Attachment 1

COLA Revision 5 Highlighted Pages for Subsection 2.5.1 RAIs

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2.5.1 BASIC GEOLOGIC AND SEISMIC INFORMATION

PTN COL 2.5-1 The geologic and seismic information presented in this subsection provides a technical basis for evaluating potential geologic hazards at the Units 6 & 7 site. This subsection summarizes the current physiography, geomorphic processes, stratigraphy, tectonic features, stress regime, and the geologic history of the region within a 200-mile (320-kilometer)[1] radius of the site. This area is known as the site region (Figure 2.5.1-201). This subsection also provides similar information about the active plate boundary between North America and the Caribbean Plates located south of the site region. Both local and distant sources contribute to the seismic hazard at the site, including sources associated with the North America-Caribbean Plate boundary, whose closest approach is about 420 miles (675 kilometers) south of the Units 6 & 7 site (Subsections 2.5.1.1.2.2 and 2.5.1.1.2.3).

Subsection 2.5.1.1 describes the regional geology. Subsection 2.5.1.1.1 contains descriptions of the geologic and tectonic characteristics of the 200-mile radius site region. Information describing the geologic and seismic characteristics beyond the 200-mile radius site region is included in Subsection 2.5.1.1.2. The description of characteristics beyond the site region focuses on the North America-Caribbean Plate boundary, including potential seismic and tsunami sources in the Gulf of Mexico and Caribbean that may impact the Units 6 & 7 site. Subsection 2.5.1.2 describes the geologic and tectonic characteristics of the site vicinity, site area, and the site.

This subsection demonstrates compliance with the requirements of 10 CFR 100.23 (c). The geologic and seismic information was developed in accordance with NRC guidance documents RG 1.206 and RG 1.208.

The following paragraphs comprise a brief overview of the geologic evolution of the North America-Caribbean Plate boundary region and are intended to provide a context for more detailed discussions of available data presented in subsequent subsections.

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1. The norm applied throughout FSAR Subsections 2.5.1, 2.5.2, and 2.5.3 regarding the presentation of English or metric units of measure is to present measurements in the units cited in the reference first, then to provide the conversion in parentheses. In general, the conversion of units is an approximation that reflects the significant figures of the original units.

Overview of Tectonic Evolution

The Units 6 & 7 site is located on the southern portion of the North America Plate along the Atlantic passive margin, approximately 420 miles (675 kilometers) north of the North America-Caribbean Plate boundary (Figure 2.5.1-202). The North America Plate has been through multiple cycles of plate-tectonic spreading and convergence, resulting in the opening and closure of ocean basins. The cycles, with a period generally in the 100-million-year range, are known as Wilson cycles (References 201 and 202). The following provides a basic overview of the regional tectonic evolution of an expanded site region as a context for subsequent discussions of physiography, stratigraphy, structures, and seismicity of specific parts of the larger region.

Paleozoic Wilson Cycle

Most of Earth's landmass amalgamated into a supercontinent, Rodinia, between 1300 and 900 million years ago (Ma) (e.g., Reference 203) (Figure 2.5.1-203) (Subsection 2.5.1.1.3.1). Beginning about 750 Ma, rifting of Rodinia formed a vast proto-Atlantic ocean known as the Iapetus ocean that separated the paleocontinents of Laurentia (ancestral North America) and Gondwana (including ancestral Africa, South America, and Florida). The final closing of this ocean late in the Paleozoic (325 to 250 Ma) led to the formation of the supercontinent Pangea (Reference 204) (Figure 2.5.1-204). This ocean closure occurred in a series of three primary collisions known as the Taconic, Acadian, and Alleghany orogenies. These tectonic events led to the deformation of a belt of rocks that in present-day North America extends from Newfoundland to Alabama and as far west as Oklahoma and Texas. This belt of deformed rocks is known as the Appalachian-Ouachita orogen (Reference 205). The final deformation event related to the closure of the intervening oceans, the late Mississippian-Pennsylvanian (Carboniferous)-Permian Alleghany orogeny, occurred more than 600 miles (~1000 kilometers) north of the Units 6 & 7 site in the southern Appalachians (Figure 2.5.1-204). Nevertheless, the Alleghany orogeny is significant because it transferred some of the crust from Gondwana to North America, in particular, the basement of the Florida and Bahama Platforms (e.g., Reference 206) (Subsections 2.5.1.1.1.2.1 and 2.5.1.1.1.2.2). This material was welded against North America along the Suwannee suture, located in southern Georgia and Alabama (References 207, 208, and 201) (Figure 2.5.1-205).

Mesozoic Break-up of Pangea

The supercontinent Pangea remained a single landmass for almost 150 m.y., from late Paleozoic (about 325 Ma) until the late Triassic-Early Jurassic (225 to 175 Ma). The earliest vestiges of rifting included widespread thinning of the underlying continental basement and intrusion of volcanics into rift valleys during the Triassic. By the Middle Jurassic (170 Ma), the supercontinent was rifted into discrete landmasses (e.g., Reference 209) (Figure 2.5.1-206) (Subsection 2.5.1.1.3.2), with present day North America moving away from present day South America and Africa. By the Late Jurassic (150 Ma), the Atlantic Ocean and Gulf of Mexico were fully opened. Prior to this rifting, the Yucatan block was located adjacent to the Texas coast, while basement underlying southern Florida may have been located farther northwest between the African and North America Plates (Reference 210) (Figure 2.5.1-206) (Subsections 2.5.1.1.2.1, 2.5.1.1.2.2, and 2.5.1.1.2.3). These blocks were transported out of the Gulf of Mexico Basin during the Mesozoic (Subsection 2.5.1.1.3.2) via rifting and/or strike-slip faulting. During the Late Triassic to Middle Jurassic, rifting and sea floor spreading formed oceanic crust in the central Gulf of Mexico while leaving the Suwannee terrane, a piece of Florida basement, stranded adjacent to the Laurentian margin (Reference 211). The Early Jurassic volcanic rocks beneath southern Florida are interpreted as transitional crust developed during this rifting (Reference 212). Marine seismic reflection profiles indicate that subsurface normal faults were active during the Middle Jurassic to Cretaceous and are interpreted to reflect this regional rifting event (Reference 211). By the Early Cretaceous, shallow water evaporites and carbonates were deposited over most of the Florida, Yucatan, and Bahama Platforms (Figures 2.5.1-207 and 2.5.1-208) (see discussion of carbonate platforms in Subsection 2.5.1.1.1.1.1.2) while the Gulf of Mexico and Atlantic coasts developed as a passive continental margin (Reference 213).

Cretaceous to Tertiary Caribbean-North America Plate Convergence

The plate tectonic regime changed in the Cretaceous when the Caribbean Plate impinged on the southern North America Plate from the west (Reference 214). An alternative hypothesis suggests in situ formation for the Caribbean Plate (Reference 608). In the Cretaceous, portions of Cuba were part of the Caribbean Plate and experienced calc-alkaline volcanism associated with the Greater Antilles volcanic arc as the North America Plate subducted southwestward beneath the Caribbean Plate (Subsection 2.5.1.1.1.3.2.4). This island arc generally was active in the Cretaceous, but volcanism occurred in different places and at different times as multiple discrete volcanic events (References 216 and 845). During subduction and subsequent collision, many of the Jurassic- to

Cretaceous-age volcanic and sedimentary strata currently exposed in central and western Cuba underwent high-pressure, low-temperature metamorphism. This metamorphism and accompanying ophiolite emplacement occurred in the mid-Cretaceous to Paleogene (References 217 and 218).

Tertiary Transfer of Cuba to the North America Plate

During the early Eocene approximately 50 Ma, the Greater Antilles volcanic arc, with a westward-dipping subduction zone, approached the Bahama Platform and subduction ceased, resulting in the transfer of Cuba from the Caribbean Plate to the North America Plate (Reference 219) (Figure 2.5.1-206) (Subsection 2.5.1.1.3.3). In western Cuba, the collision between the Caribbean and North America Plates occurred in late Paleocene to early Eocene time (Reference 220). North- to northeast-directed thrusting and contraction related to this collision was widespread in the late Paleocene to middle Eocene along the Cuban fold-and-thrust belt. A southward dip in middle Eocene and older strata in the Straits of Florida also records the overthrusting of the North America Plate by the Greater Antilles Arc (Reference 221) (Figure 2.5.1-209). After the northwestern portions of Cuba sutured to the Bahama Platform along northwest-trending contractional structures, the plate boundary shifted southward as new northeast-trending strike-slip faults initiated to the southeast (e.g., References 529 and 639). This process continued until Oligocene time, when collision ended in Cuba and the North America-Caribbean Plate boundary migrated to its approximate current position south of Cuba, where the opening of the Cayman Trough led to a more east-west oriented strike-slip system (References 217 and 222). Based on geologic mapping, tectonic reconstructions, and seismic imagery, the site region has been a largely tectonically quiescent portion of the North America Plate since the late Tertiary.

The stratigraphic and tectonic data that support an understanding of the geologic history and ongoing geologic processes affecting the site region and active plate boundary between the North America and the Caribbean Plates are the subject of this subsection. The data were developed from a review of previous reports for the existing Turkey Point units that were updated to include recent concepts, current published geologic literature, and interpretations of data obtained as part of the surface and subsurface field investigations at the Units 6 & 7 site.

2.5.1.1 Regional Geology

This subsection provides information on the physiography, geomorphic processes, stratigraphy, tectonic structures and setting, and geologic history

within the site region and those features outside of the 200-mile (320-kilometer) radius site region pertinent to geologic hazards at the site. The stratigraphic and tectonic nomenclature for Florida used in this subsection is consistent with that used by the Florida Geological Survey (FGS) (Reference 223). The stratigraphic and tectonic nomenclature for the Gulf of Mexico, Bahama Platform, and the Caribbean used in this subsection generally is consistent with that used by the U.S. Geological Survey (USGS).

The regional geologic map (Figure 2.5.1-201) contains information on the geology within the site region and is discussed in Subsection 2.5.1.1.1. Regional geologic maps (Figures 2.5.1-202 and 2.5.1-210) of the larger Caribbean region contain information requested by the NRC and are discussed in further detail in Subsection 2.5.1.1.2.

While the tectonic and seismic history of the region is the basis for evaluating seismic hazards to the site, the climate history and history of geomorphic processes recorded in the stratigraphic record and current landforms are the basis for evaluating other potential geologic hazards to the site. The Deep Sea Drilling Program (DSDP) (1966 to 1983) (Reference 802), the Oceanic Drilling Program (ODP) (1983 to 2003) (Reference 803), and the present Integrated Ocean Drilling Program (IODP) have supplied large amounts of information used in reconstructing past climates and sedimentary events across the world's oceans. Much of the initial drilling work focused on the Gulf of Mexico, the Bahamas, and Caribbean regions. The result is a growing body of knowledge regarding the climatic, geomorphic, and tectonic history of the region around the Units 6 & 7 site. See Table 2.5.1-201 and Figure 2.5.1-211 for locations of DSDP and ODP sites referenced in this subsection.

A number of events and processes, many of which are interrelated, are common to the Gulf of Mexico, Florida, Bahamas, and Caribbean regions. These events and processes include the following:

- Karstification (Subsection 2.5.1.1.1.1.1.1)
- Carbonate platform development and demise (Subsection 2.5.1.1.1.1.1.2)
- Glacial cycles and associated sea-level fluctuations (Subsection 2.5.1.1.1.1.1.1)
- Megasedimentary events (Subsection 2.5.1.1.1.1.1.3)

- Oceanic and atmospheric reorganization events
- Large igneous province (LIP) magmatic events
- Interplate creation and intraplate modification of oceanic crust

Region-specific evidence for the first four of these events and processes are described in the subsections indicated above. The tectonic evolution of the North America-Caribbean plate boundary is a case study for the remaining three processes (oceanic and atmospheric reorganization and extinction events, LIP magmatic events, and the interplate creation and intraplate modification of oceanic crust). The significance of the remaining three processes (oceanic and atmospheric reorganization and extinction events, LIP magmatic events, and the interplate creation and intraplate modification of oceanic crust) on the tectonic evolution of the Gulf of Mexico, Florida, the Bahamas, and Caribbean regions is briefly discussed below.

Oceanic and Atmospheric Reorganization and Extinction Events

The recent scientific focus on climate change has resulted in the identification of a number of large ecosystem perturbations (References 226 and 227), many of which were long known to paleontologists as mass extinction events. Some of the ecosystem perturbations are gradual but most have proven to be surprisingly catastrophic. The ecosystem perturbations resulting in major extinctions can generally be tied to a major tectonic change (e.g., emergence of the Isthmus of Panama), a magmatic episode (e.g., the Caribbean LIP), or a bolide impact (e.g., Chicxulub asteroid). While the immediate cause may be localized, the consequences may be global and the change is called a "boundary event." Causes and effects of boundary events are still poorly understood (Reference 227) but several effects pertinent to the North America-Caribbean Plate boundary region have been identified. Figure 2.5.1-212 identifies seven parameters related to climate change over the past 600 m.y. Table 2.5.1-202 provides a summary of the major Mesozoic and Tertiary boundary events and how they affected the Gulf of Mexico, Florida, the Bahamas, and Caribbean regions.

A major effect of a boundary event is to alter oceanic current patterns. Changes in oceanic currents contribute to the geomorphic processes of marine erosion and deposition. Physical barriers (islands, submarine ridges, and shoals in the Caribbean and Gulf of Mexico) play an important role in controlling the characteristics and the quantity of deep ocean waters and currents that affect the physiographic features of the region. For example, the Bahama Platform, together

with Florida and the Greater Antilles Platforms, separate the Atlantic Ocean from the Gulf of Mexico and the Caribbean. The connection between these oceans is limited to the seaways between these three landmasses. As such, the seaways act as a "valve" for the Florida Current that flows between Florida and the Bahamas into the North Atlantic Ocean, where it converges with the smaller Antilles Current to form the Gulf Stream (Reference 228) (Figure 2.5.1-213). The Florida Current is a strong surface current, producing erosion and hiatuses on the Miami Terrace and other terraces of the Florida Straits (Figure 2.5.1-214) and deposition of thick drift deposits in the deeper portions and in the lee of the current (References 229, 230, 231, and 232).

The long-term current strength is primarily controlled by paleotectonic and climatic factors (References 229, 233, and 234). Short-term sea level falls intensify the currents in the seaways because of restriction of the channel area (Reference 235). The tectonic or isostatic uplift of barriers or the filling of channels, such as has occurred for the Isthmus of Panama, the Aves Ridge, and the Suwannee Channel, can significantly affect currents (e.g., close ocean current passages or open passages for new currents, narrow or widen channels to affect the strength of currents, and modify ocean-floor depth to affect density-driven currents) (Reference 236).

LIP Magmatic Events

LIPs consisting of tholeiitic basalt lava flows, sills and dikes have formed throughout the geologic history of the earth. In general, the LIPs are major bodies of extrusive igneous rock underlain by intrusive rock with crustal thickness ranging from 20 to 40 kilometers (12 to 25 miles). The crustal structure of LIPs is comprised of an extrusive upper crust and a lower crust characterized by high seismic velocities (7.0–7.6 kilometers/second or 4.3-4.7 miles/second) and are different from "normal" oceanic or continental crust. The possible compositions of lower crustal bodies on volcanic margins are gabbroic, strongly mafic, and ultramafic rocks. Some lower crustal bodies have been explained as magmatic underplating by accumulating mantle-derived material below the original crust (Reference 237). During the initial breakup of Pangea in the Late Triassic-early Jurassic, many LIPs formed as the result of rifting, basalt extrusion, and mafic intrusions. The most notable LIPs are the Central Atlantic Magnetic Province (CAMP) and the east coast margin igneous province (ECMIP). The total volume of lava for both the ECMIP, which is the source of the east coast magnetic anomaly (ECMA), and CAMP had exceeded 2.3 million cubic kilometers (0.55 million cubic miles) (Reference 239) to 2.7 million cubic kilometers (0.65 million cubic miles) (Reference 241). However, the two large igneous province (LIPs) that are

LDP-CS536

significant to the tectonic evolution and boundary events that affected the southeastern North America and Caribbean regions are the CAMP and the Caribbean large igneous province (CLIP). LDP-CS536

The CAMP is among the largest of the continental igneous provinces on earth, emplaced synchronously with, or just prior to, the Triassic-Jurassic boundary ca. 200 Ma. Magmatism associated with the CAMP occurred from 202 to 190 Ma. Intrusive CAMP magmatism began as early as 202 Ma. Extrusive activity initiated abruptly approximately 200 Ma, reaching peak volume and intensity around 199 Ma on the African margin. There were at least two phases over approximately 1.5 Ma, with magmatism commencing along the Africa-North American margins and slightly later along the South American margin (Reference 981). The extent of CAMP during the Mesozoic as described by McHone (Reference 239) was from "within Pangea from modern central Brazil northeastward approximately 5000 kilometers (3110 miles) across western Africa, Iberia, and northwestern France, and from Africa westward for 2500 kilometers (1550 miles) through eastern and southern North America as far as Texas and the Gulf of Mexico."

The precursor to the formation of the Central Atlantic Ocean (CAO) crust and the opening of the CAO was widespread groups of dike-fed fissure eruptions and flood basalts, which started during the Early to Middle Jurassic along sections of the central Atlantic rift (References 239, 241 and 977). The occurrence of CAMP magmatism and the volcanic rift margin adjacent to the newly forming oceanic crust along the eastern margin of North America is interpreted as subaerial volcanic flows or basalt wedges. This is also indicated by seaward-dipping seismic reflectors (SDRs) (Figure 2.5.1-284, Reference 239). LDP-CS564

Within the South Georgia rift basin in the southeastern United States, the continental flood basalts overlap the SDRs. The SDRs are approximately 25-kilometer- (16-mile-) thick basalt and plutonic wedges that are approximately 55 kilometers (34 miles) wide along approximately 2000 kilometers (1240 miles) of the eastern North American margin. These SDRs comprise the ECMA, which has been referred to by Holbrook and Kelemen (Reference 976) as the ECMIP (References 239 and 979). LDP-CS564

The ECMA marks the boundary between continental and ocean crust (Figure 2.5.1-266). It forms the seaward edge of the deep, sediment-filled basins and the landward edge of normal oceanic crust (Reference 973). The location of the ECMA segmented magnetic high (200–300 nT, positive magnetic anomaly) as seen in the total field magnetic anomaly, bathymetry, free-air gravity, isostatic anomaly and reduced-to-the pole (R-T-P) anomaly maps parallels the East Coast

margin from the Blake Spur fracture zone (BSFZ) to Nova Scotia (Figures 2.5.1-382 and 2.5.1-383) (References 972, 975, 978, and 982). Holbrook and Kelemen (Reference 976) were able to create a velocity model that showed lateral changes in deep crustal structure across the ECMA margin. Lower crust velocities average 6.8 kilometers/second (4.2 miles/second) at what were believed to be rifted crust areas, whereas the velocity (*Vp*) of units below the outer continental shelf were recorded as 7.5 kilometers/second (4.7 miles/second), decreasing to 7 kilometers/second (4.3 miles/second) with lateral shift to the oceanic crust. Holbrook and Kelemen (Reference 977) were able to determine that the high-velocity (*Vp*) lower crust and SDRs comprise a 100-kilometer- (62-mile-) wide, 25-kilometer- (16-mile-) thick oceanic-continental transition zone that they interpreted to be almost entirely mafic igneous material. This created an abrupt boundary between rifted continental crust and thick igneous crust, comprising only 20 kilometers (12 miles) of the margin. Holbrook and Kelemen (Reference 976) found that the Appalachian intracrustal reflectivity largely disappears across the boundary as velocity (*Vp*) increases from 5.9 kilometers/second (3.7 miles/second) to greater than 7 kilometers/second (4.3 miles/second), implying that the reflectivity is disrupted by massive intrusion and that very little (if any) continental crust can be found east of the SDRs. LDP-CS536 LDP-CS564 LDP-CS564

The Blake Spur Magnetic Anomaly (BSMA) is a linear anomaly located near the Blake Escarpment, east of the southern portion of the ECMA (ECMA and the BSMA are conjugate anomalies, and they both coincide with the ocean-continent boundary) and north of the Little Bahama Bank (Figure 2.5.1-266) (Reference 466). The BSMA represents points of initiation of sea floor spreading between North America and northwest Africa (Reference 692) and is interpreted as a continental margin modified by a jump in the spreading center (Reference 424) during the early Callovian (middle Jurassic). Ridge jumps are possibly caused by the reheating of the lithosphere as magma penetrates it to feed near-axis volcanism (Reference 980) and can be related to plate interactions as North America separated from Gondwana (Reference 466). It has been suggested that the BSMA is the result of an eastward jump of the spreading center away from the ECMA prior to 170 Ma (Reference 466).

The CLIP formed between 95 and 69 Ma (with two magmatic pulses at 92 to 88 Ma and 76 to 72 Ma) most likely within the Pacific Ocean and most likely associated first with the East Pacific Rise spreading center and, later, with the Galapagos hotspot (References 242, 606, and 243) (Figure 2.5.1-207). Other models favor an intra-Caribbean origin; however, radiolarites of equatorial Pacific faunal affinities in Puerto Rico and the Dominican Republic are older LDP-CS536

(Reference 820) than the proposed opening of the proto-Caribbean and thus require large-scale lateral transport into their present location, consistent with formation of the CLIP in the eastern Pacific (Reference 244). The volume of the CLIP may be as large as 106 kilometers^3 (Reference 245).

The environmental impacts of LIPs are global (e.g., References 246, 247, 248, 249, and 876). Continental flood basalts and oceanic plateaus are formed by widespread and voluminous basaltic eruptions that released enormous volumes of volatiles such as CO_2, S, Cl, and F (Reference 246). These extensive eruptions likely caused massive melting of hydrates and explosive methane release where magma intrudes carbon-rich sedimentary strata along rifting continental margins (Reference 250). Courtillot and Renne (Reference 251) propose a strong correlation between LIP ages and the ages of mass extinctions or periods of global oceanic anoxia. In particular, Courtillot and Renne (Reference 251) correlate LIP events with the four largest mass extinctions in the last 260 Ma. Kerr (Reference 247) proposes a correlation between formation of the Caribbean Plateau and the 93 Ma (Cenomanian-Turonian boundary) global oceanic anoxia event (associated with worldwide occurrence of extensive, organic-rich, black-shale horizons) and the 89 Ma (end Turonian) extinction event. LIPs also appear to be associated with the occurrence of Cretaceous marine red beds, probably a secondary result of extrusion of iron-rich lava on the seafloor (Reference 252).

Interplate Creation and Intraplate Modification of Oceanic Crust

The Caribbean Plate includes basins underlain by normal oceanic crust that transition to, or is faulted against, oceanic plateau (thickened) crust (Figure 2.5.1-215). The characteristics of the two types of crust influence their behavior as they impinge on boundary terranes, such as island arcs (Hispaniola and Puerto Rico) and the South and North America Plates (Reference 253).

The boundary between the North America and Caribbean Plates and the seismogenically active areas of the Caribbean is mostly submerged; thus, data from seismic reflection profiles have been a critical tool in elucidating the stratigraphic and tectonic structures of the region. Most of what is known about the stratigraphy and structure of the submerged areas, including the abyssal basins, spreading centers, trenches, platforms, and ridges, is based on seismic stratigraphy. The following paragraphs describe the basic seismic stratigraphy of: (a) normal oceanic crust created at interplate (divergent) spreading centers (e.g., oceanic crust of the abyssal Atlantic, Gulf of Mexico, Yucatan Basin, and proto-Caribbean Basins), and (b) thickened oceanic plateau crust created at

divergent spreading centers and then modified by intraplate magmatic upwelling (e.g., oceanic plateaus of the Venezuelan and Colombia Basins, Beata Ridge, and possibly along the Atlantic margin of North America from the Bahamas to Canada).

Seismic reflection data gathered off the U.S. Atlantic coast in conjunction with early DSDP studies resulted in models of a relatively thin (about 6 kilometers or 3.7 miles thick), two-layer oceanic crust produced at mid-ocean ridge spreading centers (References 254 and 253). Following established naming conventions (A and B for the Atlantic, A' and B' for the Pacific, and A" and B" for the Caribbean), the seismic profiles of oceanic crust in the Gulf of Mexico, Yucatan, and Caribbean Basins exhibit both A" and B" horizons. A" is the higher marker horizon, signaling the change from overlying Tertiary oozes and turbidites to indurated Cretaceous sediments. B" is the lower marker horizon, signaling the change from indurated pelagic sediments and turbidites to igneous rocks (Reference 255).

The B" horizon was identified in the Gulf of Mexico abyssal basins as having a hummocky, rough lower layer (rough B"), signaling the regional basement unconformity between Cretaceous sediments and the normal-thickness oceanic basement of basalt flows intruded by diabase dikes (Reference 255). The rough character of the B" horizon is attributed to valleys separating gently tilted fault blocks, possibly similar to the ridged surface exposed today on either side of the Cayman Ridge spreading center. The rough basalt horizon is sometimes overlain by a thin sequence of mid-Mesozoic pre- and syn-rift sediments (References 255 and 256). In contrast, some of the B" horizon identified in seismic profiles of southern Caribbean Basins exhibits a smooth surface (called smooth B"). In seismic profiles analyzed by Diebold et al. (Reference 255), the rough B" horizon in the in the Venezuelan Basin-Beata Ridge transect pre-dates the emplacement of smooth B" volcanic material (Reference 255). According to Diebold et al. (Reference 255), the smooth B" represents highly mobile basalt flows that were able to spread in thin fingers over the valleys and faults of rough B" crust. In their model, Diebold et al. (References 255 and 778) postulate that the transition from smooth B" to rough B" crust coincides with shoaling of the Moho from as deep as 20 kilometers (12.4 miles) to as shallow as 10 kilometers (6.2 miles). This shoaling reflects the transition from oceanic plateau crust (Figures 2.5.1-215 and 2.5.1-216) to normal oceanic crust.

Sediments below the A" horizon are generally well laminated. Above the A" horizon, the sediments are more lenticular and hummocky for an interval defined by an overlying reflector, eM. The eM reflector correlates with the boundary between Early Miocene radiolarian ooze and Early to Middle Miocene calcareous

ooze. Above the eM reflector, the sediments exhibit continuous stratification that drapes the seafloor topography underlain by thick and thin ocean crust (Reference 253).

2.5.1.1.1 Regional Geology within the Site Region

The area within a 200-mile radius of Units 6 & 7 site is defined as the site region. The site region includes the lower half of the Florida Peninsula, a small area of northern Cuba that is part of the Greater Antilles deformed belt, and extensive marine platform areas, including parts of the Atlantic continental margin with the Blake Plateau, Florida Platform, and Bahama Platform and its western continuation through the Straits of Florida (Figure 2.5.1-201). For purposes of organizing the discussion of the physiography, geomorphic processes, stratigraphy, tectonic features, and geologic history of the site region, this subsection includes information on the Florida Peninsula, the Florida Platform, and the Florida portion of the Atlantic Continental Shelf and Slope, including the Blake Plateau, the Bahama Platform, and the entire island of Cuba. Areas of the southernmost North America Plate and areas of the North America-Caribbean Plate boundary that are not included in Subsection 2.5.1.1.1 but may be relevant to the geologic hazards at the Units 6 & 7 site are described in Subsection 2.5.1.1.2.

2.5.1.1.1.1 Regional Physiography and Geomorphic Processes

Physiography is synonymous with the term geomorphology or the study of the nature and origin of landforms. Landforms acquire their character by geomorphic processes, such as physical and chemical weathering, erosion, mass transport, biologic impacts, isostasy, and glacial impacts.

2.5.1.1.1.1.1 Florida Peninsula and Continental Margin

The Florida Peninsula and continental margin consists of three physiographic areas (Figure 2.5.1-214):

- The Florida Peninsula
- The Florida Platform
- The Atlantic Continental Shelf and Slope

The Units 6 & 7 site is situated in the southern tip of the Florida Peninsula. The site region encompasses not only part of the Florida Peninsula, but also the

southern half of the Florida Platform. The Florida Peninsula physiographic subprovinces that fall within that 200-mile radius site region are described below in Subsection 2.5.1.1.1.1.1.1. The entire Florida Platform, including the Florida Escarpment (Figure 2.5.1-214), is discussed in Subsection 2.5.1.1.1.1.1.2. The unusually broad and shallow Atlantic Continental Shelf and Slope of Florida, including the Blake Plateau, while outside of the 200-mile site radius, are discussed in Subsection 2.5.1.1.1.1.1.3 because of their importance in protecting peninsular Florida from the effects of strong waves from the Atlantic Ocean.

2.5.1.1.1.1.1.1 Florida Peninsula Physiographic Subprovinces

The Units 6 & 7 site (Figure 2.5.1-214) is located within the Atlantic Coastal Plain physiographic province, comprising low-lying, gently rolling topography. The Atlantic Coastal Plain physiographic province is the portion of the Atlantic seaboard of the U.S. that extends eastward from the Fall Line to the coastline. The Fall Line is the physiographic boundary between the Coastal Plain province and the Piedmont province. In the eastern U.S. it parallels the north-northeast to south-southwest strike of the Appalachian orogen from Maine to Georgia, and then turns west into southern Georgia and Alabama, where it delineates the northeastern boundary of the Gulf of Mexico Coastal Plain.

The Florida Peninsula is a stable carbonate platform upon which surficial, dominantly Neogene and Quaternary terrigenous sediments have accumulated in an asymmetric fashion; relatively thick on the East Coast and generally thinning toward the west coast (Figure 2.5.1-224). The carbonate platform was prograding from the Mesozoic until the late Paleogene, when the early seaways connecting the Atlantic to the Gulf of Mexico (known through time as the Suwannee Channel, the Georgia Channel, and Gulf Trough) (Figure 2.5.1-218) were finally terminated (References 257, 258, and 234). The seaway system was a dynamic barrier that prevented siliciclastic sediments, eroded from the southern Appalachian Mountains, from inundating and smothering the carbonate production to the south. Eventually the seaways were filled and the Florida Peninsula was covered by siliciclastics transported from the north via fluvial and longshore processes. The combination of antecedent topography on the carbonate bank and multiple sea level fluctuations heavily influenced distribution of the dominantly quartz sand cover (Reference 259).

Little of the ancient geologic history is revealed in the geomorphology of Florida due to its low, flat lying topography. Paleoshorelines, scarps, and terraces; present-day river drainage patterns; and karst features indicate that a combination of sea level and groundwater fluctuations and coastal geomorphic processes

have strongly shaped Florida's current terrain. The fluctuations in sea level occurred through many cycles that extended through thousands to millions of years (References 562, 749, 877, and 298). Waves, winds, and currents have eroded sands from some locations and deposited the shifting sand along the beaches. The resulting principal geomorphic features in Florida are represented by the barrier island system of the Gulf Coastal Lowlands, the Reticulated Coastal Swamps of southwest Florida, and the swales and swamps of the Everglades (Figure 2.5.1-217).

At marine high stands, the central and southern areas of the Florida Peninsula were mostly submerged (Figure 2.5.1-219), and wave action eroded coast-parallel ridges and cliffs into emergent strata. During sea level lowstands, all of the Florida Peninsula and much of the Florida Platform were subaerially exposed (Figure 2.5.1-219), and the coast-parallel ridges and cliffs controlled surface water flow to further enhance coast-parallel valleys (Reference 265). Thus, central Florida is characterized by discontinuous highlands in the form of sub-parallel ridges separated by broad valleys. Southern Florida is characterized by a broad, flat, gently sloping, and poorly drained plain that is bounded on the east by the Atlantic Ridge. These low-lying lands of southern Florida do not reach elevations as high as the Atlantic Ridge because they were not exposed to high-energy processes of the Atlantic shore, which carried sand southward from northern sources to build beaches. The high-energy environment of the Atlantic shoreline decreases as it meets the Florida Current in the Florida Straits. There, nutrients carried by the warm Florida Current feed the reef-building corals of the Florida Keys and the current redistributes sand offshore as sand ridges (Reference 265).

Eight marine terraces (Hazlehurst [formerly Brandywine], Coharie, Sunderland, Wicomico, Penholoway, Talbot, Pamlico, and Silver Bluff) are mapped across the Florida landscape (Figure 2.5.1-220) (Table 2.5.1-203). Ward et al. (Reference 260) correlate nine marine terraces found in South Carolina (laterally equivalent to those in Florida) with stranded shorelines in Australia and Alaska. The Ward et al. (Reference 260) comparison suggests four high sea levels occurred in the last 0.25 Ma and two others about 0.4 Ma. Other stages of high sea level are dated at 0.76, 1.43, 1.66, and 1.98 m.y. The authors calculated altimetric ages from correlation of high sea level data with deep-sea core stages to provide estimated ages for the Florida/South Carolina high stands (Table 2.5.1-202). The ages of the South Carolina terraces generally correlate with global sea level high stands proposed by Ward et al. (Reference 260). There is some debate regarding the appropriate methodology for age estimation of the

Florida marine terraces based on various radiometric and other age-dating techniques (Reference 261). These marine terraces are the former bottoms of shallow seas and are bounded along their inner margin by shoreline features such as relict beach ridges, swales or inner lagoons, seaward facing wave-cut scarps or sea cliffs, and offshore and bay bars (Reference 261).

The marine terraces in Table 2.5.1-203 were once thought to be the direct result of sea level fluctuations through the last glacial cycles but are now understood to be a result of complex interactions between sea level oscillation, subaerial exposure, a precipitation karstification function, and isostatic uplift (References 262 and 927). Since reefs form in a shallow marine environment, the organisms that comprise the Key Largo and Miami limestones preserve the record of Pleistocene sea level changes. These limestones in some places have been subaerially exposed. Investigators (References 928 through 933) studied the aforementioned limestones to understand the Atlantic-Caribbean sea level changes. The record of Pleistocene sea level changes is preserved in the marine sequences Q1 through Q5, from oldest to youngest, which correlate to marine isotope stages MIS 11, 9, 7, and 5e (Table 2.5.1-209) (Reference 928). The marine sequences are defined as a stratigraphic sequence of marine strata that represents a population of benthic organisms. Marine isotope stages (MIS) are alternating warm and cool periods in the earth's paleoclimate history, inferred from oxygen isotope data reflecting changes in temperature.

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Adams et al. (Reference 927) generated a model that calculates lithospheric uplift as a result of a precipitation-driven karstification function (decrease of bulk crustal density) and variations in subaerial exposure of a carbonate platform (i.e., Florida) due to oscillating sea level. The authors applied this model to north-central Florida to estimate the ages of beach ridges and depositional coastal terraces. The ages were based on the most recent estimates of sea level history since the Pliocene. The modeled ages of sea level highstands were then compared to the elevations of uplifted beach ridges and coastal terraces to evaluate plausible ages for deposition of the observed coastal geomorphic features (Reference 927). The geomorphic features were the Trail Ridge, the Penholoway Terrace, and the Talbot Terrace (Figure 2.5.1-370). The model produced the following ages for the three geomorphic features near the north Florida-southeastern Georgia Atlantic coast: (1) Trail Ridge approximately 1.44 m.y., (2) Penholoway Terrace approximately 408 k.y., and (3) Talbot Terrace approximately 120 k.y. (Table 2.5.1-203, Reference 927).

Hickey et al. (Reference 928) analyzed the $^{234}U/^{238}U$ ages of cores recovered at Grossman Ridge Rock Reef and Joe Ree Rock Reef in the Florida Everglades

and revealed additional subaerial-exposure surfaces that are used to delineate subdivisions within the five marine sequences of the Pleistocene carbonates of south Florida (Figure 2.5.1-371). These five marine sequences with Hickey et al. (Reference 928) subdivisions in parentheses are as follows: Q1 (Q1a–Q1b), Q2 (Q2a–Q2d), and Q4 (Q4a–Q4b) and Q5 (Q5e) (Figure 2.5.1-372). These subdivisions delineated by Hickey et al. (Reference 928) within units Q1 through Q5 preserve evidence of at least ten separate sea level highstands, rather than five as indicated by previous studies (Reference 928). Q5e is the youngest Pleistocene subaerial exposure surface of the Florida Keys (Figure 2.5.1-372). The fossil content and the $^{234}U/^{238}U$ radiometric ages indicate that this morphostratigraphic unit was deposited during the peak sea level of the last interglacial marine isotope substage 5e (MIS 5e). Uranium-series ages on corals from this unit from Windley Key, Upper Matecumbe Key and Key Largo range from 130 to 121 ka after corrections for calculated high initial $^{234}U/^{238}U$ content (Reference 928). A Q4a sample from Point Pleasant near the island of Key Largo has a best estimate age range of 340–300 ka, which falls into the early part of marine-isotope stage 9 (MIS 9) (Reference 928). The age of a Q4b coral sample recovered from a spoil pile in a quarry within unit Q4 on Long Key, southwest of Key Largo is approximately 235 ka (corrected for calculated high initial $^{234}U/^{238}U$). This is consistent with the early part of MIS 7. Hickey et al. (Reference 928) concludes that the Q1 through Q3 units predate MIS 9 and that their preferred interpretation is that Q3 was deposited during MIS 11 and that Q2 and Q1 represent pre-MIS 11 interglacial intervals (Figure 2.5.1-373) (Reference 928). Lastly, Muhs et al. (Reference 929) obtained ages of corals from Windley Key, the island of Key Largo, and from Long Key to Spanish Harbor Keys (middle Florida Keys) using Uranium-series dating. $^{234}U/^{238}U$ age dates are as follows: approximately 114 to 122 ka (Windley Key), approximately 120 to 123 ka (island of Key Largo), and approximately 114 ka (Long Key to Spanish Harbor Keys) (Reference 929). Thus the ages obtained by Muhs et al. (References 929 and 933) correlate to MIS 5e and are consistent with the dates obtained by Hickey et al. (Figures 2.5.1-373 and 2.5.1-377 and Table 2.5.1-209) (Reference 928).

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Although no post-Stage 5e dates have been reported from corals recovered from pits or cores from the exposed Florida Keys, several younger dates have been obtained from submerged corals recovered from the shelf to the east of the Florida Keys (References 930 and 931). These have been assigned to marine-isotope substages 5c, 5b, and 5a. These post-Q5e interglacial highstands were not high enough to flood the south Florida inner platform (Reference 928). Multer et al. (Reference 930) obtained dates for the Key Largo Limestone using thermal ionization mass-spectrometric (TIMS) uranium-thorium (U-Th) dating. The

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dates from these rocks, 112.4 to 77.8 ka, correspond to the marine-isotope substages 5c and 5a (MIS 5c and MIS 5a). These rocks were found under the shelf edge at Conch Reef, Looe Key, under Carysfort Light area and at the shelf edge near Molasses Reef (Figures 2.5.1-374, 2.5.1-375, and 2.5.1-376) (Reference 930). Toscano and Lundberg (Reference 932) also used TIMS U-Th dating and obtained dates of 7.7 +/- 0.7 ka and 8.6 +/-0.1 ka (basal Holocene) above the unconformity on the shelf edge (core SKSE) at Sand Key outlier reef (lower Keys) (Figures 2.5.1-374 and 2.5.1-375) (Reference 932). Below the unconformity, Toscano and Lundberg (Reference 931) obtained TIMS U-Th dates on corals from Sand Key outlier reef and Carysfort Light area of 86.2 +/-1.01 and 80.9 +/-1.7 ka (Figures 2.5.1-376 and 2.5.1-377). PTN RAI 02.05.01-4

The variations in orientation of shoreline features indicate variations in eustatic adjustment across the Florida Platform and Peninsula (Reference 262). Karstification effectively accomplishes the equivalent of isostatic compensation by decreasing the crustal mass within a vertical column of lithosphere. The rate of karstification (void space creation or equivalent surface lowering rate) within the north Florida Platform is about 3.5 times that of previous estimates (1 meter/11.2 thousand years [k.y.] vs. 1 meter/38 k.y.), and uplift rate is about two times higher than previously thought (0.047 millimeters/year vs. 0.024 millimeters/year) (Reference 262). This process has implications for landscape evolution in other carbonate settings and may play an underappreciated role within the global carbon cycle (Reference 262).

Karstification is the process created by chemical dissolution when weakly acidic groundwater circulates through soluble rock (Figure 2.5.1-221). Carbon dioxide from the atmosphere is fixed or converted in the soil horizon to an aqueous state, where it combines with rainwater to form carbonic acid, which readily dissolves carbonate rock. Root and microbial respiration in the soil further elevates carbon dioxide partial pressure, increasing acidity (lowering pH). In tropical and subtropical regions such as Florida, abundant vegetation, high rainfall, and high atmospheric CO_2 values favor the rapid dissolution of the preexisting limestone.

Carbonate Dissolution at Freshwater/Brine Interfaces

The freshwater/saltwater interface is defined as the location where seawater intrudes into a coastal aquifer and mixes with the discharging freshwater in a zone of mixed groundwater composition. The chemical reactivity of the mixing zone stems from the marked undersaturation with respect to carbonate minerals that develops from mixing a carbonate saturated freshwater with near surface seawater in a system closed with respect to carbon dioxide (Reference 945). PTN RAI 02.05.01-1 PTN RAI 02.05.01-17

Dissolution occurs when the two fluids of different salinities combine, even though both fluids are initially saturated with calcium carbonate (Reference 951). Because seawater saturated with calcium carbonate contains far less calcium carbonate than fresh groundwater saturated with calcium carbonate, the combined fluids become undersaturated with respect to calcium carbonate. This condition promotes dissolution of carbonate rocks. PTN RAI 02.05.01-17

Dissolution of limestone generally occurs where fresh, weakly acidic groundwater circulates through soluble carbonate rock or within zones of mixing fresh and seawater (References 263 and 965). The freshwater/saltwater interface within the Biscayne Aquifer is located approximately 9.6 kilometers (6 miles) inland from the site (Figure 2.4.12-207), groundwater at the site is saline (Table 2.4.12-210 and 2.4.12-211) and the long-term sea level rise trend at Miami Beach, Florida, as estimated based on data from 1931 to 1981, is 0.2 meter (0.78 foot) per century (Subsection 2.4.5). Therefore, the site is not a location of fresh groundwater discharge or mixing of fresh and saltwater, and the mechanism necessary to form large solution cavities does not appear to be active on or near the site. LDP-CS564 LDP-CS564

Rising sea level will increase the ocean hydrostatic head and tend to force intrusion of the freshwater/saltwater interface further inland and away from the site. Therefore, the mixing zone mechanism necessary to increase the potential for carbonate dissolution and formation of large solution cavities on or near the site will not exist. Collapse of solution cavities is generally associated with lowering of groundwater levels and withdrawal of buoyant support. A rising sea level will counter this effect.

Conversely, any potential lowering of sea level would tend to move the freshwater/saltwater interface seaward and toward the site. However, the long-term sea level rise trend at Miami Beach, Florida, as estimated based on data from 1931 to 1981, is 0.2 meter (0.78 foot) per century (Subsection 2.4.5), and sea level has been rising throughout the current interglacial stage of the Holocene. A significant lowering of sea level is not likely to occur until a future advance of continental glaciation, which is not likely to occur within the operating lifetime of Turkey Point Units 6 & 7. The magnitude of sea level lowering and the corresponding time necessary to move the interface to a location within the area of the site is not likely to occur within the operating lifetime of Turkey Point Units 6 & 7 (Subsection 2.4.5). Therefore, increased carbonate dissolution or formation of large solution cavities on or near the site due to a lowering of sea level is not likely to occur during construction or operation of the plant.

Several researchers (References 946, 947, 948, 949, and 950) indicate that carbonate dissolution associated with the mixing of freshwater and saltwater occurs predominantly at groundwater discharge sites or seafloor discharge zones. Mixing can also occur in surface water. The dissolution mechanisms are point source discharge and submarine groundwater discharge (SGD). PTN RAI 02.05.01-17

Point Source Discharge

Point source discharge is a concentrated flow of spatially constricted fresh surface water into a saltwater body. The discharge can affect the local water chemistry equilibrium with the potential to alter the rate of dissolution or deposition of carbonates within the mixing zone in its vicinity. An example of a point source discharge is surface water released to Biscayne Bay through drainage canal discharge.

The freshwater/saltwater interface at the base of the Biscayne Aquifer is located approximately 9.6 kilometers (6 miles) inland of Turkey Point Units 6 & 7, as shown on Figure 2.4.12-207. The migration of saltwater inland along the base of the aquifer occurs along the entire coastal zone and is the result of the aquifer's high permeability, the lowering of inland groundwater levels from groundwater pumping and surface drainage, and rising sea level (Subsection 2.4.5). As shown on Figure 2.4.12-207, the position of the freshwater/saltwater interface was relatively consistent between 1984 and 1995 and, in fact, provisional data from the USGS (Reference 960) showing the 2008 freshwater/saltwater interface in southeast Florida indicates a similar pattern.

Under natural conditions and before anthropogenic activity (e.g. construction of canals and enlargement of the Miami River) (References 267, 722, 955, 961, 962, and 963), the freshwater/saltwater interface in southeastern Florida was close to the coastline and freshwater discharged from springs on the floor of Biscayne Bay. In the late nineteenth century construction of flood control levees, drainage canals, and urbanization changed the position of the freshwater/saltwater interface. Canals were first dug through the Everglades to drain water from the area south of Lake Okeechobee to enable agriculture to develop (Reference 267). These canals roughly follow the transverse glades (i.e., narrow valleys or channels in which the soils (marl and sand) and vegetation are similar to those in the Everglades). By the late 1920s, major canals were constructed and rivers in the transverse glades were modified to connect Lake Okeechobee with the Gulf of Mexico and Atlantic Ocean (Figure 2.4.12-207) (References 267 and 964). In the 1930s, the government initiated flood control measures including levee construction and drainage channel modification. By the 1970s, gated control

structures were installed at the coastal end of the primary drainage canals to discharge excess water during the wet season and impede the landward movement of saltwater during the dry season. The final phase of canal development of the Everglades-South Dade conveyance system in the 1980s was constructed to meet agricultural water supply needs, control flooding, and mitigate saltwater intrusion (Reference 267).

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The increased fresh surface water discharge from the Everglades to Biscayne Bay and the Atlantic Ocean through the drainage canals and increased pumping from the freshwater aquifer has probably had an impact on coastal groundwater hydrology by contributing to inland migration of the freshwater/saltwater interface as shown in Figure 2.4.12-207. Point source discharge also may have increased the potential for dissolution of carbonate rocks in the immediate vicinity of the drainage canal outfalls. However, stratification of freshwater near the surface of the canal outfalls may limit carbonate dissolution to the near surface.

Outfalls of drainage canals closest to the site are the Model Land Canal (C107) outfall near the southeast corner of the Turkey Point cooling water canals, approximately 8 kilometers (5.0 miles) south of the site, and the Florida City Canal outfall, approximately 1.9 kilometers (1.2 miles) north of the site (Figure 2.4.1-203). Because of their distance from the site, the possible effect of freshwater stratification near the outfalls, and the effects of variable discharge from the outfalls related to operation of their control structures, variable rainfall, tidal fluctuations, and hurricanes, neither outfall is likely to induce formation of cavernous limestone with the potential for collapse at the site.

Submarine Groundwater Discharge

SGD is defined as the "phenomenon that forces groundwater to flow from beneath the seafloor into the overlying ocean regardless of its composition, whether freshwater, recirculated seawater, or a combination of both" (References 946 and 952). SGD can be subdivided into "shoreline flow" (i.e., fresh groundwater flow through an aquifer to the nearshore ocean that is driven by an inland hydraulic head) and "deep pore water upwelling" (DPU) (i.e., fresh groundwater flow beyond the shoreline on the continental shelf through deeper confined permeable shelf sediments and rocks, driven by buoyancy and pressure gradients (Reference 946). Reference 953 states "SGD per unit length of coastline could be very significant as a discharge process, due to the length of coastline where SGD occurs; whether or not rivers are present." The extent of SGD or saltwater intrusion at a given location is an issue of balance between hydraulic and density gradients in groundwater and seawater along a transect perpendicular to the

shoreline (Reference 953). The two possible modes of submarine groundwater discharge, shoreline flow and deep pore water upwelling, are discussed below.

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Shoreline Flow

As stated above, shoreline flow to the sea occurs when fresh groundwater flow through an aquifer is driven by an inland hydraulic head. As the shoreline flow nears the sea, it encounters the saltwater that has infiltrated from the ocean. The density of freshwater is lower than that of saltwater and therefore it tends to flow above the saltwater. The freshwater flowing toward the sea encounters an irregular interface where mixing of the fluids is driven by diffusion and dispersion enhanced by ocean forces (i.e., tidal pumping, wave setup, storms, buoyancy, and thermal gradients). This freshwater/saltwater circulation pattern and mixing is similar to that in surface estuaries, leading to the term subterranean estuaries. Tidal forces operating in a mixed medium (i.e., bedrock) may enhance dispersion along the freshwater/saltwater interface and the permeability and preferential flow paths may be changed by chemical reactions within the aquifer. Precipitation of solids can restrict or block some paths, while dissolution will enlarge existing paths or open new ones (Reference 946). Examples of shoreline flow are:

- Freshwater springs along Biscayne Bay (approximately 25 kilometers [16 miles] northeast of the site)
- Cave development along the Atlantic Coastal Ridge (approximately 17 kilometers [11 miles] north-northeast of the site)
- Submarine paleokarst sinkhole in the Key Largo National Marine Sanctuary (approximately 13 kilometers [8 miles] south of the site)
- Blue holes of the Bahamas in eastern South Andros Island (approximately 190 kilometers [120 miles] southeast of the site)
- Karst development on emergent carbonate islands in the Bahamas (approximately 320 kilometers [200 miles] southeast of the site)
- Karst development on the Yucatan Peninsula, Quintana Roo, Mexico (approximately 560 kilometers [350 miles] southwest of the site)

Freshwater Springs along Biscayne Bay

Fresh groundwater had discharged along the Atlantic Coastal Ridge shoreline and offshore as submarine springs before the drainage canals were built and before

substantial lowering of surface water and groundwater levels in southeast Florida. The groundwater flow conduits still exist and are dissolution features within the Biscayne Aquifer. Springs reportedly discharged near shore as freshwater boils in the shallow waters of Biscayne Bay (References 721, 954, and 955). In the late 1800s and early 1900s, springs within the Biscayne Aquifer provided a source of freshwater for sailing ships in Biscayne Bay. Parks (Reference 956) describes a freshwater spring off Coconut Grove (south of Miami) that was first documented in 1838 by Dr. Jacob Rhett Motte. Later a pump and platform was constructed to enable dories to tie up while filling wooden kegs with freshwater. This spring was marked as "freshwater" on Coast and Geodetic Survey Navigation Chart No. 166 (1896) (Reference 954). However, while many shoreline springs still exist in the bay and were formed by freshwater dissolution, salinity levels of 8 to 31 g/L (8 to 31 parts per thousand) indicate that the water quality is beyond the range for drinking water and, therefore, these groundwater discharges are no longer freshwater springs. The discharge rates from these springs are low, most likely due to blockage by sand in the conduits (Reference 954). The diminished discharge and water quality in the shoreline springs suggests that the propensity for further development of dissolution features by shoreline flow in nearshore areas of southeast Florida, including the Turkey Point Units 6 & 7 site, is diminished compared to the prevailing conditions prior to redistribution of the groundwater flow.

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Langevin (Reference 948) suggested that the drainage canals are the present focal points for groundwater discharge into Biscayne Bay, intercepting fresh groundwater that would have discharged directly to the bay. Field observations by Langevin (Reference 948) suggest that Biscayne Bay has changed from a system controlled by widespread and continuous submarine discharge and overland sheet flow to one controlled by episodic releases of surface water at the mouths of drainage canals. The canals and pumping from the freshwater aquifer have lowered the water table and, thus, submarine groundwater discharge has decreased. The Turkey Point Units 6 & 7 groundwater model is consistent with Langevin's model (Reference 948).

Cave Development along the Atlantic Coastal Ridge

Today, there are no freshwater springs discharging into Biscayne Bay. However, what do remain are the currently dry channels of past groundwater flow that were formed by freshwater dissolution. These are the caves of Miami-Dade County (Reference 955) (Figure 2.5.1-354). The 19 air-filled caves and one water-filled cave in Miami-Dade County found by Alan Cressler (Reference 955) are located along the eastern and western flanks of the Atlantic Coastal Ridge. Most caves of

southeastern Florida occur on or along the eastern flanks of the ancient Atlantic Coastal Ridge, or along the edges of transverse glades that cut through the Atlantic Coastal Ridge. According to Cressler's (Reference 955) field observations and descriptions, the caves within the Pleistocene limestones fall into four categories: (1) at least one is oriented along fractures, (2) some caves are concentrated along the margins of transverse glades, (3) some caves are composed of stratiform lateral passages, and (4) some caves have entrances along the margins of cave-roof collapse. Most of the caves discovered by Cressler (Reference 955) fall into the second category. The caves are concentrated along the margins of transverse glades. Cressler (Reference 955) hypothesized that slightly acidic water from the Everglades could be a potent agent for dissolving limestone and forming the caves in the transverse glades in the Miami Limestone.

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The most extensive karst development in Miami-Dade County lies within the boundaries of the Deering Estate County Park and Preserve (Reference 955) on the eastern flank of the Atlantic Coastal Ridge. The Deering Estate County Park and Preserve is located approximately 17.6 kilometers (11 miles) north-northeast of the site. Of the 19 air-filled caves identified by Cressler (Reference 955), seven are located in the Deering Estate. Observations in the Deering Estate indicate that variations in Pleistocene stratigraphy (i.e., Miami Limestone) may have played an important role in the origin of many small caves, including the 36.6-meter (120-foot)-long Fat Sleeper Cave. At Deering Estate, cave passages are commonly low, wide and sandwiched between crossbeds of oolitic limestone. These stratiform passages seem confined to a zone of rock with many centimeter-scale vugs related to complex burrow systems. It is hypothesized that the burrow-related porosity provided early preferential pathways for groundwater flow and concentrated dissolution. In some caves, solution pipes penetrate the upper cross-bedded limestone and connect to the land surface (References 954 and 955).

One of the most well known caves in Miami-Dade County, Palma Vista Cave, is located on Long Pine Key in the Everglades National Park (Figure 2.5.1-355). The entrance of the Palma Vista Cave probably formed by the collapse of a thin roof that spanned a stratiform cave (Reference 954). The speleothems in the cave that are underwater are important because their presence implies that they developed in Palma Vista Cave during a previous, extended dry period (i.e., sea level low stand). Such a condition would have existed when sea levels were much lower, such as the period between approximately 80,000 and 6,000 years ago (Reference 957).

The Atlantic Coastal Ridge caves formed by solution enlargement of sedimentary structures in the Miami Limestone as groundwater entered the freshwater/saltwater mixing zone and discharged as shoreline flow on the margin of the coastal ridge. The freshwater/saltwater interface is approximately 9.6 kilometers (6 miles) inland from the coast (Figure 2.4.12-207), shoreline flow at the Turkey Point Units 6 & 7 site is brackish to saline (Table 2.4.12-210 and 2.4.12-211), and the long-term sea level rise trend at Miami Beach, Florida, as estimated based on data from 1931 to 1981, is 0.2 meter (0.78 foot) per century (Subsection 2.4.5). Therefore, the mixing-zone process that formed the caves along the flanks of the Atlantic Coastal Ridge is not likely to be active currently in formation of cavernous limestone with the potential for collapse in the area of the site.

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Submarine Paleokarst Sinkhole in the Key Largo National Marine Sanctuary

A large submarine, sediment-filled paleosinkhole in the Key Largo National Marine Sanctuary off Key Largo, Florida (Figure 2.5.1-360) is described as having a 600-meter (1970-foot) diameter with a depth likely to exceed 100 meters (328 feet) (Reference 959). The Key Largo submarine paleosinkhole lies beneath 5–7 meters (16-23 feet) of water, and is bordered by Holocene reefs to the east and marine grass and carbonate sand to the west. Shinn et al. (Reference 959) jet probed to 54.5 meters (179 feet) and did not reach the bottom of the sinkhole. Patches of marine grass grow on the carbonate sands in the circular feature, but corals are absent (Reference 959). The sediments as observed from the sediment cores consist of monotonous gray aragonite mud visually lacking sedimentary laminations and fossils. The composition of the sediment as analyzed by X-ray diffraction is approximately 95 percent aragonite and 5 percent calcite. The oldest ^{14}C age (from the bottom of the jet probe sampler) is 5650 +/-90 years before present. The youngest ^{14}C age (just below the overlying carbonate sand cap) is 3260 +/-60 years before present. The high percentage of aragonite and near absence of low-magnesium calcite indicate the sediment is of marine origin and the ^{14}C dates indicate rapid deposition (Reference 959).

Shinn et al. (Reference 959) postulate that the Key Largo sinkhole is a cenote that formed during the Pleistocene. Fluctuations in sea level related to advance and retreat of continental glaciers raised and lowered the fresh groundwater/seawater shoreline mixing zone in the area of the sinkhole and facilitated dissolution of carbonate rocks to a depth near the sea level low stand. As the Wisconsinan ice sheet began to retreat and sea level began to rise 15,000 years ago, the shelf off Key Largo was at least 100 meters (328 feet) above sea level. A shallow freshwater lake would have formed at the bottom of the sinkhole. The lake would have gradually deepened as the groundwater level adjusted to the rising sea

level. By 6000 years ago, just before marine flooding of the shelf, the sinkhole would have been surrounded by wetlands. Infilling of the sinkhole most likely began with precipitated freshwater calcite muds (i.e., marl). As sea level continued to rise, fresh and brackish water were replaced by saline waters. Marine sediment began to settle into the sinkhole, at which time the sinkhole would have functioned like a giant sediment trap. The ^{14}C dates indicate that pulses of rapid sedimentation at 4.1 ka and 4.8 ka (thousand years before present) punctuated marine sedimentation. These pulses were likely the result of tropical hurricanes, which reworked and deposited the lime mud on the Florida reef tract. The lime mud sedimentation ceased and was replaced by sedimentation with skeletal carbonate sands approximately 3 ka. The eastern rim of the sinkhole is dominated by coral reefs which are assumed to be the major source of the carbonate sands that cap the muddy sediment (Reference 959). PTN RAI 02.05.01-17

In summary, the Key Largo submarine paleosinkhole began to form during the Pleistocene. Infilling of the sinkhole began approximately 15,000 years ago when sea level began to rise. The environment at the bottom of the sinkhole at that time was essentially that of a freshwater lake that became brackish and eventually evolved to the current marine environment, at which point conditions conducive for continued limestone dissolution and sinkhole formation no longer existed. At approximately 6 ka the sinkhole was inundated by seawater and became a sediment trap. Rapid pulses of sedimentation occurred approximately 4.1 ka and 4.8 ka. At approximately 3 ka, coral reefs began to accumulate on the seaward side of the sinkhole.

Because the position of the freshwater/saltwater interface is approximately 9.6 kilometers (6 miles) inland from the site (Figure 2.4.12-207), groundwater at the site is saline (Table 2.4.12-210 and 2.4.12-211), and the long-term sea level rise trend at Miami Beach, Florida, as estimated based on data from 1931 to 1981, is 0.2 meter (0.78 foot) per century (Subsection 2.4.5), there is no fresh groundwater shoreline flow near the site. Therefore, a freshwater/saltwater mixing zone that would promote carbonate dissolution at the site does not now exist and the process of shoreline flow that formed the Key Largo submarine paleokarst sinkhole is not a mechanism that is likely to produce cavernous limestone with the potential for collapse at the site. LDP-CS564

Blue Holes of the Bahamas, Eastern South Andros Island

The blue holes of the Bahamas beneath South Andros Island lead to an extensive system of underwater caves along nearshore fracture systems (Figure 2.5.1-365). Formation of the blue holes, which reach depths exceeding 100 meters (328 feet),

began during a previous eustatic sea level low stand associated with advance of continental glaciation during the Pleistocene. Groundwater circulation to the blue holes is facilitated by the fracture permeability that exists within the fracture systems in the carbonate rock. Investigations into groundwater-seawater circulation in some of the holes offshore of South Andros Island indicate a brackish mixture in the caves that readily dissolves aragonite but not calcite, producing secondary porosity. The depletion of calcium in the saline groundwater indicates precipitation of calcite cement. Bacterial processes possibly due to submarine groundwater discharge also play a significant role in driving carbonate dissolution in the Bahamas (References 946 and 950). PTN RAI 02.05.01-17

A similar nearshore fracture system has not been identified in the limestones within the area of the Turkey Point Units 6 & 7 site. As noted previously, the position of the freshwater/saltwater interface is approximately 9.6 kilometers (6 miles) inland from the site (Figure 2.4.12-207), groundwater at the site is saline (Table 2.4.12-211), the long-term sea level rise trend at Miami Beach, Florida, as estimated based on data from 1931 to 1981, is 0.2 meter (0.78 foot) per century (Subsection 2.4.5), and there is no fresh groundwater shoreline flow near the site. Therefore, a freshwater/saltwater mixing zone that would promote carbonate dissolution at the site does not now exist. For these reasons, conditions favorable for formation of dissolution features similar to the blue holes of the Bahamas do not appear to exist in the site area. LDP-CS564

Karst Development on Emergent Carbonate Islands in the Bahamas

In the Bahamas, flank margin caves (Figures 2.5.1-361 and 2.5.1-362) form on emergent carbonate islands due to the mixing of fresh and saltwater in the presence of organic matter. The presence of organic matter allows oxidation to produce carbon dioxide, which in turn produces carbonic acid that drives carbonate dissolution. This carbonate dissolution results in anoxic conditions in the mixing zone of the fresh groundwater lens. Complex oxidation/reduction reactions involving sulfur produce acids that lead to further dissolution (Reference 263). The morphology of the flank margin caves includes large, globular chambers, bedrock spans, thin bedrock partitions between chambers, tubular passages that end abruptly, and curvilinear phreatic dissolution surfaces. The flank margin caves are not conduits, but rather mixing chambers (Figure 2.5.1-362). They receive freshwater from the fresh groundwater lens in the island interior as diffuse flow, and discharge that water, after mixing, as diffuse flow to the sea. The caves develop without an external opening to the sea or the land. Current entry is possible due to surface erosion breaching into the cave (Reference 263). Examples of flank margin caves are Lighthouse Cave, San

Salvador Island, Bahamas and Salt Pond Cave, Long Island, Bahamas (Reference 263). PTN RAI 02.05.01-17

In addition to flank margin caves, there are banana holes in the Bahamas (Figure 2.5.1-362). Banana holes form inland from the flank margin caves at the top of the fresh groundwater lens where the vadose and phreatic freshwaters mix. They are smaller phreatic dissolution voids that form due to collapse of a relatively thin bedrock roof resulting in a broad, vertical-walled depression up to 10 meters (33 feet) across (Reference 263). Both the flank margin caves and banana holes are found in the Bahamas at elevations of 1 to 6 meters (3.3 to 20 feet) above sea level. These caves formed during a glacioeustatic sea level high stand that reached elevations above modern sea level. According to Mylroie and Carew (Reference 263), these caves formed approximately 125,000 years ago. The duration of this high stand above modern sea level lasted approximately 15,000 years, during which time the Bahamas consisted of islands even smaller than today because all land below 6 meters (20 feet) in elevation was below sea level. Therefore, these phreatic caves formed in small freshwater lenses in as little as 15,000 years (Reference 263).

The process of shoreline flow that formed the flank margin caves may be active in the Bahamas today, but at an elevation closer to modern sea level. However, similar processes are not likely to be active at the Turkey Point Units 6 & 7 site because of the absence of fresh groundwater shoreline flow near the site. The position of the freshwater/saltwater interface is approximately 9.6 kilometers (6 miles) inland from the site (Figure 2.4.12-207), groundwater at the site is saline (Table 2.4.12-210 and 2.4.12-211), and the long-term sea level rise trend at Miami Beach, Florida, as estimated based on data from 1931 to 1981, is 0.2 meter (0.78 foot) per century (Subsection 2.4.5). Therefore, a freshwater/saltwater mixing zone that would promote carbonate dissolution at the site does not now exist. LDP-CS564 LDP-CS564

Karst Development on the Yucatan Peninsula, Quintana Roo, Mexico

The Yucatan Peninsula is outside of the 200-mile radius "site region" but karst development there provides evidence of shoreline flow and, therefore, is discussed here. In the Yucatan Peninsula, dissolution features intermediate in size between flank margin and epigenetic continental caves form along the margin of the discharging fresh groundwater lens as a result of freshwater/saltwater mixing. Fresh groundwater discharges are very substantial on the Yucatan carbonate platform, as they are fed by a large volume of allogenic recharge (i.e., recharge of the groundwater from an outside location) from the Yucatan interior (Reference 965). Smart et al. (Reference 965) believe that the Quintana Roo LDP-CS564

caves (Figure 2.5.1-363) represent a new cave type intermediate in size between flank-margin and epigenetic continental systems.

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The Quintana Roo caves located several kilometers interior from the coast may display elements of a dendritic tributary pattern (typical of epigenetic continental caves). Downstream, this drainage passes into an extended zone characterized by a cross-linked anastomosing passage pattern that extends inland from the coast for maximum distances of 8 to 12 kilometers (5 to 7.5 miles) (Reference 965). Large isolated mixing chambers characteristic of the flank margin type caves are absent. Instead, large chambers occur as an element in the anastomosing zone and are generally associated with collapse. Rectilinear maze patterns are generally absent from the caves located in the interior; however, they do appear to be characteristic of some of the coastal caves where fractures have developed parallel to the flank margin (Reference 965).

The passage types in the Quintana Roo caves are horizontal elliptical tubes and canyon-shaped passages and are extensively modified by collapse, but many retain dissolutional wall morphology. The caves are actively enlarging because of undersaturation with respect to calcium carbonate, resulting from the mixing of fresh and saline water. However, according to Smart et al. (Reference 965), many caves in the interior are above the present mixing zone and are characterized by collapse and infill with surface-derived clays, speleothem deposits, and calcite raft sands. Cave sediment fill, speleothem, and ceiling-level data indicate multiple phases of cave development. These multiple phases are associated with glacioeustatic changes in sea level, and alternate in individual passages between active phreatic enlargement and vadose incision and sedimentation. Due to the continued accretion of carbonate rocks along the coast during the Pleistocene, caves that are now located in the interior of the Yucatan Peninsula were formerly closer to the coast and have gone through multiple phases of cave development. Collapse of the cave roofs is extensive and ubiquitous, which results in the development of crown-collapse surface cenotes. Collapse is a result of the large roof spans caused by lateral expansion of passages at the level of the mixing zone, the low strength of the poorly cemented Pleistocene limestones, and the withdrawal of buoyant support during sea level low stands (Reference 965).

Two critical conditions that control the development of multiphase Quintana Roo caves following glacioeustatic variations in sea level are:

1. When the passage segments remain connected to the underlying deep cave systems and are occupied by the present mixing zone, substantial inflow of saline water maintains the rate of mixing-driven carbonate

dissolution, and the predominantly carbonate rock is removed, allowing active passage enlargement to continue. PTN RAI 02.05.01-17

2. When the links between cave passages are absent, rates of dissolution are low, and passage enlargement ceases (Reference 965).

If the flow of freshwater through a passage is maintained by tributaries, the velocity may be sufficient to prevent accumulation of further sediments or to flush uncemented sediments from the passage and the cave void will remain open. If such freshwater flows are limited or absent due to blockage of the feeders, the passage segment will gradually become occluded by infill and roof collapse (Reference 965).

The greater topographic relief of the cenotes terrain of the Yucatan Peninsula provides a stark contrast with the flat topography seen at the Turkey Point Units 6 & 7 site and in the available bathymetric data for the near-site area of Biscayne Bay. The apparent origin of the greater topographic relief and a much more developed karst regime in the cenotes terrane relative to the Turkey Point Units 6 & 7 site and its vicinity is the relatively high rate of fresh groundwater discharge from a large inland watershed in the Yucatan that produces a more robust mixing zone and more carbonate dissolution (Reference 965). The absence of a more developed karst topography or an active mixing zone near the site (because of the location of the freshwater/saltwater interface as shown in Figure 2.4.12-207 and the presence of saline groundwater at the site as demonstrated by Table 2.4.12-210 and 2.4.12-211) suggests that t`he process of shoreline flow that is instrumental in forming the caves on the Yucatan Peninsula is not a mechanism that is likely to produce cavernous limestone with the potential for collapse at the site. LDP-CS564 LDP-CS564

Deep Pore Water Upwelling

DPU takes place beyond the shoreline on the continental shelf through advection of water through deeper, confined permeable shelf sediments and rocks driven by buoyancy and pressure gradients. Evidence of DPU is provided by the existence of offshore submarine springs. In this case, the flow may be driven by an inland hydraulic head through highly permeable confined aquifers or by the large-scale cyclic movement of water due to thermal gradients (Reference 946). Examples of deep pore water upwelling are:

- Submarine paleokarst sinkholes beneath Biscayne Bay (approximately 13 kilometers [8 miles] northeast of the site)

- Crescent Beach Spring and Red Snapper Sink, both off the coast of Crescent Beach, Florida (approximately 320 kilometers [200 miles] north of the site)

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Submarine Paleokarst Sinkholes Beneath Biscayne Bay

Cunningham and Walker (References 958 and 989) conducted a study east of the Miami Terrace using high-resolution, multichannel seismic-reflection data (Figure 2.5.1-356). The data exhibit disturbances in parallel seismic reflections that correspond to the carbonate rocks of the Floridan Aquifer system and the lower part of the overlying intermediate confining unit (Figure 2.5.1-357). The disturbances in the seismic reflections are indicative of deformation in carbonate rocks of Eocene to middle Miocene age. This deformation is interpreted to be related to collapsed paleocaves or collapsed paleocave systems and includes fractures, faults, and seismic-sag structural systems (Figure 2.5.1-358) (References 958 and 989).

In general, the seismic-sag structural systems exhibit one or more zones of vertically stacked, concave-upward arrangements of generally parallel seismic-reflection patterns (Figure 2.5.1-358) (References 958 and 989). Twelve seismic sag structural systems have been delineated on the seismic profiles of Cunningham and Walker (Reference 958). Two types of seismic-sag structural systems they have identified are "narrow" and "broad." The type of system is defined based on the measured differences in the inner sag width of the deformed seismic reflectors. The inner sag width is defined as "the distance between inflection points (i.e., where the shape of the subsidence profile changes from concave to convex) on both sides of the structural trough" (Reference 958).

Collapse related to the "narrow," seismic-sag structural systems is multistoried as shown in Figure 2.5.1-358 (Reference 958). The uppermost termination of zones of concave upward reflections displayed in many of the narrow sag structures may correspond to paleotopographic expression of the upper surface of paleosinkholes, since many are filled in with onlapping reflections. The onlapping reflections indicate passive sedimentary fill at the top of sagging reflections. This relationship is shown in zones 2 and 3 in the N1 profile in Figure 2.5.1-358. These two zones are indicative of cave collapse and suprastratal deformation during the Eocene. Cunningham and Walker (Reference 958) hypothesize that the association of narrow, seismic-sag structural systems with a possible single fault, in some cases, likely indicates a structural fabric and associated fracture/fault permeability. Although the more recent work by Cunningham and Walker confirms the existence of the seismic-sag structural systems in Biscayne Bay, the authors indicate that both faults and karst collapse systems that might cause disruption in

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confinement have only been imaged in the middle Eocene to Oligocene part of the Floridan Aquifer system (Reference 989). These faults may have a substantial control on the geographic distribution of some of the narrow seismic sag structural systems (References 958 and 989).

A major collapse event associated with the "broad" seismic-sag structural system is shown in Figure 2.5.1-359. This collapse event occurred in the Eocene based on the deformation of seismic-reflection stratigraphic layer 8 (SS8) reflections which are assigned to Eocene-age rocks. These SS8 reflectors appear to have downlapping relations onto the upper surface of the zone 2 sag structures and truncate reflectors at the top of the zone 2 structure (Reference 958).

There are three possible mechanisms for the formation of the seismic sag structures: (1) "corrosion" or dissolution by an Eocene mixed freshwater/saltwater zone associated with regional groundwater flow, (2) upward groundwater flow during the Eocene driven by Kohout convection (the circulation of relatively warm saline groundwater deep in carbonate platforms and subsequent mixing with meteoric water as it rises), and (3) upward ascension of hydrogen sulfide-charged groundwater, with the hydrogen sulfide derived from the dissolution and reduction of calcium sulfates in the deeper Eocene or Paleocene rocks (Reference 958). The potential link between the seismic sags and submarine paleosinkholes suggests the seafloor sinkholes began to form as early as the Eocene.

Regardless of the mechanism of formation, the geophysical data indicate the absence of deformation in rocks younger than middle Miocene (Figures 2.5.1-357, 2.5.1-358, and 2.5.1-359). This finding suggests that if the same mechanism had been active at the Turkey Point Units 6 & 7 site during the Eocene, none of the strata younger than middle Miocene would be deformed. These younger strata include the Miami Limestone, Key Largo Limestone, Fort Thompson Formation, Tamiami Formation and Peace River Formation. The total thickness of this section at the site is approximately 137.2 meters (450 feet) (Figure 2.5.1-332). Deformation of rocks below this depth is not likely to pose a threat of surface collapse at the site.

Crescent Beach Spring and Red Snapper Sink, Off the Coast of Northeast Florida

Crescent Beach Spring and Red Snapper Sink are located outside of the 200-mile radius site region, but the spring and sink are evidence of deep pore water upwelling and warrant discussion here. Crescent Beach Spring, a freshwater spring, is located approximately 4 kilometers (2.5 miles) east of Crescent Beach,

Florida (Figure 2.5.1-364) and is considered a first-order magnitude spring with a flow rate of greater than 40 cubic meters/second (greater than 1400 cubic feet/second) (Reference 946). The spring is located at a depth of 18 meters (59 feet) in the Atlantic Ocean, and erosion of confining strata to a depth of 38 meters (125 feet) at the mouth of the vent has enabled direct hydrologic communication of confined groundwater in the Floridan Aquifer with coastal bottom waters (Reference 946).

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The Red Snapper Sink (Figure 2.5.1-364) is located approximately 42 kilometers (26 miles) off Crescent Beach and is incised approximately 127 meters (417 feet) into the continental shelf at a water depth of 28 meters (99 feet). Divers investigating the site observed that seawater was flowing into small caves at the base of the hole, indicating possible recharge of the Floridan Aquifer, and that the water in the bottom of the hole was similar in salinity and sulfate content to ambient seawater. According to Moore (Reference 946), Red Snapper Sink was similar to Crescent Beach Spring before the piezometric head was lowered along the coast, and preservation of the feature suggests that a freshwater spring was active at this site in the recent past.

The existence of Crescent Beach Spring and, by inference, Red Snapper Sink indicates the presence of abundant fresh groundwater within confined aquifers on the continental shelf. Breaching of the confining layer overlying such aquifers by erosional or tectonic mechanisms has the potential to create similar submarine springs on the shelf off southern Florida. No capable faults that could induce a breach of the confining layer have been identified in the site vicinity (Subsection 2.5.3.6). Groundwater in the Biscayne Aquifer (the surficial aquifer) is saline (Table 2.4.12-210 and 2.4.12-211). Therefore, dissolution of carbonate rocks in the vicinity of deep pore water upwelling from this aquifer into the overlying ocean is not probable. At the site, the underlying Tamiami Formation and Hawthorne Group combined comprise more than approximately 152 meters (500 feet) of low-permeability rocks and sediments that overlie and confine the Floridan Aquifer (Figures 2.4.12-202 and 2.4.12-204). Deep pore water upwelling generally occurs well offshore, where the slope of the shelf is steeper and erosion of this thickness of confining sediments is more likely. For this reason, carbonate dissolution associated with deep pore water upwelling from the Floridan Aquifer is not likely to pose a threat of surface collapse or sinkhole hazard at the site.

Sea Level Changes and Migration of Freshwater/Saltwater Interface and Conclusion

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The freshwater/saltwater interface and the zone of diffusion (potential dissolution) is located approximately 10 to 13 kilometers (6 to 8 miles) inland from the Turkey Point Units 6 & 7 site. The site, therefore, occupies a saline hydrogeochemical environment that is different from the locations of active carbonate dissolution and karst development discussed above. This difference provides an explanation for the apparent absence of large dissolution features.

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Since a potential rise in sea level at the Turkey Point Units 6 & 7 site would cause the freshwater/saltwater interface to migrate inland and away from the site, a rise in sea level is not expected to result in dissolution of the subsurface limestones. Conversely, a fall in sea level would cause the freshwater/saltwater interface to migrate toward the site and could result in carbonate dissolution (Reference 956).

Karstification can occur at the ground surface resulting in sinkholes. These sinkholes create depressions that can be filled with water or sediment. There are three main types of sinkholes common to Florida (Reference 264):

- Solution sinkholes occur where limestone is exposed at the ground surface or is covered with a thin mantle of material. Dissolution is concentrated at the surface and along joints, fractures, or other openings in the rock. The development of such features is accomplished by a slow drop of the ground surface that results in the formation of a depression that is commonly filled with organic-rich sediments. These sinkholes typically manifest themselves as bowl-shaped depressions at the ground surface.

- Cover-collapse sinkholes occur where a solution cavity develops in the limestone to a size such that the overlying material cannot support its own weight. The result is generally a sudden collapse of the overburden into the cavity. These sinkholes are common in areas where limestone is close to the ground surface and under water-table conditions, with accelerated dissolution occurring in limestone zones at and just below the water table.

- Cover-subsidence sinkholes occur where the overburden is comprised of unconsolidated and permeable sands. They form when the sand slowly moves downward into space formerly occupied by other sediments, which have already moved downward into space formerly occupied by limestone that has been removed by dissolution. These sinkholes generally develop gradually.

The Florida Geological Survey (FGS) classifies sinkhole occurrences into four type areas (Reference 264) (Figure 2.5.1-222):

- Area I: Bare or thinly covered limestone. Sinkholes are few, generally shallow and broad, and develop gradually. Solution sinkholes dominate. The Units 6 & 7 site is located in Area I (Subsection 2.5.1.2.4).

- Area II: Cover is 9 to 60 meters (30 to 200 feet) thick and consists mainly of incohesive and permeable sand. Sinkholes are few, shallow, of small diameter, and develop gradually. Cover-subsidence sinkholes dominate.

- Area III: Cover is 9 to 60 meters (30 to 200 feet) thick and consists mainly of cohesive clayey sediments of low permeability. Sinkholes are numerous, of varying size, and develop abruptly. Cover-collapse sinkholes dominate.

- Area IV: Cover is more than 60 meters (200 feet) thick and consists of cohesive sediments interlayered with discontinuous carbonate beds. Sinkholes are few in number. However, the ones that do occur are generally large in diameter and deep. Cover-collapse sinkholes dominate.

Finally, karstification has a significant effect on the strength and density of thick carbonate sequences. Over time, limestone strata can become highly porous and weakened by karstification. The structural stability of extensive carbonate platforms in the Caribbean can be affected by karstification (Reference 263). The weakened platforms have undergone extensive escarpment collapse, similar to that described in Subsection 2.5.1.1.1.1.1.2. In addition, karstification during glacial periods of low sea level decreases the density of carbonate sediments by significant amounts and enhances the vertical tectonics of isostatic rebound (Reference 262), as discussed in Subsection 2.5.1.1.1.1.1.1 with respect to glacial cycles and sea-level fluctuations.

Recent work by the Florida Department of Environmental Protection and the FGS now favors an organization by primary and secondary physiographic provinces. This subsection reflects the new organization and does not refer to the three former physiographic zones of Florida. The hierarchy of primary and secondary physiographic zones pertinent to the site region is outlined in the table below.

Florida's Secondary Physiography Provinces within the 200-Mile Radius Site Region[1]

Atlantic Coastal Lowlands	Intermediate Coastal Lowlands	Gulf Coastal Lowlands	Central Highlands
• Atlantic Coastal Ridge • Ten-Mile Ridge • Eastern Valley • Osceola Plain • Crescent City Ridge • Bombing Range Ridge	• Florida Keys, including Coral and Oolite Keys • Florida Bay Mangrove Islands • Everglades • Big Cyprus Spur • Immokalee Rise • Southwestern Slope • Southern Slope • Okeechobee Plain • Caloosahatchee Valley	• Reticulate Coastal Swamps and Ten Thousand Islands • Gulf Coastal Swamps and Drowned Coastal Karst • Gulf Coastal Lagoons and Barrier Chains • Gulf Coastal Terraces • Gulf Coastal Estuaries • DeSoto Plain • Polk Upland	• Lake Wales Ridge • Winter Haven Ridge • Lake Henry Ridge • Lakeland Ridge

Note: modified from Reference 266

The physiography of the Florida Peninsula and continental margin is divided by Schmidt (Reference 266) into seven primary physiographic provinces; Atlantic Coastal Lowlands, Intermediate Coastal Lowlands, Gulf Coastal Lowlands, Central Highlands, Northern Highlands, and the Marianna Lowlands. Two of these, the Northern Highlands and the Marianna Lowlands, are in northern Florida and are outside of the 200-mile radius site region.

Florida's terrestrial geomorphology (i.e., not including the Atlantic Continental Shelf and Slope) is further subdivided by Schmidt (Reference 266) into a number of secondary and tertiary physiographic provinces. The secondary physiographic provinces associated with each primary province are listed in the above table and are shown in Figure 2.5.1-217. The tertiary physiographic provinces are not described in Section 2.5. The following subsections provide a description of geomorphologic features of the southern and central Florida Peninsula within the 200-mile radius site region. The paragraphs address the physiographic subprovinces of the Atlantic Coastal Lowlands, followed by those of the Intermediate Coastal Lowlands, the Gulf Coastal Lowlands, and, finally, the Central Highlands.

1. Small or scattered physiographic provinces may not be identified on Figure 2.5.1-217.

Atlantic Coastal Lowlands Primary Physiographic Province

Atlantic Coastal Ridge Physiographic Subprovince

The Atlantic Coastal Ridge physiographic subprovince is a ridge of sand overlying limestone that ranges in elevation from approximately 10 to 50 feet (3 to 15 meters) above sea level (Figure 2.5.1-217). It averages approximately 5 miles (8 kilometers) wide and is breached in places by shallow sloughs (Reference 267). It is comprised of single and multiple relict beach ridges and bars (Reference 265).

The Atlantic Coastal Ridge (Figure 2.5.1-217) most likely formed during Pamlico time when sea level was approximately 30 feet (9 meters) higher than today. The eastern slope of the ridge mimics the present continental slope because of a sea level regression caused by a rapid onset of glaciation (see discussion of coastal features related to glacial cycles and sea-level fluctuations in Subsection 2.5.1.1.1.1.1.1). Most of the eastern coast of Florida is an erosional shoreline rather than a prograding one (Reference 265).

In southern Florida, the narrow part of the Atlantic Coastal Ridge is a relict beach ridge that surmounts the crest of the remnant Pamlico offshore scarp (steep slope). The ridge and scarp are essentially preserved in their original form (Reference 265). The wider and higher relief of the Atlantic Coastal Ridge is due in part to deposition from terrigenous local sources. The terrigenous local sources consist of quartz sand deposits with broken fragments of contemporaneous shells in the beach sands. The Pamlico Scarp contributes quartzose sand along the Atlantic Coastal Ridge (Reference 265). The Atlantic Coastal Ridge disappears south of Florida City but reappears in the lower Florida Keys (Figure 2.5.1-217). The ridge was probably formed at approximately the same time as the Talbot terrace as an irregular limey bar (see discussion of dating of marine terraces related to glacial cycles and sea-level fluctuations in Subsection 2.5.1.1.1.1.1.1) (Reference 268).

The Atlantic Coastal Ridge, lagoons, and barrier chain is wider in central Florida than it is in southern Florida. The province is made of relict beach ridges and bars, sometimes single and sometimes multiple. The Atlantic Coastal Ridge appears to be almost wholly a product a time when sea level was about 30 feet (9 meters) higher than it now is. The eastern slope of the Atlantic Coastal Ridge closely resembles the present submarine slope that is very uniform offshore from the ocean beaches throughout the length of the Atlantic Coastal Plain. These offshore submarine slopes drop off seaward, steeply at first and then more gently until they are mostly flat at a depth of about 30 feet (9 meters). Because this slope is so

persistent, the Pamlico offshore profile appears to have emerged rapidly without having been altered by later sea level changes. White (Reference 265) speculates that the clean, undamaged emergence may be due to the fact that marine regression from the Pamlico level was caused by rapid onset of glaciation.

Ten-Mile Ridge Physiographic Subprovince

Ten-Mile Ridge comprises the narrow, discontinuous remnants of a former coast ridge a few miles inland from the Atlantic Coastal Ridge of central Florida (Figure 2.5.1-217). Ten Mile Ridge is one of five distinct landscapes in Florida that preserves scrub vegetation, a rare and vanishing ecosystem. The scrub vegetation landscape is associated with ridges of well drained to moderately well drained soils. The five Florida landscapes preserving scrub vegetation are: (a) the recent barrier islands including Cape Canaveral, (b) Merritt Island, (c) the Atlantic Coastal Ridge, (d) the Ten Mile Ridge, and (e) a small ridge in the southwest corner of Brevard County. These areas differ in age and topography, but all have a similar origin as coastal dunes (Reference 269).

Ten-Mile Ridge was once a line of dunes but now serves as the elevation upon which Interstate 95 runs. Although topographically low, it is sufficient to prevent water from draining from the interior into the Indian River lagoon. The result is the formation of a marsh west of the ridge that is some 20 miles (32 kilometers) wide and 30 miles (48 kilometers) long. The Ten-Mile Ridge is part of the Pamlico Terrace (Reference 270) (Figure 2.5.1-220), from the Late Pleistocene (Sangamonian) interglacial period (Reference 271).

Eastern Valley Physiographic Subprovince

The southern limit of the Eastern Valley province lies just north of the Atlantic Coastal Ridge and east of the Everglades subprovince of southern Florida (Figure 2.5.1-217). The Eastern Valley is a broad, flat valley with elevations of about 25 feet (7.6 meters), slightly lower than the level of the Pamlico stand. There are indications that the topographic surface was higher and has been reduced by solution of the constituent shell fraction (Reference 265). Thus, the area may precede the Pamlico stand. There are relict beach ridges throughout much of the length and width of the valley, indicating that the landform is a regressional or progradational beach ridge plain (Reference 265).

The Eastern Valley physiographic subprovince of central Florida (Figure 2.5.1-217) is a broad flat valley in Seminole and Indian River counties. Its elevation varies from 20 to 30 feet (6 to 9 meters) above sea level. There are

relicts of beach ridges that at one time constituted a regressional or progradational beach ridge plain (Reference 265).

The head of the St. Johns River consists of a broad swampy valley with lakes. The river flows through each lake along its longest axis. This suggests that at one time there was a standing body of water that has been filled with sediments and vegetation between the upper levels of the lakes that eventually formed the flat, swampy flood plain; the unfilled places became the current chain of lakes in the St. Johns River's headwaters (Reference 265).

Southward of the St. Johns River, the topography has approximately 5 feet (1.5 meters) of local relief throughout the area. This topography is bounded by the headwaters of the St. Johns River at the north, the bounding scarp of the Eastern Valley on the west, the St. Lucie Canal on the south, and Ten Mile and Atlantic Coastal Ridges on the east (Figure 2.5.1-217). The surface of the entire area has elevations close to 25 to 30 feet (7.6 to 9 meters) (Reference 265).

Osceola Plain Physiographic Subprovince

The Osceola Plain physiographic subprovince (Figure 2.5.1-217) extends southeasterly through eastern Okeechobee County, extreme southwestern St. Lucie County, and into western Martin County. It is bounded on the west and northwest by the Lake Wales Ridge and the southern ends of the Mount Dora and Orlando Ridges. On the northeast, east, and south it is bounded by an outward-facing erosional ridge (Reference 265). The Osceola Plain reaches approximately 90 to 95 feet (27 to 29 meters) in elevation near its northern edge. It reaches an elevation of 80 feet (24 meters) east and northeast of Lake Kissimmee. Its local relief is very small, with variations of 10 feet (3 meters) across the entire subprovince (Reference 265).

The Kissimmee River passes roughly west of the Osceola Plain. The river is confined to a valley for 25 miles (40 kilometers) south of Lake Kissimmee. North of Lake Kissimmee, several lakes occupy most of the Osceola Plain (Reference 265). The Arbuckle Creek on the western side of the Osceola Plain (west of the Bombing Range Ridge) drains Lake Arbuckle into Lake Istokpoga below the southern bounding scarp of the Osceola Plain (Reference 265) (Figure 2.5.1-217).

Crescent City Ridge Physiographic Subprovince

Crescent City Ridge (Figure 2.5.1-217) is located in southeast Putnam County and northwest Volusia County in east-central Florida. The altitude of the land

surface ranges from near mean sea level at Crescent Lake to approximately 120 feet (37 meters) above mean sea level in the sandy ridges. The topography is characterized by a series of terraces, or step-like surfaces of increasing elevation, which are the result of wave erosion and deposition during the advance and retreat of sea level during the Pleistocene glacial maxima. These marine terraces have been dissected by varying degrees of erosion and are capped by thin surficial sands (Reference 272).

The Crescent City Ridge also exhibits karst topographic features resulting from the dissolution of the underlying limestone formations. Such areas are characterized by high local relief, a lack of surface drainage features, subsurface drainage, sinkholes, and sinkhole-related lakes and springs. Due to the lack of surface drainage in the upland sandy ridges, almost all of the precipitation falling in these areas is either lost to evapotranspiration or drains downward through the permeable soils or sinkholes (Reference 272).

Bombing Range Ridge Physiographic Subprovince

Bombing Range Ridge is a small subprovince located east of elongate Lake Wales Ridge (Figure 2.5.1-217). The lakes of the Bombing Range Ridge and the northern Lake Wales Ridge are darker colored with higher nutrients than the lakes found on the southern Lake Wales Ridge. Elevations are 70 to 130 feet (~21 to 40 meters), and there are more extensive areas of poorly drained soils, such as the Satellite and Basinger series. Peaty muck Samsula soils border many of the lakes (Reference 291).

Intermediate Coastal Lowlands Primary Physiographic Province

Florida Keys Physiographic Subprovince

The Florida Keys physiographic subprovince (Figure 2.5.1-217) is a narrow chain of small islands at the southern tip of the Florida Peninsula. The Florida Keys consist of the High Coral Keys, Low Coral Keys, and the Oolite Keys. The keys are composed of Pleistocene reef sediments (Key Largo and Miami Limestones). There is not a clear dividing line between the High and Low Coral Keys; however, the Key Largo Limestone predominates in the High and Low Coral Keys whereas the Miami Limestone dominates in the Oolite Keys (Reference 274).

The Florida Keys extends for approximately 150 miles (240 kilometers) from Miami southwest to Key West. It is bounded by the Atlantic Ocean and the Bay of Florida, inland waters, and the Gulf of Mexico (Reference 274). The edge of the continental shelf parallels the Florida Keys approximately 7 miles (11.3 kilometers)

offshore. Maximum elevations reach 10 to 12 feet (~3 to 4 meters) above mean sea level. On the offshore islands, elevations are less than 10 feet (3 meters) above mean sea level. Average depths in Hawk Channel, which parallels the shore in Florida State waters, are 13 to 15 feet (~4 to 5 meters). In the Florida Keys coral replaces the sand in barrier bars and islands along the southern coast of Florida (Reference 273).

A topographic high beneath Key Largo was the focus of reef growth (Reference 275). The late Tertiary siliciclastic sediments underlying the Quaternary carbonate rocks appears to control the position and arc shape of the recent shelf and slope of southern Florida (Reference 275). Additionally, the arc pattern of the Florida Keys is related to the bathymetry of the shelf edge and the Florida current. The growth of patch reefs depends on nutrient availability, sea level, and topography.

Florida Bay Mangrove Islands Physiographic Subprovince

The island and shoals of Florida Bay (Figure 2.5.1-217) are made of lime mud and their surfaces usually lie within a foot above or below normal water level. The westernmost of these banks are sandy and extend southeastward from East Cape near Cape Sable. These banks face the deeper, open waters of the Gulf of Mexico on their west side. The remaining banks further east are made of mud and form a network of long, narrow vegetation-covered shoals and mangrove-covered islands (Reference 271).

Everglades Physiographic Subprovince

The Everglades physiographic subprovince (Figure 2.5.1-217) extends from Lake Okeechobee southward toward Florida Bay. It consists of the Everglades, Big Cypress Spur, Florida Bay, and coastal mangroves (References 267 and 276). The Everglades is a wetland prairie created by the overflow of Lake Okeechobee, whose water spreads in a slow-moving sheet flow across a slope of less than 2 inches/mile from Lake Okeechobee to the mangrove lined margins of southwest Florida near Florida Bay and the Gulf of Mexico (References 267 and 277).

Elevations in the Everglades range from 14 feet (4 meters) near Lake Okeechobee to sea level at Florida Bay. Before development of the canals in the northern portion of the Everglades, water discharged from the Everglades into the Florida Bay, the Gulf of Mexico, and the Atlantic Ocean through small rivers in the Atlantic Coastal Ridge or as seepage and spring flow into Biscayne Bay (see

discussion of changes in drainage related to glacial cycles and sea-level fluctuations in Subsection 2.5.1.1.1.1.1.1) (Reference 267).

The vegetative wetland community and landscapes consist of a central core of peat that extended from Lake Okeechobee to Florida Bay. Organic soils (peat) overlie the limestone throughout most of the Everglades. The fibrous peat accumulates on limestones because the limestone can be dissolved down to the water table. This results in swampy conditions for the growth of fibrous swamp plants and their preservation as fibrous peat (Reference 265).

The water flowing through the Everglades is only a few inches deep, but is 50 miles (80 kilometers) wide. The predominant vegetation, sawgrass, is flooded during the wet season (summer) and burned (parched) during the dry season (winter/spring). During flood periods, the movement of water causes tree islands to develop an alignment pattern parallel to the lines of surface water flow (Reference 267). Interspersed through the sawgrass are deeper water sloughs where tree islands or hammocks appear. The coastal mangrove swamps form very thick, dense thickets at the coastline where the sawgrass and cypress swamps meet (References 267 and 276).

Big Cypress Spur Physiographic Subprovince

The Big Cypress Spur physiographic subprovince (Figure 2.5.1-217) is located west of the Everglades. The land surface is flat except for low-mounded limestone outcrops and small, oval, elongated depressions in the limestone. Water drains to the south and southwest through cypress strands into the coastal mangroves (Reference 267). The elevations are below 16 feet (5 meters), and the physiography consists of prairies, marshes, and stunted cypress (Reference 278). The subprovince is dominated by cypress, pine, and wet prairie. The Big Cypress Spur receives approximately 50 inches of rainfall per year but does not have overland flow similar to that of the Everglades (References 267 and 276).

Immokalee Rise Physiographic Subprovince

The Immokalee Rise physiographic subprovince (Figure 2.5.1-217) is north of the Big Cypress Spur, west of the Everglades, and south of the Caloosahatchee Valley. It appears to be a southern extension of Pamlico marine sand because it exhibits several relicts of Pamlico shoreline features (see discussion of marine terraces related to glacial cycles and sea-level fluctuations in Subsection 2.5.1.1.1.1.1.1). According to White (Reference 265), the Immokalee Rise was built as a submarine shoal that extended southward from a mainland

cape at the south end of the DeSoto Plain. The Immokalee Rise is ringed with small peripheral lakes that formed as a result of limestone dissolution (Reference 265). Elevations range between 30 and 40 feet (9 to 12 meters) (Reference 278).

The Immokalee Rise includes most of Hendry County and eastern Lee County. It is about 8 meters (25 feet) in elevation, but can peak at 11 meters (36 feet) and 13 meters (43 feet) in some areas. All soils are deep, nearly level, and poorly drained, with a water table less than 25 centimeters (10 inches) from the surface during at least part of the year (Reference 279).

Southwestern Slope Physiographic Subprovince

The Southwestern Slope physiographic subprovince (Figure 2.5.1-217) is located along the eastern shore of the Gulf of Mexico, west of the Immokalee Rise and Big Cypress Spur. The landscape is relatively flat and underlain with an uneven bedrock surface that is usually covered by a veneer of soils. The soils are relatively modern and in the process of formation from surficial sediments such as sand and calcareous marl mixing with organic peat and muck components. Tidal marshes, sea grass beds, and mangroves develop in this region of Florida because of the low wave energy (Reference 267).

Southern Slope Physiographic Subprovince

The Southern Slope physiographic subprovince (Figure 2.5.1-217) is located along the southern shore Florida Bay, south of the Everglades and the Big Cypress Spur. The topography is at or near sea level, and the area consists of broad bands of swamps and marshes that are flooded by tides or by freshwater run-off. Tidal marshes, seagrass beds, and mangroves develop in this region of Florida because of the low wave energy (Reference 267).

Okeechobee Plain Physiographic Subprovince

The Okeechobee Plain physiographic subprovince (Figure 2.5.1-217) is within Okeechobee County and includes part of Lake Okeechobee. The southern part of this plain abuts the Everglades with Lake Okeechobee bisecting the plain. The Okeechobee Plain is divisible from the Everglades by its slightly better drainage and slightly steeper slope and a higher mineral content in its soils. The Okeechobee Plain slopes gradually south to approximately elevation 20 feet (6 meters) at the northern shore of Lake Okeechobee (Reference 265).

Caloosahatchee Valley Physiographic Subprovince

The Caloosahatchee Valley physiographic subprovince is part of the "Caloosahatchee Incline." This broad gentle incline forms the valley wall of the Caloosahatchee River, which runs through the valley. The Caloosahatchee incline slopes eastward from the eastern toe of the Lake Wales Ridge at 50 to 60 feet (15 to 18 meters) elevation down to 30 to 35 feet (9 to 10.7 meters) at the edge of the Okeechobee Plain. The Caloosahatchee incline is a remnant of a submarine shoal (Reference 265).

The Caloosahatchee River subprovince encompasses the Caloosahatchee River watershed, the lower Charlotte Harbor estuarine system (which includes the San Carlos Bay, Matlacha Pass, and Pine Island Sound), the Estero Bay estuary and watershed, and the Immokalee Rise. This area is approximately 516,000 hectares (1.28 million acres), and includes most of Lee County, the southeastern portion of Charlotte County, western Hendry County, and southern Glades County. The major physiographic provinces of the subregion are the Caloosahatchee Valley, Gulf Coast Lowlands, DeSoto Plain, and the Immokalee Rise (Reference 279).

Gulf Coastal Lowlands Primary Physiographic Province

Reticulate Coastal Swamps Physiographic Subprovince

Reticulate Coastal Swamps (Figure 2.5.1-217) consist of a broad band of swamps and marshes south of the Everglades and the Big Cypress Swamp. The land is at or near sea level and is often flooded by tides or by freshwater runoff. Salinities range from freshwater to hypersaline, depending on the amount rainfall and runoff and tide levels. The gradual slope of the land continues offshore across the broad west Florida Platform into the Gulf of Mexico. Much of the southern Florida Gulf Coast receives low wave energy that is favorable to the development of tidal marshes, seagrass beds, and mangrove forests (Reference 267).

Gulf Coastal Swamps and Drowned Karst Physiographic Subprovince

The Gulf Coastal swamps and drowned coastal karst include coastal areas with too little sand to build beaches. The areas start in the south at Florida Bay and continue northward as far as Naples and then from Tarpon Springs to the west side of Apalachee Bay. Coastal lagoon areas include large areas of swamp where they have accumulated sediment and covered with vegetation. Continuous areas of drowned karst along the Gulf Coast of Florida are most prevalent north of Tampa Bay, but small, isolated areas occur south of Tampa Bay to Florida Bay (not shown on Figure 2.5.1-217).

Gulf Coastal Lagoons and Barrier Chains Physiographic Subprovince

The Gulf coastal lagoons and barrier chains are found the entire length of the central Gulf Coast area north of the Caloosahatchee River estuary and continuing northward along the Florida Panhandle. The barrier lagoon system of the Gulf Coast of Florida differs from those on the Atlantic coast in that the sands that build the barriers appear to have been locally derived by erosion of the headlands (Reference 265). The barriers appear to have been developed by erosion of the coastal prominences that separate the large estuaries. White (Reference 265) notes that the process of building spit-like barriers from headlands at the drainage divides appears to have been the habit of this part of the Gulf Coast for some time.

Gulf Coast Terraces Physiographic Subprovince

The Gulf Coastal Terraces physiographic subprovince (Figure 2.5.1-220) can be subdivided by the terraces created by marine regression of the Gulf of Mexico (see discussion of marine terraces related to glacial cycles and sea-level fluctuations in Subsection 2.5.1.1.1.1.1.1). The topography is dominated by broad marine plains and gentle depositional slopes; the regional slope steepens and narrows northward and becomes terraced. The main terraces in the Gulf Coast Terraces province, from oldest to youngest, are the Wicomico, Penholoway, Talbot, and Pamlico (Reference 281). According to Walker and Coleman (Reference 282), these four terraces and three intervening cemented ridges represent the late Pleistocene near shore environments. However, it has been suggested by Alt and Brooks (Reference 283), Lichtler (Reference 284), and Healy (Reference 261) that the elevation of the terraces above 100 to 170 feet (30 to 52 meters) within Florida are not representative of the Pleistocene marine terraces but of older Pliocene and upper Miocene deposits.

Gulf Coast Estuaries Physiographic Subprovince

Estuaries are found along the length of coastal areas of the western Florida Peninsula. The most noteworthy estuaries of southwest Florida are the Charlotte Harbor, Lemon Bay, Coastal Venice, Pine Island Sound, the Caloosahatchee River, Estero Bay, Wiggins Pass/Cocohatchee River, Naples Bay, Rookery Bay, and the Ten Thousand Islands estuaries. Estuaries are semi-enclosed areas, such as bays and lagoons, where fresh water meets and mixes with salty ocean waters. Estuaries are dynamic systems with constantly changing tides and temperatures where salinity varies temporally and spatially. The rivers and streams that drain into the estuaries bring nutrients from the uplands. The

nutrients are a food source for a wide range of species, from floating phytoplankton and large algae attached to the estuary floor, to rooted plants including mangrove, marsh grasses, and seagrass. As these endemic plants and animals die, they decompose and provide food for fish, crustaceans and shellfish for which estuaries serve as nurseries (Reference 285).

DeSoto Plain Physiographic Subprovince

The DeSoto Plain physiographic subprovince (Figure 2.5.1-217) in DeSoto County is approximately 45 to 50 miles (72 to 80 kilometers) in length and varies in width from 25 miles (40 kilometers) at the southern edge to 50 miles (80 kilometers) at the northern edge. The northern edge is adjacent to the Polk Uplands. This subprovince has low relief; the northern edge has elevations of 75 to 90 feet (23 to 27 meters); the southern edge has an elevation of 60 feet (18 meters) (Reference 265). The DeSoto Plain and the Osceola Plain represented the southernmost edges of the Florida Peninsula during the Pliocene. The two shorelines were separated by a shallow embayment, the Kissimmee Embayment, which is still recognizable as the Okeechobee Plains lowlands (Reference 286).

The Peace River transverses the DeSoto Plain and is entrenched into the plain 30 to 40 feet (9 to 12 meters). The DeSoto Plain consists of a line of elongated cypress swamps underlain by clay deposits (Reference 278). The relict depositional environment is that of a lagoon that existed during the emergence of the DeSoto Plain. The plain consists of Long Island Marsh, Rainsy Slough, and Valley of Fisheating Creek (Reference 265).

Polk Upland Physiographic Subprovince

The Polk Upland physiographic subprovince (Figure 2.5.1-217) is within Polk County. It is surrounded by the DeSoto Plain on the south, the Gulf Coastal Lowland on the west, the valley of the Hillsborough River and upper Withlacoochee River on the north, and by the Lake Wales Ridge on the east. The Winter Haven and Lakeland Ridges rise from its surface from the northeast. The elevation of the Polk Upland ranges from 100 to 130 feet (30 to 40 meters) (Reference 265).

From the south, a ridge separates the Polk Upland from the DeSoto Plain. The ridge turns 90° N at the southwestern corner of the Polk Upland and terminates approximately halfway on its western side. The edge of the ridge is at an elevation of about 75 to 85 feet (23 to 26 meters), and the crest is at an elevation of about 100 feet (30 meters) (Reference 265). A second ridge is located at the northern

end of the western boundary of the Polk Upland and a third irregular ridge is located at the southern boundary. These ridges are most likely erosional marine scarps formed as Gulf of Mexico shorelines during Wicomico sea level (see discussion of marine scarp features related to glacial cycles and sea-level fluctuations in Subsection 2.5.1.1.1.1.1.1) (Reference 265).

Central Highlands Primary Physiographic Province

The Central Highlands are comprised of coast-parallel scarps and constructed sand ridges (see discussion of shoreline variations related to glacial cycles and sea-level fluctuations in Subsection 2.5.1.1.1.1.1.1) (References 265 and 287). The ridges are all long, narrow, and elongated in the same orientation of the relict beach ridges along the eastern shore of the peninsula. The region consists of xeric (arid) residual sand hills, beach ridges, and dune fields interspersed with numerous sinkhole lakes and basins caused by the dissolution of the underlying limestone bedrock (Reference 265). Millions of years ago, these ridges were formed by rising and falling sea levels. During the periods when the sea level was high and flooded most of peninsular Florida, these ancient islands became refuges for plants and animals. Populations were isolated from the mainland for thousands of years and evolved within these small, sandy habitats. The central inland ridges are older, having remained islands while coastal ridges were flooded, and have a greater concentration of endemic species (Reference 288).

Lake Wales Ridge Physiographic Subprovince

The Lakes Wales Ridge physiographic subprovince (Figure 2.5.1-217) is a unique mosaic of elevated sandy ridges encompassing an area from about the southern Highlands County boundary 160 kilometers (99 miles) north to near Orlando. The Lake Wales Ridge averages about 7.5 kilometers (4.6 miles) wide (Reference 289). Though the name implies a single physiographic area, the Lake Wales Ridge actually consists of three elevated sandy ridges that were once the beach and dune systems of Miocene, Pliocene, and early Pleistocene seas (Reference 290). These relict dunes and the deep, sandy, well-drained soils support a number of plant communities that have adapted to xeric conditions over millions of years. Due to the elevation and geologic age of the soils of Lake Wales Ridge scrubs, it has been estimated that the highest hilltops in this area have supported upland vegetation for about 25 million years. On the Lake Wales Ridge, an estimated 200 ancient scrub islands have been identified (Reference 290). Between ridges and at the base of hills, the soils become fine and compacted and often retain surface water, forming wetlands and lakes. Rainfall, seepage, and elevated water tables provide the sources of water for these aquatic systems.

Combined with the aquatic and wetland communities that now exist between and within the ridges, this subregion consists of a complex mosaic of habitats, some unique to Florida (Reference 291).

Winter Haven Ridge and Lake Henry Ridge Physiographic Subprovinces

The Winter Haven Ridge and Lake Henry Ridge physiographic subprovinces (Figure 2.5.1-217) are an upland karst area that is 130 to 170 feet (40 to 52 meters) in elevation with an abundance of small- to medium-sized lakes. Candler-Tavares-Apopka is the soil association of the well-drained upland area, with longleaf pine and xerophytic oak natural vegetation. Pliocene pebbly quartz sand and the phosphatic Bone Valley Member of the Peace River Formation comprise the underlying geology. The lakes of the area can be characterized as alkaline, moderately hardwater lakes of relatively high mineral content, and are eutrophic (Reference 291).

Lakeland Ridge Physiographic Subprovince

The Lakeland Ridge physiographic subprovince (Figure 2.5.1-217) includes the sand hills of the Lakeland Ridge. These sand hills are covered by phosphatic sand or clayey sand from the Miocene-Pliocene Bone Valley Member of the Peace River Formation. The region generally encompasses the area of most intensive phosphate mining, but phosphate deposits and mining activities are also found south of this region. The dominant characteristic of all lakes in this region is high phosphorus, nitrogen, and chlorophyll-a values. The lakes are alkaline, with some receiving limestone-influenced groundwater (Reference 291).

2.5.1.1.1.1.1.2 Florida Platform

This subsection describes the submerged portion of the west coast of Florida projecting into the Gulf of Mexico (Figure 2.5.1-201). This submerged platform is underlain in its entirety by a large carbonate province of middle Jurassic to Late Cretaceous age, which in turn rests upon what is believed to be highly extended transitional continental-oceanic crust. The west Florida coastal zone is characterized by a series of barrier beaches and back barrier lagoons. The subsiding platform shelf with spotty carbonate production and the quantity of siliciclastic sediments moving along the west Florida coast are evidence that the carbonate platform has been “drowned” by climatic and tectonic changes that occurred since the Miocene.

The siliciclastic-dominated, west-central barrier island coast of peninsular Florida and its adjacent inner continental shelf lie at the center of a huge, ancient

carbonate platform that forms the proximal portion of the western Florida shelf/ slope system. This platform's east-to-west lateral dimensions extend 310 miles (500 kilometers) from the 650 feet (200 meters) water depth shelf break off the east coast of Florida to the 5000 feet (1500 meters) top of the Florida Escarpment, which drops precipitously nearly another 6000 feet (1800 meters) into the deep Gulf of Mexico (Reference 292) (Figures 2.5.1-201 and 2.5.1-214). For the purpose of this discussion, the Florida Platform is defined as that area between the western coastline of the Florida Peninsula to the edge of the Florida Escarpment (Figure 2.5.1-214). The distinction of where to draw the eastern boundary of the Florida Platform is somewhat arbitrary in that the thick sequence of Upper Jurassic carbonate deposits extends inland to the edge of the Paleozoic rocks forming the core of the Florida Peninsular Arch.

"Carbonate platforms" is a morphological term for a three-dimensional structure and a stratigraphic term for thick sequences of shallow-water carbonates (Reference 293); in this subsection carbonate platforms are referred to in the morphological sense. All carbonate platforms are uniquely sensitive environments that have formed and disappeared many times throughout the Proterozoic and Phanerozoic. The most important controls on carbonate platform development are a shallow substrate, temperature, salinity, and light intensity (Reference 293).

Beginning in the Jurassic, a carbonate platform began to form along the length of the Appalachian-Ouachitan suture/Pangean rift zone. The thinned continental crust foundered to create a long, shallow basin. Seawater incursions were sporadic at first, leaving accumulations of evaporites. Eventually, seawater flooded the shallow basins, building a carbonate megaplatform that deposited 7- to 11-kilometer (4.3- to 6.8-mile) thicknesses of limestones, dolomites, and evaporites (Reference 294). The megaplatform stretched continuously from the Bahamas along the entire Atlantic margin in Late Jurassic to Early Cretaceous time (Reference 424), eventually extending with little or no interruption over 3700 miles (6000 kilometers) from the Gulf of Mexico to the Grand Banks. This carbonate "gigaplatform" probably was one of the largest carbonate platform systems in earth's history (Reference 296).

Because of their sensitivity to depth, temperature, salinity, and light intensity, carbonate platforms are excellent markers for climate change induced by tectonics, igneous activity, reorganization of ocean currents, and other causes. The two types of events that can immediately shut down carbonate production globally or over very wide regions, namely rapid transgression and carbonate crashes, are discussed in the following paragraphs.

Rapid transgression (drowning events) may submerge the carbonate platform below the photic zone and subsequently shut down carbonate production. The rate of transgression must exceed the rate of carbonate accumulation for this to occur. These events may create unconformities that transition from carbonate to siliciclastic deposition. Platform drowning may also occur from episodic rapid subsidence induced by tectonism. If the depth of drowning is moderate, sediment accumulation may allow the seafloor to rise into the photic zone and again begin carbonate production. While reefs may keep up with sea level rise due to high accumulation rates, shallow tidal carbonates, with lower growth rates, may be drowned more easily thus forming a rimmed platform (Reference 297).

A "carbonate crash," a time of increased regional carbonate dissolution, occurred in the Gulf of Mexico and Caribbean region at the middle to late Miocene transition (Figure 2.5.1-223). Miller et al. (Reference 298) proposes that the early Eocene peak in global warmth and sea level was due not only to slightly higher ocean-crust production but also to a late Paleocene-early Eocene tectonic reorganization caused by emplacement of geographic barriers. The Miocene crash is marked by five dissolution episodes, occurring from 12 to 10 Ma and characterized by significant reductions in the mass accumulation rates of carbonate, noncarbonated, and bulk sediments. Antarctic Intermediate Water filled the Caribbean to abyssal depths during the Quaternary interglacial stages and initiated the North Atlantic Deep Water production. The combination of the initiation of North Atlantic Deep Water, the partial closing of the Isthmus of Panama, and the opening of the Pedro Channel in the northern Nicaraguan Rise led to a change in the global thermohaline circulation. The increase of return flow that passes through the Caribbean may have brought corrosive Antarctic Intermediate Water into the Caribbean, which then caused dissolution of carbonate sediment at the sea floor (References 299, 236, 300, and 302) (Figure 2.5.1-213).

The carbonate crash occurred on both sides of the Isthmus of Panama and in the northwest Caribbean and thus is related to a major reorganization of the ocean circulation during the late Neogene and possibly to the first establishment of a pattern of global thermohaline ocean circulation approaching that of today (Reference 301) (Figure 2.5.1-213). The opening of major seaways would have caused a major reorganization of the deep and intermediate oceanic circulation and maybe the initial production of the North Atlantic Deep Water. The influx into the Caribbean Basins of Antarctic Intermediate Water would have been initiated at this time to replenish the waters sinking in the northern latitudes of the North Atlantic Ocean. This influx of southern source intermediate waters that were

corrosive toward carbonate sediments would explain the occurrence of the systematic carbonate dissolution observed in the Caribbean Basin at the middle to late Miocene transition (References 299, 236, and 302).

The geomorphic processes acting on the Florida Platform today include the scouring action of the Loop Current and the influx of terrigenous sediments at the mouths of rivers, generally represented by large embayments, most with backwater lagoons and estuaries (e.g., the Waccasassa, Tampa, Sarasota, and Charlotte Bays). In general, siliciclastics are introduced into the embayments where they may be trapped by persistent carbonate collapse-related sedimentary basins (e.g., Tampa Bay) (Reference 303) or, more likely, the siliciclastics are redistributed by the Loop Current to deep-water basins (see discussion of currents related to changes in oceanic circulation patterns in Subsection 2.5.1.1.1.1.1.2). Very little clastic deposition (<66 feet or 20 meters of accumulated Neogene and Quaternary sediments) occurred in the northern half of the Florida Platform at this time. The southern portion of the Florida Platform is characterized by thick (>330 feet or 100 meters) to very thick (>660 feet or 200 meters) Neogene and Quaternary siliciclastic and carbonate deposition. At intermediate water depths, coarse sand clastics move along the eroded carbonate base of the Florida Platform as coast-parallel sand ridges (Reference 304).

The Florida Peninsula is the emergent portion of the wide, relatively flat geologic feature called the Florida Platform, which forms a rampart between the deep waters of the Gulf of Mexico and the Atlantic Ocean (Figures 2.5.1-201 and 2.5.1-214). The Florida Peninsula is located on the eastern side of the platform. The western edge of the platform lies over 100 miles (161 kilometers) west of Tampa, while on the east side of Florida it lies only 3 or 4 miles (5 to 6.4 kilometers) off the coast from Miami to Palm Beach. Within relatively short distances from the basinal edge of the Florida Platform water depths increase sharply, eventually reaching "abyssal" depths of over 10,000 feet (3050 meters). This relatively sharp break in seabed topography is the Florida Escarpment. The Florida Escarpment is analogous to the Campeche Escarpment and the Bahamas Escarpment, representing the margin of a once actively prograding carbonate platform at a passive margin with an ocean basin (e.g., the two spreading centers in the Gulf of Mexico on the northwestern and northeastern edges of the Yucatan Platform, discussed in Subsection 2.5.1.1.2.1.1, and the mid-Atlantic Ridge east of the Bahama Platform). As discussed in Subsection 2.5.1.1.1.1.2, none of the escarpments formed as the result of Quaternary faulting, and all are retreating on their seaward edges by ocean current erosion and mass wasting into the nearby abyssal basins (Reference 305). Thus, the former spreading center (inactive)

faults are likely many kilometers basinward from the current edge of the escarpments.

Diving expeditions with the deep submersible *Alvin* along the escarpment west of Tampa found the escarpment there consisted of a limestone cliff that rose over 6000 feet (1830 meters) above the 10,700-foot (3260-meter) deep Gulf of Mexico floor. Based on seismic evidence from petroleum exploration, carbonate and evaporitic rocks may underlie the southern tip of Florida at depths greater than 20,000 feet (6100 meters) below ground surface (Reference 287).

The Florida Platform is part of the larger Florida-Bahama Platform, which is approximately 600 miles (900 kilometers) long, 620 miles (1000 kilometers) wide, and over 7.5 miles (12 kilometers) thick (Reference 307). The Florida-Bahama Platform represents an enormous shallow-water, carbonate sedimentary province (Figures 2.5.1-201, 2.5.1-202, and 2.5.1-224). The marginal-reef facies along the Florida Platform have been eroded back by as much as 4 to 16 miles (6 to 26 kilometers) since the close of the Oligocene (References 305 and 308).

The Florida Platform is a broad smooth shelf with widths of nearly 124 miles (200 kilometers) and little morphologic variability. Vertical relief is shown in the coral algae ridges (Reference 282). Facies distribution across the Florida Platform is shown in Figure 2.5.1-225. Before the Florida Platform was drowned in the Late Jurassic, the facies distribution of active carbonate accretion was very similar to that of the current Bahama and Yucatan Platforms. Today, carbonate production on the Florida Platform is much reduced (see discussion of carbonate platforms in Subsection 2.5.1.1.1.1.1.2). The "drowning" of the Florida Platform did not occur as a single event. Carbonate production kept pace with subsidence across the platform through the Early Cretaceous (until about 100 Ma). Siliciclastic sedimentation on the area near the Florida Panhandle began to replace carbonate production sometime in the mid-Cretaceous. Carbonate production in the southern portion of the Florida Platform kept pace with subsidence until early Late Cretaceous (Cenomanian) time, when the Florida Platform and the other carbonate banks of the Gulf of Mexico were drowned and covered by neritic marl containing some siliciclastic debris. Only the south-southwestern most parts of the west Florida and Bahama Platforms continued to be areas of active carbonate sedimentation (Reference 309). Elsewhere, the juxtaposition of underlying mollusk-rich, shallow-water limestones with the overlying pelagic facies indicates that the drowning was rapid and terminal (Reference 309). The drowning event is recognized in the stratigraphy across the Gulf of Mexico and West Florida area and is known as the mid-Cretaceous sequence boundary or unconformity (see discussion of the oceanic and atmospheric reorganization events in

Subsection 2.5.1.1). The general distribution of facies prior to drowning is shown in Figure 2.5.1-225. Like other drowned carbonate platforms, today's Florida Platform has subsided to bathy depths, forming a deep-water plateau as the increasingly deeper-water sedimentary facies were unable to keep pace with subsidence (Reference 305). By the end of the Late Cretaceous, water depths exceeded 900 meters (2950 feet) over the marginal parts of the former shallow-water carbonate bank. Currently, the Gulf Coast continental shelf and slope of the Florida Platform constitutes a ramp system that connects the Gulf of Mexico abyssal plain with the Florida Peninsula. Since the end of Late Cretaceous (Maastrichtian) time (66 Ma), the Florida Platform has undergone phases of aggradation and progradation with sea-level fluctuations. The sea-level fluctuations, together with the action of the Loop Current, are reflected in the character and continuity of Cenozoic sediments that overlie the platform.

2.5.1.1.1.1.1.3 Atlantic Offshore Continental Shelf and Slope

The Atlantic Continental Shelf is generally a continuation of the coastal plain as it dips beneath sea level at a low angle (0.1 to 0.5°) to a depth of about 150 meters (500 feet). The shelf is characterized by a shallow gradient to the southeast (Reference 313) and many shallow water features that are relicts of lower sea levels. Below 150 meters of water depth, the shelf abruptly transitions to a steeper continental slope that dips seaward at an angle of between 3 and 6° to a depth of about 1500 meters (5000 feet). At the base of the continental slope is a fan of turbidite sediments that make up the continental rise, with a slope of 0.5 to 1.0° (Reference 310), and extends to depths of 1500 to 5000 meters (5000 to 16,500 feet), which in turn transitions to the abyssal plain between water depths of 3000 and 10,000 meters (10,000 to 33,000 feet) (Reference 311).

The Atlantic coast of Florida, however, does not fit easily into the morphological characterization described above. Here, the continental shelf (defined based on water depth) is represented by the narrow Florida-Hatteras Shelf extending to 80 meters (260 feet) water depth, after which the seafloor drops precipitously to several hundred meters depth. At a depth of 400 to 1250 meters (1300 to 4100 feet), a broad, flat carbonate plateau, the Blake Plateau, forms part of the Atlantic Continental Slope. The plateau is an 8- to 12-kilometer thick sequence of flat-lying Middle Jurassic and lower Cretaceous carbonate rocks that are continuous with the limestone platform rocks underlying the Florida, Bahama, and Yucatan Platforms. The Blake Plateau and Escarpment separate the wedge of Upper Cretaceous and Cenozoic siliciclastic sediments from the Atlantic Abyssal Plain (Reference 487). The Blake Plateau represents a carbonate platform that "drowned," or ceased carbonate production in the early Cretaceous (see

discussion in Subsection 2.5.1.1.1.1.1.2) (Reference 312). In contrast to the Blake Plateau, carbonate production kept pace with subsidence across most of the Bahama Platform, and the platform's northern shallow water areas now form part of the continental shelf of the North America Plate. Therefore, the Blake Plateau and the canyons and channels that cut the Bahama Platform are considered part of the Atlantic Continental Slope, whereas the shallower portions of the Bahama Platform are considered to be part of the Atlantic Continental Shelf (Figure 2.5.1-214).

The continental shelf/slope along Florida's Atlantic coast has four distinct physiographic areas illustrated in Figure 2.5.1-214:

- Florida-Hatteras Shelf, from the shoreline to 80 meters (260 feet) water depth.
- Florida-Hatteras Slope, from 80 to 400 meters (260 to 1300 feet) water depth.
- Blake Plateau, from 400 to 1200 meters (1300 to 4000 feet) water depth.
- Blake Escarpment, the nearly vertical escarpment beginning at 1200 meters (4000 feet) water depth, formed by erosional retreat of the Blake carbonate platform (Reference 305). The Blake Escarpment connects the Blake Plateau to the Atlantic's abyssal Blake-Bahamas Basin, found at 4000 to 5400 meters (13,000 to 17,700 feet) water depth (Figure 2.5.1-214).

Sedimentary processes and sediment instability on the Florida-Hatteras Shelf and Slope appear to be related to evidence of potential megasedimentary events that have occurred from shallow water sources into deeper basins and troughs in the Gulf of Mexico, Straits of Florida, Florida Atlantic continental margin, Bahama Platform, and Caribbean region. There are indications that the “normal” processes seen today have, in the past, occurred on a much larger scale than has been often witnessed in human history. Based on evidence gathered by the DSDP and ODP, turbidite megasedimentation events have occurred with relative frequency in the Caribbean and Atlantic Ocean Basins (Reference 315).

Pilkey (Reference 315) roughly estimates the probability of recurrence of such megaturbidite events at once every 50,000 years in the largest ocean basins and more frequently in smaller basins. In the basin plains studied by Pilkey (Reference 315), hemipelagic sediments usually make up less than 20 percent of the sediment volume. Convulsive megasedimentation events flatten out the topographic irregularities formed by the deposition of small gravity flows between the large events. Large-scale events and the sediments they deposit maintain the

flat plain floor. In some instances, a single giant flow may arrive on the basin simultaneously from geographically widespread basin entry points, indicating that the initiating mechanism, probably an earthquake, was regional in scope (Reference 315). The water displacement associated with these megaturbidites likely creates tsunamis that may have affected any nearby coastal areas. The largest event measured by Pilkey in modern basin plains is the Black Shell Turbidite of the Hatteras Abyssal Plain. The Black Shell Turbidite is at least 100 kilometers^3 (24 miles^3) in volume and occurred about 16,900 years ago, based on radiocarbon dating of the youngest shell material incorporated in the deposit (Reference 316). Major events such as the Black Shell Turbidite should not be assumed to be restricted to times of lowered sea level. The 1929 Grand Banks slump, which occurred in the current interglacial period, may have involved as much as 400 kilometers^3 (96 miles^3) of material (Reference 317). This slump resulted in a turbidity current that traveled 500 kilometers (300 miles) to the Sohm Abyssal Plain, but the full areal extent of the resulting turbidite is still unknown (Reference 315). In this case, the slump and resulting tsunami were induced by the M_w (moment magnitude) 7.2 earthquake that occurred on November 18, 1929 (Reference 318). See further discussion of the Black Shell Turbidite in Subsection 2.5.1.1.5.

Subsections 2.5.1.1.2.1.1 and 2.5.1.1.2.2.3 provide discussions of megasedimentation events in the Gulf of Mexico and Caribbean regions. Subsections 2.5.1.1.5 and 2.4.6 provide discussions of tsunami hazards from seismic or non-seismogenic submarine landslides.

There is no evidence to indicate that a feature similar to the Florida-Hatteras Slope existed prior to the Eocene. Paleocene strata are deeply eroded in the subsurface, and this erosion may mark the initial appearance of the Gulf Stream in this region (see discussion of changes in oceanic circulation patterns in Subsection 2.5.1.1.1.1.1.2). During the Eocene and Oligocene, another wedge of shelf sediments prograded across the Blake Plateau but were interrupted by erosion at the end of the Oligocene. A prograding wedge of Miocene to Holocene sediments covers the erosional unconformity at the base of the Miocene. The accumulation of post-Paleocene sediment at the foot of the Florida-Hatteras Slope and seaward onto the Blake Plateau has been very slow, due in part to the reduction in sediment supply as well as to erosion by the modern Gulf Stream (Reference 487).

The Blake Plateau underlies the continental shelf off the east coast of Florida, northeast of the northwest-striking Jacksonville Fracture Zone interpreted from magnetic anomaly maps (Reference 212). The plateau is dominantly underlain by

Jurassic to Cretaceous carbonates and thought to be built upon transitional crust of African affinity, like the Florida Platform (Reference 307). The east edge of the plateau is near the Blake Spur magnetic anomaly, which may mark the boundary between transition and oceanic crust (Reference 307). The Jurassic and younger strata of the Blake Plateau are unfaulted, though an apparent left-lateral offset in the M-25 magnetic anomaly may reflect dextral shearing along the Jacksonville Fracture Zone (called the Bahama Platform-Blake Plateau boundary or Great Abaco Fracture Zone) (Reference 307) (Figures 2.5.1-206 and 2.5.1-229). Faulting here is pre-Miocene age.

At the northern terminus of the Blake Plateau lies the Blake Ridge contourite drift. The Blake Ridge contourite deposits (Figure 2.5.1-226), located off the U.S. coast, east of Charleston, South Carolina, and Savannah, Georgia, represent one of the largest methane hydrate provinces on Earth. Hornbach et al. (Reference 319) analyze high-resolution three-dimensional seismic data to map seismic indicators of concentrated hydrate and fluid flow. Their analysis reveals that the Blake Ridge gas hydrate system is significantly more dynamic than previous studies suggest, and hypothesize that fluctuating sedimentation and erosion patterns cause hydrate phase-boundary instability that triggers fluid flow. The surface morphology of Blake Ridge is controlled by the western boundary undercurrent, which erodes sediment from the eastern flank of the ridge and redeposits it on the western half. The Blake Ridge contourite deposits apparently are affected by complex ocean currents created by the overriding Gulf Stream mixing with the Western Boundary Undercurrent at intermediate bottom waters. Therefore, what appears to be an anomalously linear western margin of the Blake Ridge is caused by circulating currents, erosion, and deposition and not by faulting (Subsections 2.5.1.1.1.3.2.2 and 2.5.1.1.1.3.2.3).

Twitchell et al. (Reference 320) report numerous underwater landslide scars on the sea floor off the U.S. Atlantic coast (Figure 2.4.6-202). They find that landslide scars cover 13 percent of the sea floor on the Atlantic Continental Slope and Rise between Cape Hatteras and the Blake Plateau. Landslides can be divided into two categories based on their source areas: those with sources located in submarine canyons and those with sources located on the open Atlantic Continental Slope and Rise. The deposits from both landslide categories are generally thin (mostly 20 to 40 meters [65 to 131 feet] thick) and primarily comprise Quaternary material, but the volumes of the open-slope sourced landslide deposits can be larger (1 to 392 kilometers^3) than the canyon sourced ones (1 to 10 kilometers^3) (Reference 320).

Canyons are absent south of Cape Hatteras. Two large submarine landslides are identified south of Cape Hatteras, the Cape Fear slide and an older Cape Lookout slide that is crossed by the Cape Fear slide (References 320 and 321) (Figure 2.4.6-202). Both slides have been correlated with salt diapirism and methane hydrate layers. The Cape Fear landslide is about 100 meters (330 feet) thick, 25-kilometer (15.5-mile) wide, involving an area of about 5000 kilometers^2 (Reference 322), with a volume that is likely in excess of 200 kilometers^3 (Reference 321). Using C^{14} of sediment retrieved from the uppermost Cape Fear Slide plain, the slide is dated at early- to mid-Holocene (between 3800 to 10,000 years before present) (Reference 322). Based on multibeam bathymetry data and seismic Chirp data, Hornbach et al. (Reference 323) interpret structural controls between the salt diapirs near the Cape Fear landslide and the diapirs associated with the Blake Ridge gas hydrate field (Figure 2.5.1-226). A 40-kilometer (25-mile) long fault, imaged using Chirp data, stretches between the diapirs and may serve as a point of failure on the headwall of submarine slides.

Submarine slope failures are caused by lower sediment strength and an increase in shear stresses (Reference 324). The dissociation of natural gas hydrates found in the seafloor can cause a reduction in sediment shear strength and an increase in pore water pressure along the potential failure plane (Reference 324). Both of these mechanisms can contribute to seafloor slope failures. Hornbach et al. (Reference 323) report that at least 10 major (>100 kilometers^3) (>24 miles^3) mass wasting events have been documented in the North Atlantic Ocean during the past 40 k.y. They indicate that potential triggers of submarine slope failures may act alone or in concert include earthquakes, mechanical failure of overpressured sediments, storm waves, groundwater seepage, failure of oversteepened slopes, gas hydrate dissociation, and sea level change.

Gas hydrates are most susceptible to decomposition in response to lowered sea level if they occur in 200- to 600-meter (650- to 1970-feet) water depths. The Blake Ridge contourite is in water depths of over 2000 meters (6500 feet) (Figure 2.5.1-226). However, Paull et al. (Reference 325) and Tappin (Reference 326) indicate that an increase in submarine slumping occurred during the last glacial maximum that is consistent with the prediction that a causative relation exists between gas hydrate decomposition induced by lowered sea level and slumping frequency. A further consideration is that the Blake Ridge contourite deposits are located offshore of the Charleston, South Carolina seismic zone and large mass wasting events associated with gas hydrate decomposition could represent an additional consequence of earthquake activity there. Twichell et al. (Reference 320) conclude that large landslides may be related to the upward