault GM-Z to the southeast and fault GM-AF to the northwest.

The growth fault projections derived from the Dodge and Posey (Reference 2.5.1-238) and Galloway et al. (Reference 2.5.1-239) cross sections are mapped as single points and show a similar clustering as the Geomap (Reference 2.5.1-123) data. However, it is difficult to correlate faults from the cross sections to faults identified in the Geomap (Reference 2.5.1-123) data. This apparent lack of correlation is primarily due to the uncertainty in the two sets of projections and their derivation from unique datasets. However, the number of growth faults identified proximal to the site is largest in the Geomap (Reference 2.5.1-123) dataset, which is the most recent and presumably most comprehensive dataset, suggesting that it is unlikely the Geomap (Reference 2.5.1-123) dataset lacks any growth faults identified in either of the two other cross sections.

Analysis of LiDAR-derived topographic data within the site vicinity identified numerous topographic lineaments that are parallel to the general NE-SW-trend of the growth faults identified in the subsurface data (Figure 2.5.1-44). LiDAR data was not available for Goliad County in the northwest region of the site vicinity, so no lineaments are identified in that region. The greatest density of lineaments occurs in the central and northwestern part of the site vicinity. Based on their character and geomorphic context, LiDAR lineaments were assessed to be potentially related to Quaternary (post-Beaumont Formation) growth fault activity, or to be the result of non-tectonic depositional, fluvial, and/or erosional processes unrelated to growth fault activity (Figure 2.5.1-44). Criteria used to assess the lineaments include:

- The degree of linearity and consistency of expression. Slope breaks associated with growth faults in the site vicinity have a distinct linear expression on LiDAR data that contrasts strongly with surrounding topography, and they almost exclusively face toward the southeast. In contrast, fluvial features are generally fainter, more discontinuous, more difficult to distinguish from the surrounding topography, and lack a consistent southeast facing direction.

- The degree of lateral continuity. Strongly linear topographic features that can be traced for many thousands of feet to miles are more likely to be associated with linear geologic structures than non-tectonic features, particularly if the lineaments trend at a high angle to local streams, drainage patterns and other fluvial geomorphic features.

- Cross-cutting relationships. Linear topographic features that cross boundaries between non-tectonic landforms such as terraces, stream margins, levees, etc., post-date the landforms and indicate different (non-fluvial) processes of formation.

- Deflected or otherwise modified fluvial systems. Lineaments associated with topographic slope breaks that clearly deflect or influence drainage development are potentially related to growth faults.

These criteria were used to classify the lineaments as either: (1) potentially related to growth faults, or (2) probably related to fluvial process as shown in Figure 2.5.1-44.

From inspection and comparison of the data presented in Figures 2.5.1-36 and 2.5.1-44, there is a close spatial association between the LiDAR growth fault lineaments and the projected surface traces of growth faults, which supports the interpretation that these lineaments reflect growth fault activity (Table 2.5.1-1). For example, the updip projections of faults GM-Z and GM-T in the northwest group are locally coincident with northeast-trending topographic lineaments assessed as potentially related to growth fault activity. The few LiDAR lineaments potentially related to growth fault activity that are not closely or obviously associated with the projected faults shown in Figure 2.5.1-36 (e.g., the two lineaments furthest and directly east of the site) are associated with growth faults mapped at depth by Geomap (Reference 2.5.1-123). The lack of surface projections for these faults is due to incomplete and discontinuous fault traces within the Geomap horizons, which precluded the derivation of surface projections.

The topographic lineaments potentially associated with growth fault activity generally do not exhibit the same lateral continuity as the surface projections of faults from the subsurface (Reference 2.5.1-123). This observation suggests that activity is restricted to short segments of growth faults relative to the length of the fault observed at depth.

All potential growth fault lineaments identified from LiDAR within the site vicinity were assessed in the field during ground and aerial reconnaissance efforts. Aerial reconnaissance was conducted from a small fixed wing aircraft from an altitude of approximately 1400 feet (427 meters) during low sun angle conditions. None of the lineaments were observable features during the aerial reconnaissance most likely due to the extremely subtle nature of the topographic features (see description in Subsection 2.5.1.2.4.2.3.2). Ground reconnaissance of lineaments was conducted where property access could be obtained (approximately two-thirds of the lineaments are on property where access was not obtained). Of those features that were accessible, only GM-A, GM-D, GM-E, GM-T, GM-U, and GM-V had a topographic expression that was observable on the ground. The lack of observable expression on the ground is most likely due to the extremely subtle relief of the features and the presence of vegetation, which in many cases was at least as tall as the topographic relief.

2.5.1.2.4.2.3 Site Area Growth Faults

Growth faults in the site area were identified from publications indicating the presence of growth faults in the subsurface (References 2.5.1-238 and 2.5.1-239), the subsurface mapping of the Geomap Company (Reference 2.5.1-123), the LiDAR data from the Texas Natural Resources Information Systems (References 2.5.1-249 and 2.5.1-250), aerial photography, and seismic reflection data licensed for the VCS ESP, all described in Subsection 2.5.1.2.4.2.1. A compilation of growth faults identified within subsurface data is shown in Figure 2.5.1-40 and listed in Table 2.5.1-1.

All the topographic lineaments identified from the LiDAR data is shown in Figure 2.5.1-42 and those lineaments potentially representing growth faults are also shown in Figure 2.5.1-40. Topographic lineaments identified from aerial photography are shown in Figure 2.5.1-37. Growth faults identified in the seismic reflection data is described later in this subsection.

Within the site area the growth faults identified within the Geomap data (Reference 2.5.1-123) is used as the reference set of faults to which other indicators of faults (i.e., the growth faults of Dodge and Posey (Reference 2.5.1-238), the growth faults Galloway et al. (Reference 2.5.1-239), and the LiDAR lineaments potentially associated with growth faults) are referenced. Five growth faults identified and mapped in the subsurface by Geomap (Reference 2.5.1-123) are present within the site area (Figure 2.5.1-40). From southeast to northwest, the faults include: GM-E, GM-D, GM-K, GM-L, and GM-N. Faults GM-A and GM-B occur just outside of the site area. The Geomap mapping (Reference 2.5.1-123) and the resultant surface projections of the faults presented here show that GM-D is a regional-scale growth fault that extends throughout the site vicinity (Figures 2.5.1-36 and 2.5.1-240). Fault GM-E is an approximately 5-mile-long southward-branching splay of fault GM-D. Similarly, fault GM-L is a regional-scale growth fault extending throughout the site vicinity, and fault GM-K is a splay of this regional fault (Figures 2.5.1-36 and 2.5.1-40). Fault GM-N is an approximately 25-mile-long growth fault that is roughly confined to the site vicinity to the west of the site and merges with the regional fault GM-L north of the site (Figures 2.5.1-36 and 2.5.1-40). The surface projections of the Geomap faults (Reference 2.5.1-123) show these faults as crossing over one another, while the subsurface data licensed from Geomap (Reference 2.5.1-123) shows these faults merging at depth. This apparent discrepancy is an artifact of the uncertainty in the projection locations of the Geomap faults.

Dodge and Posey (Reference 2.5.1-238) identified two growth faults within the site area: fault DP-6 to the southeast of the site and fault DP-7 to the northwest of the site (Figure 2.5.1-40). Galloway et al. (Reference 2.5.1-239) identified one growth fault within the site area, fault G-13 to the southwest of the site. Given the uncertainty in the position of the surface projections of the faults it is difficult to positively correlate DP-6, DP-7, and G-13 with any particular Geomap fault. As described in Subsection 2.5.1.2.4.2.2, there are several Geomap faults with which each of these faults could be correlated. Therefore, it is likely these growth faults are represented by the Geomap faults (Reference 2.5.1-123), so DP-6, DP-7, and G-13 are not described further.

The LiDAR lineaments within the site area previously described in Subsection 2.5.1.2.4.2.2 are shown in Figure 2.5.1-42. In Subsection 2.5.1.2.4.2.2 these lineaments were classified as either potentially related to growth fault activity or probably related to fluvial and erosional processes. Within the site area there are only two sets of LiDAR lineaments that may be related to growth faulting. Stereoscopic analysis of aerial photography within the site area also identified lineaments based on subtle tonal and vegetation changes as well as topographic features (Figure 2.5.1-37). In many cases, these photo lineaments are coincident with lineaments identified with the LiDAR data

(Figure 2.5.1-42). Using the criteria presented in Subsection 2.5.1.2.4.2.2 to distinguish between potential growth-fault related and other lineaments, it was determined that the only aerial photograph lineaments that represent growth fault activity are those correlative to growth fault related LiDAR lineaments shown in Figure 2.5.1-42.

A comparison of the Geomap fault projections and the lineaments potentially related to growth faults within the site area shows that of the five identified growth faults within the site area, only faults GM-D and GM-E are spatially associated with anomalous topographic lineaments or features (Figure 2.5.1-40). The topographic lineament associated with the surface projection of fault GM-E is a localized down-to-the-southeast inflection of the land surface in the southeastern part of the site area (Figure 2.5.1-39). Towards the central and western parts of the lineament, the surface projection of growth fault GM-E trends northward away from the lineament and does not cross the San Antonio River valley. In contrast, the lineament is present in the upper surface of the Pleistocene Beaumont Formation trending to the southwest across the San Antonio River valley (Figure 2.5.1-39). This apparent discrepancy in the correlation between the lineament and fault projection is most likely due to uncertainties in the fault projection from depth. The raw Geomap data (Reference 2.5.1-123) shows the subsurface trace of GM-E crossing the San Antonio River valley along a trend similar to the lineament. Therefore, it is reasonable to conclude that the lineament and correlated surface projection of growth fault GM-E represent the same growth fault. For simplicity, this fault is referred to as fault E.

The topographic lineament of fault E is clearly discernable west of the San Antonio river valley and cuts across an abandoned oxbow incised in the upper surface of the Beaumont Formation. East of the San Antonio River valley, the LiDAR lineament splits into two short (approximately 0.25 mile or 0.4 km) branches with the lineament extending further east from between these branches (Figure 2.5.1-39). Immediately east of the fork the lineament is associated with a jog or deflection in the channel of Kuy Creek (Figure 2.5.1-39). Two short tributary branches of Kuy Creek appear to be just south of and aligned parallel to the lineament. Geologic field reconnaissance conducted for the VCS ESP application study confirmed the presence of the southeast-facing topographic break associated with accessible portions of the lineament. In particular, expression of the lineament is obvious where it crosses SR 239, FM 445, and between the crossing of the Kuy Creek main stem and the previously mentioned tributaries.

The topographic lineament associated with the surface projection of fault GM-D is located between Kuy Creek on the southwest and the Guadalupe River valley on the northeast. The lineament is expressed as a subtle down-to-the-south topographic feature that is difficult to identify in places, but is interpreted to extend continuously for several miles (Figures 2.5.1-40 and 2.5.1-38). In contrast to the relatively linear surface projection of fault GM-D, the LiDAR lineament is concave toward the southeast and it has several second-order curves in its trace just south of the VCS site; none of this complexity is reflected in the original subsurface mapping (Reference 2.5.1-123) from which the

surface projection is derived. However, the strong spatial correlation between the lineament and the surface projection of fault GM-D leads to the conclusion that the surface projection and lineament represent the same fault. For simplicity this fault is referred to as fault D.

The fault D lineament extends from just west of the Guadalupe River valley westward south of the site. The closest approach of the integrated zone of deformation to the planned VCS power block area is approximately 509 feet (155 meters) south of the south corner of the planned power block area (Figure 2.5.1-43). Just west of the site the lineament curves southward and continues subparallel to U.S. Highway 77. The topographic expression of the fault D lineament is much more subtle in the LiDAR data (Figure 2.5.1-38) than the fault E lineament (Figure 2.5.1-39) (see description in Subsection 2.5.1.2.4.2.3.2 for detailed analysis of the characteristic of the lineaments), and it is difficult to discern without a high-resolution, shaded-relief image. Field reconnaissance of the fault D lineament confirmed the presence of a subtle southeast-down inflection of the land surface associated with the lineament. Due to the subtle nature of the lineament and the relatively tall, grassy vegetation throughout the site area, the lineament was not obvious along its entire extent as defined by the LiDAR data, but was observed in the field at several locations.

2.5.1.2.4.2.3.1 Seismic Reflection Data

Analysis of publicly available data (References 2.5.1-238 and 2.5.1-239), proprietary subsurface mapping (Reference 2.5.1-123), LiDAR-derived topography (References 2.5.1-249 and 2.5.1-250), and aerial photographs indicates that there are five growth faults within the site area and only two of those faults have potentially caused deformation of the Late Pleistocene Beaumont Formation within the site area. The envelope of interpreted zone of deformation for Fault D approaches within approximately 509 feet (155 meters) of the power block area and has a subtle topographic lineament. Fault E approaches within approximately 2.6 miles (4.2 km) of the site and has a more distinct topographic expression. All of the remaining growth faults within the site area do not have any associated topographic lineaments and are at greater distances from the site.

Given the nearness of fault D to the site and the potential Quaternary surface deformation associated with the fault, proprietary seismic reflection data was licensed from a seismic data broker, as described in Subsection 2.5.1.2.4.2.1.5, to further document the structural characteristics of fault D and better characterize the potential for surface deformation from Quaternary activity on the fault.

2.5.1.2.4.2.3.1.1 Interpretation Methodology

As described in Subsection 2.5.1.2.4.2.1.5, Exelon licensed an array of seismic reflection lines from Seismic Exchange, Inc. This data consisted of: two “dip lines” (TGS and GDI) oriented roughly perpendicular to faults D and E, one oblique line (GSI) oriented roughly north-south, and one “strike line” (PLJ) oriented roughly perpendicular to the dip lines and located near the surface projection of the fault D. The raw data from these seismic lines was processed by Excel Geophysical Services,

Inc. to generate a series of reflection profiles for each line (three profiles in the time domain [unfiltered data, migrated data, final stacked data] and one depth migration). The processed data was analyzed and interpreted using Seismic Micro-Technology, Inc.'s Kingdom Suite software (Version 8.2) to display and interpret the data. The interpretation methodology included the following steps:

1. Examining all seismic profiles for correlative seismic reflectors and faults at points where one line crosses another. Due to differences in data acquisition and processing methods, sequences of correlative reflectors were located at slightly different record times or depths on different seismic lines. This data "mis-ties" were corrected by applying bulk shifts in time or depth for internal consistency of data throughout the seismic array.

2. Identifying and mapping of distinct marker horizons and fault surfaces on individual seismic lines, and correlating these features throughout the array.

3. Measuring the amount of displacement of the mapped marker horizons where they were offset by faults.

4. Examining each seismic line in detail where it crosses topographic lineaments potentially related to growth faults, with emphasis on characterizing the presence and style of shallow deformation, if any, associated with Quaternary activity.

5. Importing additional map data to facilitate interpretation. Map data incorporated in the analysis included the locations of potential VCS structures, LiDAR derived topographic data (References 2.5.1-249 and 2.5.1-250), Geomap (Reference 2.5.1-123) structural contour maps, growth fault surface projections developed from the Geomap data, and LiDAR lineaments potentially related to growth fault activity.

While all four profiles from each line were analyzed, only interpretations of the final stack, time-migrated sections are presented here for TGS, GDI, and GSI because these profiles provide the best imaging of the site area structure. In addition, a section of the depth-migrated profile for GDI is presented to provide detail of fault D in the relatively shallow subsurface near the site and to allow for projection of the apparent zone of deformation associated with fault D to the surface. No profiles from line PLJ are presented because the line was used primarily to correlate marker horizons between the other lines and is poorly oriented to image the structure of growth faults within the site area.

2.5.1.2.4.2.3.1.2 Identification and Mapping of Stratigraphic Marker Horizons

Prominent and laterally continuous high-amplitude reflectors are readily observable in the upper parts of the reflection profiles. Four distinct reflectors were chosen as key horizons for constraining the subsurface structure primarily based on the depth to the horizon and continuity of the horizon through the lines. These marker horizons were mapped on all profiles and are referred to, from deepest to

shallowest, as Horizon 1, Horizon 2, Horizon 3, and Horizon 4 (Table 2.5.1-3) (Figures 2.5.1-45, 2.5.1-46, 2.5.1-47, and 2.5.1-49).

The deepest marker horizons (Horizon 1 and Horizon 2) (Figures 2.5.1-45, 2.5.1-46, and 2.5.1-47) correlate with the stratigraphic horizons depicted in the regional cross sections of Dodge and Posey (Reference 2.5.1-238) that indicate, respectively, the top of the Vicksburg Formation and the top of the Frio Formation. Within the site area Dodge and Posey (Reference 2.5.1-238) map the top of the Vicksburg at approximately 5500 to 6500 feet (1676 to 1981 meters) below sea level. In the seismic profiles, Horizon 1 occurs at depths between 5090 and 7320 feet (1550 and 2230 meters) as seen in the depth migrated profiles, or equivalently 1.357 to 1.838 seconds in the time-migrated profiles (Table 2.5.1-3). This correlation suggests that Horizon 1 seen in the reflection profiles is at or near the top of the Vicksburg. Within the site area Dodge and Posey (Reference 2.5.1-238) map the top of the top of the Frio Formation at approximately 3000 to 4000 feet (914 to 1219 meters) below sea level. In the seismic profiles, Horizon 2 occurs at depths between 3575 and 4450 feet (1090 and 1356 meters) as seen in the depth migrated profiles, or equivalently 0.986 to 1.184 seconds (Table 2.5.1-3). This correlation suggests that Horizon 2 seen in the reflection profiles is at or near the top of the Frio. Horizon 1 and Horizon 2 may represent deeper-water finer-grained sediments (e.g., shale) deposited during marine transgressions (Reference 2.5.1-128). This hypothesis is consistent with observations that suggest the seismic velocity of the horizons differs markedly from that of overlying and underlying sediments, which may be more sand-rich, thus producing an impedance contrast that gives rise to a distinct reflector in the seismic data.

Horizon 3 occurs in the seismic profiles at a depth of 1150 to 1790 feet (351 to 546 meters), or equivalently 0.357 to 0.533 seconds (Figure 2.5.1-3). This horizon does not appear within regional cross sections drawn within the site vicinity (References 2.5.1-238 and 2.5.1-239) because these cross sections did not identify deposits stratigraphically above the Frio. However, a regional cross section from Baker (Reference 2.5.1-251) within the site vicinity and east of the site area projects the base of the Goliad Formation to similar depths. Based on this correlation, Horizon 3 is interpreted to be a finer-grained unit, potentially shale, underlying the relatively sandier Goliad Formation. The deposit marking this horizon was potentially deposited during a latest Miocene or Early Pliocene marine transgression.

Horizon 4 occurs in the seismic profiles at a depth of 650 to 1340 feet (198 to 408 meters), or equivalently 0.209 to 0.385 seconds (Table 2.5.1-3). This horizon is the shallowest laterally continuous reflector imaged in the seismic array (Figures 2.5.1-45, 2.5.1-46, 2.5.1-47, and 2.5.1-48). The exact nature and age of the stratigraphic boundary represented by this marker is unknown.

2.5.1.2.4.2.3.1.3 Growth Fault Structure

The primary geologic structures imaged within the seismic array are a series of southeast-dipping normal faults and smaller, second-order synthetic and antithetic normal faults that are present in the

hanging walls of the major southeast-dipping faults (Figures 2.5.1-45, 2.5.1-46, 2.5.1-47, and 2.5.1-48). These faults were identified based on the abrupt lateral termination of reflectors, abrupt changes in reflector apparent dip, disturbed reflectors along apparent bedding planes, and relations among geologic structures and stratigraphy common to growth fault systems (e.g., Reference 2.5.1-224). The faults are appropriately classified as growth faults based on the presence of thickened sedimentary sections on the downthrown sides of the faults. Many of the faults identified on the seismic lines are correlative with faults identified within the Geomap data (Reference 2.5.1-123), including faults GM-L, GM-K, GM-E, GM-D, and GM-A (Figure 2.5.1-36). Other faults identified in the seismic profiles were given a number designation with the prefix SR that increases sequentially from north to south, respectively (Figures 2.5.1-45, 2.5.1-46, and 2.5.1-47). The discrepancy between faults identified in the reflection data and the Geomap data does not reflect significant inconstancies between the two datasets. Faults identified within the reflection data yet not within the Geomap data generally does not propagate upwards to the stratigraphic horizon used in the Geomap mapping, preventing their identification by Geomap. Also, these faults not identified in the Geomap data sometimes occur as subsidiary splays very proximal to other faults identified by Geomap making them difficult to resolve within the well-log data used by Geomap to identify faults.

The major southeast-dipping faults exhibit a listric geometry; i.e., they dip steeply at their upward terminations and progressively flatten downward to sole into or terminate against sub-horizontal detachment horizons, as is characteristic of Vicksburg growth faults (References 2.5.1-135, 2.5.1-125, 2.5.1-94, 2.5.1-128, 2.5.1-134, and 2.5.1-28). The deepest and most laterally extensive detachment horizon is consistently imaged in the time profiles for each line at a depth of approximately 3.9 to 4.5 seconds (Figures 2.5.1-45, 2.5.1-46, 2.5.1-47). This sub-horizontal fault is the main detachment within the site area and is likely regional in extent. Similar detachments are characteristic of the Texas Gulf Coastal Plains growth faults systems, including the Vicksburg (References 2.5.1-252, 2.5.1-224, 2.5.1-253, 2.5.1-94, 2.5.1-122, 2.5.1-131, and 2.5.1-28). As shown in the interpretations of the time domain 2D reflection profiles (Figures 2.5.1-45, 2.5.1-46, 2.5.1-47), faults SR-01, GM-L and GM-K terminate downward and sole into the main detachment. Given the characteristics of other growth faults observed throughout the Texas Gulf Coastal Plains (References 2.5.1-135, 2.5.1-125, 2.5.1-94, 2.5.1-51, 2.5.1-128, 2.5.1-134, and 2.5.1-28), it is likely that the other major faults to the southeast of fault GM-K also root into the main detachment or another shallower detachment horizon, but the seismic data does not extend far enough to the southeast to image these relationships.

As previously mentioned, the patterns of layered reflectors present in the profiles are consistent with the interpretation that the southeast-dipping faults are growth faults. In the depth range between Horizon 1 and the main detachment, the layered reflectors dip consistently toward the northwest into the listric growth fault surfaces. The reflectors in this depth range exhibit a downward fanning pattern above the faults (i.e. they become progressively steeper with depth). All of these relationships are

indicative of deposition occurring while the faults were active (Reference 2.5.1-51). Some reflectors form convex patterns, indicating the presence of “rollover anticlines” that may develop with sufficiently large normal displacements on the faults and underlying detachment.

Fault GM-E, which is associated with an anomalous topographic lineament, is not visible in the seismic profiles. This is not surprising because none of the lines cross the topographic lineament associated with the fault, and only line TGS crosses the surface projection of the fault. Line TGS was carefully examined for evidence of fault GM-E, but no discernable signature of the fault was observed. The point at which line TGS crosses the subsurface trace of fault GM-E is relatively close to the end of the line and near the lateral termination of the subsurface trace. The lack of a signature of fault GM-E is primarily attributed to the reduced imaging capability at the end of the line and the possibility that the fault has largely died out within the stratigraphic section imaged by the line.

2.5.1.2.4.2.3.1.4 Stratigraphic and Structural Relations

The patterns of reflectors relative to the mapped horizons and growth faults (Figures 2.5.1-45, 2.5.1-46, and 2.5.1-47) indicate that the majority of growth fault activity in the site area occurred before deposition of Horizon 1, and thus before the deposition of the top of the Vicksburg Formation in Middle Oligocene time (see Table 2.5.1-4 for a summary of updip fault extents and horizon offsets for each fault). Below Horizon 1, in the depth range of about 2.5 to 4.0 seconds, layered reflectors of Gulf Coastal Plains strata dip consistently to the north and have distinct downward fanning geometries. The layered reflectors dip much less steeply above a depth of about 2.2 seconds, and with minor local exceptions, the reflectors are sub-horizontal just below Horizon 1, indicating that growth fault activity of the entire system had either ceased or decreased to very low rates of movement by upper Vicksburg time. Faults SR-01 and SR-03 are overlain by an undeformed Horizon 1 marker, and thus have not been active since Middle Oligocene time. In contrast, some faults (e.g., faults GM-K and GM-D) deform bedding above Horizon 1, but with much smaller offsets. This timing of growth fault movement and the characteristic of some faults having minor activity beyond the Middle Oligocene is consistent with these faults belonging to the Vicksburg growth fault zone (see description in Subsection 2.5.1.1.4.3.4.2).

The faults that exhibit evidence for post-Vicksburg activity are of primary importance for the VCS site. Fault GM-L appears to offset the Horizon 1 marker very slightly on seismic line GDI (Figure 2.5.1-47), but displacement on the fault does not reach as high as Horizon 1 in the stratigraphic section on lines TGS (Figure 2.5.1-45) and GSI (Figure 2.5.1-46). Fault GM-K extends above Horizon 1 on all three dip lines to offset Horizon 2, indicating some activity following deposition of the Frio (Late Oligocene to Early Miocene). However, Horizon 3 is undeformed above fault GM-K, demonstrating the absence of activity since Early Pliocene time. The upward terminations of faults SR-04, SR-05, SR-06, and SR-07 vary from seismic line to seismic line. For example, displacement on fault SR-05 terminates below Horizon 1 on seismic line TGS (Figure 2.5.1-45), but extends slightly above Horizon 1 on line

GDI (Figure 2.5.1-47). Similarly, faults SR-06 and SR-07 can be traced above Horizon 1 and Horizon 2, respectively, but both die out below Horizon 3. Because all of these faults do not project near the site at the surface (Figure 2.5.1-47) and have no apparent activity since Early Pliocene time, they are do not have the potential to cause permanent ground deformation at the site (see Subsection 2.5.3).

Fault GM-D is the only structure in the site area that exhibits evidence in the seismic data for post Horizon 3 (post Early Pliocene) displacement. Fault GM-D can be traced upwards through Horizon 3, and it is observed to cause down-to-the-southeast displacement of Horizon 4 on all of the profiles except those of line GSI (Figures 2.5.1-45, 2.5.1-46, 2.5.1-47, and 2.5.1-48) (Table 2.5.1-4), which crosses the fault where there is no lineament potentially indicative of surface deformation. Tracing discrete displacement of reflectors associated with fault GM-D above Horizon 4 on lines TGS and GDI is difficult because Quaternary deposition of Gulf Coastal Plains sediments occurred in a near-shore fluvial-deltaic environment (e.g., Reference 2.5.1-40), which is less amenable to the development of laterally continuous reflecting horizons than the shelf and deeper marine environments that characterize the underlying stratigraphic section. Additionally, the seismic acquisition parameters used when collecting the original data in the late 1970s to mid-1980s were designed to optimize imaging for petroleum exploration, which primarily occurs at depths of several thousand feet and greater in the site area. Given these caveats, the pattern of reflectors in the seismic data above Horizon 4 suggest that post-Horizon 4 activity of fault GM-D has produced distributed down-to-the-southeast tilting or folding of strata within a triangular zone that widens upward from a point just above Horizon 3. As interpreted on a section of the depth-migrated profile of line GDI (Figure 2.5.1-48), the width of the zone of deformation at the top of the seismic record section is approximately 1600 feet (488 meters). When compared to a detailed topographic profile along the seismic line extracted from LiDAR data, the triangular zone projects upward to a southeast-down break in slope with about 5 feet (1.5 meters) of total relief on the upper surface of the Beaumont Formation (Figure 2.5.1-48).

The southeast-down tilting of post Horizon 4 reflectors in the hanging wall of fault GM-D is interpreted to be folding related to relatively young activity of the fault at depth, and possibly upward propagation of the fault tip through unfaulted shallow sediments above Horizon 4. Post Horizon 4 activity on fault GM-D may have occurred at such a low rate that the tip of the fault has not been able to propagate upward through the most recent accumulation of Quaternary sediment. Although the fault may not actually break the surface, movement on the fault at depth could potentially produce southeast-down displacement of the overlying sediments and land surface in the hanging wall, which would be expressed as a local southeast-down tilting of the land surface above the fault. The zone of tilting would predictably narrow downward to the point at where discrete displacement is occurring on the fault at depth, and where beds are offset rather than tilted or folded. The process of tilting or folding of the material above the tip of a buried fault is generally referred to as “fault-propagation folding” (Reference 2.5.1-254). Fault-propagation folding has been observed to occur within triangular zones,

called “trishear zones,” updip of the fault tip (Reference 2.5.1-255). The tilting of discontinuous reflectors above Horizon 4 in the shallow subsurface suggests that trishear fault-propagation folding, or some other mechanism of distributed southeast-down tilting, is the primary mode of Quaternary surface deformation related to activity of fault GM-D rather than discrete surface faulting.

2.5.1.2.4.2.3.2 Surface Deformation Associated with Fault D

The updip projection of fault D from the seismic reflection data intersects the ground surface at the topographic lineament associated with fault D identified within the LiDAR data, and the zone of deformation updip of the fault D tip as observed in the reflection data correlates with anomalous tilting of the land surface (see topographic profile in Figure 2.5.1-48 extracted from the LiDAR data along the extent of seismic line GDI). Based on the spatial correlation between deformation in the subsurface and the lineament, it is concluded that the lineament and the associated southeast-facing slope break represent surface deformation associated with movement on fault D since deposition of the middle to late Pleistocene Beaumont Formation.

Over 90 topographic profiles perpendicular to the lineament were extracted from the LiDAR data to fully characterize the style, extent and magnitude of surface deformation associated with activity on fault D. A subset of characteristic profiles is presented in Figures 2.5.1-50a through 2.5.1-50c to document the style of surface deformation associated with fault D (see Figure 2.5.1-49 for profile locations). Overlain on these profiles is the site geology as presented in Figure 2.5.1-4.

Topographic profile 4, located approximately 6000 feet (1829 meters) south of the western portion of the VCS power block area, documents a clear, southeast-down step in the upper surface of the Beaumont Formation. Total relief across the slope break is approximately 4 feet (1.2 meters) over 820 feet (250 meters). Profile 6, located approximately 4000 feet (1219 meters) east of the eastern portion of the power block area, also images a southeast-facing slope break with approximately 3.3 feet/410 feet or 1 meter/125 meters of total relief. Although the large vertical exaggeration of the topographic profiles makes the surface deformation look like a discrete fault scarp, the slope breaks in these profiles actually represents very subtle localized increases in the regional gradient of the land surface. For example, in profile 4 the surface slope of the tilted surface of the Beaumont Formation associated with the slope break is only approximately 0.28 degrees (i.e., 4 feet/820 feet or 1.2 meters/250 meters), and in profile 6 the tilted surface slope of the slope break is only approximately 0.46 degrees (i.e., 3.3 feet/410 feet or 1 meter/125 meters). The subtle nature of this tilting of the land surface was confirmed in the field where in many cases the tilting could not be discerned from other variations in topography.

These detailed topographic profiles are consistent with the interpretation of the seismic reflection data in Subsection 2.5.1.2.4.2.3.1 in that deformation of the land surface related to Quaternary activity of fault D primarily is characterized by down-to-the-southeast tilting or folding, rather than

discrete surface faulting. The lateral extent of tilting measured at the surface ranges from approximately 200 to 1300 feet (61 to 396 meters), which is similar to the width of the panel of southeast-tilted reflectors in the hanging wall of fault D as inferred from analysis of seismic line GDI (Figure 2.5.1-48). From these relationships, it is concluded that Quaternary surface deformation associated with fault D is characterized by monoclinal fault-propagation folding. The folding is recorded as localized tilting of the Beaumont Formation and the formation of a slope break.

Topographic profiles within 0.6 mile (0.97 km) of the site generally document down-to-the-southeast tilting or folding of the upper surface of the Beaumont Formation along the lineament of fault D, but they also reveal variations in the topographic expression of the deformation due to localized modification of the tilted surface from geomorphic processes (e.g., localized runoff along Kuy Creek) and cultural modifications (e.g., roads, pipelines) (Figures 2.5.1-49 and 2.5.1-50a through 2.5.1-50c). For example, profile 2, which passes through the power block area and profile 8, which passes through the center of the eastern portion of the power block area, both reveal distinct southeast-facing tilting of the Beaumont surface with relief of approximately 4 feet (1.2 meters) occurring over hundreds of feet. On these and similar profiles the zone of potential deformation associated with fault D (i.e., the tilting associated with the slope break) is obvious. In contrast, profiles 1, 5, and 7 also show relief in the land surface of approximately 4 feet (1.2 meters), but this relief occurs over approximately 1000 feet or more. In addition, there are other topographic signals with relief also on the order of several feet that make identifying the tilting and associated slope break more difficult. Given the proximity of profiles 2 and 8, which display the break in slope and distinct tilting, to profiles 1, 5, and 7, where the tilting and slope-break are less obvious, it is reasonable to conclude that the northwest-to-southeast decrease in surface elevation on profiles 1, 5, and 7 is also due to southeast-down movement on fault D, even though as discrete of a slope break cannot be discerned. Further analysis of the profiles, described below, supports this conclusion.

The lack of an apparent slope break in profiles 1, 5, and 7 is attributed to erosional processes associated with late Quaternary drainage development on the exposed surface of the Beaumont Formation and cultural modifications made to the land surface that have degraded the original slope break. The site geologic map (Figure 2.5.1-5) shows that the footprint of the potential power block area is located within the “meander belt” unit of the Beaumont Formation (map unit Qbs), which represents older fluvial channel deposits associated with braided and meandering streams. The Qbs unit trends northwest-southeast (Figure 2.5.1-5) and is flanked on the east and west by map unit Qbc, which represents levee and overbank deposits adjacent to the Beaumont fluvial channels. A southwest-northeast topographic profile through the VCS power block area (profile 3) (Figures 2.5.1-50a through 2.5.1-50c) shows that the Qbs unit is associated with a broad and shallow topographic low that is 1 to 2 feet (0.3 to 0.6 meters) lower in elevation than the surrounding Qbc units. This broad low in Qbs is interpreted as inherited topography potentially reflecting a relict stream channel active during the final stages of deposition of the Beaumont Formation. In an area of

extremely low relief with surface slopes near zero as in the site area, subtle, preexisting lows in topography may preferentially gather and distribute meteoric run-off and thus influence the development of subsequent erosional drainage networks.

It is concluded that the subdued expression of the slope break on profiles 1, 5, and 7 is due to surface run-off that is preferentially captured and directed into the relict Qbs topographic low. Despite the extremely low surface gradients of the tilted surfaces (less than 0.5 degrees), this concentrated ruff-off is preferentially degrading the southeast-facing slope break within the Qbs unit by laying back the tilting of the land surface over a larger horizontal distance. Profiles 1 and 2 provide a clear example of this degradation. Profile 2 runs through the center of the Qbs topographic low, the region of the degraded slope break, and profile 1 runs along the edge of the Qbs and Qbc units in the region expected to have experienced less degradation (Figures 2.5.1-4, 2.5.1-49, 2.5.1-50a through 2.5.1-50c). A comparison of the profiles shows that the profiles have a remarkably similar form with the major difference in topographic shape occurring at the points of highest curvature of the slope break in profile 2. At highest end of the profile 2 slope break profile 1 is lower in elevation; at the lowest end of the profile 2 slope break profile 1 is higher in elevation. This relationship between the slope breaks on both profiles suggests that material has been eroded near the “top,” or northwestern end, of the slope break in profile 1 and deposited near the “bottom,” or southeast, end as a “bench” of eroded material. This process effectively decreases the tilting observed in the topographic profile across the slope break. In this example, the cross sectional area of material that has apparently been removed in profile 2 from the uphill end of the slope break is comparable to the cross-sectional area of material apparently added to the downhill end.

Also apparent in profile 1 is a distinct decrease and increase in the topographic profile of approximately 3 feet (1 meter) to the northwest (uphill) of where the power block area projects into the profile. This step in the profile partially masks the tilting of the Beaumont Formation because the relief of the step is on the same order as that of the tilting associated with fault D. From analysis of the LiDAR data, it is apparent that this step reflects modifications of the land surface from a road or pipeline (Figure 2.5.1-49).

Topographic profiles 7 and 8 trend through the power block area, and also show the characteristic differences between a zone of degraded tilting and one with less modification (Figure 2.5.1-50c). Profile 8 has a distinct slope break and tilted surface between a profile distance of approximately 1.2 to 1.3 miles (1900 to 2150 meters). This region is interpreted as a largely unmodified slope break delineating the extent of surface deformation related to fault D activity. This zone is over 980 feet (300 meters) from the eastern portion of the power block area. In contrast, profile 7 does not have as sharp of a slope break in the region of the fault D lineament; the slope break in profile 7 is more subtle, occurs over a larger profile distance that includes the western portion of the power block area, and it has several relatively abrupt steps in topography. The increased extent of the subtler slope break is due to the localized erosional degradation of the zone of titling, and the abrupt steps in

topography are due to cultural features. Despite degradation of the slope break, using detailed analysis of the LiDAR data, the region of potential deformation associated with fault D can be constrained to an area that does not extend to the western portion of the power block area. In particular, the zone of potential deformation along profile 7 is defined based on identifying the characteristic depositional benching and erosional degradation features identified in profile 1, extrapolating the projected zone of deformation from the seismic data (Figure 2.5.1-48), identifying cultural features modifying the land surface, and extrapolating the zone of deformation from neighboring topographic profiles. The resultant interpreted zone of deformation is over 820 feet (250 meters) from the power block area.

A suite of over 90 topographic profiles was analyzed to define a zone enveloping the interpreted extent of tilting or folding associated with post-Beaumont activity on fault D. As shown in the eight representative profiles compiled in Figures 2.5.1-50a through 2.5.1-50c, for each profile the interpreted zone of deformation was defined by identifying the “uphill” and “downhill” extent of the deformation (see the zones of potential deformation identified in Figures 2.5.1-50a through 2.5.1-50c). This zone was defined for each profile using a “top” and “bottom” point defined by identifying the top and bottom of the slope break, as defined above, while taking into account potential masking of the zone of deformation from depositional benching, erosional degradation, and cultural modifications of the land surface. For each profile these top and bottom points of the interpreted zone of growth fault related deformation are shown in Figures 2.5.1-43 and 2.5.1-49. An envelope was drawn around these points to delimit the maximum extent of interpreted deformation related to post-Beaumont activity of fault D. In places where the zone of tilting was significantly degraded, the envelope is dashed yet drawn well outside of the interpreted extent of growth-fault related tilting. These few regions correlate to areas where the fault D lineament was poorly defined. The closest approach of the zone of interpreted surface deformation associated with fault D to the power block area is 509 feet (155 meters) (Figure 2.5.1-43).

2.5.1.2.4.2.3.3 Activity Rates of Growth Fault D

Long-term average rates of surface deformation associated with growth fault D can be estimated from the age of the deformed upper surface of the Beaumont Formation and the total surface relief. Based on analysis of topographic profiles, the separation of the upper surface of the Beaumont Formation across fault D ranges from approximately 1.5 feet to 4.5 feet (about 0.5 to 1.5 meters). As described in Subsection 2.5.1.2, the precise age of the Beaumont Formation is uncertain. Current estimates of the age of the Beaumont vary between 350 ka and 100 ka (References 2.5.1-218, 2.5.1-220, 2.5.1-40, 2.5.1-132, and 2.5.1-219). From the extremes in the range of relief and ages, the corresponding range in long-term average separation rates across fault D is approximately 5.1×10^{-5} inches per year to 5.4×10^{-4} inches per year. If it is assumed that fault D slips continuously and uniformly at these rates, then the maximum down-to-the-southeast displacement of the land surface across the fault in 100 years will be about 1/18th of an inch.

These estimates assume that the observed relief in the surface of the Beaumont Formation has occurred through continuous and uniform movement on the fault. It is possible that movement occurs episodically in response to changes in fluid pressure associated with hydrocarbon migration, localized loading of the growth fault system by pulses of sedimentation in the offshore region, or other natural phenomena. If this is the case, short-term activity rates will be higher than the long-term average rates determined above.

2.5.1.2.4.2.4 Growth Fault E

Growth fault E is over 2.6 miles (4.2 km) from the VCS, and as such, any activity on the fault will not affect the site. Despite this fact, fault E is still a potentially significant structure because it is the only fault besides fault D that has an associated topographic lineament within the site area. As described in Subsection 2.5.1.2.4.2.3.1.3, fault E is not apparent in the seismic reflection data because the seismic profiles do not extend far enough to cross the fault and give sufficient resolution of the fault at depth. Despite the lack of reflection imaging, the distinct topographic lineament apparent in the LiDAR data and its spatial correlation with the surface projection of fault E strongly suggests that fault E has been active and formed the slope break causing the lineament sometime in the Quaternary.

As described in Subsection 2.5.1.2.4.2.3, fault E crosses a variety of features including the deposits of the Beaumont Formation, younger Pleistocene stream terrace deposits, and man-made features (i.e., FM 445, U.S. Highway 77, SR 239) (Figures 2.5.1-4 and 2.5.1-39). Field reconnaissance of the fault across these features was unable to provide any refinements on the timing of activity other than that movement has occurred since deposition of the Beaumont, similar to the constraints on timing of fault D activity. Topographic profiles of the fault along FM 445 derived from the LiDAR data reveal that the slope break associated with the fault has the same general characteristics as the non-degraded profiles of fault D (e.g., profile 4 and 8): a distinct inflection of the ground surface at the location of the lineament with the southeast side down. For fault E the relief across the tilted surface is approximately 4.9 feet (1.5 meters) over 980 feet (300 meters), or equivalently an increase in surface slope to approximately 0.29 degrees. As with fault D, the age of the Beaumont Formation provides the only constraint on the rate of deformation for fault E. Again, assuming the Beaumont was deposited between 350 ka and 100 ka, long-term deformation rates for fault E are between 1.7×10^{-4} inches per year and 5.9×10^{-4} inches per year. This vertical relief and implied deformation rates are similar to those observed for fault D. These similarities between the two faults could either be coincidental or may suggest that the mechanisms, rates, and characteristics of growth fault activity within the site area are fairly uniform.

2.5.1.2.4.2.4.1 Growth Fault Summary

Exelon conducted a comprehensive and multidisciplinary study of growth faults within the VCS site vicinity using publicly available information, proprietary commercial subsurface mapping, aerial photography, LiDAR data, commercial seismic reflection data, and field reconnaissance. This study

identified numerous growth faults within the site vicinity, a subset of which exhibit potential evidence for Quaternary deformation. Review of publicly available reports, proprietary subsurface mapping (Reference 2.5.1-123), and analysis of proprietary seismic reflection data demonstrated that the site area is underlain by the Vicksburg system of growth faults. Of the growth faults present in the subsurface of the site area, only faults GM-D and GM-E are associated with anomalous southeast-facing slope breaks that are potentially indicative of Quaternary growth fault activity. Key stratigraphic and structural relationships visible in seismic reflection data document that all other growth faults identified within the reflection data in the site area subsurface have not been active since Early Pliocene or earlier.

Growth fault E is within the site area, but is not visible in the seismic reflection data due to the limited extent of the reflection profiles. Growth fault E is over 2.6 miles from the site and does not have the potential to affect the site. The topographic lineament associated with fault D approaches within approximately 509 feet (155 meters) of the power block area. Seismic reflection profiles confirm that the subsurface trace of fault D offsets a probable Quaternary stratigraphic marker and projects directly updip to the lineament and corresponding southeast-facing slope break at the surface. Based on interpretation of the reflection data, shallow deformation associated with fault D is characterized by distributed monoclinal fault-propagation folding within an upward-widening triangular zone in the hanging wall. This potential zone of deformation was mapped out along the fault D lineament using topographic profiles from LiDAR data to define the extent and magnitude of the slope break. The closest approach of the zone of interpreted surface deformation associated with fault D to the power block area is approximately 509 feet (155 meters) (Figure 2.5.1-43). The long-term average deformation rate is extremely slow (approximately 1/18th of an inch of southeast-down motion every 100 years); it is unknown if deformation occurs continuously or episodically.

2.5.1.2.5 Site Area Geologic Hazard Evaluation

No geologic hazards have been identified within the VCS site area. No geologic units at the site are subject to dissolution. No deformation zones were encountered in the site investigation for VCS.

Volcanic activity typically is associated with subduction zones or “hot spots” in the earth’s mantle, neither of which are present within the VCS site region. Therefore, no volcanic activity is anticipated in the region.

The site area and site vicinity were investigated for evidence of prehistoric earthquakes in the form of paleoliquefaction and other anomalous geomorphic features. This investigation included aerial and ground reconnaissance within the site vicinity, analysis of stereo-paired aerial photos within the greater site area, analysis of LiDAR derived topography within the site vicinity, and reviews of published literature. This investigation focused on identifying any anomalous geomorphic feature that may be related to strong ground shaking, including sand blows and boils, lateral spreading, and

ground cracks. During this investigation particular emphasis was placed on areas with younger, Holocene deposits (i.e., valley fill deposits along the San Antonio and Guadalupe rivers) (Figures 2.5.1-4 and 2.5.1-23), but other Pleistocene deposits were examined as well. No evidence of prehistoric earthquakes or paleoliquefaction was observed within the site area or site vicinity during this investigation.

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 described in Subsection 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 be 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 Prior Earthquake Effects

Studies of the available outcrops examined during the VCS investigations have not indicated any evidence for prior earthquake activity that affected Pleistocene deposits.

2.5.1.2.6.4 Effects of Human Activities

Man's activities, including mineral mining, withdrawal of oil and gas often accompanied with associated saltwater, and pumping of groundwater from near surface aquifers often results in surface movements in the vicinity of the activity.

2.5.1.2.6.4.1 Site Vicinity Petroleum

Figure 2.5.1-51 presents locations of known oil and gas wells in southern Victoria County. Table 2.5.1-5 illustrates the active wells on the VCS site.

The mineral rights on VCS property are leased by four entities: Apache Corporation, Sanchez Oil & Gas, Texcom, and Americo. Figure 2.5.1-52 illustrates the approximate leases for these four corporations, as of December 2007 (Landcom Services, Inc, 2007).

There are approximately 130 oil and gas borings (personal communication, Apache Corporation) on the approximate 11,500 acres (46.5 km^2) that comprise the VCS site. Of these, 21 are considered

active. Of these active wells, seven produce from strata between 1700 and 1950 feet (518 and 594 meters) bgs, three from strata between 2200 and 3000 feet (671 and 914 meters) bgs, and eleven from strata between 3400 and 6400 feet (1036 and 1951 meters) bgs. All of these active wells produce gas and only three produce any oil. Production in the area is at least as old as the mid-1960s.

There is little published information about the producing formations in the Kay Creek Field. The perforated intervals and estimated stratigraphic column suggest that the production is in the Frio sands and equivalents. Burns et al. (Reference 2.5.1-259a) describe the nearby North McFaddin field as "…dominantly structurally controlled…" with reservoir terminations forming important stratigraphic traps. Individual reservoir units are typically 5 to 10 feet (1.5 to 3 meters) thick and are separated by "non-reservoir" facies. According to Burns et al. (Reference 2.5.1-259a) the reservoirs are laterally discontinuous lobate sheets, at least 5000 to 6000 feet (1524 to 1829 km) in width, and are typically oriented in a northeast-southwest direction. These thin sand reservoirs are difficult to evaluate using conventional logging devices resulting in many beds capable of production being passed by (Reference 2.5.1-259a). The Kay Creek Field producing from the same formations has similar trapping mechanisms and thin reservoirs.

None of the wells are located in the power block area, although a few are relatively close. Available records suggest that these production wells yield formation water as well as gas, and very little oil. Formation water pumping volume is relatively small, amounting to 0.25 gallons per minute (gpm) maximum per well, based on production figures. None of the wells are being pumped, all are natural flow, so there is very little stress placed on the reservoir.

Ratzlaff (Reference 2.5.1-259b, Figure 6) shows that there was subsidence related to the Kay Creek and McFaddin North fields, but the total over the period 1918-1973 was less than 0.5 feet (15.2 cm). Ratzlaff attributed this subsidence to withdrawal of oil and gas because there were very few water wells in the area in 1973, which is still the case in 2008. The Ratzlaff report is the most recent publicly available study on subsidence in Victoria County.

2.5.1.2.6.4.2 Groundwater and Subsidence

A detailed description of the groundwater is presented in Subsection 2.4.12.

The surface formation (Beaumont) is considered an aquitard by the Texas Water Development Board (TWDB), although there is a domestic water supply well on site for the ranch house and there are stock wells that are pumped occasionally to water the stock that are screened in the sand found at about 130 feet (40 meters) bgs. The deep observation wells installed as part of the site investigation are screened in this sand as well. The static water level in these wells is approximately 50 feet (15.24 meters) bgs. Construction dewatering wells are presently planned to be installed in this sand unit to lower the water table appropriately during construction.

Presently, the only water supply well(s) pumping from the Evangeline aquifer are those wells of the DuPont chemical plant, approximately 5 miles (8 km) northeast of the proposed power block area location. Water supply wells for the non-cooling water at the VCS site will be obtained from wells drilled into the Evangeline aquifer.

Subsidence often results from withdrawal of fluids such as petroleum and/or groundwater over a long period of time. The weight of the overlying sediments is supported, in part, by the fluids in the underlying sediments. The fluid loss in the sand aquifer impacts the overlying clays because the gradient from the clay to the aquifer increases as dewatering proceeds, so the clay is dewatered as well as the sand. Lithostatic pressure compacts the clay irrevocably so that even if groundwater levels are restored to pre-pumping levels, the clays will not rehydrate. Subsidence is common in many parts of the county with the common connection found in long-term pumping of groundwater from unconsolidated sands and gravels with clay interbeds. Ratzlaff (Reference 2.5.1-259b) states that land surface subsidence in southern Victoria County is less than 0.5 feet (15.24 cm) over a period of 55 years (1918-1973). This is the most recent, publicly available data.

Normal practice in calculating potential subsidence is that the compression (ΔB) is equal to the storage coefficient (S) multiplied by the change in hydraulic head (Δf) due to pumping (Edgar et al., 2000).

$$\Delta B = S \times \Delta f$$

The storage coefficient at VCS has been determined to be 0.0005 (dimensionless) through slug and pumping tests (Subsection 2.4.12), and the maximum drawdown due to construction dewatering is expected to be approximately 50 feet (15.2 meters) (Subsection 2.5.4.6.2) based on an assumed embedment depth of approximately 66 feet (20.1 meters). Therefore the anticipated subsidence at VCS due to construction dewatering is between 0.02 and 0.03 feet or about 0.3 inch (0.76 cm). Because there are other considerations, such as filling the cooling basin and stormwater infiltration, it is unlikely that the clay strata underlying the power block area and cooling basin will dry out and compact over the 4 years that construction dewatering is anticipated to be required.

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Table 2.5.1-1 (Sheet 1 of 2)
Growth Faults within Site Vicinity

Growth Fault Name	Source	Dip Direction	Depth of Projection (ft)[a]	Surface Expression[b]
DP-1	Dodge and Posey (Reference 2.5.1-238)	South	6700	Unknown
DP-2	Dodge and Posey (Reference 2.5.1-238)	South	6200	Unknown
DP-3*	Dodge and Posey (Reference 2.5.1-238)	North	6100	Unknown
DP-4	Dodge and Posey (Reference 2.5.1-238)	South	5900	Unknown
DP-5	Dodge and Posey (Reference 2.5.1-238)	South	4100	Unknown
DP-6	Dodge and Posey (Reference 2.5.1-238)	South	3300	Unknown
DP-7*	Dodge and Posey (Reference 2.5.1-238)	South	3300	Unknown
DP-8	Dodge and Posey (Reference 2.5.1-238)	South	2200	Unknown
DP-9	Dodge and Posey (Reference 2.5.1-238)	South	4100+	Unknown
G-1	Galloway et al. (Reference 2.5.1-239)	North	5800	Unknown
G-2	Galloway et al. (Reference 2.5.1-239)	South	5900	Unknown
G-3	Galloway et al. (Reference 2.5.1-239)	South	5900	Unknown
G-4	Galloway et al. (Reference 2.5.1-239)	North	5800	Unknown
G-5	Galloway et al. (Reference 2.5.1-239)	Unknown	5800	Unknown
G-6	Galloway et al. (Reference 2.5.1-239)	South	5600	Unknown
G-7	Galloway et al. (Reference 2.5.1-239)	Unknown	5600	Unknown
G-8	Galloway et al. (Reference 2.5.1-239)	North	5500	Unknown
G-9	Galloway et al. (Reference 2.5.1-239)	South	5400	Unknown
G-10	Galloway et al. (Reference 2.5.1-239)	South	5300	Unknown
G-11	Galloway et al. (Reference 2.5.1-239)	South	3800	Unknown
G-12	Galloway et al. (Reference 2.5.1-239)	South	5000+	Unknown
G-13	Galloway et al. (Reference 2.5.1-239)	South	3800	Unknown
G-14	Galloway et al. (Reference 2.5.1-239)	South	3200	Unknown
G-15	Galloway et al. (Reference 2.5.1-238)	South	3000	Unknown
G-16	Galloway et al. (Reference 2.5.1-239)	South	2300	Unknown
G-17	Galloway et al. (Reference 2.5.1-239)	South	1800	Unknown
G-18	Galloway et al. (Reference 2.5.1-239)	South	2800+	Unknown
G-19	Galloway et al. (Reference 2.5.1-239)	South	2700+	Unknown
GM-A	Geomap (Reference 2.5.1-123)	Southeast	4450-4870	Yes
GM-B	Geomap (Reference 2.5.1-123)	East	4240-4520	Yes
GM-C	Geomap (Reference 2.5.1-123)	South	4270-4500	No
GM-D	Geomap (Reference 2.5.1-123)	Southeast	4100-5300	Yes
GM-E	Geomap (Reference 2.5.1-123)	South	4240-4500	Yes
GM-F	Geomap (Reference 2.5.1-123)	East	4230-4660	No
GM-G	Geomap (Reference 2.5.1-123)	Southeast	4030-4570	Yes
GM-H	Geomap (Reference 2.5.1-123)	East	4180-4350	No
GM-I	Geomap (Reference 2.5.1-123)	East	4180-4700	Yes

Table 2.5.1-1 (Sheet 2 of 2)
Growth Faults within Site Vicinity

Growth Fault Name	Source	Dip Direction	Depth of Projection (ft)[(a)]	Surface Expression[(b)]
GM-J	Geomap (Reference 2.5.1-123)	East	5020-5430	No
GM-K	Geomap (Reference 2.5.1-123)	East	3990-4200	No
GM-L	Geomap (Reference 2.5.1-123)	East	3840-4500	Yes
GM-M	Geomap (Reference 2.5.1-123)	South	4100-4970	No
GM-N	Geomap (Reference 2.5.1-123)	East	3420-3880	No
GM-O	Geomap (Reference 2.5.1-123)	South	3510-3570	No
GM-P	Geomap (Reference 2.5.1-123)	South	3880-4140	No
GM-Q	Geomap (Reference 2.5.1-123	East	3530-3940	Yes
GM-R	Geomap (Reference 2.5.1-123)	South	3500-3600	No
GM-S	Geomap (Reference 2.5.1-123)	North	3320-3480	No
GM-T	Geomap (Reference 2.5.1-123)	South	2630-3620	Yes
GM-U	Geomap (Reference 2.5.1-123)	East	3000-3470	Yes
GM-V	Geomap (Reference 2.5.1-123)	South	2880-3300	Yes
GM-W	Geomap (Reference 2.5.1-123)	East	3120-3160	No
GM-Y	Geomap (Reference 2.5.1-123)	South	2450-2690	No
GM-Z	Geomap (Reference 2.5.1-123)	South	2180-5450	Yes
GM-AA	Geomap (Reference 2.5.1-123)	South	4800-5340	Yes
GM-AB	Geomap (Reference 2.5.1-123)	East	4620-4960	No
GM-AC	Geomap (Reference 2.5.1-123)	South	4940-5320	No
GM-AD	Geomap (Reference 2.5.1-123)	North	5540-6200	No
GM-AE	Geomap (Reference 2.5.1-123)	South	5370-6450	No
GM-AF	Geomap (Reference 2.5.1-123)	South	4620-4740	No
GM-AG	Geomap (Reference 2.5.1-123)	North	7110-7700	No
GM-AH	Geomap (Reference 2.5.1-123)	South	7140-8000	No

(a) Depth from which growth fault surface trace was projected. For Geomap (Reference 2.5.1-123) growth faults, the range of depths of the upper structural horizon within which the growth fault was identified is given. For all other growth faults, the depth is the approximate shallowest observed depth of the fault tip as determined from published cross sections.

(b) Whether or not surficial expression of the growth fault was observed.

* Indicates fault projected from published cross section was originally drawn to represented two individual faults.

+ Published cross section shows fault terminated at horizon with this depth.

Table 2.5.1-2
Summary of Meers Fault Characterizations from Existing Literature

	Ramelli and others	Madole	Crone and Luza	Swan and others
Age of events				
Young Holocene	Within several thousand yr.	1280 yr. BP (uncalibrated C-14 age)	1200 to 1300 cal. yr. BP	1300 to 1400 cal. yr. BP
Old Holocene	NA	NA	NA	2100 to 2900 cal. yr. BP
Pre-Holocene	NA	NA	Greater than 100,000 yr. BP	Greater than 200,000 to 500,000 yr. BP
Style of faulting	Left oblique slip with lateral to vertical ratio of 2:1 to 4:1	NA	Left oblique slip with lateral to vertical ratio of 1.6:1 to 3.3:1	Left oblique slip with lateral to vertical ratio of 1.3:1
Length of surface rupture	37 km	NA	26 to 37 km	26 to 37 km
Event displacement	NA	NA	3.1 to 5.9 m	Average 1.75 to 3 m; maximum 3.5 to 5.25 m
Slip rate				
Holocene	NA	NA	NA	1 to 5 mm/yr.
Quaternary	NA	NA	NA	10^{-4} to 10^{-5} mm/yr.
Clustered behavior	NA	NA	NA	Yes, cannot assume out of cluster
Event magnitude	Ms 6.75 to 7.25	NA	Approximately Ms 7	Ms 6.75 to 7.25

Preferred values identified by the study authors are given when available; otherwise the range of possible values from the study is presented. NA indicates that a study did not address a topic.

Table 2.5.1-3
Seismic Reflection Horizon Depths

Horizon	Approximate Time Range (seconds, from final stacked time profiles)	Approximate Depth Range (feet below sea level, from depth migration)
Horizon 1	1.357–1.838	5090–7320
Horizon 2	0.986–1.184	3575–4450
Horizon 3	0.357–0.533	1150–1790
Horizon 4	0.209–0.385	650–1340

Table 2.5.1-4
Updip Fault Terminations and Horizon Offsets Observed in Seismic Lines

Fault	Updip Fault Termination (s)	Updip Fault Termination (ft)	Offset of Horizon 1 (s)	Offset of Horizon 1 (ft)	Offset of Horizon 2 (s)	Offset of Horizon 2 (ft)	Offset of Horizon 3 (s)	Offset of Horizon 3 (ft)	Offset of Horizon 4 (s)	Offset of Horizon 4 (ft)
Line GSI										
GM-D	2.087	6965	—	—	—	—	—	—	—	—
GM-K	0.809	3054	0.097	30	0.046	0	—	—	—	—
GM-L	1.750	6653	—	—	—	—	—	—	—	—
SR-01	2.128	6297	—	—	—	—	—	—	—	—
SR-03	1.945	7670	—	—	—	—	—	—	—	—
SR-04	1.750	6854	—	—	—	—	—	—	—	—
SR-07	0.985	3937	0.018	149	0.011	37	—	—	—	—
Line GDI										
GM-D	0.307	1101	0.082	375	0.054	184	0.018	75	0.022	72
GM-K	0.872	3402	0.107	439	0.037	152	—	—	—	—
GM-L	1.442	5870	0	0	—	—	—	—	—	—
SR-01	2.416	7956	—	—	—	—	—	—	—	—
SR-03	1.688	6950	—	—	—	—	—	—	—	—
SR-04	1.812	7580	—	—	—	—	—	—	—	—
SR-05	1.326	5295	0.02	60	NA	—	—	—	—	—
Line TGS										
GM-A	1.408	5335	0.02	48	—	—	—	—	—	—
GM-D	0.482	1630	0.082	158	0.037	148	0.027	74	0.02	66
GM-E	0.831	5085	0.033	315	0.021	—	—	—	—	—
GM-K	0.973	4288	0.049	352	0.045	—	—	—	—	—
GM-L	3.138	9273	—	—	—	—	—	—	—	—
SR-01	2.017	7958	—	—	—	—	—	—	—	—
SR-03	2.510	10,107	—	—	—	—	—	—	—	—
SR-04	1.456	5864	0.01	27	—	—	—	—	—	—
SR-05	1.922	7513	—	—	—	—	—	—	—	—

Note: All offsets and updip terminations reported in seconds were determined from final stack time migrated seismic reflection profiles, and all offsets and updip terminations reported in feet were determined from final depth migrated profiles.

Table 2.5.1-5 (Sheet 1 of 2)
Active Wells Victoria County Station Site

Master I.D.	Lease Name	Well No.	Field Name	API Number	Operator Name	O/G	Upper Perf - Ft	Lower Perf - Ft	Prod St Yr - Mo	Prod End Yr - Mo	Peak Oil Bopd	Peak Gas Mcfd	Oil Cum Mbbl	Gas Cum Mmcf	Status	Zone	Lat/Long	TD FT
1	Mccan	T7	McFaddin Nor	4246934001	Texcom Opera	O	5,232	5,234	7-Jan	7-Nov	43	94	9	19	A	Frio 5350	286234309700478	6,224
2	Mcfaddin A	2	Kay Creek	46901563	Apache Corpo	G	2,230	2,250	65-Jan	7-Apr	—	722	—	3,517	A	Catahoula 22	285672409700952	7,403
3	Mcfaddin A	41	Kay Creek	46932738	Apache Corpo	G	2,202	2,210	91-June	6-Dec	—	175	—	123	A	Catahoula 22	285808109701950	6,512
4	Mcfaddin A	61	McFaddin Nor	46932959	Americo Ener	G	4,898	4,902	93-May	7-Nov	—	310	—	619	A	Frio 4800	286296209701687	5,260
5	Mcfaddin A	29	McFaddin	46932664	Apache Corpo	G	1,722	1,734	97-May	6-Dec	—	536	—	499	A	1800/ miocene	285835009701625	4,200
6	Mcfaddin A	18	McFaddin	46932019	Apache Corpo	G	1,738	1,773	97-June	7-Nov	—	406	—	717	A	1800/ miocene	285769209699868	5,500
7	Mcfaddin	69	McFaddin Nor	46933631	Americo Ener	G	4,831	4,835	2-Feb	7-Nov	—	230	—	171	A	4950/frio	285931209694612	6,300
8	Mcfaddin A	1	Kay Creek	46901545	Apache Corpo	G	1,932	1,941	3-Feb	7-Nov	—	105	—	69	A	1950/ miocene	285772409700360	5,300
9	Mcfaddin A	25F	Kay Creek	46932355	Apache Corpo	G	1,745	1,750	3-Mar	7-Nov	—	155	—	152	A	1750/ miocene	285648909700092	4,000
10	Mcfaddin J A	A44 C	McFaddin	46933719	Apache Corpo	G	3,448	3,570	3-Jul	7-Nov	4	336	—	258	A	3430/frio	285695409699703	3,751
11	Mcfaddin J A	A44T	McFaddin	46933719	Apache Corpo	G	3,448	3,570	3-Jul	7-Nov	3	388	—	295	A	3500/miocene	285695409699703	3,751
12	Mcfaddin A	45L	Kay Creek	46933769	Apache Corpo	G	2,837	3,696	3-Oct	7-Nov	—	429	—	484	A	Catahoula	285745709700827	3,835
13	Mcfaddin A	43	Kay Creek	46933588	Apache Corpo	G	3,413	3,427	5-Jan	7-Nov	—	284	—	160	A	3400/ miocene	285695009699498	3,900
14	Mcfaddin A-	46	Kay Creek	46933894	Apache Corpo	G	1,904	1,918	5-May	7-Nov	—	200	—	110	A	1950/ miocene	285783509700839	5,250
15	Mcfaddin A	17	Kay Creek	46932002	Apache Corpo	G	1,702	1,710	5-Aug	7-Nov	—	189	—	77	A	1750/ miocene	285681909700035	6,697
16	Mcfaddin A		Kay Creek	46932681	Apache Corpo	G	1,722	1,729	5-Nov	7-Nov	—	133	—	75	A	1750/ miocene	285624509700524	5,500
17	Mcfaddin J A A	41	Kay Creek	46932738	Apache Corpo	G	5,660	5,663	7-Jan	7-May	—	162	—	11	A	Frio 5600	285808109701950	6,512

Table 2.5.1-5 (Sheet 2 of 2)
Active Wells Victoria County Station Site

Master I.D.	Lease Name	Well No.	Field Name	API Number	Operator Name	O/G	Upper Perf - Ft	Lower Perf - Ft	Prod St Yr - Mo	Prod End Yr - Mo	Peak Oil Bopd	Peak Gas Mcfd	Oil Cum Mbbl	Gas Cum Mmcf	Status	Zone	Lat/Long	TD FT
18	Mccan	T4	McFaddin Nor	46933970	Texcom Opera	G	5,582	5,584	6-Dec	7-Nov	—	536	—	171	A	5600	286197709699359	6,320
19	Mccan	T3	McFaddin Nor	46933950	Texcom Opera	G	5,920	5,922	6-Dec	7-Nov	—	46	—	6	A	6350/frio	286222709699289	6,388
20	Mccan	T6	McFaddin Nor	46933987	Texcom Opera	G	—	—	6-Oct	7-Nov	—	207	—	50	A	5680/ unknown	286218709699742	—
21	Mccan	T5	McFaddin	46933980	Texcom Opera	G	5,404	5,406	6-Jun	7-Nov	—	236	—	106	A	Unknown/ gret	286203309698931	6,387