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Regional Controls on the Formation of the Ancestral DeSoto Canyon by the Chicxulub Impact

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ABSTRACT

The DeSoto Canyon, a subsea canyon in the northeastern Gulf of Mexico, has a history of erosion dating back to the Cretaceous/Paleogene boundary (KPgB). The canyon resides within the DeSoto Canyon Salt Basin (DCSB), a large, basement-controlled graben formed by Jurassic rifting. The basin was the focus of deposition throughout the Late Jurassic and Early Cretaceous, although seismic profiles indicate that sedimentation was nearly uniform by the end of the Jurassic. The area was dominated by carbonate reef production during the Early Cretaceous, transitioning to siliciclastic deposition north of the canyon and pelagic, carbonate mud south of the canyon during the Cenomanian. Initial incision of the canyon is visible on reflection seismic profiles as a series of truncated reflections that outline a canyon-shaped feature on the Upper Cretaceous isochore map. The downcutting surface ties to the KPgB reflection in nearby industry wells. Although much of the early canyon was buried by Cenozoic siliciclastic deposition, it remained a zone of instability, characterized by chaotic seismic facies and common truncation of internal reflections. Smaller than the ancestral canyon, the modern DeSoto Canyon remains within the confines of the initial KPgB incision.

It is hypothesized that differential subsidence preferentially induced faulting and fracturing of the carbonate margin in front of the DCSB. Chicxulub impact-induced seismicity caused the fractured margin to collapse at the KPgB, enabling unconsolidated Upper Cretaceous sediments to fail and the ancestral canyon to form. There is no evidence that the Suwannee Current played a significant role in formation of the canyon, although there is a possibility that the channel acted as a funnel for an impact-induced tsunami that removed Upper Cretaceous sediments from within the Suwannee Channel and deposited them on the Blake Plateau to the east.

INTRODUCTION

The impact that formed the 93-mi (150-km) diameter Chicxulub crater (PASSC, 2013) on the Yucatan Peninsula is typically associated with the Cretaceous/Paleogene (K/Pg) mass extinction (Hildebrand et al., 1991). This impact also triggered a magnitude 10 to 11 earthquake (Day and Maslin, 2005) that caused the collapse of the northwestern margin of the Yucatan carbonate platform into the adjacent Bay of Campeche. The platform collapse produced a carbonate breccia averaging 1000 ft (300 m) in thickness over a 3100-mi^2 (8000-km^2) area that is the primary reservoir for the giant Cantarell oil field (Grajales-Nishimura et al., 2000; 2003; 2009). The Cuban and eastern Yucatan carbonate platform margins were also hypothesized to have undergone partial collapse due to the Chicxulub impact based on carbonate breccias deposited in the proto-Caribbean and then thrusted onto Cuba (Alegret et al., 2005; Goto et al., 2008a, 2008b; Kiyokawa et al., 2002; Tada et al., 2003; Takayama et al., 2000). Landslide and mass-wasting deposits in the western North Atlantic (Klaus et al., 2000; Norris et al., 2000; Norris and Firth, 2002) and the Baja California margin (Busby et al., 2002) were similarly hypothesized to be impact related, demonstrating the considerable extent of the seismicity. Computer modeling estimated that the Florida carbonate platform underwent more than 30 ft (10 m) of displacement based on its position about 300 mi (~500 km) from the Chicxulub impact site (Boslough et al., 1996). However, no association between collapse of the western Florida margin and the K/Pg boundary (KPgB) has been established, even though there is considerable evidence for erosion and slope failure along the Florida Escarpment (e.g., Mullins et al., 1986; Twichell et al., 1990).

One of the larger erosional features along the western Florida margin is the DeSoto Canyon, at the northern limit of the Florida Escarpment, southwest of Pensacola, Florida (Fig. 1). The sigmoidal modern canyon is approximately 25 mi (41 km) in length with a gradient ranging from 0.25° to 9° and a relief from 140 to 820 ft (43–250 m) (Harbison, 1968). Several studies have investigated the salt basin underlying the canyon (e.g., Dobson and Buffler, 1991, 1997; MacRae and Watkins, 1993, 1995, 1996), but none have directly addressed formation of the canyon. Although it has been posited that the modern canyon is merely a topographic feature created by the convergence of the Florida Escarpment and the Mississippi Fan (Coleman et al., 1991), examination of seismic profiles in the region identified a prominent, canyon-shaped erosional feature beneath the modern DeSoto

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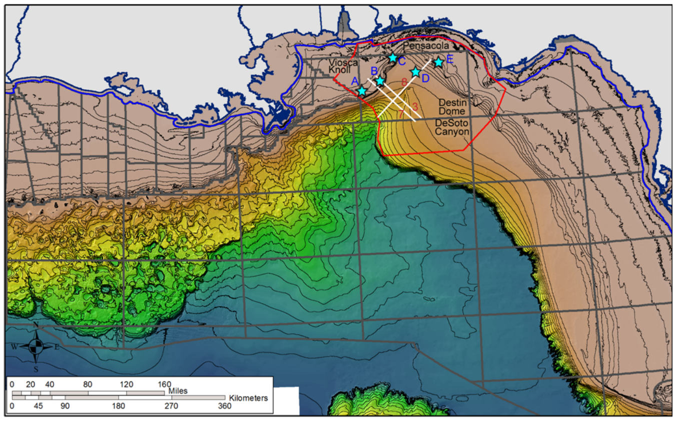

Figure 1. Bathymetric map of northeastern Gulf of Mexico with study area outlined in red. Well locations marked by stars: A, Destin Dome (DD) 529 #1; B, DD 360 #1; C, DD 56 #1; D, DD 284 #1; and E, DD 160 #1. Seismic profiles shown in Figures 3, 7, and 8 outlined in white. Bathymetric data courtesy of NOAA Coastal Data Center.

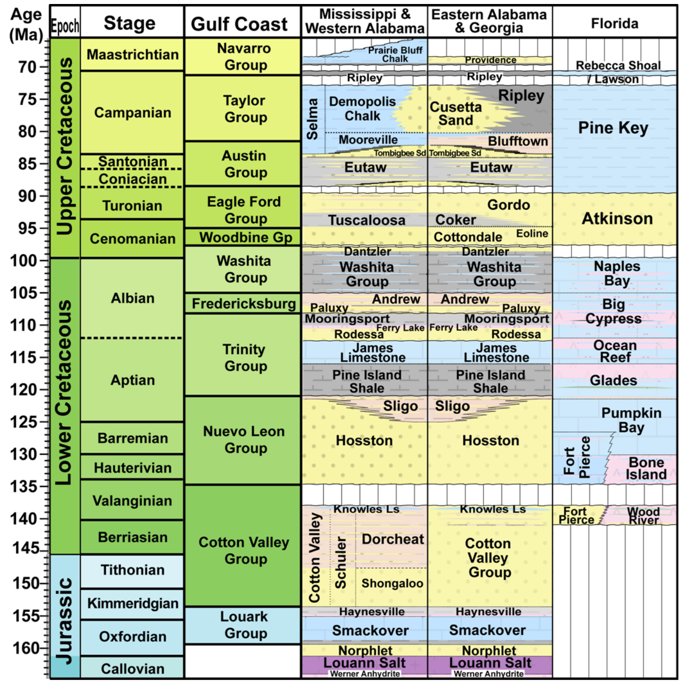

Figure 2. Absolute ages, chronostratigraphic units, and lithostratigraphic units for the Mesozoic of onshore areas in the vicinity of the DeSoto Canyon. Absolute ages from Ogg et al. (2008). Lithostratigraphy compiled from Applin and Applin (1967), Mancini and Puckett (2005), McFarlan and Menes (1991), and Sohl et al. (1991).

Canyon that correlates to the KPgB in nearby industry wells. This study of the structural and sedimentation history of the DeSoto Canyon area was initiated to determine if there is a relationship between the DeSoto Canyon and the KPgB Chicxulub impact event, and was expanded to include the Florida Escarpment and the Suwannee Channel after a connection between these features and formation of the DeSoto Canyon was identified. The results of this study suggest that seismicity generated by the Chicxulub impact caused portions of the Florida Escarpment to collapse, which induced sediment failure on the northern Florida Platform and formation of the ancestral DeSoto Canyon.

METHODS AND MATERIALS

The study area lies within portions of the Pensacola, Destin Dome, DeSoto Canyon, and Viosca Knoll protraction areas of the northeastern Gulf of Mexico (Fig. 1). Approximately 25,000 mi (~40,000 km) of seismic from 320 2D seismic lines with northwest-southeast and southwest-northeast orientations and an average spacing of approximately 1.9 mi (3 km) were utilized for the study, with infill lines at some localities. To determine when the DeSoto Canyon formed, eight surfaces were correlated: (1) the base of salt or equivalent (BSE of MacRae and Watkins, 1993), which corresponds to the basement of Buffler and Sawyer (1985) and Dobson and Buffler (1997), (2) Haynesville top, (3) Cotton Valley siliciclastics top, (4) Trinity top, (5) Albian top, (6) KPgB, (7) water bottom, and (8) top of salt (Fig. 2). These surfaces were chosen because their corresponding seismic reflections typically have strong, continuous amplitudes. Two-way-time (TWT) interpretations were imported into ArcGIS™ and then gridded and contoured using Priemere Power Tools for ArcGIS. To ascertain thickness relationships for the intervals discussed, it became apparent that isochron (time thickness) mapping of intervals might be misleading compared to isochore (vertical depth thickness) mapping. A velocity analysis was conducted with the available well information, and a simple V_0K depth conversion was applied to domain convert the TWT maps to an untied depth map. The maps were tied to five wells in the northern half of the study area to create isochore maps, with the assumption that the regional velocities analyzed with the well data are representative of the entire study area. Five isochore maps plus a basement structure map were created.

Biostratigraphic and petrophysical data released by the Bureau of Ocean Management (BOEM) and Bureau of Safety and Environmental Enforcement (BSEE) from five industry wells in the Destin Dome (DD) protraction area (DD 529 #1, DD 360 #1, DD 284 #1, DD 56 #2, and DD 160 #1 wells) were utilized for the study (https://www.data.boem.gov/homepg/data_center/paleo/paleo.asp; https://www.data.bsee.gov/homepg/data_center/other/WebStore/master.asp) (Fig. 1). The well data were tied to the seismic with synthetic seismograms created using the recorded sonic (modified by replacing water velocity and a normal compaction trend velocity down to the first sampled depth) and bulk density logs (Fig. 3).

RESULTS

Upper Jurassic

The Upper Jurassic isochore map represents the succession between the BSE and the top of the Cotton Valley siliciclastics (Fig. 4). The BSE is a high-amplitude, relatively continuous reflection separating the pre-rift sediments or crystalline basement from overlying Callovian evaporates (MacRae and Watkins, 1993). During rifting, an extensional half-graben formed within the study area, creating the DeSoto Canyon Salt Basin (DCSB) (Apalachicola Basin of Dobson and Buffler, 1991) between the Wiggins and Pensacola Arch to the north and the Southern Platform and Middle Ground Arch to the south (Figs. 3 and 5–8). Initial transgression into the DCSB during the Callovian enabled deposition of the Werner Anhydrite and Louann Salt, followed by Norphlet eolian sands and related nearshore and continental deposits and Smackover carbonates during the Oxfordian (Salvador, 1991). Salt movement occurred relatively early, as evidenced by varying thicknesses of Norphlet and Smackover related to salt roller growth (MacRae and Watkins, 1996), although Norphlet thicknesses may also be related to dune/interdune deposition. Both of these formations are generally restricted to the basement lows and onlap adjacent highs (Dobson and Buffler, 1997).

Basement topography and movement continued to exert control over sedimentation, but its influence was considerably lessened by the end of Smackover deposition (Dobson and Buffler, 1991; MacRae and Watkins, 1993). The overlying Haynesville and Cotton Valley, although thickest within the DCSB, represent the first relatively continuous deposition over the study

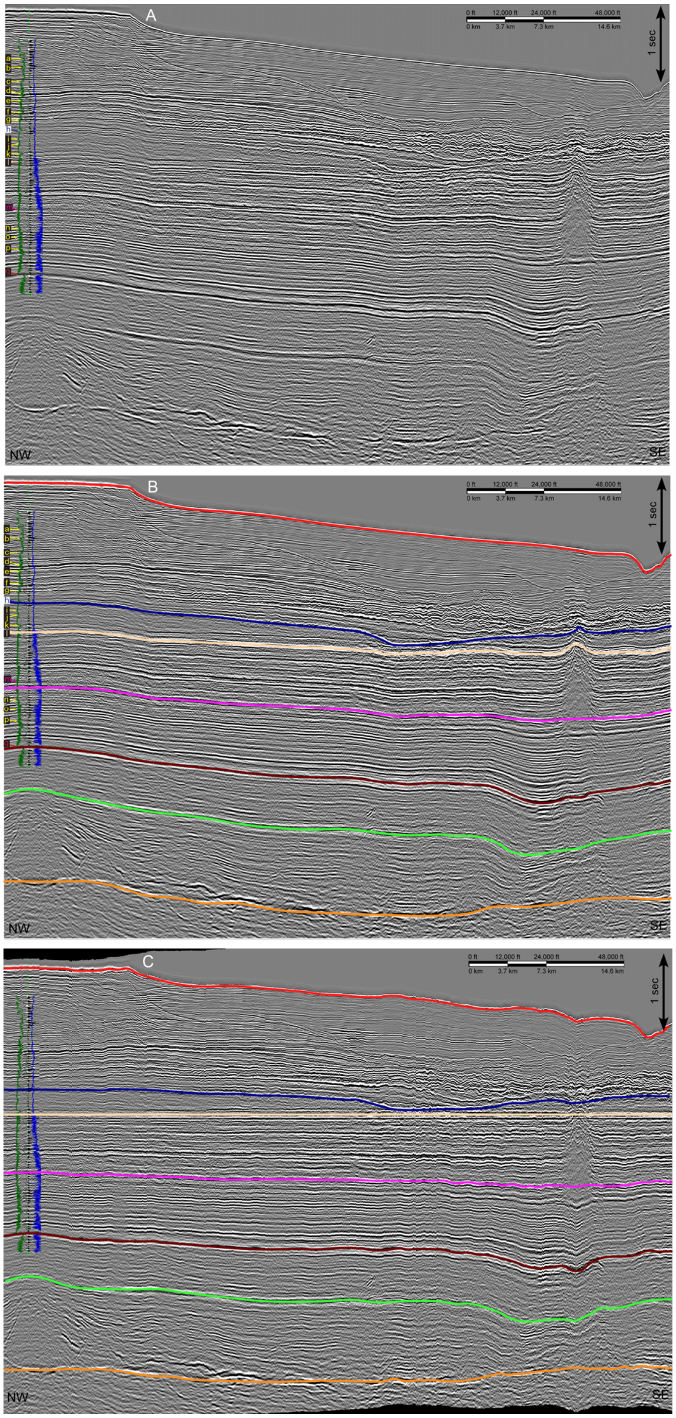

Figure 3. Northwest-southeast strike seismic profile across the DeSoto Canyon (location shown in Figure 1). (A) Uninterpreted line. (B) Interpreted line. (C) Interpreted line flattened on top of Albian. Log data tied to the line is from DD 360 #1 well: gamma ray log (green at right), synthetic seismogram (center), and resistivity log (blue at left). Chronostratigraphic and lithostratigraphic horizons identified in the well are: a, Pliocene; b, upper Miocene; c, middle Miocene; d, lower Miocene; e, Oligocene; f, upper Eocene; g, middle Eocene; h, KPgB (blue); i, Campanian; j, Turonian (Tuscaloosa); k, Washita; l, Albian (tan); m, Trinity (magenta); n, Pine Island; o, Hauterivian; p, Valanginian; and q, Cotton Valley siliciclastics (maroon). Correlated horizons are BSE (base of salt or equivalent) in orange, Haynesville top in green, Cotton Valley siliciclastics top in maroon, Trinity top in magenta, Albian top in tan, KPgB in blue, and water bottom in red. Seismic courtesy of Spectrum Geo Inc.

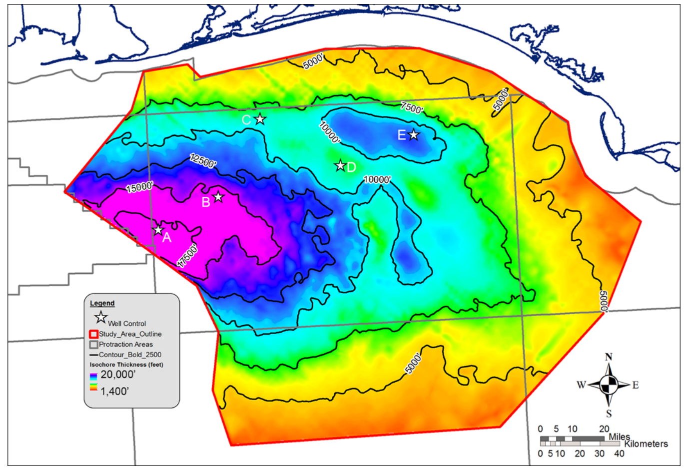

Figure 4. Isochore map of Upper Jurassic, bounded by BSE (base of salt or equivalent) and top of Cotton Valley siliciclastics. Contours in 2500-ft (760-m) intervals. Maximum thickness of 20,000 ft (6100 m) (magenta); minimum thickness of 1400 ft (425 m) (red). Well control marked by stars: A, Destin Dome (DD) 529 #1; B, DD 360 #1; C, DD 56 #1; D, DD 284 #1; and E, DD 160 #1.

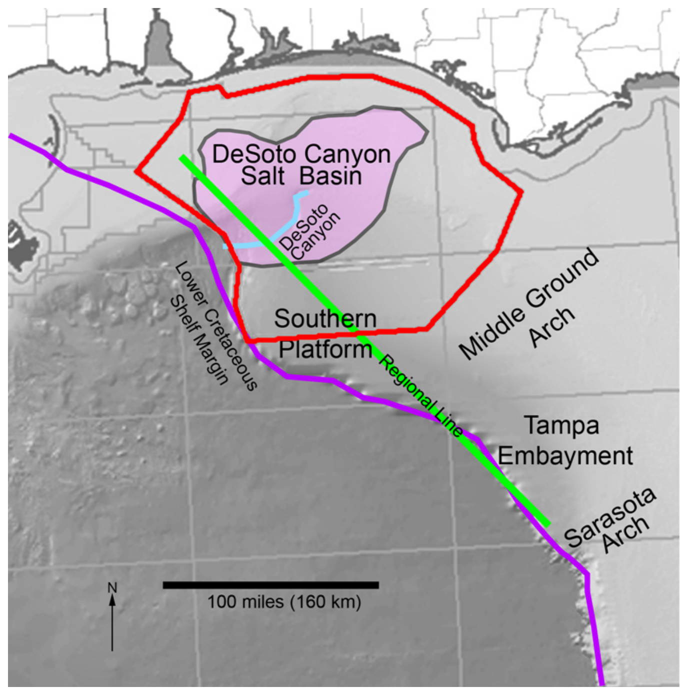

Figure 5. Map of regional structural framework in the vicinity of the DeSoto Canyon. Study area outlined in red, thalweg of the modern DeSoto Canyon in light blue, Lower Cretaceous shelf margin in purple, and regional cross-section line (Fig. 6) in green. Compiled from Dobson and Buffler (1991) and MacRae and Watkins (1996). Bathymetric data courtesy of National Oceanic and Atmospheric Administration Coastal Services Center (www.csc.noaa.gov).

area, indicating that the DCSB had been filled to the level of the Middle Ground Arch and Southern Platform by the end of the Oxfordian (Fig. 6). Although the DCSB was a topographic low during the Upper Jurassic, it was a time of rapid deposition with no sign of canyon formation.

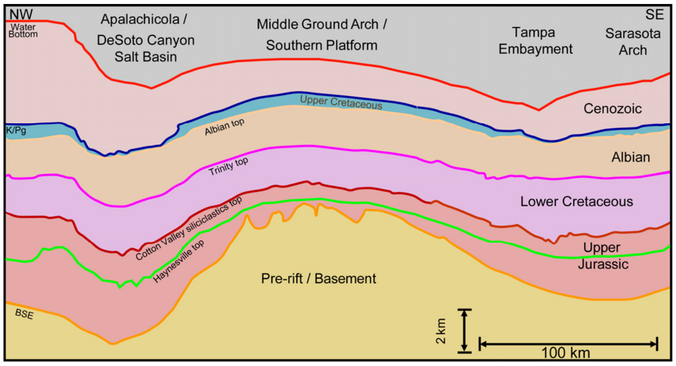

Figure 6. Northwest-southeast schematic cross-section showing positions of regional basins and topographic highs and mapping horizons utilized for isochore maps (line of section shown in Figure 5). Correlated horizons are BSE (base of salt or equivalent) in orange, Haynesville top in green, Cotton Valley siliciclastics top in maroon, Trinity top in magenta, Albian top in tan, KPgB in blue, and water bottom in red.

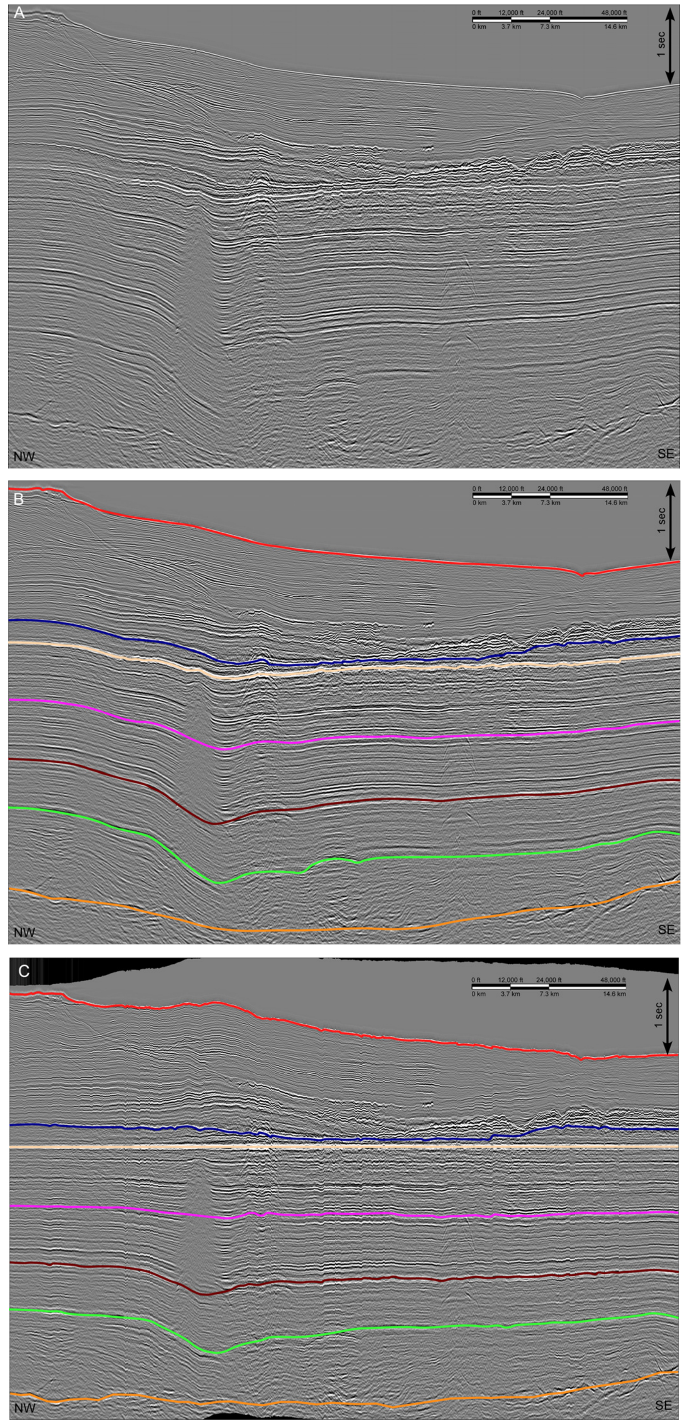

Figure 7. Northwest-southeast strike seismic profile across the DeSoto Canyon (location shown in Figure 1). (A) Uninterpreted line. (B) Interpreted line. (C) Interpreted line flattened on top of Albian. Correlated horizons are BSE (base of salt or equivalent) in orange, Haynesville top in green, Cotton Valley siliciclastics top in maroon, Trinity top in magenta, Albian top in tan, KPgB in blue, and water bottom in red. Seismic courtesy of Spectrum Geo Inc.

Lower Cretaceous

The sediments of the Knowles Limestone, Hosston, Sligo, Pine Island, James, Rodessa, Ferry Lake, and Mooringsport formations (Fig. 9) were deposited during the Berriasian to the earliest Albian (Fig. 2). Relative sea-level fell significantly during the Valanginian, producing an unconformity with subaerial exposure prior to deposition of the Hosston siliciclastics and subsequent initiation of the build-up of the Florida Platform (Sligo-equivalent) carbonate reef during the Aptian (Mancini and Puckett, 2005). Overall, sedimentation rates were high throughout the area, with thicknesses ranging from about 3000 ft (~1000 m) deposited on structural highs to more than 6000 ft (>2000 m) within the DCSB. Of note is the change of the Destin Dome anticline from a structural low in the Late Jurassic to a minor structural high in the Early Cretaceous due to downdip salt flow (MacRae and Watkins, 1992), and a shift of the depositional thick from the western edge of the study area (Fig. 4) eastward to the center of the study area within the DCSB (Fig. 9). As with the Late Jurassic, there is no seismic evidence for erosion or sediment bypass during the Early Cretaceous.

Albian

Rapid sedimentation persisted throughout the study area during the Albian, with about 3000 to 6500 ft (~1000–2000 m) of Paluxy, Andrew, and Washita sediments being deposited (Fig. 10). The carbonate platform continued to build, producing the outer margin of the Florida Platform. Although the thickest deposits remained within the DCSB, there is a linear depositional trend roughly parallel to the Florida Escarpment discernible within the Albian deposits that was only weakly developed during the Aptian (Fig. 9). There is no indication of the presence of a canyon or erosion during the Albian.

Upper Cretaceous

The Washita/Tuscaloosa boundary (middle Cenomanian) represents a profound transformation in depositional systems throughout the northern Gulf Coast (Buffler, 1991). A drop in relative sea-level during the middle Cenomanian exposed much of the shelf which was followed by incursion of the oxygen-depleted waters of the Eagle Ford/Tuscaloosa. The Florida Carbonate Platform was drowned which effectively ended carbonate production by the reef (Schlager et al., 1984). Progradation of Tuscaloosa siliciclastics from the north produced a thick sediment wedge that reached the northwestern portion of the study area (Fig. 11). Southeast of this siliciclastic wedge the sediments have been described as unconsolidated, foraminiferal/coccolith carbonate muds (Mitchum, 1978; Addy and Buffler, 1984; Gardulski et al., 1991), which are much thinner and relatively uniform in extent.

A pronounced, canyon-shaped feature is evident on the isochore map, where the Upper Cretaceous is very thin or missing

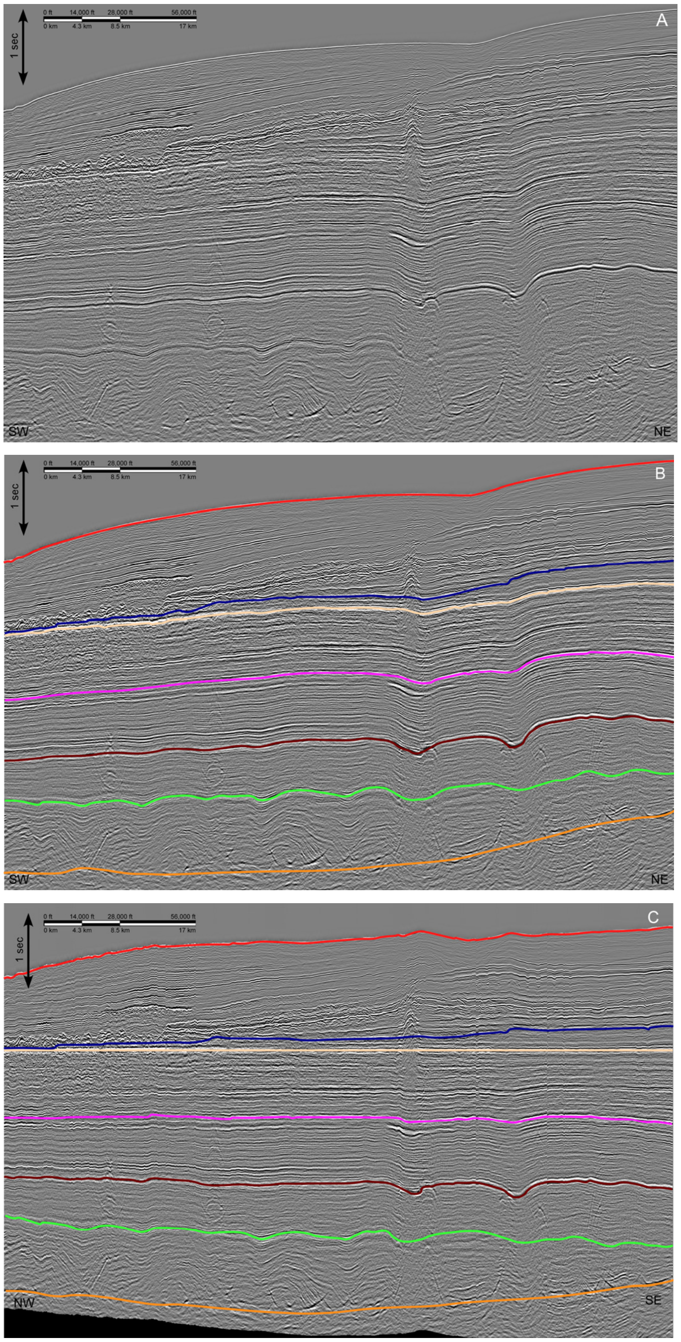

Figure 8. Northeast-southwest dip seismic profile through the DeSoto Canyon (location shown in Figure 1). (A) Uninterpreted line. (B) Interpreted line. (C) Interpreted line flattened on top of Albian. Correlated horizons are BSE (base of salt or equivalent) in orange, Haynesville top in green, Cotton Valley siliciclastics top in maroon, Trinity top in magenta, Albian top in tan, KPgB in blue, and water bottom in red. Seismic courtesy of Spectrum Geo Inc.

(Fig. 11). Truncation of Upper Cretaceous seismic reflections by the KPgB reflection can be seen on all of the seismic profiles that cross the feature (Figs. 3, 7, and 8). This feature and its associated truncation were also noted by Mitchum (1978). As the points of truncation form the general outline of the feature (Fig. 11), it is assumed to be erosional. Flattening on the underlying Albian reflection yields an estimate of 1300 ft (400 m) of missing section in the center of the feature (Fig. 12). Based on its morphology and erosional nature, this feature is interpreted to be a canyon with its initial incision occurring at the KPgB.

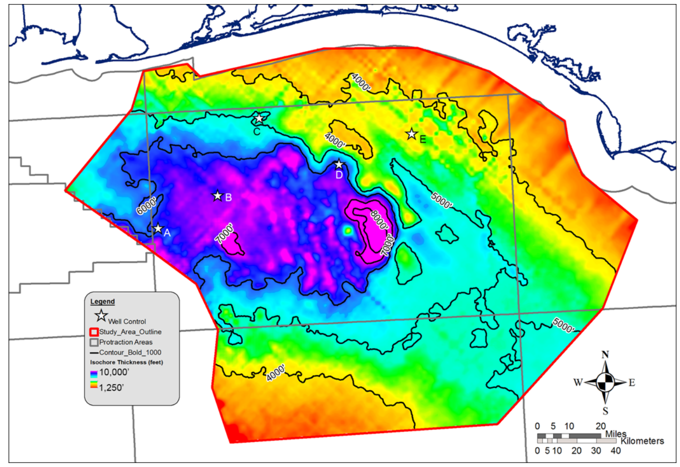

Figure 9. Isochore map of Lower Cretaceous, bounded by top of Cotton Valley siliciclastics and top of Trinity. Contours in 1000-ft (300-m) intervals. Maximum thickness of 10,000 ft (3050 m) (magenta); minimum thickness of 1250 ft (380 m) (red). Well control marked by stars: A, Destin Dome (DD) 529 #1; B, DD 360 #1; C, DD 56 #1; D, DD 284 #1; and E, DD 160 #1.

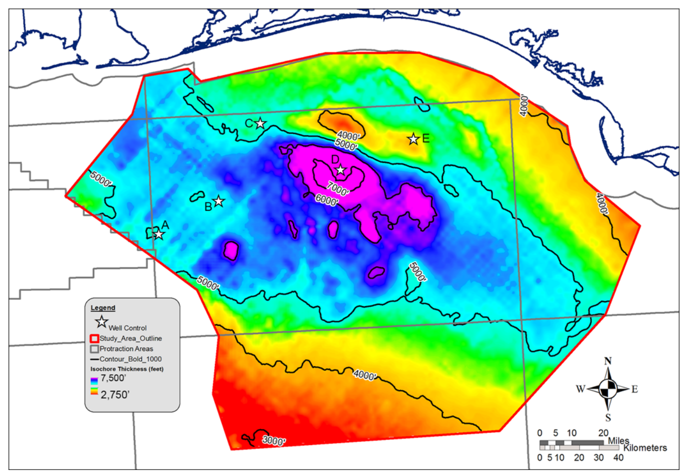

Figure 10. Isochore map of Albian, bounded by top of Trinity and top of Albian. Contours in 1000-ft (300-m) intervals. Maximum thickness of 7500 ft (2300 m) (magenta); minimum thickness of 2750 ft (840 m) (red). Well control marked by stars: A, Destin Dome (DD) 529 #1; B, DD 360 #1; C, DD 56 #1; D, DD 284 #1; and E, DD 160 #1.

Cenozoic

Deposition of plankton-rich carbonate muds continued from the Danian to the Oligocene, followed by siliciclastic deposition from the proto-Mississippi during the Neogene (Mitchum, 1978). These sediments onlap the KPgB, and buried the Florida Escarpment in the northwestern part of the study area where more than 8000 ft (>2400 m) of Cenozoic sediments are found (Fig. 13). The seismic profiles suggest that the western side of the KPgB canyon was filled by the end of the middle Eocene (Figs. 3 and 7) based on the chronostratigraphic data from the DD 360 #1 well and Mitchum (1978). The seismic reflections above the KPgB reflection are often chaotic with multiple truncation surfaces and are indicative of instability and erosion (Figs. 3, 7, and 8). Although the lower canyon shifted to the southeast (Harbison, 1968) and the overall size of the canyon shrank, the canyon system remained within the confines of the initial canyon's outer margins with the canyon head nearly stationary.

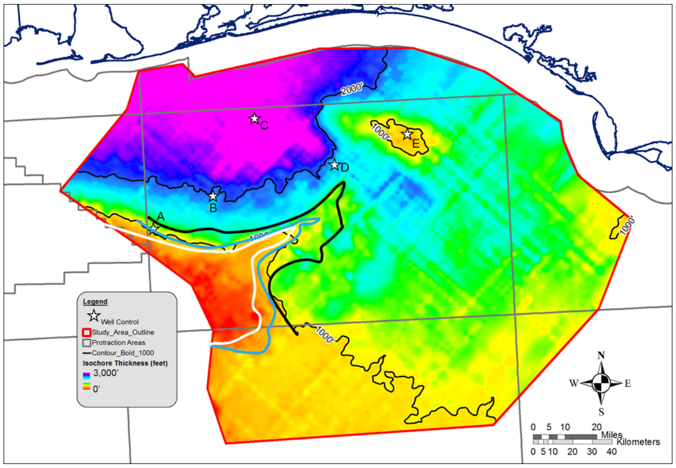

Figure 11. Isochore map of Upper Cretaceous, bounded by top of Albian and KPgB. Contours in 1000-ft (300-m) intervals. Maximum thickness of 3000 ft (915 m) (magenta); minimum thickness of 0 ft (0 m) (red). Truncation levels of Upper Cretaceous reflections marked in black (uppermost), blue (middle), and white (basal). Well control marked by stars: A, Destin Dome (DD) 529 #1; B, DD 360 #1; C, DD 56 #1; D, DD 284 #1; and E, DD 160 #1.

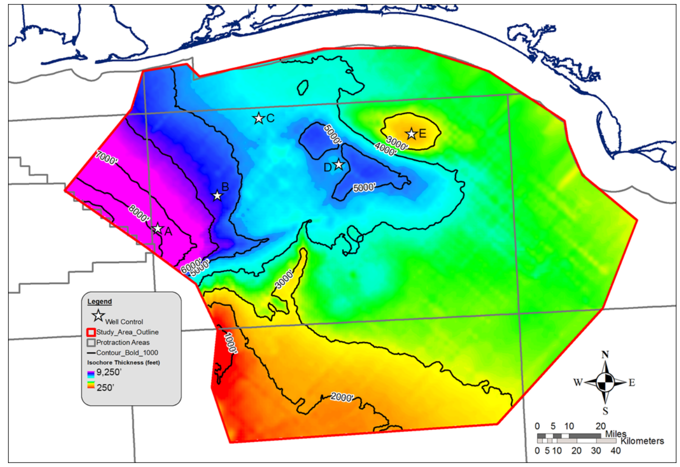

Figure 13. Isochore map of Cenozoic, bounded by KPgB and water bottom. Contours in 1000-ft (300-m) intervals. Maximum thickness of 9250 ft (2800 m) (magenta); minimum thickness of 250 ft (75 m) (red). Well control marked by stars: A, Destin Dome (DD) 529 #1; B, DD 360 #1; C, DD 56 #1; D, DD 284 #1; and E, DD 160 #1.

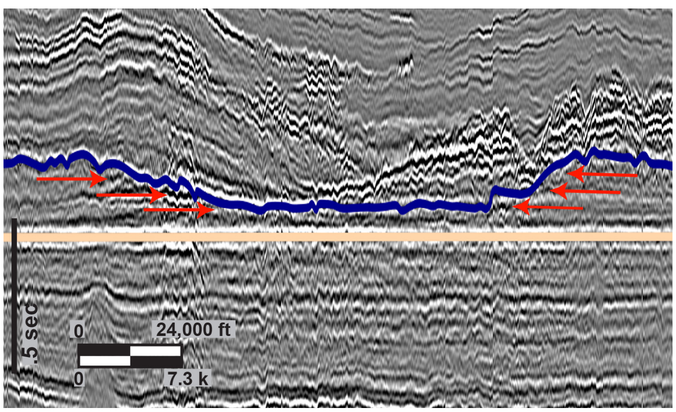

Figure 12. Magnification of strike seismic profile shown in Figure 7, flattened on Albian reflection (tan). Truncation of Upper Cretaceous reflections by KPgB reflection (blue) marked by red arrows. Seismic courtesy of Spectrum Geo Inc.

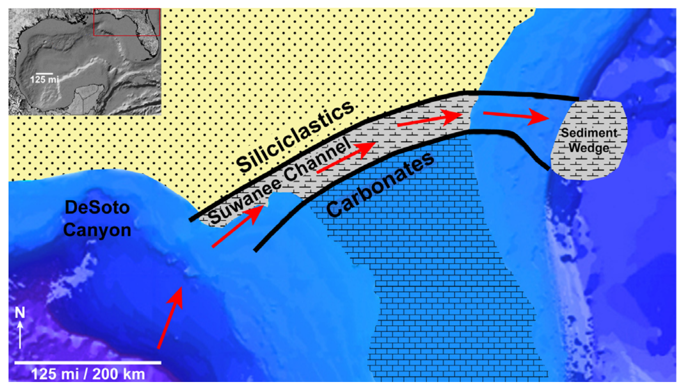

Figure 14. Map of approximate location of Suwannee Channel, dividing carbonate domain to the south and siliciclastic domain to the north, and Maastrichtian/Campanian sediment wedge on Blake Plateau (modified after Pinet and Popenoe, 1985a). Interpreted direction of Suwannee Current shown by red arrows. Bathymetric data and inset map courtesy of National Oceanic and Atmospheric Administration Coastal Services Center (www.csc.noaa.gov).

DISCUSSION

Suwannee Channel

Mitchum (1978) speculated that the DeSoto Canyon has been an area of non-deposition and erosion since the Late Cretaceous due to currents originating from the Suwannee Strait, also referred to as the Suwannee Channel (Chen, 1965) and the Suwannee Saddle (Applin and Applin, 1967). This northeast-southwest trending feature, located near the Georgia/Florida boundary, connected the Apalachicola Embayment on the Gulf of Mexico side to the Georgia Embayment on the Atlantic side (Fig. 14). The younger (Eocene to Miocene) Gulf Trough found north of the Suwannee Channel (Popenoe et al., 1987) also resides within the South Georgia Rift, leading to the more inclusive term "Georgia Channel Seaway System" (Huddlestun, 1993). The Upper Cretaceous to Eocene section within the channel is considerably thinner than corresponding sections to the north and south (Applin, 1952), especially the Navarro (Maastrichtian) section, which is missing within the channel (Hull, 1962; Applin and Applin, 1967).

Several interpretations for the presence of the Suwannee Channel have been proposed. Applin (1952) originally attributed the thinner section to erosion, which was later amended to erosion and non-deposition due to the channel being a hypothetical topographic high (Applin and Applin, 1967). Most later studies interpreted the channel as an area of slow to non-deposition, due either to the presence of a boundary between siliciclastic facies to the north and carbonate facies to the south (Hull, 1962; Chen, 1965; McKinney, 1984), or the so-called Suwannee Current flowing eastward from the Gulf of Mexico to the Atlantic, originally through the Suwannee Channel and then the Gulf Trough (Pinet and Popenoe, 1985a; Dillon and Popenoe, 1988; Huddlestun, 1993). The Suwannee Current purportedly prevented sedimentation within the channel (Chen, 1965) by moving sediments through the channel and then dumping them onto the Blake Plateau (Fig. 14). The primary evidence for this is a more than 650-ft (>200-m) thick wedge of Maastrichtian/Campanian sediments identified on shallow seismic profiles offshore of the Florida/Georgia border (Pinet and Popenoe, 1985b, their Figure 6).

Whereas a combination of the Suwannee Current and the carbonate/siliciclastic domain boundary may have prevented sedimentation within the Suwannee Channel, it does not fully explain why the Maastrichtian is missing while other formations are present, and why a thick sediment wedge is found at the mouth of the channel only within Maastrichtian/Campanian strata. There are no direct penetrations of the wedge, so its age was correlated from the Continental Offshore Stratigraphic Test (COST) GE–1 well drilled closer to shore (Pinet and Popenoe, 1985b). Spherical glass beads interpreted as possible pyroclastic grains were identified in the cuttings samples near the Maastrichtian/Campanian boundary (Rhodehamel, 1979). Their description as being perfectly spherical is somewhat suspicious, so there is a possibility that they are contamination from the mud system. However, if these beads were proven to be glass spherules derived from the Chicxulub impact and not contamination, it is possible that an impact-induced tsunami was funneled into the Suwannee Channel, eroded unconsolidated Maastrichtian sediments, and then deposited the sediment wedge as the tsunami reached the deeper, unconfined Blake Plateau.

There are also potential difficulties with Mitchum's (1978) conclusion that Upper Cretaceous sediments in the DeSoto Canyon were similarly thin due to erosion by the Suwannee Current and inhibition of carbonate productivity by adjacent siliciclastics. With cessation of reefal production during the Cenomanian, subsidence began to prevail over sedimentation, producing a rapid shift to deep-water conditions (Mitchum, 1978). Although siliciclastics may be able to "suppress" carbonate productivity in a reefal setting (Walker 1983; McKinney, 1984), the Upper Cretaceous carbonates at DeSoto Canyon are deep-water foraminiferal/coccolith ooze, which can be diluted by siliciclastic input but not "suppressed." Admittedly, the basal seismic reflections of the Tuscaloosa downlap onto the Albian reflection and some may pinch out as they reach the vicinity of the canyon (Figs. 3, 7, and 8) (Mitchum, 1978), but this does not account for the truncation, steep walls, and overall canyon morphology observed on the seismic profiles. Likewise, the Suwannee Current probably had little effect on the area, as there is no evidence that it impinged on the western Florida seafloor. It is interpreted to have been neither as powerful nor as voluminous as the modern Loop Current, and it flowed in the opposite direction of the canyon (Gardulski et al., 1991).

Erosion of the Florida Escarpment

The Florida Escarpment is the outer margin of the Florida Platform, with 3300 to 6500 ft (1000–2000 m) of relief and a slope of 40° (Twichell et al., 1991). The initial declivity was formed by buildup of carbonate reefs during the Aptian and Albian, as were similar reef margins in Texas (McFarlan and Menes, 1991). Aptian and Albian samples taken from along the escarpment were interpreted to have been deposited in low-energy environments, not the high-energy environment found at the reef margin (Freeman-Lynde, 1983; Paull et al., 1990a). Freeman-Lynde (1983) estimated that 3 to 6 mi (5–10 km) of the platform margin front had been removed by erosion based on widths of depositional facies, whereas Corso et al.'s (1987) seismic analyses yielded an estimate of 4 mi (6 km). These estimates do not include the many ravines and canyons that have incised the escarpment, particularly along the southern margin.

A number of mechanisms have been proposed for erosion of carbonate platform escarpments, mostly based on the premise of undercutting of the escarpment's base. Proposed processes for undercutting include submarine spring sapping with acidic or saline brines (Paull and Neumann, 1987; Paull et al., 1990b; Twichell et al., 1990, 1991), acidic bottom waters below the carbonate compensation depth (CCD) (Paull and Dillon, 1980), and bottom water currents (Mullins and Hine, 1989). Other mechanisms that have been suggested are self-erosion due to excess sediment (Schlager and Camber, 1986), sediment gravity flows and turbidity currents (Lindsay et al., 1975), exfoliation along exposed joints (Freeman-Lynde and Ryan, 1985), and collapse due to gravitational instability (Mullins et al., 1986) or earthquakes (Mullins and Hine, 1989).

Spring sapping was cited by several studies as the most likely source of canyon incision of the southern Florida Escarpment, although it was noted that this interpretation was based primarily on sonar-based canyon morphology and not the actual presence of active or fossil springs (Paull et al., 1990b; Twichell et al., 1990, 1991). A similar dissolution-collapse mechanism was originally hypothesized for the Yucatan and Chiapas-Tabasco platforms (Grajales-Nishimura et al., 2009), but they have now been demonstrated to have experienced collapse induced by Chicxulub impact seismicity (Smit, 1999). As mentioned above, computer simulations estimated that the magnitude 10 to 11 earthquake caused by the Chicxulub impact (Day and Maslin, 2005) produced 30 ft (10 m) of vertical displacement along the Florida Escarpment (Boslough et al., 1996). Therefore, the KPgB Chicxulub impact provides a mechanism for an earthquake-induced gravity collapse along the Florida Escarpment, similar to that hypothesized as a cause for "scalloped" margins in tectonically active margins (Mullins and Hine, 1989).

The principal evidence for Chicxulub-induced collapse of the Yucatan, Chiapas-Tabasco, and Cuban platforms are carbonate breccias containing shallow-water fossils found at the base of deep-water KPgB deposits (Smit, 1999) from southeastern Mexico and the Bay of Campeche (Grajales-Nishimura et al., 2000, 2003, 2009), Guatemala (Stinnesbeck et al., 1997), Belize (Smit, 1999), and Cuba (Alegret et al., 2005; Goto et al., 2008a; 2008b; Kiyokawa et al., 2002; Tada et al., 2003; Takayama et al, 2000). Underneath the spherule-rich KPgB layer at Deep Sea Drilling Project (DSDP) Site 540 is a 150-ft (45-m) thick deposit containing shallow-water limestone clasts with Cenomanian fossils (Premoli Silva and McNulty, 1984) described as the "pebbly mudstone" unit (Alvarez et al., 1992) (Fig. 15). Bralower et al. (1998) tentatively included the "pebbly mudstone" unit as part of the KPgB deposit, but this connection has been questioned due to the unit's lack of impact ejecta (Goto et al., 2008a). Smit (1999) speculated that the "pebbly mudstone" unit originated from mass wasting of the Campeche (eastern Yucatan) platform, but this is unlikely as Site 540 is on the Florida side of the Straits of Florida. A thick, acoustically transparent "sediment wedge" identified on a seismic profile shot across the Florida Escarpment and through Site 540 (Line SF–4, Schlager et al., 1984, their Figure 14) lies directly underneath the mid-Cretaceous unconformity ("MCU") is thickest below the escarpment, and pinches out in the vicinity of Site 540. Therefore, the "pebbly mudstone" unit of Site 540 is considered to be the downdip portion of a "sediment wedge" originating from collapse of the Florida Escarpment at the KPgB, corresponding to the carbonate breccias associated with the KPgB collapse of the Yucatan, Chiapas-Tabasco, and Cuban platforms.

The connection between Gulf of Mexico erosion and the KPgB has been obscured by the erroneous labeling of the high-amplitude KPgB deep-water seismic reflection as the "MCU" (Dohmen, 2002; Denne at al., 2013). Along the margins and basement highs of the Gulf of Mexico, including the Florida Escarpment, the "MCU" reflection is characterized by truncation of underlying reflections and onlap of overlying reflections, signifying a "turning point" in the Gulf of Mexico's sedimentary history (Schlager et al., 1984). Corso et al. (1987) found that the truncation of reflections they interpreted as toe-of-slope facies at the base of the Florida Escarpment were all associated with the "MCU" reflection, with no sign of truncation of the "MCU" by younger reflections. Faust (1990) found that truncation of underlying reflections by the "MCU" extended into the basin approximately 125 mi (200 km) beyond the Florida Escarpment. Although the "MCU" reflection is concordant with underlying reflections in the basin center (Faust, 1990), an unconformity representing a hiatus of at least 9 m.y. is present at all locations drilled by industry that have fully penetrated the KPgB deposit (Denne et al., 2013).

Erosion of the Florida Escarpment has likely been produced by several mechanisms, such as spring sapping. Considerable erosion occurred after the KPgB (Mullins et al., 1986; Paull et al., 1991), as made evident by deposition of possible mass-wasting deposits directly above the KPgB (Freeman-Lynde, 2002) and collapse and erosion during the Miocene due to gravi-

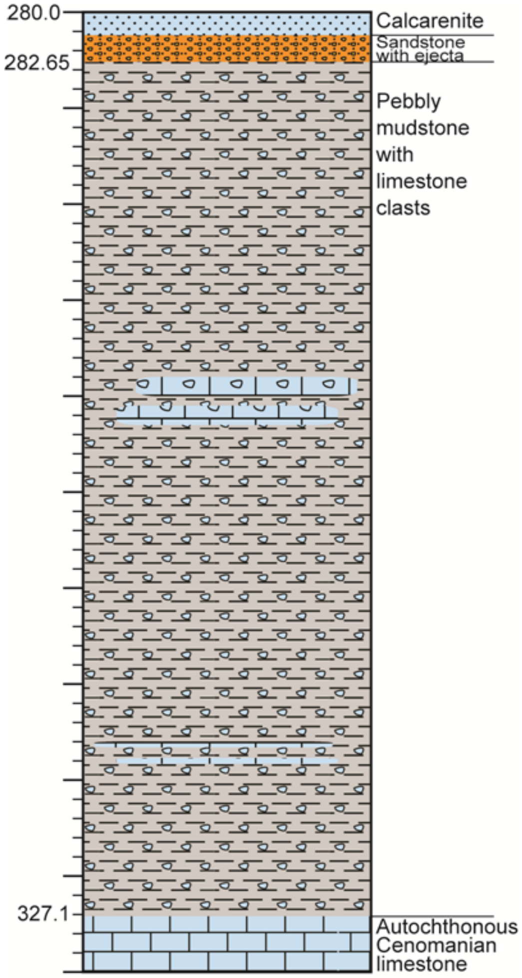

Figure 15. Lithology of the KPgB deposit from DSDP Site 540 based on core descriptions in Buffler et al. (1984), Alvarez et al. (1992), and Goto et al. (2008a). Core depths are in meters.

tational instability and accelerated current flow (Mullins et al., 1986; Snedden et al., 2012). Differential subsidence may also play a role in the extent and morphology of erosion along the escarpment (Twichell et al., 1996). However, based on the widespread erosion found at the KPgB throughout the Gulf of Mexico, the collapse of similar carbonate platform margins at the KPgB, and the presence of a carbonate breccia at the base of the KPgB deposit at Site 540 that ties to a thick "sediment wedge" originating from the Florida Escarpment, it is hypothesized that seismic shock waves induced by the Chicxulub impact caused extensive collapse of the Florida Platform.

Formation of the DeSoto Canyon

The DeSoto Canyon area was a transition zone between the predominantly progradational, narrow Early Cretaceous shelf, with dips of less than 10°, to the northwest and the steeply-dipping, aggradational, wide shelf to the southeast (Corso et al., 1989). The Cretaceous margin near the DeSoto Canyon also has pervasive shallow faulting (e.g., Corso et al., 1989, their Figure 7). These faults penetrate the KPgB reflection but not overlying reflections, so they were active at the KPgB but had become inactive before Cenozoic sediments began to bury the margin. Twichell et al. (1996) concluded that similar faulting of the margin in the Tampa Embayment and South Florida Basin was due to fracturing of the carbonates by differential subsidence, which was responsible for the greater amount of erosion found along those portions of the margin. The considerably thicker sediment load deposited within the DCSB also caused it to subside at a greater rate than adjacent highs, producing a broad topographic low (Fig. 6) and the faulted margin.

No evidence for widespread erosion and re-deposition of unconsolidated sediments exists on the Florida Platform at the KPgB, unlike on the slope and basin floor (Denne et al., 2013). Therefore, the erosion of as much as 1300 ft (400 m) of unconsolidated Upper Cretaceous carbonate mud within the canyon is unlikely. Upper Cretaceous rocks are uniformly thin close to the margin along the escarpment (Fig. 11), as are Cenozoic strata (Fig. 13). Therefore, this thinning may be a depositional and not an erosional feature. To account for the erosional focus at DeSoto Canyon during the KPgB requires a combination of factors produced by regional tectonics, sediment loading, depositional systems, and paleoceanography.

The following scenario is hypothesized for the formation of the DeSoto Canyon. Rifting of the Gulf of Mexico created accommodation within the DCSB for large amounts of Jurassic sediments while the adjacent Southern Platform, Middle Ground Arch, Pensacola Arch, and Wiggins Arch remained sediment starved until the Early Cretaceous (Figs. 4 and 6). Thick Lower Cretaceous siliciclastics and carbonates were deposited over the entire study area, but the DCSB continued to be the focus of deposition due to subsidence and salt movement (Figs. 9 and 10). The western Florida Platform south of the DeSoto Canyon was dominated by aggrading carbonate reefs during the Early Cretaceous, producing a relatively flat-lying platform behind the steep, rimmed, carbonate margin known as the Florida Escarpment. A mixture of carbonates and siliciclastics produced a margin with gentler dips to the north of the canyon, with the transition between these two margin styles occurring in front of the DCSB (Corso et al., 1989). Subaerial exposure and drowning of the platform during the Cenomanian/Turonian transgression produced a change in depositional systems, ending reefal carbonate production on the outer platform (Schlager et al., 1984) and allowing subsidence to outstrip sedimentation. The region north of the DeSoto Canyon was initially loaded with Tuscaloosa siliciclastics prior to pelagic deposition (Fig. 11), whereas south of the canyon pelagic, plankton-rich carbonate muds and thin turbidites dominated (Gardulski et al., 1991). The DCSB, loaded with thick Jurassic to Lower Cretaceous sediments, subsided at a greater rate than nearby topographic highs, producing a topographic low and causing the carbonate margin to preferentially fracture and then fault (Fig 16a). The platform experienced 30 ft (10 m) of vertical movement produced by seismic shock waves from the Chicxulub impact on the nearby Yucatan Platform at the KPgB (Boslough et al., 1996). Although the entire western platform may have undergone some margin collapse, the greater density of fracture and faults in front of the DCSB amplified the effects of the shock waves, causing substantial collapse of the margin. The unconsolidated Upper Cretaceous sediments at DeSoto Canyon failed due to collapse of the margin front and seismic-induced instability. This did not occur on the adjacent highs because of the lower dips behind rimmed margins and lower incidence of margin collapse. A tsunami may have been funneled into the Suwannee Channel updip of the DeSoto Canyon, but its erosive effects are interpreted to have been restricted to the channel. Headward erosion caused the canyon to retrograde (Pratson and Coakley, 1996; Tripsanas et al., 2008) back to the relatively flat-lying portion of the platform in front of the Destin Dome (Fig. 16b), producing the steep scarps observed on seismic profiles (Figs. 3, 7, and 8). Erosion or sediment bypass continued until the Eocene, when clastic sediments from the north began to fill in the northern side of the canyon, shifting the thalweg southward. The canyon remained a zone of instability throughout the Cenozoic, supplemented by ongoing subsidence and high sedimentation rates during the Neogene, prolonging its existence to the Recent (Figs. 13 and 16c).

SUMMARY AND CONCLUSIONS

Analysis of a seismic profile grid tied to data from industry wells in the vicinity of the modern DeSoto Canyon revealed that the canyon is the remnant of an older, much larger, erosional

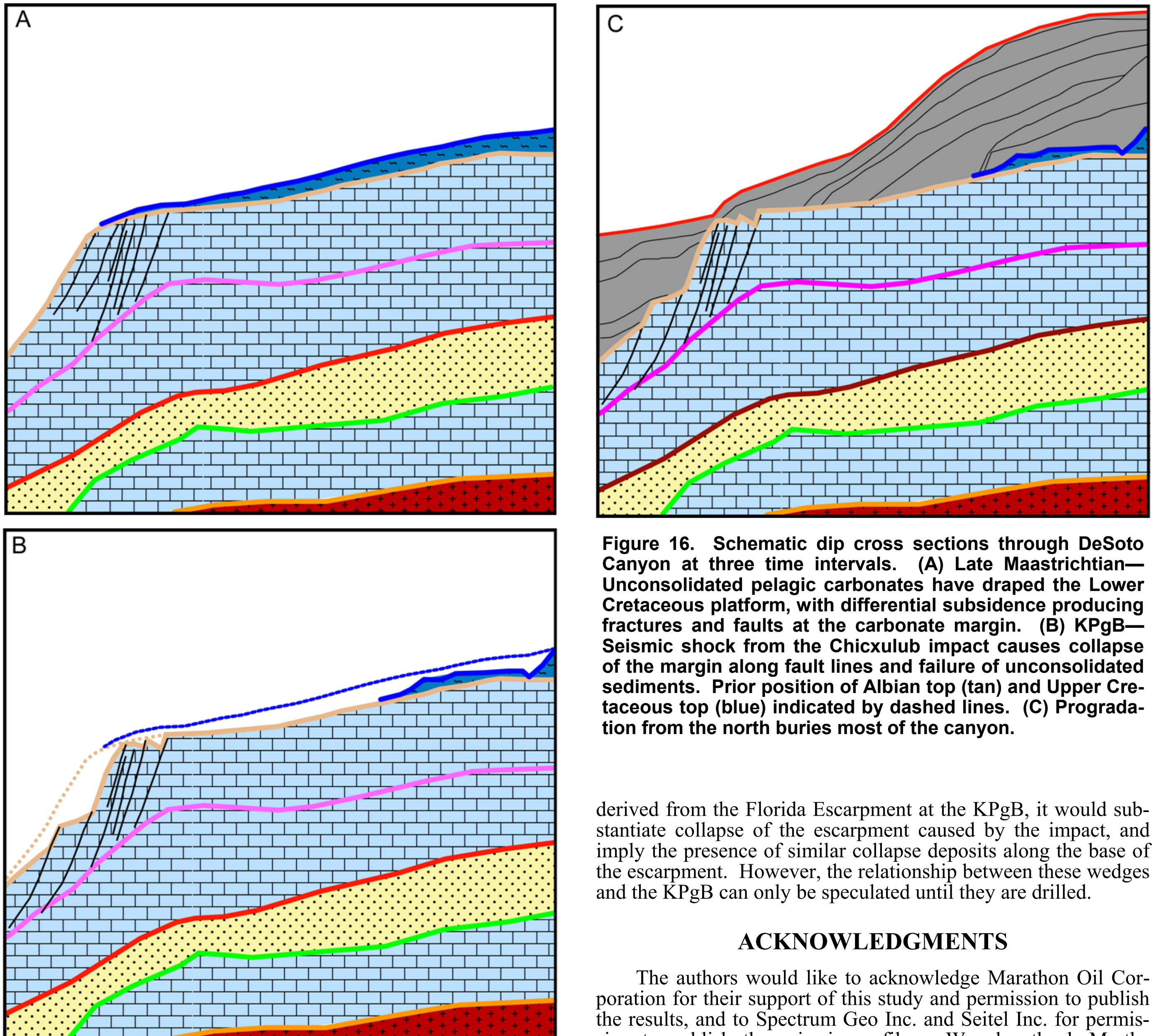

Figure 16. Schematic dip cross sections through DeSoto Canyon at three time intervals. (A) Late Maastrichtian—Unconsolidated pelagic carbonates have draped the Lower Cretaceous platform, with differential subsidence producing fractures and faults at the carbonate margin. (B) KPgB—Seismic shock from the Chicxulub impact causes collapse of the margin along fault lines and failure of unconsolidated sediments. Prior position of Albian top (tan) and Upper Cretaceous top (blue) indicated by dashed lines. (C) Progradation from the north buries most of the canyon.

canyon, and not merely a recent topographic feature related to the convergence of the Florida Escarpment and the Mississippi Fan (Coleman et al., 1991). Previously, this erosion was thought to be related to either the Suwannee Current or the siliciclastic/carbonate boundary (Mitchum, 1978). These factors may have played a role in the formation of the ancestral canyon, as did basin morphology and differential subsidence, but it is concluded here that it is not a coincidence that the initial incision of the DeSoto Canyon corresponds to the KPgB, and therefore is a product of seismicity induced by the Chicxulub impact.

In the scenario hypothesized above, assumptions were made about the association between the formation of sediment wedges identified on seismic by previous researchers and the KPgB. This includes the sediment wedge downdip of the Suwannee Channel on the Blake Plateau (Pinet and Popenoe, 1985b) and the wedge updip of DSDP Site 540 in the Florida Straits (Schlager et al., 1984). Although not essential to this scenario, these wedges are of considerable importance in understanding the effects of the Chicxulub impact. Proving a relationship between the Suwannee wedge and the KPgB would offer insight on the Suwannee Channel and could provide convincing evidence of a mega-tsunami. If the Florida Straits wedge is shown to have derived from the Florida Escarpment at the KPgB, it would substantiate collapse of the escarpment caused by the impact, and imply the presence of similar collapse deposits along the base of the escarpment. However, the relationship between these wedges and the KPgB can only be speculated until they are drilled.

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