OCS Study
BOEM 2020-004

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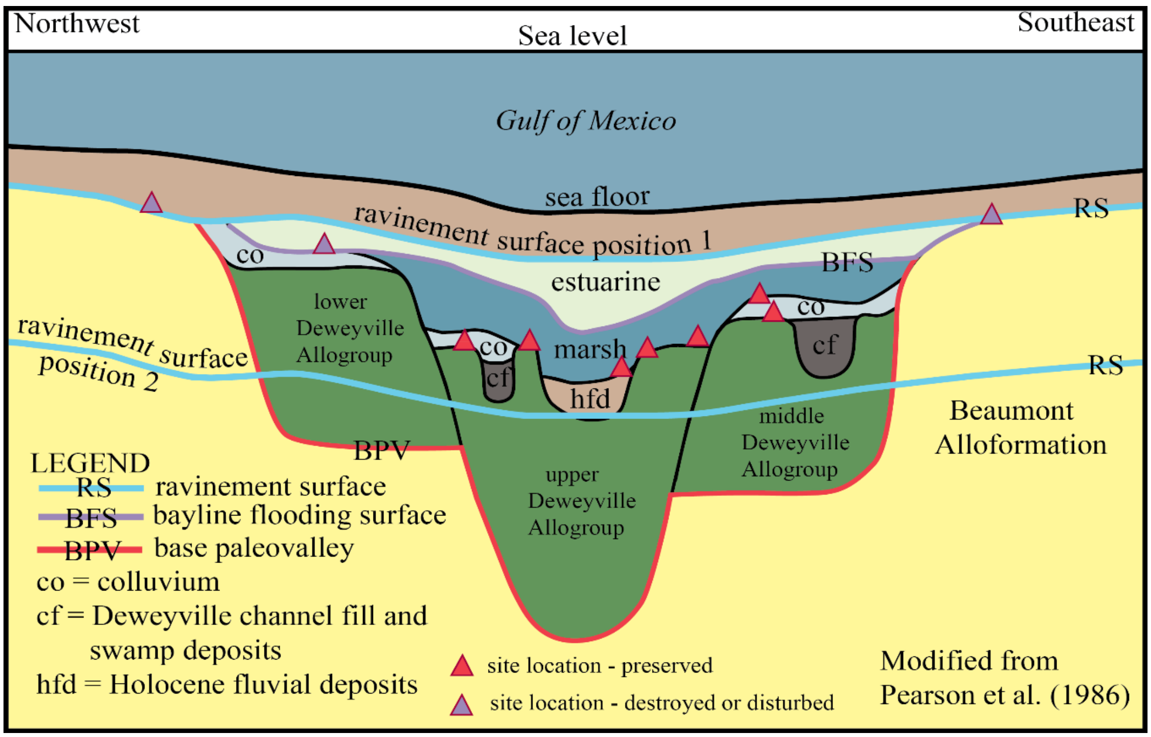
Response of Later Quaternary Valley Systems to Holocene Sea Level Rise on the Continental Shelf Offshore Louisiana: Preservation Potential of Paleolandscapes



U.S. Department of the Interior
Bureau of Ocean Energy Management
Gulf of Mexico OCS Region

Cooperative Agreement
Coastal Marine Institute
Louisiana State University

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Response of Later Quaternary Valley Systems to Holocene Sea Level Rise on the Continental Shelf Offshore Louisiana: Preservation Potential of Paleolandscapes

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DISCLAIMER

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REPORT AVAILABILITY

To download a PDF file of this report, go to the US Department of the Interior, Bureau of Ocean Energy Management Data and Information Systems webpage (http://www.boem.gov/Environmental-Studies-EnvData/), click on the link for the Environmental Studies Program Information System (ESPIS), and search on 2020-004. The report is also available at the National Technical Reports Library at https://ntrl.ntis.gov/NTRL/.

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1. Introduction

Understanding the complex response of coastal landforms to sea level rise is important for evaluating preservation potential of paleolandscapes and identifying sand resources on the Outer Continental Shelf (OCS). During the last glacial maximum, which ended approximately 18,000 years before present, sea level in the northern Gulf of Mexico was approximately 120 m lower than it is today and the lowstand shoreline was positioned approximately coincident with the shelf edge. As a result, intrabasinal coastal streams extended across the shelf within deep incised valleys separated by interfluves. As sea level rose, the incised valleys were infilled and coastal systems migrated landward across the shelf. This valley infilling consisted of both aggrading fluvial deposits and backstepping (migrating landward with sea-level rise) estuarine systems that developed within the flooded river valleys. As sea level rose, the retreat of coastal and bay shorefaces ultimately encountered and significantly eroded valley fill deposits and flanking interfluve surfaces. This produced regionally persistent ravinement surfaces. Also, shoreface retreat and erosion significantly affected the surfaces and sediments of any attendant deltaic systems. Typically, the complex erosional and depositional responses to sea level rise associated with shoreface retreat resulted in all but the basal portions of coastal and estuarine deposits being truncated and reworked by wave and tidal currents. There is little to no preservation potential of coastal lithosomes seaward of the modern shoreface. However, there may be unique situations associated with fluvial aggradation and subsequent estuarine deposition within paleovalleys that could potentially provide an opportunity for preservation of buried geomorphic surfaces and intact prehistoric landscapes. Preserved fluvial and estuarine sediments as part of paleovalley fills on the OCS is important because they present rare opportunities for preservation of paleolandscapes and associated cultural sites that survived sea level rise because they accumulated at elevations below the depths of shoreface ravinement. Moreover, the fluvial sediments underlying these terraces are inferred to be locally sand-prone channel bed and point bar deposits, a potential sand source for coastal restoration projects.

Since the mid-Pleistocene (approximately 900 ka), the periodic growth of continental ice sheets has reduced the volume of seawater in the ocean basins. These reductions in the volume and thermal contraction of seawater have repeatedly lowered global, eustatic sea levels on the order of 120 m below present levels during periods of maximum glaciation (Chappell et al. 1996) (Figure 1.1). Due to the effect of glacial isostatic adjustment, relative sea level was lowered on the order of about 90 m below the present sea level (Bart and Ghoshal 2003, Roberts et al. 2004). During the last glacioeustatic cycle, Marine Isotope Stages (MIS) 5 to 1, sea level fell and rose within the Gulf of Mexico as ice sheets in North and South hemispheres expanded and shrank. At the beginning of the last glacioeustatic cycle about 120 ka (MIS 5), sea level in the Gulf of Mexico was at a maximum highstand of ~6 m above present. Subsequently, it episodically fell until the middle of MIS 4, at about 70 ka when it was approximately 80 m below present sea level. Then sea level rose to about 15 m below present sea level during the early part of MIS 3. Following this intermediate highstand, sea level again episodically fell to about 120m below present by about 22 ka during MIS 2. This lowstand lasted from about 22 to 16 ka. Between 16 and 4 ka, sea level rose over 100m as continental ice sheets melted (Anderson et al. 1992 and 2004, Blum and Price 1998) (Figure 1.2).

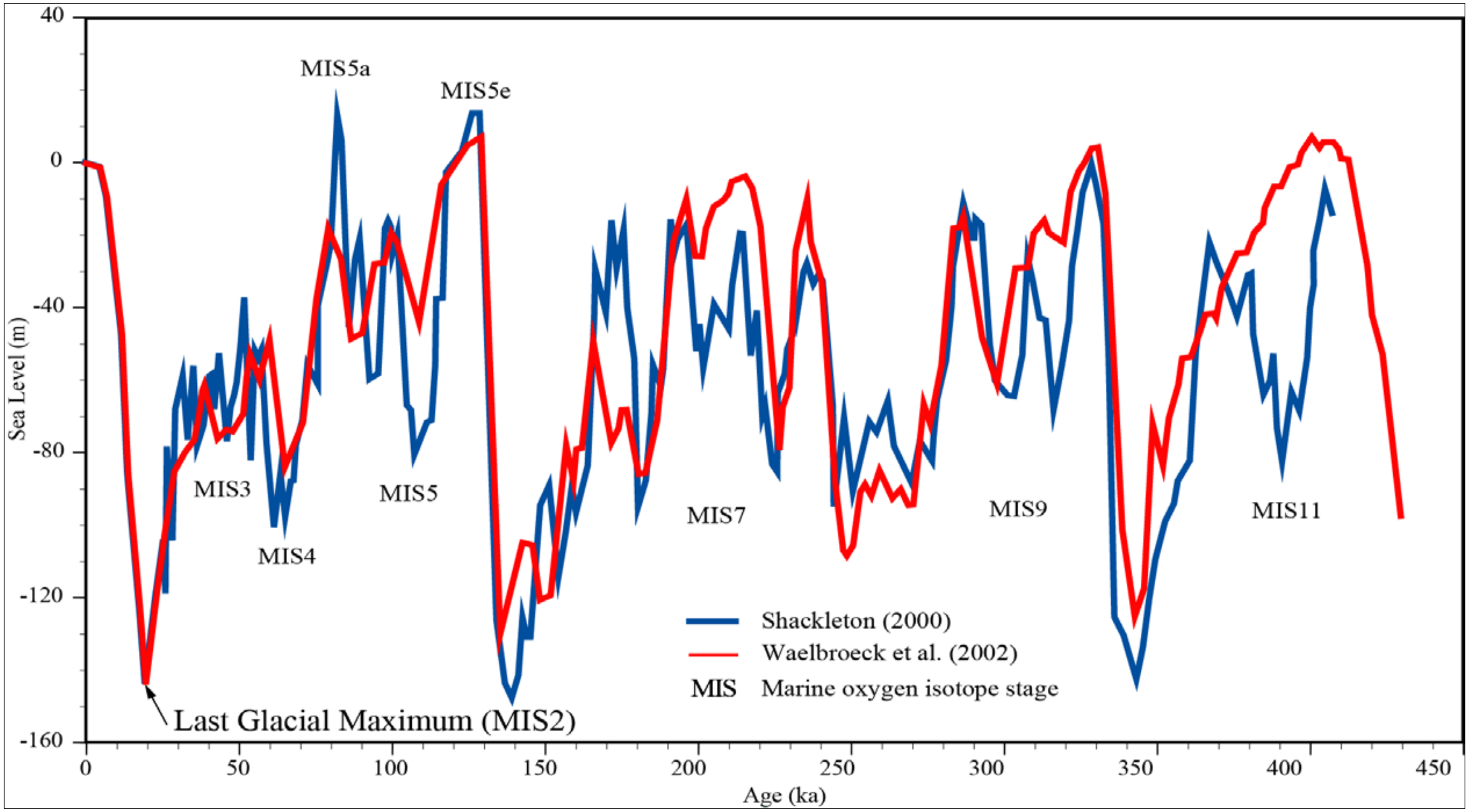

Figure 1.1. Late Pleistocene eustatic sea level curve constructed from Pacific Gas and Electric Company (2011).

Shackleton (2000)'s curve is based on marine oxygen isotope record corrected using Vostok ice core oxygen isotope record. The curve of Waelbroeck et al. (2002) is based on regression analysis of relative sea-level data from coral reef and oxygen isotope data.

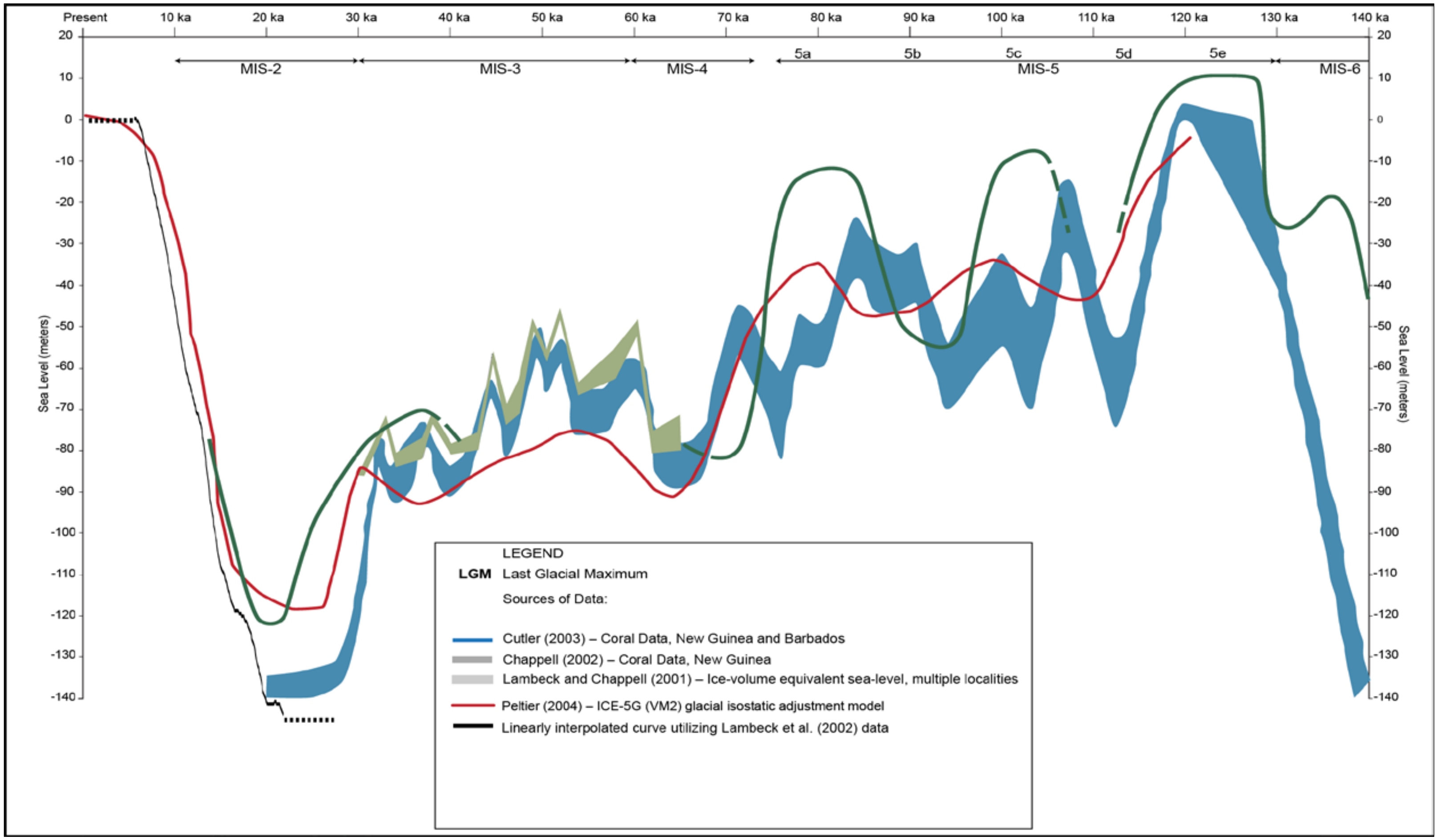

Figure 1.2. Eustatic sea level curve for the last glacial-interglacial cycle according to, compiled, and modified from Lambeck and Chappell (2001), Chappell (2002), Lambeck et al. (2002), Cutler (2003), Peltier (2004), and Pacific Gas and Electric Company (2011).

Within the Gulf of Mexico, these sea-level fluctuations have resulted in drastic changes in the position of the shoreline. During the peak of the MIS 5e highstand about 120 ka the Gulf of Mexico shoreline was several kilometers inland of the present shoreline along the Ingleside relict shoreline trend in Texas and Southwest Louisiana. In Southeast Louisiana, its location is uncertain, although during MIS 5a the Gulf of Mexico's shoreline lay along the Ponchatoula strandplain. Following MIS 5, the alternating lowstands and highstands have resulted in the development of entrenched stream valleys and deposition of shelf-phase deltas on the Louisiana Continental Shelf. These and lesser fluctuations of base level also have resulted in well-defined unconformity-bounded, allostratigraphic, packages of fluvial sediments and their associated terrace surfaces within these entrenched valleys (Anderson et al. 1992 and 2004, Blum and Price 1998, Blum and Törnqvist 2000).

Although their ages of deposition predate the established timing of prehistoric human occupation in North America, many of the terraces within coastal plain valleys and paleovalleys were subaerial landscapes open to prehistoric occupation during MIS 2. In some cases, fluvial, estuarine, and eventually marine deposits buried these paleolandscapes and associated floodplains as sea level rose between 16 and 4 ka. Lying within entrenched valleys beneath shoreface erosion and the resulting ravinement surface, these surfaces and any associated archaeological deposits might have survived the marine transgression associated with postglacial sea-level rise (Blum and Price 1998, Gagliano et al. 1982, Pearson et al. 1986, Stright 1990 and 1995).

In addition, the Pleistocene fluvial packages that underlie or are laterally bounded by these terrace deposits are a potential sand resource for long-term coastal restoration and beach nourishment programs. The fluvial sediments underlying these terraces are inferred to be locally sand-prone channel bed and point bar deposits. Therefore, some potential sources of sand from the Louisiana-Texas Continental Shelf are also areas that potentially contain preserved, buried paleolandscapes. Thus, it is important for archaeologists studying the continental shelf to understand the stratigraphy, geology, geochronology, and geoarchaeology of the deposits filling the incised valleys and the attendant erosional surfaces pervasive on the Louisiana-Texas Continental Shelf (Gagliano et al. 1982, Pearson et al. 1986, Stright 1990 and 1995).

1.1 Project Overview and Objectives

The objective of the study was to develop a conceptual model for Late Pleistocene to recent modification of the northern Gulf of Mexico coastal plain during the Holocene transgression. The purpose of this model is to evaluate the preservation potential of coastal plain paleosurfaces, e.g., uplands, channel belts, floodplains, terraces, and barrier islands as sea level submerged the Late Pleistocene coastal plain and transformed it into the modern continental shelf. Eventually, it should be possible to map preserved paleosurfaces and evaluate the potential existence of prehistoric cultural resources on the OCS. In addition, the same conceptual model can be used to evaluate the potential existence of sand resources on the OCS suitable for restoration projects and quantitatively predict their distribution. Essentially the same geologic factors that govern the preservation or destruction of paleosurfaces within the OCS also influence distribution of sand within shelf-coastal systems. Finally, this study developed a basic stratigraphic framework model that can be used to inform future geologic and high resolution geophysical data collection to address questions about paleosurfaces and sand resources.

1.2 BOEM Information Needs Addressed

BOEM is tasked with making available OCS resources for expeditious and orderly development in a manner that safeguards environmental and cultural resources. A number of uses of Gulf of Mexico resources, such as fishing, energy development, recreation, sand dredging, affect underwater cultural resources, such as historic shipwrecks and submerged prehistoric sites. However, little is known about early human occupation on the continental shelf and the potential for site preservation. Sedimentological and geoarchaeological studies have shown that deposits on continental shelves respond to modern hydrological processes. Moreover, during marine transgression significant erosion by waves and tides as the coastal system migrates up the shelf results in significant erosion of the shelf. This observation renders invalid the concept that the continental shelves consist of an unmodified, drowned coastal plain. However, sedimentary deposits and subaerial paleosurfaces, with which intact and potential significant cultural deposits might be associated, can be preserved under predictable circumstances (Coastal Environments Inc. 1977, Belknap and Kraft 1981, Kraft et al. 1983, Pearson et al. 1986, and Dix et al. 2008). Using such circumstances, this study provides a methodology for managing conflicts between the preservation of prehistoric cultural resources and resource extraction by providing a basis for identifying and eventually mapping paleosurfaces with which significant cultural resources might be associated. The same methodology also provides a basis for mapping potential sand resources for use in coastal restoration.

Furthermore, from an academic, archaeological point of view, the need to reconstruct submerged paleolandscapes is multi-fold (Dix et al. 2008). First, for the accurate interpretation of the terrestrial record of the modern coastal plain, it is essential that archaeologists understand the contemporaneous paleogeography, including paleoshoreline locations and configurations, of now-submerged portions of prehistoric coastal plains. Second, the archaeological, paleontological and paleobotanical potential of the submerged paleolandscapes must be understood both in terms of primary cultural deposits that are preserved as a result of submergence and secondary cultural materials that have been re-worked as the result of the latest cycle of exposure and marine inundation.

2 Methodology

2.1 Background and Archival research

The first phase of this study consisted of collecting the existing subsurface data for the Southwest Louisiana Outer Continental Shelf (OCS) study area (Figure 2.1). The data consisted of data collected by Louisiana Geological Survey (LGS), Texas A&M University (TAMU), the Bureau of Ocean Energy Management (BOEM), Louisiana Department of Natural Resources (LaDNR), Louisiana Coastal Protection and Restoration Authority (CPRA), and US Geological Survey (USGS) since the early 1980s for this region. It included LGS and LaDNR sand resource survey studies; CPRA data for site specific sand borrow area prospecting; GIS files from Heinrich et al. (2015); USGS seismic data collected in 1994 and 1995; 2010 TAMU-BOEM vibracore data from the Sabine incised valley area; Rice University data; block surveys conducted for geologic hazards and cultural resource surveys; and bathymetry, and offshore platform and onshore geotechnical borings. Key geologic data (sediment cores and geotechnical borings) from these data were used to constrain two-way travel time compared to true depth relationships and in evaluating stratigraphic and other interpretations made from the seismic data.

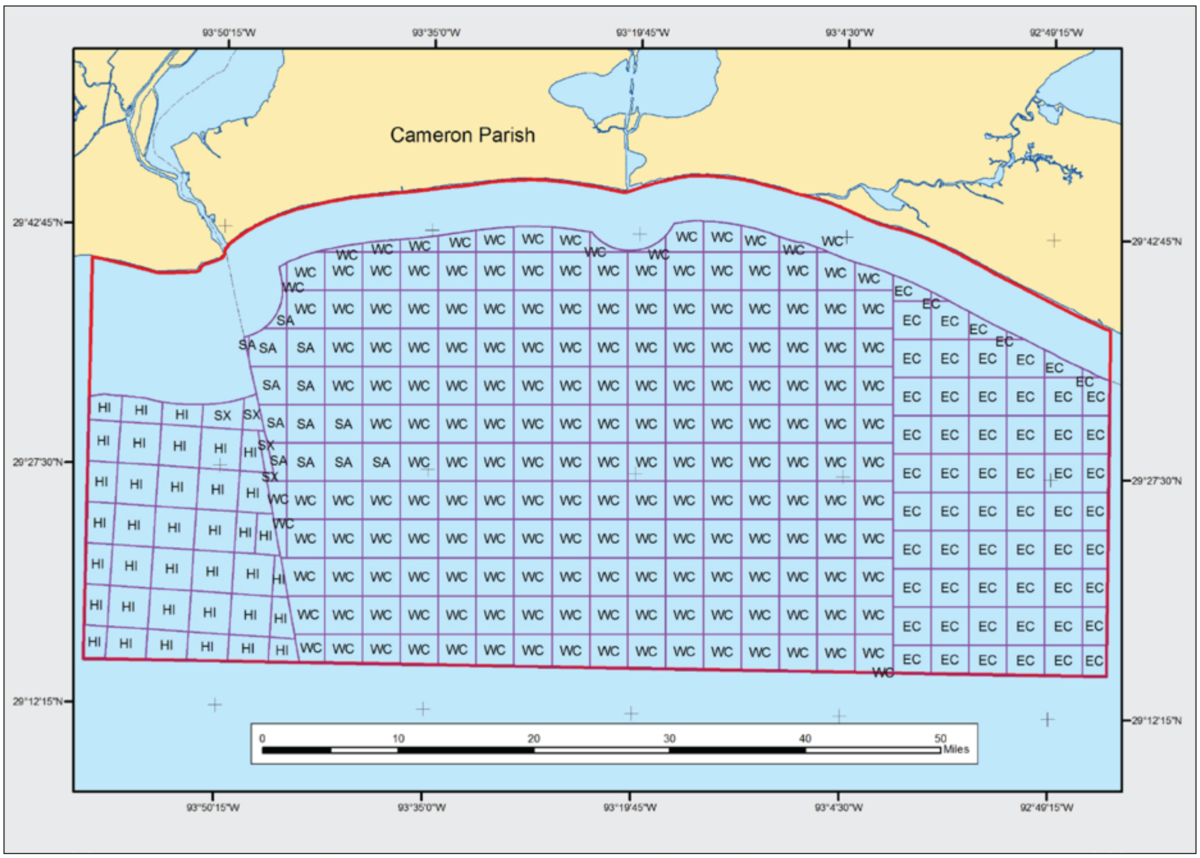

Figure 2.1. Southwest Louisiana Outer Continental Shelf Study Area.

2.2 Literature Review and Synthesis

In the next phase of this study, an intensive examination was made of published literature concerning the geoarchaeology of the Louisiana OCS region and adjacent coastal plain. This material included peer-reviewed literature, theses, dissertations, gray literature cultural resource management reports, and BOEM reports. Where found, useful paleogeographic and isopach maps were copied and later digitized for use in this project. The information gained from these sources was used to interpret the paleogeography, depositional-erosional history, and sedimentary processes associated with the accumulation of sediments within the study areas. This information, along with available core and seismic data, was also used in the construction of cross-sections and correlation of allostratigraphic units from onshore to offshore and to interpret depositional facies, potential for sand resources, and the nature of contacts between allostratigraphic units. Finally, conceptual models for paleosurface preservation potential and paleovalley sand resource occurrence on the northern Gulf of Mexico OCS were produced based on the available data and concepts of sequence stratigraphy.

BOEM requires that surveys in the Gulf of Mexico use US measurement units (feet and inches). Because of the prevalence of oil and gas industry-related research in the Gulf, many of the references, maps, and data were initially obtained and/or reported in feet. Formatting guidelines for BOEM studies require International or metric units as the primary convention. All measurements in this report, therefore, are given in metric (SI) units. In most cases, the original unit was feet; metric conversions were calculated specifically for this report.

2.3 Offshore Lease Block Hazard Survey Reports Database Development

An integral part of this project was development of a GIS geodatabase for visual representation, data inventory tracking, and computational support. The GIS application used in this project is Environmental Systems Research Institutes (Esri®) ArcMap® v10.1. Digital map data were loaded into GIS for interpretive analysis. Base and thematic data were acquired from state and federal sources. See Appendix A for a list of layers. The majority of the GIS work was to convert portable document Format (PDF) raster maps to digital spatial data for incorporation into the GIS. These documents were included in Lease Block Hazards Survey Reports submitted to BOEM for compliance. The reports contain maps produced from data interpretations and no original data products (e.g., navigation files, geophysical profiles). Three hundred and fifty interpreted hazards and cultural resource maps were made available for this study area and spanned over three decades of survey data, dating to the late 1970s. Accompanying 28 of the PDF files were original computer aided design (CAD) files used to generate the PDF maps for inclusion in the GIS project.

The process of developing GIS data from Portable Document Format (PDF) Geophysical Hazards Maps was challenging because of inconsistent reporting formats, drafting styles, and nomenclature used to describe geologic features, and compilation of the data is detail-intensive. Initial steps involved determining which PDF map files were relevant to this study on shelf paleovalleys and paleochannels The relevant information consists of numerous paleovalleys and a few scattered paleochannels and channel belts that were mapped as a part of individual lease block hazard surveys (Fig 2.1). These surveys often only cover a 3 x 3 NM area of the shelf coincident with an OCS lease block that must be cleared for hazards and potential cultural resources prior to development by the lessee. Relevant PDF map files were exported from Adobe® Acrobat® as Tagged Image File Format (TIFF) files. A spreadsheet was created to document the content of each PDF file and track whether it was digitized for the study.

Each TIFF file was assigned spatial reference defining where the features are located in the real world. This task was accomplished by using the georeferencing toolbar in ArcMap®. One hundred and twenty-one TIFF files were digitized within the study area in East Cameron, West Cameron, and Sabine Addition Areas offshore western Louisiana and Texas. After georeferencing each map, line or point, shapefiles were created with ArcCatalog applying the respective spatial reference. Digitizing the lines and/or points was done manually while in edit mode on each file. Much effort was devoted to ensuring features were represented as they appear on the maps. The vectorization tool in ArcMap®, which automatically traces linework, was of little use because of extraneous line data on the maps such as track lines or deeper geologic features not of interest to this study. Digitized lines were further developed into polygons using the Polyline to Polygon tool in ArcToolbox®. These polygon data typically represent the paleochannel and/or paleovalley edges, while the point data represent the tops and bases of the channels.

Attribute tables were developed after digitization of the features. Unfortunately, most of the maps were unique and shared little legend continuity, which created issues with nomenclature development. This is likely a result of the different interpreters and interpretation dates and lack of an accepted standard for geologic feature nomenclature. The descriptions are recorded in the attribute tables (Table 1) along with other pertinent data. These include the lease block number assigned by BOEM. Descript column describes the attribute as defined in the maps legend. As mentioned, there are many different descriptions within these geophysical maps. The tables are populated with the descriptions as listed in Appendix B.

In total, 350 offshore-block hazard surveys were reviewed of which 122 were digitally synthesized and 228 were found to be unusable. Despite the problematic variability of the source information, its compilation and availability in one GIS should have value as a reference resource (Figure 2.2).

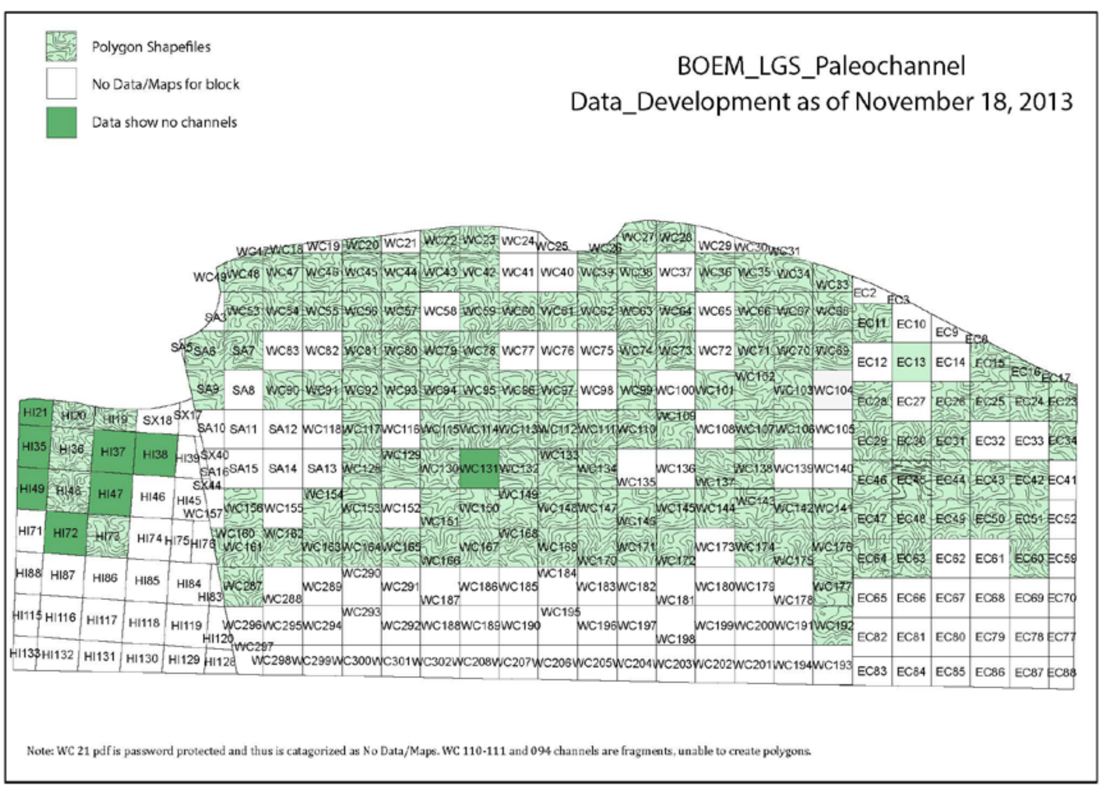

Figure 2.2. Data Distribution Southwest Louisiana Outer Continental Shelf Study Area.

Most of the mapped features in the final compilation appear to be smaller valleys of upland tributaries draining interfluves, though some portions of larger and/or wider features were identified: 1) in the western portion of the southwestern study area, correlating with parts of the Sabine and Calcasieu paleovalleys, and 2) in the eastern portion of the southwestern study area, for which we are aware of no obvious correlation with known features and the basis of which remains unclear (Figure 2.3).

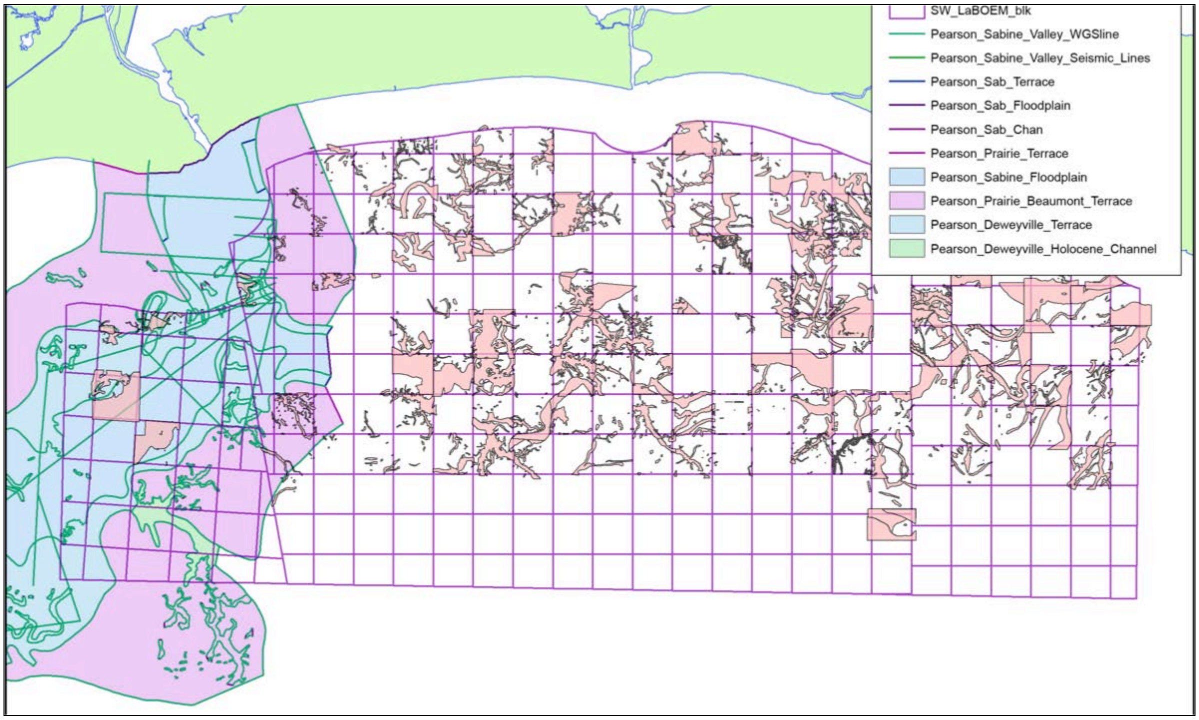

Figure 2.3. Distribution of paleolandforms in southwest Louisiana Outer Continental Shelf Study Area, as mapped by OSC Hazard studies and Pearson et al. (1986).

3. Northern Gulf of Mexico Geology and Sea Level History

3.1 Geologic Setting

The study area encompasses the northern and middle portions of the Southwest Louisiana Continental Shelf adjacent to the Southwest Louisiana shoreline from Joseph Harbor Bayou (Humble Canal), Cameron Parish west to about 6 km west of Sabine Pass along the shore of Chambers County, Texas. This study area lies within the coastal plain and continental shelf that forms the northwestern margin of the Gulf of Mexico.

The Gulf of Mexico Basin, as used in this report, is a roughly circular structural basin that is approximately 1,500 km in diameter. It is filled with up to 15.1 km (9.4 mi) thick sedimentary deposits ranging from Jurassic to Holocene in age. Locally, Triassic rift basins along its edges contain an additional 1 km or more of Triassic sedimentary units (Bryant et al. 1991, Salvador 1991). The Gulf of Mexico Basin is the result of rifting within the North American plate as it drifted away from the African and South American plates (Salvador 1991). As currently understood, the first phase consisted of Late Triassic–Early Jurassic (235–174 Ma) rifting that resulted in the opening of the Gulf of Mexico in a NW-SE extension direction. This rifting also created a broad zone of thinned, continental crust along the northern margin of the Gulf of Mexico that now underlies the northern salt basins of Texas, Louisiana, and Mississippi. However, it failed to produce in this area a parallel and contiguous zone of oceanic crust. During the Middle Jurassic, marine flooding of the proto-Gulf of Mexico basin deposited extensive evaporite, mainly salt, strata. After the accumulation of thick evaporites, a second phase of Late Jurassic (156–145 Ma) extension occurred along a highly arcuate, slow spreading ridge system. This ridge system created a large expanse of salt-free, oceanic crust that underlies the deepwater shared by the US, Mexico, and Cuba (Nguyen and Mann 2016). During the Early Cretaceous, carbonate platforms and shelf margins rimmed the Gulf of Mexico and thick accumulations occurred along its northwestern rim (Winker and Buffler 1988). Carbonate sedimentation ceased as the result of continental-scale drainage reorganization that occurred between Middle Cretaceous, Aptian-Cenomanian, and Paleocene. This reorganization expanded the area drained into the Gulf of Mexico from largely the Appalachian Mountains to include fluvial systems of western and midcontinental North America that were routed either directly into the Gulf of Mexico or the Mississippi Embayment (Blum and Pecha 2014). As a result, km-thick sequences of sandstones and mudstones buried the Cretaceous carbonates and extended basinward to the base of the modern continental slope along the northwestern rim of the Gulf of Mexico over the Cenozoic (Bentley et al. 2016).

The current, Quaternary coastal plain of Texas and Louisiana consists of the strata that were deposited by fluvial systems of various types, sizes, and sediment composition (Young et al. 2012). Despite these variations, these fluvial systems can be divided into extrabasinal and intrabasinal rivers. Extrabasinal rivers are rivers that have large drainage basins that extend well beyond the coastal plain. Over the Cenozoic, such rivers have persistently occupied major embayments within the coastal plain as they currently do. For example, the Rio Grande, Colorado-Brazos, and Mississippi rivers, respectively, occupy the Rio Grande, Houston, and Mississippi embayments. The point where an extrabasinal river enters the coastal plain is stable owing to the entrenchment of its valley across the slightly uplifted margin of the coastal zone (Winker 1979). Gulfward from this entry point, extrabasinal fluvial systems freely migrate laterally across the coastal plain and, thus, construct alluvial plains composed of sand-rich channel-fill facies and mud-rich floodplain facies (Figure 3.1) (Morton and McGowen 1980, Galloway 1981; Blum and Price 1998, Anderson and Fillon 2004).

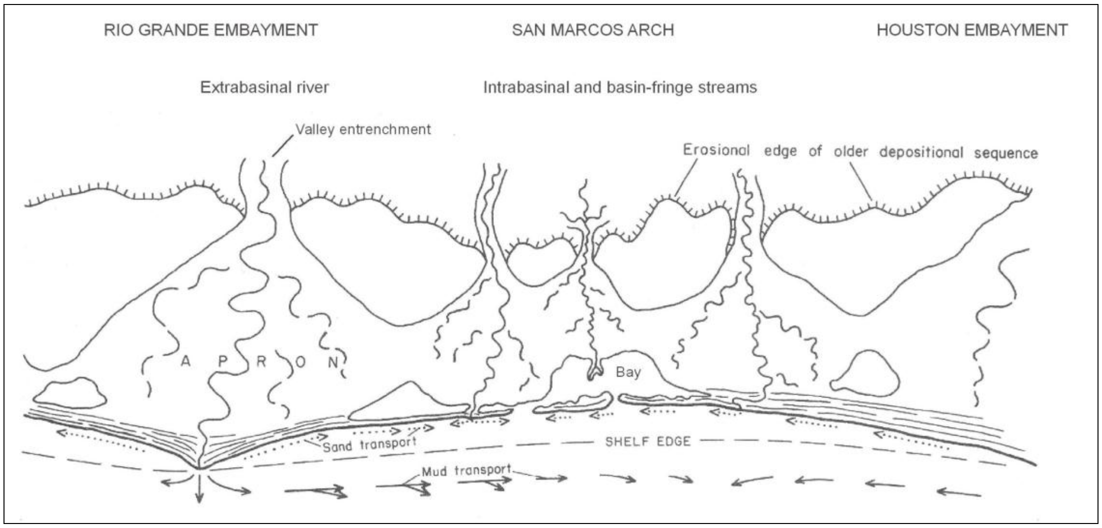

Figure 3.1. Diagrammatic interpretation of Quaternary depositional systems of the Texas Coastal Plain.

Reprinted with permission from Young et al. (2012).

In contrast, intrabasinal rivers are rivers that have drainage basins marginal to and within the coastal plain. Within both East Texas and Southwest Louisiana, the overall drainage pattern and individual courses of intrabasinal rivers are strongly controlled by the relict antecedent topography of major extrabasinal rivers and coast-parallel terraces. In the Houston region, intrabasinal rivers and streams flow roughly parallel to highstand, interglacial meander-belt ridges within interridge lowlands and along the gulfward edges of coastal plain segments (Van Siclen 1985). In Southwest Louisiana, the courses of the Mermentau River and its tributaries occupied interchannel lows that lie between relict Red River fluvial and beheaded deltaic distributary ridges.

The Quaternary alluvial aprons grade basinward into deltaic and shore-zone depositional systems. In the Rio Grande and Houston embayments, the major extrabasinal rivers have constructed sand-rich deltaic headlands on the shore of the modern coastal plain. Within and gulfward of the Mississippi embayment, the Mississippi River has constructed a major deltaic plain. Intrabasinal rivers empty into underfilled estuaries with bayhead deltas at their apex. Elsewhere along the Gulf of Mexico, coastal plain, lagoon and barrier island systems, and the Louisiana Chenier Plain form the coastline (Morton and McGowen 1980, Galloway 1981).

The Cenozoic strata underlying the northwest coastal plain of the Gulf of Mexico consist of a heterogeneous mixture of sand, sandstone, mud, and mudstone that lacks any distinctive lithologic subdivisions (Galloway et al. 1991). These sediments have been subdivided using a combination of: 1) biostratigraphic zonation; 2) fluvial-deltaic depositional models based on Quaternary examples; and 3) the definition of regionally cyclic depositional episodes (Galloway et al. 2000, Young et al. 2012). The extinction points of biostratigraphic zonation are known as “last occurrence or datum top” of calcareous nannofossils and planktonic foraminifers and other marine microfossils (Galloway et al. 1991, Lawless et al. 1997, Fillon and Lawless 2000). Because of the absence of marine fossils in fluvial sediments, stratigraphic subdivisions are extended downdip from outcrop using volcanic ash layers, well log correlation techniques, and limited nonmarine (vertebrate faunal) biostratigraphy (Lundelius 1972, Tedford and Hunter 1984, Baskin and Hulbert 2008, Young et al. 2012).

3.2 Sea Level Change: An Overview

Change in relative sea level is the result of the complex and independent interaction between the movements of two different surfaces: the surface of the sea and either the land surface or sea floor. The surface of the ocean is spatially and temporally variable related to total mass of ocean water compared to water that is sequestered as continental ice sheets, the eustatic component. Tectonic deformation of either land or the sea floor is spatially and temporally variable and arises, in part, due to changes in ice loading of the crust such as during a during a glacial cycle, subsidence induced by sediment compaction, or tectonic loading (Douglas 2001, Dix et al. 2008).

As discussed by Dix et al. (2008), the diverse factors that control fluctuations in both the elevation of the either the land surface or sea floor and either the sea or ocean surface interact on a wide range of spatial and temporal scales. They vary from astronomical factors that operate from outside the Earth, such as variations in the planet's angular velocity, to other factors that operate on the Earth's surface, like the global distributions of glaciers and meltwater. Finally, there are factors working within the Earth's interior, such as displacements of mantle material (Pirazzoli 1993, Douglas 2001). These factors are often interconnected and they operate in conjunction with each other. The dominant influence on relative sea level change at a specific place and time depends on the temporal and spatial scale on which it is observed (Dix et al. 2008).

3.2.1 Eustatic controls on sea level

During the Quaternary, the dominant mechanism responsible for sea-level changes has been the cyclic addition and subtraction of water from global oceans from the progressive buildup and decay of continental-scale ice sheets in response to Milankovitch forcing insolation changes. Minor input and outputs to the volume of water stored in the world's oceans and seas have come from atmospheric water, rivers, lakes, peat bogs, and so forth as part of the hydrological cycle. However, variations in storage in these sources can contribute only relatively little–a layer of water some 58.3 cm thick evenly distributed worldwide–to present ocean volume (Dix et al. 2008).

Throughout the Quaternary, the interplay between variations in the Earth's orbit and axial tilt; ice sheet dynamics; and ocean circulation have resulted in a long-term cycle of climate change consisting of alternating glacial and interglacial phases. These cycles were about 41,000 years in the early Pleistocene and Pliocene and about 100,000 years from about 800 ka BP to present. The removal of waters from global oceans during the formation of continental ice sheets resulted in sea level lowstands and melting of continental ice sheets resulted in sea level highstands (Zachos et al. 2001, Lambeck et al. 2002, Dix et al. 2008).

Such relative changes in global sea level, which were on the order of 100 m, were termed "eustatic" by Suess (1906). Thus, eustatic sea-level change is defined as the global shift in the height of the ocean surface that occurs, in the absence of gravitational and rotational effects, ocean dynamics, and solid Earth deformation, due to the input of meltwater to the ocean. Because eustatic sea-level change is directly equivalent to the changing mass of the global ice sheets, it provides an independent method for calibrating the marine oxygen isotope record and the total volume of continental ice at the Last Glacial Maximum (Farrell and Clark 1976, Dix et al. 2008, Whitehouse and Bradley 2013). Since the end of the Last Glacial Maximum at about 20 ka BP, sea-level change has been dominated by the addition of meltwater into the oceans associated with the retreat and shrinkage of the major ice sheets. Thus, it has been postulated that locally, relative sea-level change in tectonically stable regions far from the major ice sheets, known as far-field sites, approximates eustatic sea-level change since the Last Glacial Maximum (Milliman and Emery 1968).

3.2.2 Vertical displacement of oceanic and continental crust

Active displacement, both uplift and subsidence, of the Earth's crust is a quite common occurrence anywhere on the Earth's surface, including the Gulf of Mexico coastal plain and continental shelf. Both subsidence and uplift, respectively, are associated with isostatic adjustments as the result of increased or decreased loads on the earth's surface. It is called "glacio-isostasy" when the load is ice, "volcano-isostasy" when it is extruded magma, "sedimento-isostasy" when it is sediment deposition or erosion, and "hydro-isostasy" when it is a layer of water (Pirazzoli 1991). Finally, either temperature and/or density changes of materials inside the Earth's interior can cause uplift and subsidence of the Earth's crust. It is called thermo-isostasy when temperature is the predominant factor, e.g., near a hot spot (Pirazzoli and Grant 1987). Along the northern Gulf of Mexico, for the late Quaternary, glacio-isostasy and sediment-isostasy have been demonstrated to influence relative sea-level (Wolstencroft et al. 2014).

3.2.1.1 Glacio-isostasy

The cyclic transfer of large volumes of water from the oceans to land ice and back that characterized the Quaternary changes not only created the eustatic component of sea level, it also resulted in increased or decreased loads on the Earth's surface as continental ice sheets grew and shrank in area and volume. This redistribution of surface water mass is large enough to have caused a significant, cyclic glacio-isostatic deformation, known as "glacial isostatic adjustment" of the solid Earth. It is one of the key processes that contributed to glaciation-induced sea level change. During the glacial maximums, ice sheets reached thicknesses of several kilometers in places. The accumulated mass of glacial ice sheet was sufficient to depress the underlying crust by the order of hundreds of meters. Peripheral to areas of depression that underlie the ice sheets, there was corresponding uplift of the crust to form a forebulge in front of it (Chappell 1974, Peltier and Andrews 1976, Mitrovica and Peltier 1991, Milne and Shennan 2013).

During the retreat and disappearance of the ice sheets from North America and Eurasia during the early to mid-Holocene, there was, and still remains, a significant uplift in these regions. This is the result of the relatively high viscosity of mantle material and the resulting slow rate at which the Earth's crust is attaining isostatic equilibrium following the removal of ice sheets. There is also a corresponding belt of relatively slow isostatic subsidence that result from the collapse of the forebulges surrounding the regions depressed by glacial ice sheets (Peltier and Andrews 1976, Mitrovica and Peltier 1991, Lambeck 1993, Milne and Shennan 2013).

It has often been assumed that glacial isostatic adjustment was insignificant within the Southwest Louisiana-Southeast Texas coastal plain and continental shelf (Gagliano et al. 1982; Anderson and Bissett 2015). However, modeling of glacial isostatic adjustment demonstrates that it had some effect on relative sea level (Simms et al. 2007, Milne and Shennan 2013). The magnitude of the effect of glacial isostatic adjustment on relative sea level, as discussed later, can be seen in the maximum MIS 2 lowstand of relative sea level in the northern Gulf of Mexico of between -80 and -90 m when the eustatic sea level was about -120 m from between about 29 and 18 ka (Waelbroeck et al. 2002).

3.2.1.2 Hydro-isostasy and continental levering

Another process that affects relative sea level by causing significant deformation of the Earth's crust is hydro-isostasy. It is the deformation of the Earth's crust as a coastal plain is submerged beneath rising sea level. The weight of the water loading the crust causes it to subside even further. In addition, as the crust is loaded by rising sea level, it not only subsides, it also forces mantle landward beneath crust, causing it to be uplifted. This uplift is called "continental levering." The effects of hydro-isostasy and associated continental levering is most discernable in regions, "far-field" regions, far removed from the major ice centers where the effects of glacial isostatic adjustment are considerably reduced, as shown by the small uplift around the coasts of Australia and Africa (Clark et al. 1978, Milne and Shennan 2013).

3.2.1.3 Ocean syphoning

During deglaciation, both hydro-isostasy and glacial isostatic adjustment increase the volume of the world's oceans by lowering sea floor and deepening the oceans within collapsing forebulges and subsiding continental shelves. This deepening of the ocean floor will increase volume of the ocean and its capacity to store water in these areas and draw ("syphon") sea water away from more stable far field, largely equatorial, areas and cause a fall in relative sea level within them. The magnitude of the global mean sea-level fall due to syphoning during deglaciation can be a few 10s of meters from the Last Glacial Maximum to present (Mitrovica and Peltier 1991, Mitrovica and Milne 2002).

3.2.1.4 Sedimento-isostasy

Another source of tectonic displacement of the crust within the northwestern Gulf of Mexico is sedimento-isostasy. Before anthropogenic modification o f the Mississippi River, its tributaries and drainage basin steadily supplied large sediment loads to the continental margin with little long-term sediment storage within the lower Mississippi River Valley. This process has resulted in the construction and extension of the Mississippi River Delta Plain southward into the Gulf of Mexico Basin, which began in the early Paleocene. The loading of the continental shelf with fluvial sediments was greatly accelerated during Pleistocene glacial periods, most recently during MIS 4 through MIS 2. Later, during the slowdown in deglacial sea level rise about 8000 ka, near-shore aggradation of this fluvial sediment load allowed the development of the Mississippi River Deltaic Plain (Coleman and Smith 1964). The weight of this load of accumulating deltaic sediments induced subsidence due to downward flexure within the underlying crust (Bowie 1927, Fisk and McFarlan 1955, Blum et al. 2008). In this process, the Earth's crust deforms elastically in that it deforms under the stress, but it reverts to its original form after the strain ceases. In contrast, the mantle acts as a highly viscous fluid and permanently deforms under the stress of the accumulated sediments.

The accumulation sediments on the underlying lithosphere (i.e., the crust and upper mantle) causes mean downward surface displacement of the underlying lithosphere, i.e., the crust and upper mantle. If the mean increase in elevation due to sediment aggradation over an affected area is less than the downward displacement of the surface then subsidence and eventual submergence will occur. If the mean accumulation of sediment over an affected area is equal to or greater than the downward displacement of the surface the surface of the affected area will remain stable and the excess sediment will be transported seaward and eventually result in gulfward progradation and forced regression of the shoreline. Sediment loading differs from sediment compaction because it refers to subsidence in the material underlying a sediment load rather than subsidence resulting from volume decrease within the sediments themselves (Fisk and McFarlan 1955, Coleman and Smith 1964, Diegel et al. 1995, Dokka et al. 2006).

The process of growth faulting has also been attributed, in part, to sedimento-isostasy resulting from the construction and extension of the Mississippi River Delta Plain southward into the Gulf of Mexico Basin. This process began loading with fluvial sediment in the early Paleocene. It has been argued that flexure of the continental crust has reactivated growth faults with general east-west orientations, perpendicular to the direction of delta growth. Also, growth faults are created as the delta progrades down an inclined surface. In time, the prograding delta front overextends and detachment and the formation of fault zones, which slip by breakaway, and gravitational slumping, occurs (Berman 2005, Diegel et al. 1995, Dokka et al. 2006).

Although the Mississippi River Delta is well recognized for its associated sedimento-isostasy subsidence, there is evidence of sedimento-isostasy subsidence association with the valley fills and deltas of the Colorado, Brazos, and Rio Grande rivers. As noted by Graf (1966) and Winker (1990), the elevation of beach ridges associated with the Ingleside shoreline systematically varies and can be interpreted as reflecting subsidence associated with the valleys of Brazos and Colorado rivers and the Mississippi alluvial valley and delta and uplift between them.

3.2.3 Halokinesis and surface deformation

Halokinesis, salt tectonics, also can cause deformation, either subsidence or uplift of the surface of the Gulf of Mexico coastal plain and sea bottom. Large pillars of salt intrude upward through kilometers of the largely unlithified sediments that underlie the surface of the coastal plain and sea bottom as the result of buoyancy effects. The upward relative movement of salt domes can result in the uplift of local, roughly circular areas, e.g., the Five Islands and the subsidence of a moat-like area around an uplifted area. The movement of salt domes can create radial faults and downthrown grabens. Subsidence resulting from halokinesis would occur very slowly over geologic timescales (> 1000 yrs). The spatial scale would either be the areal diameter of the diapir, 1 to 100 km^2, or regional as halokinesis causes large blocks to slide gulfward along regional growth faults (Berman 2005, Diegel et al. 1995, Dokka et al. 2006, Reed and Yuill 2009).

3.2.4 Sediment compaction and subsidence

In coastal environments, newly deposited fluvial and marine sediment may compact in time as the result of various physical, biological, and chemical processes (van Asselen et al. 2009). Within coastal Louisiana, including the Mississippi River Delta, physical compaction is the most commonly cited compaction-related cause of subsidence. The process of compaction includes both compression and consolidation of the sediments. Compression is the decrease in soil volume due to a restructuring of internal grain alignment as a result of an applied stress. This restructuring of the sediment causes the sediment to become more tightly packed. Unlike consolidation, compression occurs in relatively short timescales upon application of an applied load and is generally controlled by the geotechnical properties of the soil. Consolidation relates to the time-dependent expulsion of pore water in response to an applied stress, which reduces the internal pore pressure causing pore collapse. Unlike compression, consolidation is a gradual process that is controlled by soil-water interactions. In sediments composed of peat and other organic-rich sediments, biological (microbial decay of organic material) or chemical (oxidation of organic carbon) processes can result in significant compaction (van Asselen et al. 2009, Reed and Yuill 2009).

3.3 Relative Sea Level History

Beginning about 2.6 million years ago, continental ice sheets became a prominent feature in the Northern Hemisphere. Later they became widespread elsewhere over the Earth’s surface in response to the cooling global climate. During the Quaternary Period, which is divided into the Pleistocene and the Holocene epochs, these continental ice sheets episodically covered millions of square kilometers and many were up to 3 km thick. The water that produced the huge quantities of snow that created these continental ice sheets came from the world’s oceans. The result was a global drop in sea level during glacial epochs that lowered global (eustatic) sea level by up to 130 m (Figure 1.1). The periodic drop in global sea level caused repeated exposure of part of the coastal plain of what is now the Gulf of Mexico continental shelf. The carbonate platforms of Florida and Yucatan were also periodically exposed with the development of karstic surfaces with innumerable sinkholes. Of these glacial–interglacial cycles, this study is concerned only with the last one, from MIS 5 to MIS 1, because that is when landforms and landscapes that form the background to this study were created (Figure 1.2) (Bentley et al. 2016).

3.3.1 MIS 5, Last Interglacial

The last interglacial period, MIS 5, spans the period from about 130 to 71 ka. This period was the last interval of the Pleistocene when global ice volumes were small, and ice-equivalent sea levels oscillated from +6 to -45 m, with a mean value of -27 m. Regionally, the MIS 5 sea levels departed from these globally averaged values because of GIA contributions and vertical tectonic displacements (Waelbroeck et al. 2002). During MIS 5e, ice-equivalent sea levels were 6–9 m higher than present (Kopp et al. 2009 and 2013, Dutton and Lambeck 2012, Anderson et al. 2014, Bentley et al. 2016).

Offshore shorelines for periods of sea level fall after MIS 5e are poorly constrained. Simms et al. (2009) concluded that the MIS 5a paleoshoreline occurs at -11 ± 2 m water depth on the east Texas continental shelf. They also refined earlier estimates by Rodriguez et al. (2000) of the depth of the MIS 3 paleoshoreline and concluded that it might have been as high as -20 m, after correcting for local glaciohydro-isostatic effects (Figure 1.2).

3.3.2 Last Interglacial highstands, MIS5, Gulf of Mexico

During the early part of the last interglacial, MIS 5e, sea level was at its maximum highstand position, a few meters above present sea level. It was at this time that an ancient beach-ridge complex, referred to as the Ingleside Paleoshoreline, was created. It marks the location of the MIS 5e shoreline from eastern Louisiana to just south of the city of Corpus Christi, Texas. In East Texas and Louisiana, the Ingleside Paleoshoreline is located 35 to 60 km inland of the modern shoreline. Its preservation demonstrates that it was never subjected to either transgressive or regressive shoreface erosion and it likely marks the landward extent (maximum flooding shoreline) of the Gulf of Mexico during MIS 5 (Graf 1966; Barrilleaux 1986, Winker 1990).

The Western Louisiana segment of the Ingleside Paleoshoreline, the Houston Ridge, is a 32 km-long, narrow, 500 to 900 m wide, east-west trending ridge. The crest of this ridge ranges in elevation between 7.5 and 11 m above mean sea level and lies 2 to 4.5 m above the surface of the adjacent Beaumont Alloformation. The development of innumerable pimple mounds has extensively modified the surfaces of both the Houston Ridge and other parts of the Beaumont Alloformation. However, using LIDAR DEMs, remnants of beach ridges can be observed within the eastern end of the Houston Ridge. The Ingleside Paleoshoreline is absent locally where removed by fluvial processes (Figure 3.2) (Heinrich 2007).

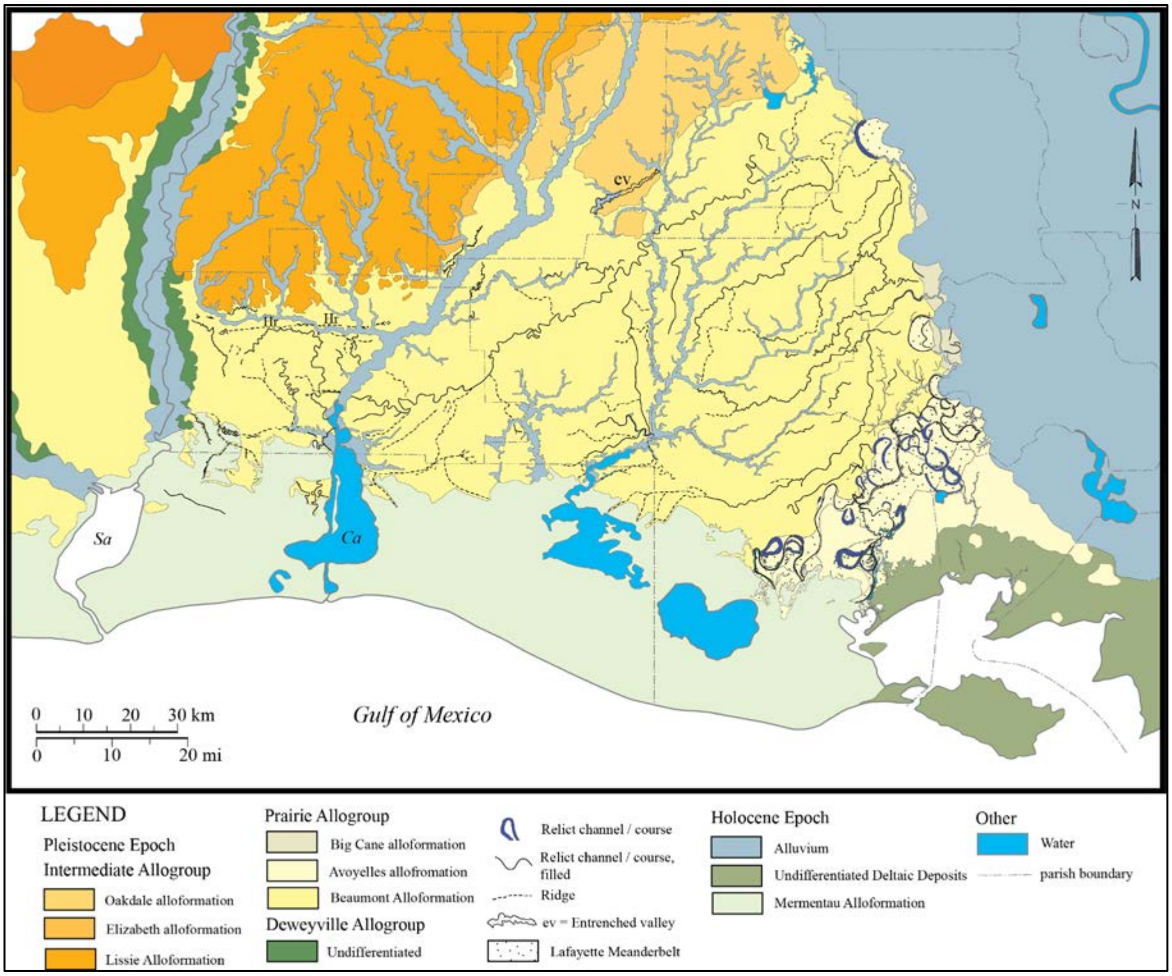

Figure 3.2. Geomorphic map of Southwest Louisiana Pleistocene paleochannels.

Compiled and generalized from McCulloh and Heinrich (2002), Saucier and Snead (1989), and various published and unpublished 1:100,000 scale Louisiana geologic quadrangles (Heinrich 2005a).

Despite initial attempts to directly date this paleoshoreline providing mixed results (e.g. Shideler 1986, Otvos and Howat 1997, Blum et al. 2002), the Ingleside Paleoshoreline has long been regarded as representing the ancient shoreline of the last interglacial highstand (Price 1933, Graf 1966, Barrilleaux 1986, Morton and Price 1987, Paine 1993, Anderson et al. 2004). More recent research by Simms et al. (2013) yielded six optically stimulated luminescence ages for segments of the Ingleside Paleoshoreline near Smith Point east of Galveston Bay and Port O'Conner, Texas. The ages obtained by Simms et al. (2013) cluster into groups of 119 ka–127 ka BP, 54 ka–57 ka BP, and 1.3 ka BP. The two groups of younger ages appear to be related to two periods of regional eolian reworking of coastal plain surface (Otvos 2004, Forman et al. 2009, Simms et al. 2013). The ages of 119–127 ka BP support earlier interpretations that the Ingleside Paleoshoreline is an MIS 5e shoreline deposit (Figure 1.2) (Anderson et al. 2004 and 2016).

After the last interglacial MIS 5e, global sea level experienced significant oscillations. During MIS 5, these fluctuations were MIS 5d to MIS 5a, of which MIS 5c and MIS 5a correspond with each interstadial sea level highstands. From the interglacial highstand of MIS 5e, sea level fell abruptly about 115 ka BP down to -60 m. Afterwards, sea level rose to about -15 m around 103 to 98 ka BP and then fell again to about -60 m during MIS 5b. During MIS 5a about 85 ka BP, sea level rose again 15–25 m of its present level. Sea level was about 65 m below present at the end of MIS-5a about 75 ka BP (Lambeck and Chappell 2001, Lambeck et al. 2002, Cutler et al. 2003, Potter et al. 2004, Hopley et al. 2007).

The surface characteristics of the Beaumont Alloformation north of the Ingleside Paleoshoreline varies greatly. Within Southeast Texas, its surface consists of a mosaic of imbricated offlapping alluvial and deltaic plains that exhibit relict fluvial ridges and paleochannels of meander belts and deltaic distributaries. Since their abandonment, an inversion of topography has occurred in which modern drainages have come to occupy the interchannel areas between fluvial ridges and deltaic distributaries and the boundaries of individual offlapping coastal plains. Based on stratigraphic and geomorphic considerations, except for terraces along modern drainages, these features are inferred to predate MIS 5e (Van Siclen 1985 and 1991, Aronow 1971). Within Southwest Louisiana, the narrow belt of coast-parallel terrace lying between Ingleside Paleoshoreline and where it onlaps the surface of the Lissie Alloformation consists of a typically flat plain that is featureless except for the occasional paleochannels of a small-scale meandering channel belt (Heinrich 2007).

South and east of the Ingleside Paleoshoreline within the Louisiana Coastal Plain, the surface of the Beaumont Alloformation is characterized by two different groups of relict fluvial-deltaic depositional topography (Figure 3.2) (Fisk 1948, Saucier 1977, Aronow 2004). East of Bayou Choupique, the coast-parallel terrace of the Beaumont Alloformation exhibits paleochannels and depositional ridges of Red River channel belts and deltaic distributaries. These features range in degree of preservation from the well-defined meander loops to either fragments of channels and detached, discontinuous loops or featureless distributary or fluvial ridges. These relict fluvial-deltaic features, especially east of the Calcasieu River, impart a northeast-southwest grain to the drainage patterns and, similar to the area in Texas discussed above, there has been an inversion of topography with the modern drainage now occupying interchannel lows between paleochannels. Some of the paleochannels have distinctly incised their channels, often leaving beheaded deltaic distributaries on both sides of them. There is an apparent lack of continuity in trend and scale between the paleochannels and channel belts observed on the surface of the Beaumont Alloformation and the “fluvial channels” of Suter (1986 and 1987); the “distributary channels” (“incised valleys”) of Wellner (2001); and the “channel belts” (“channel complexes”) of Anderson et al. (2016). Finally, on the surface of the Beaumont Alloformation, a lower relief and more poorly defined set of possible beach ridges extend for a distance of about 15 km from the east end of Houston Ridge at Sam Houston State Park to the west valley wall of the Calcasieu River.

The western portion of the terrace surface of the Beaumont Alloformation exhibits innumerable poorly preserved, rather obscure, and fragmentary meander loops of paleochannels. These fragmentary paleochannels become better defined and more continuous to the northwest, towards Starks, Louisiana. They also impart a northwest-southeast grain, which controls the minor drainages, to the landscape. Bernard and LeBlanc (1965) and Aronow (2004) infer that this portion of the terrace surface of the Beaumont Alloformation was laid down by the Sabine River.

The terrace surface and surficial sediments south and east of the Ingleside Paleoshoreline postdate the MIS 5e highstand based on cross-cutting relationships. In addition, the contemporaneous incision of fluvial channels, dissection of deltas and beheading and stranding of their distributaries are indicative of forced regression brought on by falling relative sea level.

As summarized by Shen et al. (2012), the age and correlation of the coast-parallel surface and associated sediments that form the upper contact of the Beaumont Alloformation have been the subject of considerable speculation. The first optically stimulated luminescence ages for the uppermost, coast-parallel sediments of the Beaumont Alloformation by Otvos (2005) yielded ages of MIS 5 to MIS 3. However, as discussed by Shen et al. (2012), Otvos (2005) lacks the necessary documentation regarding either the dating methodology or the stratigraphic context of dated samples. Commenting on the inconsistencies in these dates noted by Heinrich (2006), Shen et al. (2012) proposed that many of these optically stimulated luminescence ages underestimate the age of the overbank deposits due to pedogenic and biogenic disturbance. Well-documented by Miller et al. (1985), Schumacher et al. (1988), and Autin and Aslan (2001), the surficial fluvial deposits of the Beaumont Alloformation are capped by a well-developed paleosol and are mixed with the overlying Peoria Loess in the eastern part of Southwest Louisiana. In this part of Southwest Louisiana, this surface was subaerially exposed for at least thousands of years during which significant pedoturbation of the uppermost few meters of these sediments occurred (Miller et al. 1985, Autin and Aslan 2001). In the western part that lacks a loess cover, the surface would have been exposed to pedoturbation for tens of thousands of years and at least to two periods of regional eolian reworking of coastal plain surface (Otvos 2004, Forman et al. 2009) as evidenced by optically stimulated luminescence ages published by Simms et al. (2013) from the Ingleside Paleoshoreline. In addition, there likely was significant influx to and mixing of wind-blown dust into the surface of the Beaumont Alloformation. As a result, there certainly has been significant pedoturbation and addition of eolian dust to the uppermost few meters of its sediments since the surface of the Beaumont Alloformation was abandoned.

According to Shen et al. (2012 and 2013), the upper Beaumont Alloformation aggraded in association with a MIS 5a sea-level highstand and subsequent abandonment of the coastal plain is not only demonstrated by their optically stimulated luminescence ages, but also the optically stimulated luminescence ages from and stratigraphic position of the younger Lafayette Meander Belt of the Avoyelles alloformation (Mateo 2005). Also, as noted by Shen et al. (2012 and 2013), the longitudinal profiles of the Mississippi River valley train from MIS 4 and 3 occur at elevations about 10 m below the MIS 5a long profile of the surface of the Beaumont Alloformation (Rittenour et al. 2007). Finally, the longitudinal profiles of fluvial terraces of the MIS 4 and 3 Deweyville Allogroup along the Calcasieu, Mermentau, and Sabine rivers also lie well below the elevation of the surface of the Beaumont Alloformation (Blum et al. 1995 and 2013, Heinrich et al. 2002 and 2003, Heinrich 2005b and 2006).

Optically stimulated luminescence ages from the Freeport Rocks Bathymetric High, off of the Central Texas coast, show that their age is about 91 ka BP. Based on their relationship to former relative sea levels, Simms et al. (2009) concluded that MIS 5a sea levels within the northwestern Gulf of Mexico were around -11 ± 2 m at this time. This is consistent with an estimated MIS 5a sea level highstand of around -7 to -9 m about 98 ka BP for the southwestern Louisiana continental shelf according to the ICE5G_120550_Full model of Martini (2015).

3.3.3 Falling-stage sea level: MIS 4 and 3

After MIS 5a, global sea level fell irregularly through MIS 4 and 3 into the Last Glacial Maximum, MIS 2. During this period, global sea level rapidly fluctuated, with cycles of ~15 m amplitude recurring every 6–7 ka. Global sea level rose to close to -40 m midway through MIS 4, then fell to about -70 m by 65 ka BP, and rose again to about - 50 m by the start of MIS 3. During MIS 3, 70–30 ka BP, sea level fluctuated between highstand and lowstands between -45 m and -90 m several times (Figs 1.1 and 1.3; Chappell, 2002, Lambeck et al. 2002, Anderson et al. 2016).

Within the southwest Louisiana continental shelf, the ICE5G_120550_Full model of Martini (2015) estimates that during MIS 4, relative sea level fell to just less than -50 m about 56 ka BP from around -7 m to -9 m about 98 ka BP (MIS 5). For the MIS 4 lowstand, the same model estimates that the highest that relative sea level rose during the MIS 3 highstand was about -45 m and the lowest that it dropped was about -90 m.

Based on stratigraphic studies, the older ("Early Wisconsin") channel belts and associated shelf-margin deltas (mapped by Suter 1987) formed during an informal "falling-stage sea level" period that consists of MIS 5d to 5a, 4 and 3 (Figures 3.4 and 3.5; Coleman and Roberts 1988a and 1988b), Wellner et al. 2004, Anderson et al. 2004, 2014, and 2016). Suter and Berryhill (1985), Berryhill (1987), and Suter (1987) conducted the original, detailed studies of these "falling stage" channel belts and shelf margin deltas mapping fluvial channel belts on the western Louisiana and Texas continental shelves and concluded that they formed during "Early Wisconsin" time.

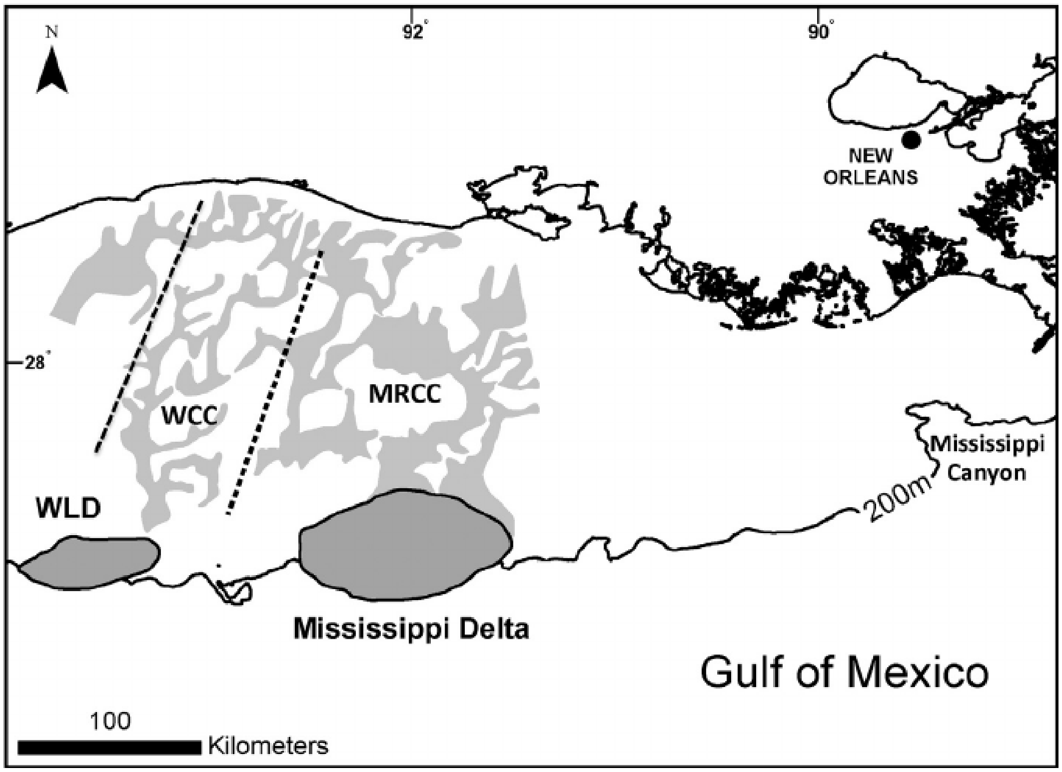

Figure 3.3 MIS 5–3 channel belts of the western Louisiana continental shelf, as modified from Suter and Berryhill (1985) by Anderson et al. (2016).

Anderson et al. (2015) subdivides them into three channel belts complexes. The eastern channel belt complex merges seaward with the Mississippi shelf margin delta. A second, narrower western channel belt complex merges with MIS 3 Western Louisiana Delta (WLD). The western most channel belt complex they assign to the the Calcasieu and Sabine paleovalleys. MRCC = Mississippi River channel complex; WCC = Western channel complex. Reprinted from Anderson et al. (2016).

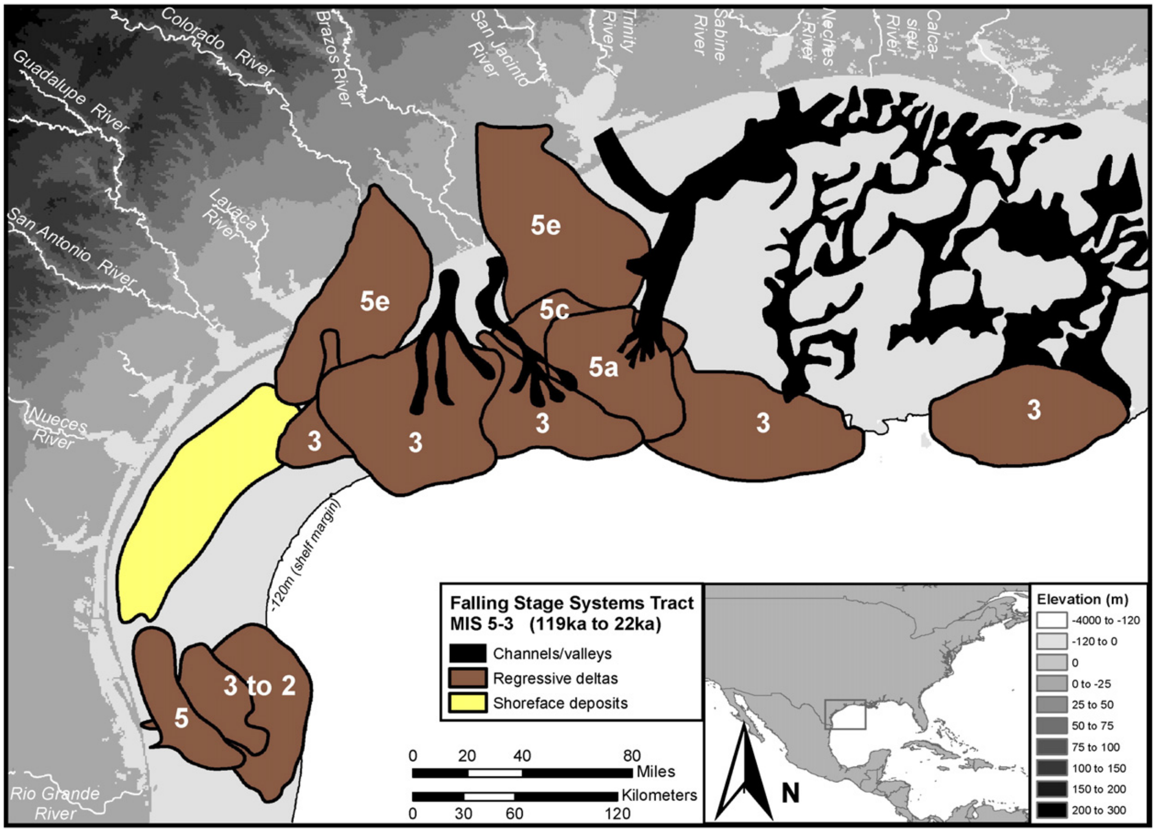

Figure 3.4. Paleogeographic map showing major depositional systems that existed on the shelf 120 ka to 22 ka (stages 5e–3 highstand).

The different lobes of individual deltas are numbered in chronological order. RGD = Rio Grande Delta, CD = Colorado Delta, BD = Brazos Delta, WLD = Western Louisiana Delta, LD = Lagniappe Delta, EMD = Eastern Mobile Delta, WFLAD = West Florida–Alabama Delta, and AD = Apalachicola Delta. Reprinted from Anderson et al. (2004)

Anderson et al. (2016) divided the channel belts of Berryhill (1987) and Suter (1987) into two separate drainage systems. They recognized an eastern channel belt complex, which they inferred to be a paleo-Mississippi River channel belt complex that is characterized by somewhat wider, more closely spaced channel belts. This eastern set of channel belts displays lateral accretion, generally less than a kilometer, indicating modest channel sinuosity, and occupied an area at least 150-km wide. It is attached to an eastern shelf margin delta, which Suter and Berryhill (1985) called the "Mississippi Delta". This delta has not been directly dated but is assumed by Anderson et al. (2016) to be an MIS 3 feature because the Mississippi River is known to have fed into the Mississippi Canyon during MIS 2.

As defined by Anderson et al. (2016), the western channel belt complex consists of channel belts that converge seaward and occupy an area about 80 km wide. Individual channel belts are more than 35 m deep, and have width-to-depth ratios generally greater than 30:1 (Suter 1987). It is inferred that these channel belts lay on the surface of the coastal plain without any associated paleovalleys because these channel belts are single, not multi-story. One of the channel belts appears to be an offshore extension of one branch of the paleo-Mississippi Lafayette Meander Belt and associated Avoyelles alloformation, which ends in a large shelf phase and shelf margin delta, known as the "Western Louisiana Delta." This delta has been radiocarbon-dated by Wellner (2004) at 30 ka BP. However, radiocarbon dates in that age range have sometimes proved unreliable and misleading within the Gulf of Mexico, as noted by Stapor and Tanner (1973), and should be corroborated using optically stimulated luminescence dating.

3.3.4 Last Glacial Maximum sea level MIS 2–LGM

During the last glacial cycle, the maximum lowstand of sea level occurred during the Last Glacial Maximum (MIS 2), between 30 and 19 ka BP. Sedimentological and age data from the Huron Peninsula, New Guinea, indicated that sea level fall to this lowstand began about 32 ka with a rapid drop of 30 to 50 m over a period of 1 to 2 ka (Lambeck et al. 2002). The exact amount of global sea level fall is controversial, but consensus based on ice-volume models and the Barbados coral record is near -125±5 m (Nakada and Lambeck 1989, Fairbanks 1989, Bard et al. 1990, Fleming et al. 1998).

The elevation of the relative sea level lowstand in the Gulf of Mexico during MIS 2 is still poorly constrained. Using glacial isostatic adjustment modeling, Törnqvist et al. (2006) estimated that relative sea level at this time was 34 m to 40 m "below shelf edge." Bart and Ghoshal (2003) concluded, from an examination of the shelf-margin delta geomorpology within the northeastern Gulf of Mexico, that relative sea level during the MIS 2 was about -65 m to -85 m depending on location along the shelf edge. Along the southeastern Louisiana continental shelf, Roberts et al. (2004) argued that relative sea level was at between -85 m and -89 m during the formation of the Lagniappe delta about 19 ka. They also reported a still inexplicable drop in relative sea level of sea level of 30 m over a 1,000 to 3,000 period about 16–18 ka. The ICE5G_120550_Full model of Martini (2015) produces relative sea level values similar to those reported by Bart and Ghoshal (2003) and for 19 ka by Roberts et al. (2004).

On the basis of limited data, it appears that the paleoshoreline in the northern Gulf was situated at or near the shelf break (approximately -65 m to -85 m water depth) during the MIS 2 glacial lowstand of sea level (Bart and Ghoshal 2003, Roberts et al. 2004). As a result, almost the entire continental shelf was a subaerially exposed coastal plain and available for occupation by humans. Prominent features of the lowstand coastal plain included fluvial valleys and lowstand, shelf margin deltas. Offshore, they fed large amounts of sediment to delta-fan complexes, slope fans, and sediment-gravity-flow deposits, along with hemipelagic drapes and contourites on the continental slope (Anderson et al. 2004, 2014, and 2016). The exact duration of time for which this lowstand coastal plain was subaerially exposed is still uncertain, but it lasted only a few thousand years. This surface, buried by younger transgressive deposits, forms a prominent bounding unconformity known by a variety of names including "Base of Holocene, Late Wisconsinan unconformity," and "Pleistocene-Holocene surface." Simms et al. (2007) have constructed a paleogeographic map of this surface (Figure 3.5).

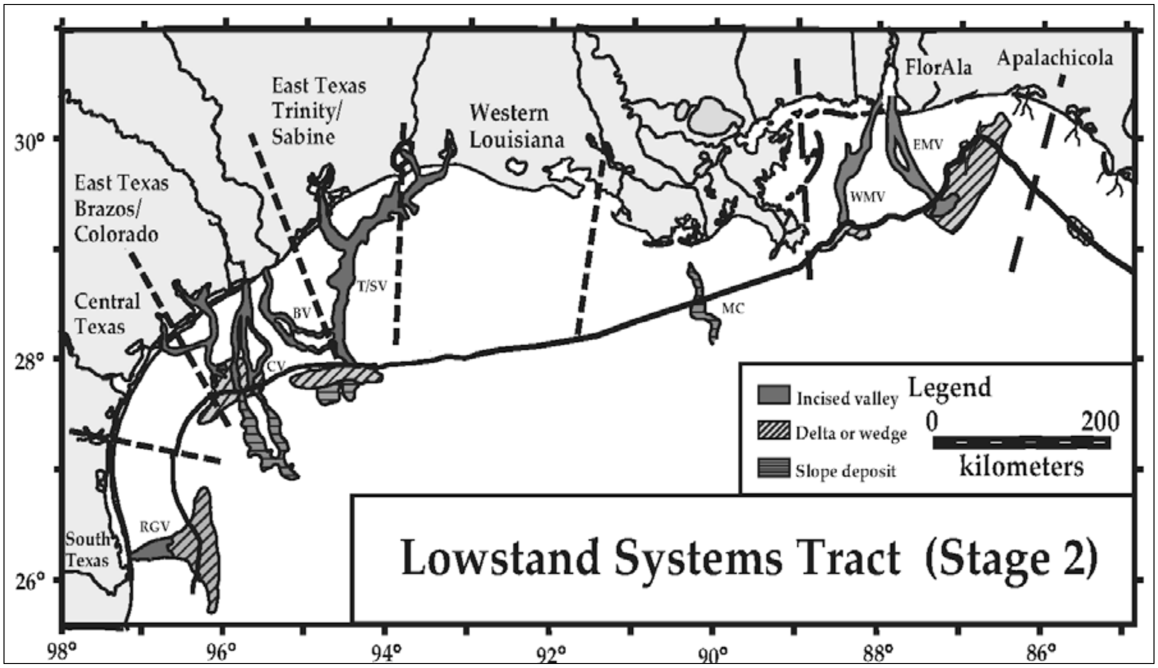

Figure 3.5. Paleogeographic map showing the major depositional systems that existed on the shelf and upper slope during the 22 ka to 16 ka (MIS 2) lowstand.

Lowstand paleovalleys are labeled as follows. RGV = Rio Grande, CV = Colorado, BV = Brazos, T/SV = Trinity–Sabine, WMV = west Mobile, and EMV = east Mobile. MC = Mississippi Canyon. Reprinted from Anderson et al. (2004)

During the MIS 2 sea level lowstand, fluvial systems within the northeastern Gulf of Mexico coastal plain extended their valleys to the shelf edge and incised their valleys deeply into the surface of the coastal plain. Because of the low gradients of the east Texas and western Louisiana continental shelves, the lowstand paleovalleys of the Calcasieu, Mermentau, Sabine, and Trinity rivers become broader and shallower in a gulfward direction (Figure 3.6; Anderson et al. 2016). Typically, these paleovalleys are 40 m deep at the present shoreline (Blum and Price 1998). Within the paleovalleys of the Sabine and Trinity rivers, well-defined fluvial terraces have been found (Thomas and Anderson 1994, Blum et al. 1995, Rodriguez et al. 2005). The presence of similar terraces within the onshore valleys of the Calcasieu and Mermentau rivers indicates that, depending on transgressive shoreface and tidal ravinement depths, terraces likely are preserved in the offshore paleovalleys of these rivers. They also suggest that the incision of the onshore valleys, and presumably the equivalent offshore paleovalleys of the Calcasieu, Mermentau, Sabine, and Trinity initially occurred during either MIS 3 or even MIS 4.

3.3.5 Holocene Sea Level: MIS 1

Starting about 19 to 18 ka and lasting until about 6 ka, rapid deglaciation of continental ice sheets occurred. As a result, at the end of the Last Glacial Maximum, about 19 ka, global sea level correspondingly rose rapidly by around 125±5 m to near present about 6 ka. After about 6 ka, the surviving ice sheets have stayed relatively stable and, as a result, rates of sea-level rise decelerated significantly to an average rate of 1.4 mm/yr (Lambeck et al. 2002).

However, it is well documented that global sea level during the last deglaciation did not rise continuously, but in a series of steps with different rates. There was a distinct and short period of stability during the Younger Dryas, between 12.5 and 11.5 ka, when ice melt briefly ceased (Chappell and Polach 1991, Lambeck et al. 2002). Also, episodes of extremely rapid global sea level rise that are related to accelerated ice-sheet decay and melt-water discharge have also been recognized. The two most widely recognized and accepted are referred to as melt-water pulse 1A (MWP-1A) and melt-water pulse 1B (MWP-1B) (Fairbanks 1989; Bard et al. 1990). Although the precise timing, duration, amplitude, and rates of global sea-level rise associated with these events remain a subject of continuing discussion, MWP-1A is generally accepted as having started at 14.5 ka, after which global sea level rose by 40–55mm/ yr for about 500 years (Edwards et al. 1993). In the case of the more ambiguous and less certain MWP-1B, it is speculated that it began about 11.5 ka, lasted for about 200 years, and produced global rates of sea-level rise comparable to those of MWP-1A (Fairbanks 1989). Additional melt-water pulses have been reported as beginning at 19 ka (Clark et al. 2004) and 7.6 ka (Blanchon and Shaw 1995); however, their significance and extent have been disputed (Woodroffe and Horton 2005).

As discussed by Anderson et al. (2008) and Anderson et al. (2016), the relative sea level history of the northeastern Gulf of Mexico is well constrained for the terminal Pleistocene (latter part of MIS 2), and the Holocene (MIS 1). Starting about 19 ka the rate of sea level rise is unconstrained, but was rapid and punctuated with intervals of very rapid sea level rise until about 10 ka. Between about 10ka and 7.5 ka, the mean rate of sea level rise was about 4.2 mm/yr. At about 7.5 ka, the mean rate of sea-level rise slowed to 1.4 mm/yr until about 4.0 ka. At that time, the current highstand began when the rate of sea-level rise slowed to ~0.4 to 0.6 mm/yr. It was at about 4±1 ka when most of the current strandplains, barrier islands, peninsulas, and chenier plains began to form (Anderson et al. 2014 and 2016, Milliken et al. 2008a).

During the late MIS2 and MIS 1 marine transgression, preservation of the coastal deposits was minimal (e.g., Anderson et al. 2016; Rodriguez et al. 2004; Penland et al. 1988; Belknap and Kraft 1981). Some fluvial deposits, such as channels, floodplains, and terraces and associated bayhead delta, estuarine, and tidal deposits are preserved and comprise the fill within major paleovalleys, such as the Sabine-Trinity system (e.g., Thomas and Anderson 1994, Rodriguez et al. 2005, Simms et al. 2006). As will be discussed in more detail, low preservation potential outside of the major paleovalleys is the result of deep shoreface and tidal ravinement processes (Goff et al. 2015, McBride et al. 2004, Penland et al. 1988). As a result, the record of paleosurfaces and coastal evolution on the continental shelf remains fragmentary outside of paleovalleys (Nelson and Bray 1970, Thomas and Anderson 1994, Anderson et al. 2014 and 2016, Simms et al. 2006).

4. Shelf Morpho-sedimentary Response to Sea Level Change

A variety of processes act to modify paleolandscapes of a coastal plain as changes in relative sea level and sediment supply result in either the landward movement of the shoreline, a transgression, or the seaward movement of the shoreline, a regression. As indicated by the previously discussed sea level history of the last glacial–interglacial cycle, MIS 5 through 1, changes in shoreline can be subdivided into four general modes of coastal plain and continental shelf response to relative sea level change: 1) continuous transgression, 2) episodic transgression, 3) normal regression, and 4) forced regression.

4.1 Transgressive Surfaces

The significant modification of a subaerial coastal plain occurs during both continuous and episodic transgression as the result of relative sea level rise and resulting formation of transgressive surfaces. The surface formed is either a flooding or ravinement surface. A flooding surface is a disconformity produced by transgression that marks an abrupt increase in water depth. For example, this surface can be the abrupt upward change from either the subaerial exposed surface of a coastal plain to marsh and eventually lagoonal sediments or from shallow-water sediments into those that accumulated in deeper water. If this surface exhibits significant erosion, it is known as a ravinement surface. A ravinement surface is an erosional surface produced in the coastal zone, by either wave erosion or by tidal scour in channels within estuaries, during marine transgression, often of a formerly subaerial environment. An in-depth review of transgressive surfaces can be found in Cattaneo and Steel (2003).

4.1.1 Coastline flooding (ravinement) surfaces

Of significance to the preservation of paleolandscapes and archaeological sites is what is designated in this report as a coastal flooding surface. A coastal flooding surface is a disconformity produced by transgression that marks an abrupt change from the terrestrial sediments of a coastal plain to either lagoonal or marine sediments. This disconformity can be either the relatively intact surface of a coast-wise terrace, fluvial terrace, floodplain, or delta plain or, more commonly, eroded to form a minor ravinement surface.

Based on their environment of formation, three types of coastal flooding surfaces are end members of a continuum that grade into each other and form a single, regional coastal flooding surface. They are: 1) bayline flooding (ravinement) surface; 2) lagoonal flooding surface; and 3) soundline flooding (ravinement) surface. The bayline flooding surface is the disconformity produced by transgression of a bay over floodplains, bayhead deltas, and terraces within either an estuary or ria. Depending on the size of the estuary involved, there may not be enough fetch for significant wave action and the resulting erosion and ravinement development to take place. In addition, the accumulation of marsh or swamp deposits also can bury the valley floor and associated terraces under a protective layer of sediment. A lagoonal flooding surface is the disconformity produced by transgression of a lagoon over a coastal plain consisting of coast-wise terraces and non-incised fluvial channel-belts. In case of a narrow lagoon behind barrier islands that would characterize such a coast compared with a wider sound, the fetch would be narrow enough to limit wave erosion and marsh accumulation along the landward edge of a lagoon and would offer some protection to the surface of the subsiding coastal plain and associated channel belts. Finally, a soundline flooding surface is the disconformity produced by transgression of a sound over either a coastal plain or delta plain. The fetch of a sound would be large enough to allow significant, but still limited, wave erosion. As a result, a soundline flooding surface typically is a minor ravinement surface with erosion on a magnitude of a meter or so. An example of a soundline flooding surface is the eroded surface of the St. Bernard delta lobe underlying Chandeleur Sound between the Chandeleur Islands and west to the eroding edge of the subsiding Mississippi Delta Plain.

4.1.2 Shoreface ravinement surface

A major transgressive surface is the shoreface ravinement surface. The shoreface ravinement surface is a well-defined erosion surface produced in the coastal zone, by wave erosion within the shoreface during marine transgression along the seaward edge of either a barrier island chain or coastal plain shoreline. As the shoreline moves landward, the shoreface truncates back-barrier coastal sediments and underlying coastal plain sediments to depths of 8 to 10 m before the wave base is exceeded (Wallace et al. 2010). If the rate of sea level rise is relatively low, the depth of erosion is deep enough that it typically erodes through the coastal flooding surface and the marsh, lagoonal, beach, and other back-barrier coastal sediments that lie landward of the shoreface and into the underlying coastal plain (Figure 4.1). If the rate of relative sea level rise and sedimentation is relatively rapid, the shoreface might not be deep enough to reach the coastal flooding surface and allow preservation of it and some back-barrier coastal sediments overlying the coastal flooding surface (Figure 4.2) (Belknap and Kraft 1981 and 1985, Swift and Thorne 1991, Thorne and Swift 1991).

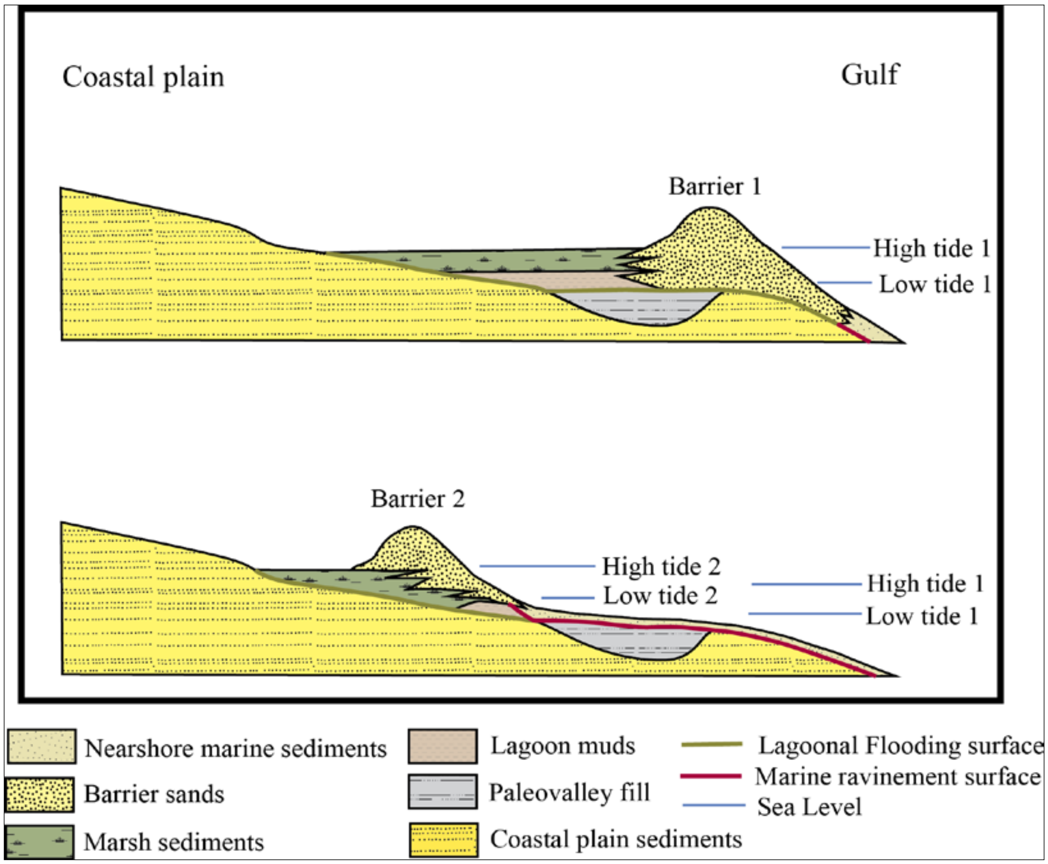

Figure 4.1. Marine transgression by shoreface retreat in which shoreface erosion associated with a rise in sea level removes older coastal deposits, except within the paleovalley, and lagoonal flooding surface and creates marine ravinement surface.

The marine ravinement surface truncates older coastal plain surface and destroys any associated cultural deposits lying on it. The high-tide level [HT] and the low-tide level [LT] are shown at times 1 and 2. Modified after Waters (1992).

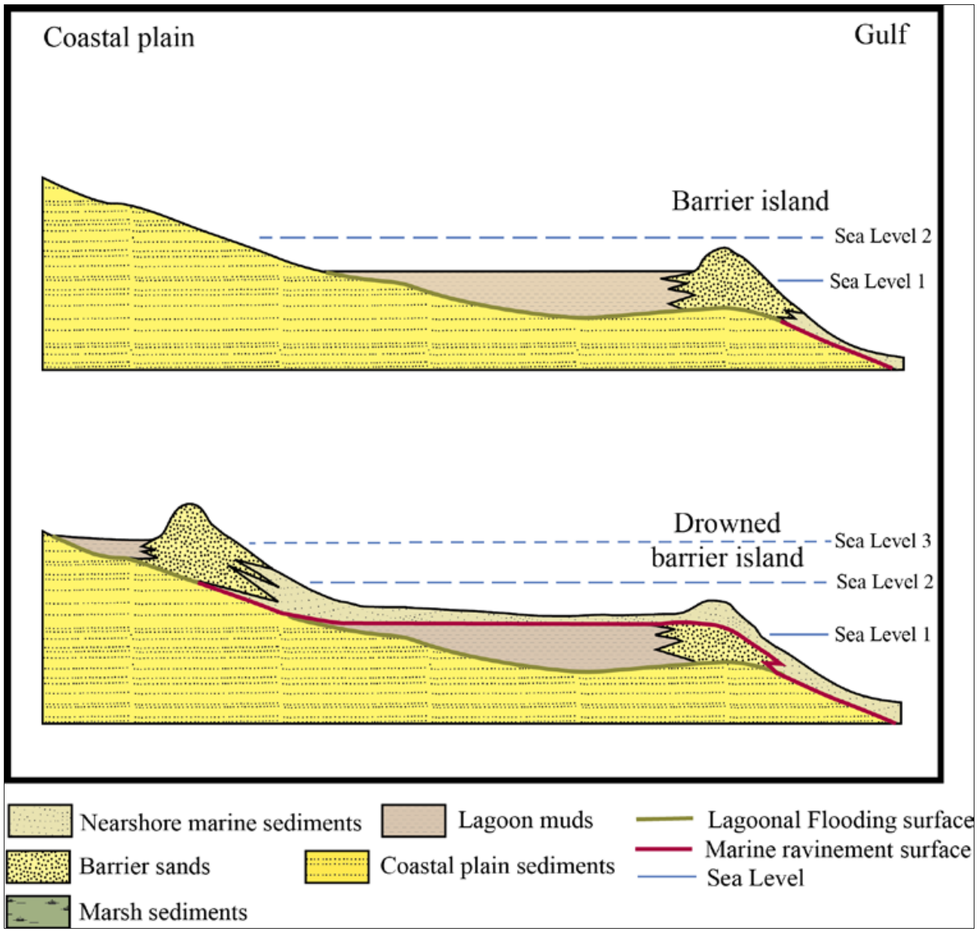

Figure 4.2. Marine transgression by in-place drowning during period of rapid relative sea level rise.

The shoreline jumps landward as relative sea level rises rapidly from a lower position to a higher position. Although a large block of coastal sediments and a segment of lagoonal flooding surface are preserved because the surf and swash zones have little time to erode the older barrier island sequence, the surface of the older barrier island is reworked by surf and swash zones and tidal currents. Modified after Waters (1992).

The shoreface ravinement surface is created by waves hitting the shoreface and interacting with longshore, geostrophic, and tidal currents. Waves are regarded as an overwhelmingly destructive force in terms of terrestrial paleosurfaces and associated archeological resources. The depth of erosion by shoreface erosion varies greatly. Most commonly, it appears to be within the 8 m to 10 m range (Belknap and Kraft 1981 and 1985).

The formation of a shoreface ravinement surface by wave action and associated longshore currents has important ramifications for the preservation of paleosurfaces and associated archaeological deposits. First, the surfaces and sediments of barrier, e.g., beach ridges, dunes, and beaches, and back-barrier environments, are typically prone to being destroyed by shoreface erosion, except under the most unusual circumstances. Beach ridges and dunes, unless they consisted of either carbonates or mixed carbonate-clastic sediments cemented into beachrock, are typically destroyed by shoreface erosion during submergence. Second, the barrier and backbarrier sediments, coastal flooding-ravinement surface, and the underlying Pleistocene sediments are vulnerable to erosion depending on the depth of erosion. In general, the slower relative sea level rises from a combination of subsidence and eustatic sea level rise, the longer the shoreface is subject to the erosive power of waves and the deeper it cuts into the barrier and backbarrier sediments. If the rate of relative sea level rise is slow enough, the shoreface eventually erodes through the barrier and backbarrier sediments and coastal flooding-ravinement surface and into the underlying Pleistocene sediments. The faster relative sea level rises, the shorter the shoreface is subject to the erosive power of waves and it might not erode completely through the backbarrier sediments and leave the coastal flooding-ravinement surface and any underlying deposits intact. Finally, the lighter materials found within any barrier, backbarrier, and underlying sediments that are eroded will be transported either in the direction of littoral drift or shoreward onto the beach. Denser and larger material, e.g., shells, artifacts, and pedogenic nodules, will tend to remain in place, as the size and weight for materials that can be transportable by wave action and longshore currents are limited. This process will concentrate these clasts, including artifacts, as an erosional lag associated with and overlying the shoreface ravinement surface (Belknap and Kraft 1981 and 1985, Murphy 1990).

Geostrophic currents are a dominant process on the Texas-Louisiana continental shelf. They are currents that are sustained by a balance between a pressure gradient force, e.g., either the Gulf Stream or hurricane storm surge, and the Coriolis force. Geostrophic currents are insufficient as an erosional force in the surf zone. However, they occur in combination with combined flow to transport sand offshore on the shelf during storms. They will have the same general effects as wave action on shoreface erosion and associated clast (artifact) transport and concentration (Snedden 1985, Snedden et al. 1988, Siringan and Anderson 1994).

4.1.3 Tidal ravinement surfaces

Another transgressive surface that is often important in terms of the preservation and destruction of paleosurfaces is the tidal ravinement surface. In a wave- and tide-dominated environment, the geometry and extent of a tidal ravinement surface is a function of the tidal ravinement processes, which characterize either an estuary or tidal inlet along a barrier shoreline.

The tidal inlets associated with barrier island chains of a subsiding delta lobe can be more or less anchored by the paleochannels of deltaic distributaries (Kulp et al. 2007). In case of such tidal inlets, which Féniès et al. (2010) refers to as a “anchored tidal ravinement,” the inlet remains in place and the cross-section of the original distributary channel is enlarged with time as the bay behind it and associated tidal prism enlarges. This has the effect of eroding and eventually destroying the original channel and the natural levees on either side of a deltaic distributary’s paleochannel (Boyd et al. 2006).

Other tidal inlets are associated with barrier islands within broad estuaries; the tidal inlet often migrates in response to lateral spit accretion. These processes create a laterally persistent, coast-wise tidal ravinement surface (Miner et al. 2007). Such tidal ravinement surfaces often cut into, across, and cap the fills of paleovalleys and adjacent interfluves along former shoreline positions. Depending on the depth and lateral extent of the paleovalley, all or part of the fills can be removed by tidal ravinement. As a result, the preservation of the bayline (coastline) flooding surface, terrace surfaces, and other paleosurfaces will depend on the depth of erosion by the tidal surface. All else being equal, the thickness of estuarine and fluvial deposits, which is a function of the size and depth of the paleovalley and size of the drainage basin, will be a major factor (Boyd et al. 2006, Féniès et al. 2010).

4.2 Paleovalleys and Channel Belts

As previously noted, formation of paleovalleys as the result of forced regression created by glacial and stadial sea level fall and lowstand has repeatedly occurred within the northwestern Gulf of Mexico. During the subsequent interstadial or interglacial sea level rise and highstand, these paleovalleys started to fill with sediments before the subsequent migration of the shoreline over them. This creates an aggradational package of sediments that often has a significant potential for the preservation of paleosurfaces (Anderson et al. 2004, 2014, and 2016, Suter 1987, Suter et al. 1987).

4.2.1 Forced regression, paleovalleys, and channel belts

Falling relative sea level causes basinward retreat of the shoreline, a process termed “forced regression” by Plint (1996). The fall in relative sea level that results from a forced regression exposes the continental shelf as coastal plain to subaerial erosion. Within the northwestern Gulf of Mexico where the gradient of the continental shelf is nearly equal to or greater than the gradient of its river course, as is typical of major, extrabasinal river systems, a river valley will vertically incise (downcutting) as its mouth migrates seaward. Depending on the degree of relative sea level fall and gradient of the continental shelf, its course continually grades downward to progressively lower sea levels and creates a deeply incised valley on the exposed shelf (Anderson et al. 2004, 2014, and 2016, Suter 1987, Suter et al. 1987).

If the forced regression episodically alternates with periods of either stability or sea level rise, then channel belts will likely be created that fill the valley at each period by lateral migration of the contemporaneous channel. At the same time lateral planation by river scour creates an erosional surface that is buried as soon as it is created. This is called a “regional composite scour surface” and forms the bottom of a valley. The resumption of relative sea level fall will typically result in incision of the former channel belt and associated floodplain and formation of subaerial paleosurfaces in the form of terraces lying above and flanking a newly developed floodplain and channel belt. If the fluvial incision is deep enough, the lateral planantion of the new channel belt will cut through and destroy both the older fluvial deposits and its composite scour surface. As a result, the bottom of the valley will be a composite of regional composite scour surfaces that never existed as a single subaerial topographic surface and that formed diachronously at the channel-belt scale over the entire fall to rise of a base-level cycle (Holbrook and Bhattacharya 2012).

As is typically the case of intrabasinal river systems with significantly smaller drainage basins and river gradients that are largely equal to those of the continental shelf exposed during the forced regression, either no or relatively insignificant entrenchment will occur (Anderson et al. 2004, 2014, and 2016, Suter 1987, Suter et al. 1987). Instead, the river will form a channel belt that is generally level with or occupies a fluvial ridge lying above the level of the coastal plain as observed for the surface of the MIS 5 coast-parallel terraces in East Texas by Aronow (1971) and Van Siclen (1985 and 1991).

In case of Southwest Louisiana, an incised, post-MIS 5e Red River paleochannel has been observed near Lake Charles, Louisiana in interpreted LiDAR DEMs. In this case, it appears that the lowering of relative sea level during the latter part of MIS 5 was at a rate that allowed for the downcutting of a paleochannels, but prohibited the development of a paleovalley by lateral migration. The entrenchment of this channel appears to have beheaded what are likely fluvial or deltaic distributary ridges as it extended its mouth gulfward.

Modern valleys are defined as elongate topographic lows that contain rivers, which lack the discharge to routinely flood beyond the valley incision and onto the interfluves. However, this criterion often cannot be applied effectively in identifying buried, prehistoric paleovalleys or differentiating them from unincised channel belts that once were part of the relatively flat surface of submerged coastal plain. To differentiate a channel belt lying at the bottom of a paleovalley and a channel belt lying unincised on the surface of a coastal plain, a proxy for incision associated with a channel belt is required (Holbrook and Bhattacharya 2012). Thus, the recognition of a paleovalley requires either (a) the preservation of recognizable valley walls that extend above the level of the floodplain or (b) a sand body lying beneath a ravinement surface that consists of two or more sets (stories) of fluvial sands. In the case of single-story channel belts with upper surfaces truncated by regional ravinement surfaces, differentiating between coastal plain channel belts and deeply eroded paleovalleys might be difficult, if not impossible.

4.2.2 Transgressions, paleovalleys, and channel belts

The preservation potential of a channel belt during a marine transgression varies greatly depending on its position relative to the surface of a coastal plain. The paleosurfaces of channel belts that either lie on the surface of a coastal plain or are part of a fluvial ridge are the least likely to survive a marine transgression, because they will be directly exposed to shoreline erosion as the coastal plain is submerged. In contrast, both lying below the level of shoreface erosion and the transgressive backfilling of paleovalleys with a prism of sediments often protects the surface of channel belts, their associated terraces, and other paleosurfaces during a marine transgression (Pearson et al. 1986).

As described by Blum and Törnqvist (2000), a rise in relative sea level results in channel shortening, decreases in the distance over which sediments can be stored and, in most cases, a flattening of the channel slope. Even though discharge is conserved, these changes result in corresponding downstream decreases in stream power, sediment transport rates, and, ultimately in net aggradation. As a result, rising sea level causes a prism of initially fluvial and later estuarine sediments to accumulate within a valley system that extends and tapers upstream from the contemporaneous shoreline. As relative sea level rises, this sedimentary prism buries the surfaces of channel belts, floodplain, and in some cases even low-lying terraces (Blum and Törnqvist 2000, Shen et al. 2012, Blum et al. 2013).

During the periodic episodes of abrupt relative sea level rise, rapid flooding of an estuary will occur. Such episodes cause the overstepping and flooding of bayhead deltas and segments of fluvial terraces and landward shift in estuarine facies and the formation of flooding surfaces. If sea level rise occurs fast enough, the surface of bayhead deltas and terraces are buried and preserved as paleosurfaces within the fill of a paleovalley (Rodriguez et al. 2005, Milliken et al. 2008b and 2008c). If sedimentation rates are high and the paleovalley deep enough, these paleosurfaces can survive subsequent destruction by shoreface erosion and formation of a ravinement surface (Pearson et al. 1986).

As seen in aerial photographs and LiDAR DEMs, the surface of the MIS 5 coastal-parallel terrace of southwest Louisiana and southeast Texas is characterized by numerous relict Pleistocene channel belts and fluvial and distributary ridges along with younger Holocene drainages (Fisk 1948, Bernard and LeBlanc 1965, Aronow 1971 and 2004, Van Siclen 1985 and 1991, Heinrich 2007). During a marine transgression the erosion of the upper 8 m to 10 m of the MIS 5 coastal-parallel terrace would destroy distributary and fluvial ridges and constructional surfaces of relict channel belts. At best, only the truncated coarse-grained basal channel fills of the larger intrabasinal rivers and streams would remain. Although, the surface of the MIS 5 coastal-parallel terrace might survive the initial flooding by backbarrier lagoons and burial by lagoonal sediments, it would be eventually destroyed from shoreface erosion and formation of a ravinement surface.

4.3 Deltas and Chenier Plain

During either stillstands or generally low rates of relative sea level rise, segments of the northwestern Gulf of Mexico coastal plain have experienced extensive periods of normal regression (Figure 4.3). In this part of the Gulf of Mexico, normal regression is the seaward advance of the coastline as a result of the progressive addition of sediment to either the front of a shoreline, as in case of the Louisiana Chenier Plain, or a delta system. In the case of major delta systems, normal regression is typically associated with aggradation of the delta plain because of subsidence-induced relative sea level rise. When delta lobes are abandoned because of delta switching, deltas are locally and periodically subject to marine transgression (Coleman and Roberts 1998a and 1998b).

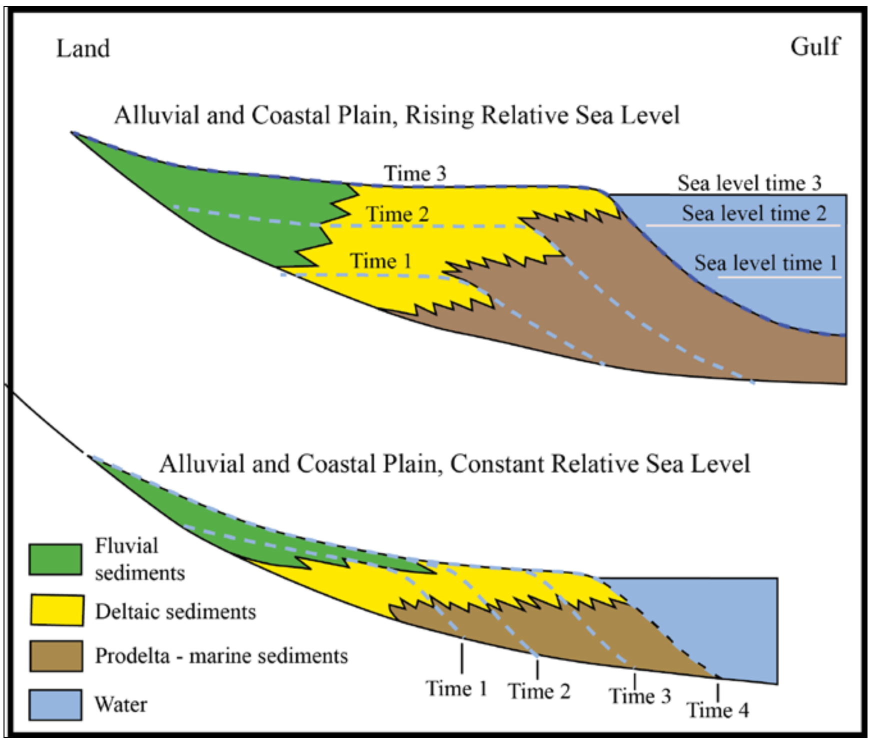

Figure 4.3 Normal regression in which coastline progrades gulfward in response to sediment accumulation during period of constant or slowly rising relative sea level.

The creation of new and productive delta or chenier plains may promote the preservation of cultural resources.

4.3.1 Regressive and transgressive phases of deltas

Although deltas like the Mississippi River Delta have, until historic times, been actively prograding outward into the Gulf of Mexico, each delta lobe evolves through a cycle of events that begins with sedimentation and rapid growth as part of a constructional or regressive phase. The constructional or regressive phase of delta growth typically consists of the rapid infilling of lakes and bays, followed by the local gulfward progradation of the coastline and the deposition of a thick wedge of terrigenous clastic sediment on the adjacent shelf. During this phase, progradation creates a relatively thin shelf delta with numerous distributary channels, merged distributary mouth bar sands, and distributary channels that cut below the base of delta deposit (Penland et al. 1988, Roberts 1997, Coleman and Roberts 1988a and 1988 b).

The constructional or regressive phase is followed by the abandonment of a delta lobe by its distributary channel, which cuts off water and sediment flowing to it, and results in the deltaic deterioration of its destructional or transgressive phase. During this phase, distributaries of the previously active delta lobe become erosional headlands as marine transgression and associated reworking results in the landward migration of the shoreline. Sandy sediments eroded from distributaries that form the headlands is reworked laterally by waves to form flanking barrier islands that enclose restricted interdistributary bays. As the deltaic plain subsides below sea level, the transgressive barrier island arc is separated from the retreating mainland shoreline and eventually forms an inner-shelf shoal. After submergence, the inner-shelf shoal continues to be reworked into a marine sand body on the inner continental shelf. In addition, organic sediments would continue to accumulate in the interior and distal portions of the delta plain relative to the shoreline. This process would bury natural levees and other surfaces within these portions of the delta plain until submerged by relative sea level rise (Penland et al. 1988, Roberts 1997, Coleman and Roberts 1988a and 1988b).

4.3.2 Regressive and transgressive phases of Louisiana Chenier Plain

The Louisiana Chenier Plain within Southwest Louisiana consists of multiple, shore-parallel, sand-rich ridges that range in elevation from 2 m to 6 m. They rest on and are separated from one another by finer-grained, clay-rich sediment. They evolved during the Holocene as the result of the periodic progradation of mudflats that were interrupted by regressional episodes that eroded and reworked existing mudflats and concentrated the coarse sediment fraction into sandy or shelly ridges. The mudflats reflect the accumulation of fine-grained sediment discharged by the Mississippi River and transported westward. During episodes of non-deposition due to eastward shifts of the Mississippi River deposition, coarse-grained sediment and shells are concentrated along the shore forming chenier ridges. Renewed mudflat deposition and isolation of the chenier ridges occurs when a new supply of sediment from the Mississippi River distributaries is again transported westward. Currently, mudflats along segments of the Louisiana Chenier Plain are presently prograding from sediments being transported westward from the Atchafalaya River (Gould and McFarlan 1959, Penland and Suter 1989).

Because of their low relief, the surfaces of both the Mississippi River Delta and the Louisiana Chenier Plain, like the beach ridges of barrier islands and delta plains, would be susceptible to rapid submergence and in-place drowning from a period of rapid, episodic sea level rise. In-place drowning would cause an abrupt lateral shift of the shoreline, called “overstepping,” from its current shoreline to the boundary between these plains and the Pleistocene coast-parallel terraces. Although this would likely prevent complete destruction from shoreface erosion of the bulk of the deposits underlying these plains, their surfaces would still be significantly disturbed and, in case of the cheniers, reworked into inner-shelf shoals similar to the proposed origins for Herald and Sabine banks.

4.3.3 Forced regression, deltas, and beach ridges

During the last glacial-interglacial cycle, forced regression impacted delta and beach ridge systems before the human occupation of the study region during MIS 5b through MIS 3 (Figure 4.4). Forced regressions have resulted in abandonment and stranding of the relict Houston Ridge segment of the Ingleside Paleoshoreline within Calcasieu Parish, Louisiana, and the beach ridges, including Milton Island, of the Ponchatoula beach system adjacent to the north shore of Lake Pontchartrain in southeast Louisiana. Also, off-lapping shelf deltas of the West Louisiana Delta within the southwest Louisiana continental shelf, are attributed to forced regression (Figure 3.4).

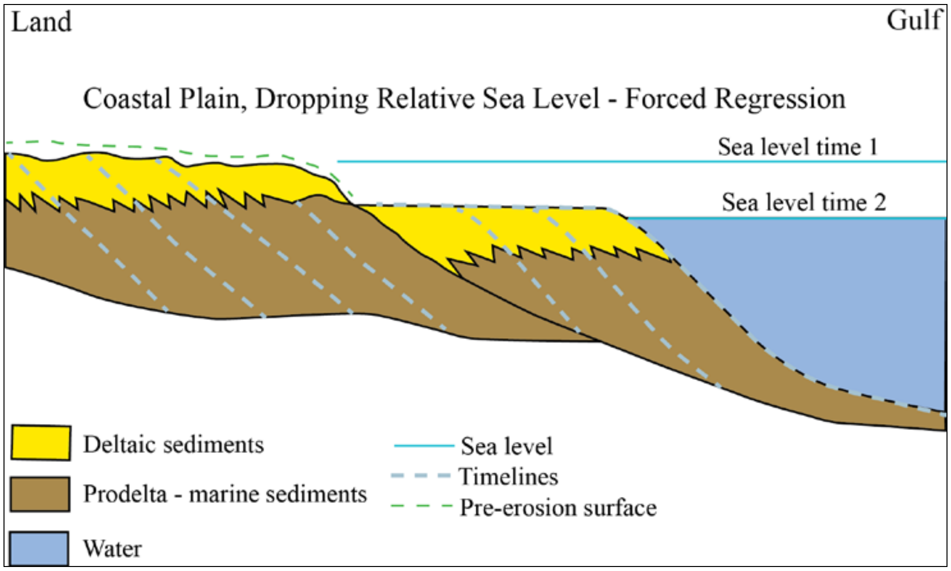

Figure 4.4. Forced regression in which coastline moves gulfward in response to relative sea level fall.

Falling sea level allows for the preservation of cultural resources and opening of new land for occupation.

5 Preservation of Paleolandscapes and Cultural Resources on the Northern Gulf of Mexico Continental Shelf

5.1 Landscape: Cultural Resources Associations

A number of studies–including McIntire (1958, 1971), Coastal Environments, Inc. (1977), Gagliano et al. (1982), Gagliano (1984), Goodwin et al. (1991), Weinstein et al. (1992), Dunbar et al. (1991), Faught and Donoghue (1997), and Faught (2004)–on the northern Gulf of Mexico plain have documented a very close relationship between the pattern of prehistoric settlement and modern geomorphic surfaces and their associated landforms. Within the northwestern Gulf of Mexico coastal plain, these geomorphic surfaces can subdivided into the Mississippi Delta Plain; incised fluvial valleys of the Sabine, Trinity, and other rivers; the chenier plain, barrier islands, salt dome "islands;" and upland coast-parallel terraces. In the northeastern Gulf of Mexico Coastal Plain, another geomorphic terrain, the karst uplands dominates large parts of the landscape.

The Mississippi Delta Plain was widely used by prehistoric cultures as documented by a number of studies. The numerous archaeological studies and cultural resource management surveys have shown that archaeological sites have specific associations with various deltaic landforms. Deltaic archaeological sites are largely associated with the crests of major and minor natural levees. Specifically, preferred locations include crests of natural levees along trunk channels; natural levees of crevasse distributaries near swamp-marsh interfaces; on distributary levees at juncture with crevasse subdeltas; on distributary levees at the heads of major subdelta lobes; and natural levees near the mouths of active distributaries (Figure. 5.1). Other preferred locations for cultural deposits associated with deltas include the shores of bays and lakes within deltas and the beach ridges of barrier arcs and distributary mouth accretion ridges.

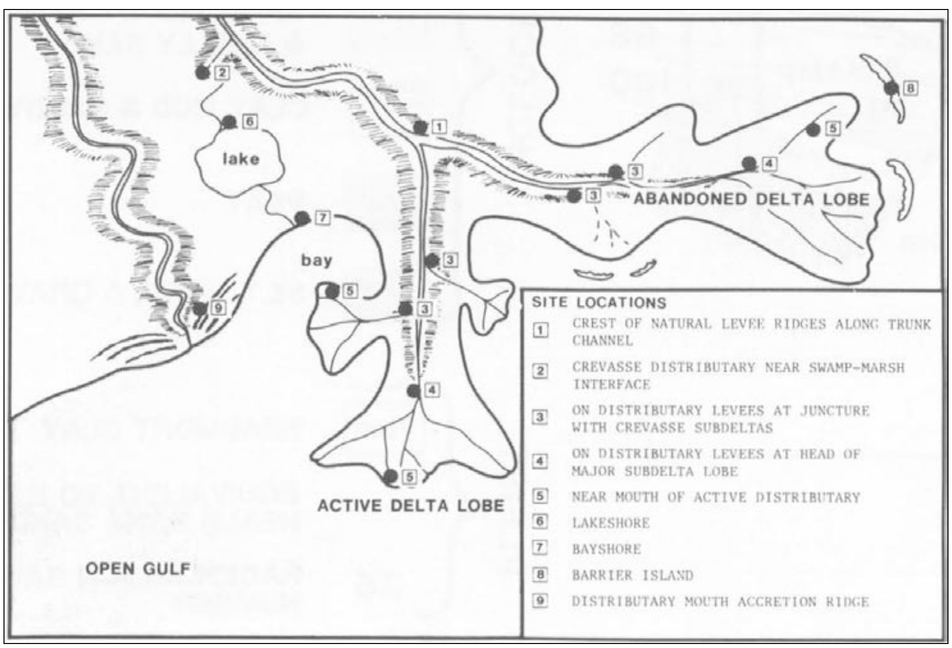

Figure 5.1. Preferred prehistoric site locations in the Mississippi Deltaic Plain.
Reprinted after Gagliano et al. (1982).

Within deltas, the largest accumulations of cultural deposits are typically associated with the crests of natural levees typically either along the cutbank side of meander loop cutoffs, at junctions of tributaries and distributaries, or along crevasse distributaries. Most commonly, the cultural deposits lie on the surface of the natural levee and, thus, accumulated after the distributary had been abandoned as an active course of the parent river system. This is typical of the major cultural deposits associated with intensive occupation areas, burials, and so forth, found within the Mississippi River Delta (Coastal Environments Inc. 1977, Gagliano et al. 1982, Weinstein and Gagliano 1985, Weinstein et al. 1992). Gagliano (1984) noted that cultural deposits also have been found interbedded with sediments of a natural levee, indicating that the site was occupied while the distributary paleochannel was active and within the adjacent channel fill that accumulated after the distributary had been abandoned (Figure 5.2). Gagliano (1984) briefly noted that cultural materials might be found interbedded with backswamp clays and peats where they accumulated either while a distributary was active or after it had been abandoned. Such cultural deposits are usually deposited in standing water and may contain well-preserved perishable artifacts and ecofacts.

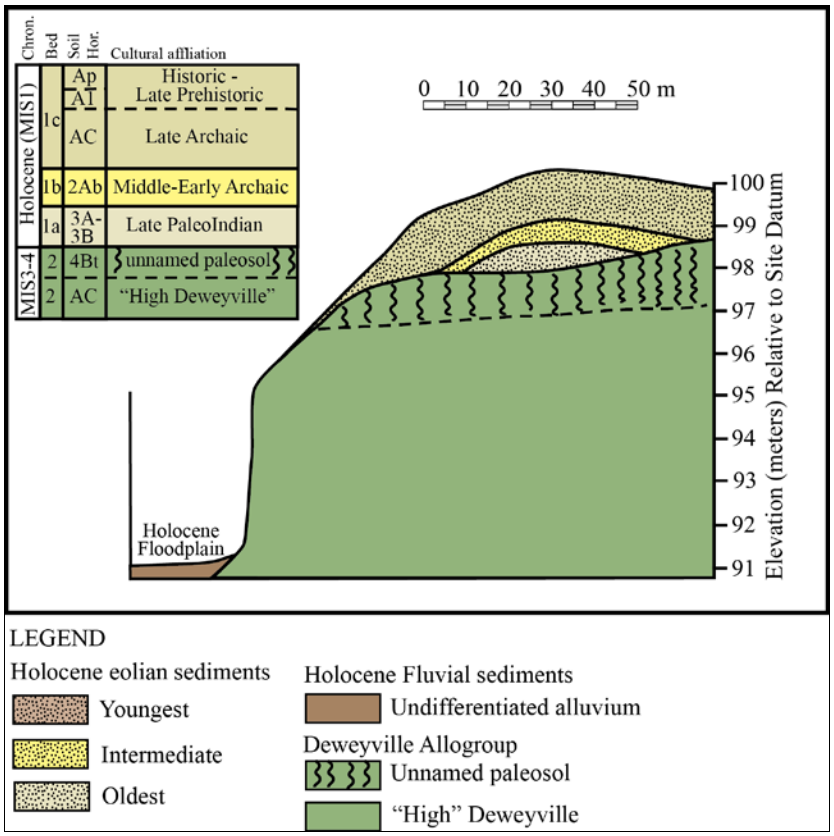

Figure 5.2. Cross section of terrace of "High" Deweyville at McNeill Ranch Site, Victoria County, Texas.

Modified from Aiuvalasit (2007).

As discussed by Coastal Environments, Inc. (1977) and Gagliano et al. (1982), the beach ridges of the Louisiana Chenier Plain, Texas barrier islands, barrier arcs of subsiding delta lobes, and distributary mouth accretion ridges are visually prominent, well-drained, and suited landforms for prehistoric habitation. As a result, significant accumulations of cultural deposits are commonly associated with them. A prime location for the accumulation of cultural deposits is the ends of beach ridges either at the mouth of chenier plain bayous or within barrier-spits adjacent to tidal channels. Because of the nature of the depositional environments, cultural deposits are typically surficial in nature and of restricted thickness (Coastal Environments, Inc. 1977, Gagliano et al. 1982, Gagliano 1984).

The distribution of archaeological sites within fluvial valleys of the major rivers within Southeast Texas and the Louisiana Coastal Plain has been studied by Gagliano et al. (1982), Pearson et al. (1986), Abbott (2001), and Aiuvalasit (2007). They found that, as within deltas, major sites can be found on the crests of natural levees found within active floodplains. Based on research within the Mississippi River alluvial valley, Weinstein (1981) argues that prehistoric cultures preferentially occupied the natural levees bordering abandoned channel segments, with the largest sites being located adjacent to oxbow lakes within these segments, and avoided the natural levees bordering active channels. However, it is important to note that Gagliano et al. (1982), Pearson et al. (1986), Abbott (2001), and Aiuvalasit (2007) conclude that for incised fluvial valley systems, such as the Calcasieu-Sabine-Trinity, the margins of terraces and valley walls nearest active channels were the preferred location for prehistoric occupations and accumulation of significant cultural deposits, and not the natural levees, as observed in the delta plain. In southeast Texas, Abbott (2001) and Aiuvalasit (2007) found that the margins of terraces were also the location for the accumulation of wind-blown sediment periodically during the Holocene. The periodic accumulations of eolian sediments lead to the creation of well-stratified cultural deposits (Figure 5.2). During the flooding of alluvial valleys by sea level rise, the margins of valley walls and terraces overlooking an estuary also were the locations of prehistoric occupations and the accumulation of significant cultural deposits (Figure 5.3) (Gagliano et al. 1982).

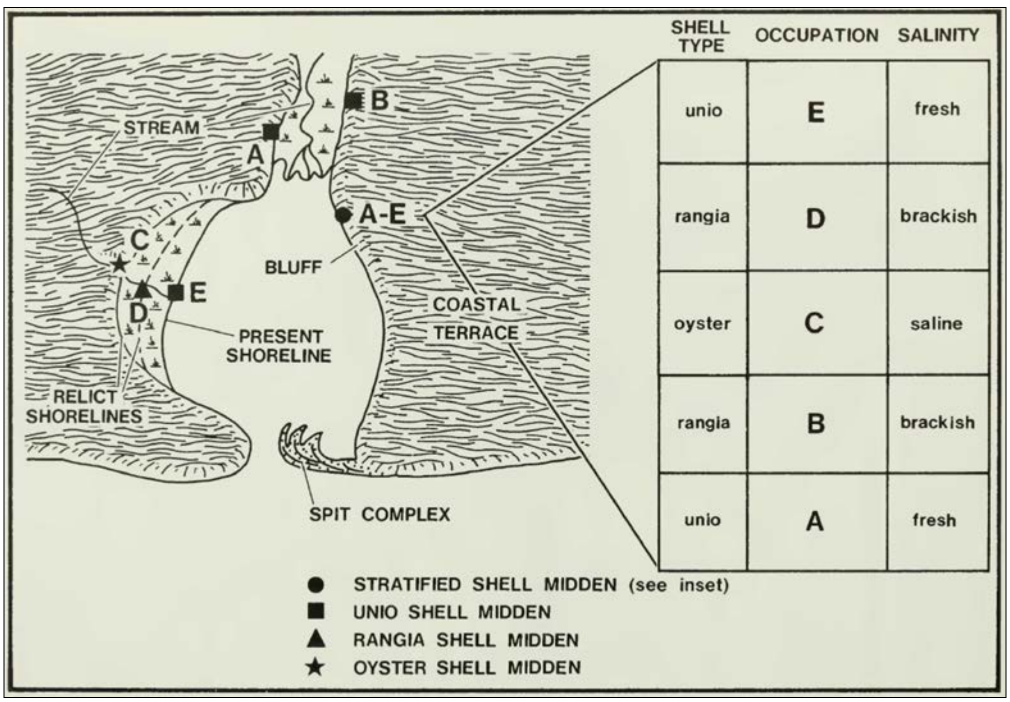

Figure 5.3. Location of shell middens and shell content in a hypothetical coastal estuary.

Also, changes in shell content of a hypothetical shell midden resulting from sea level fluctuations. Reversal in apparent salinity within the stratified site results from a rise in sea level (A–C), followed by stream progradation (D–E). A: oldest occupation; E: youngest occupation. Reprinted from Coastal Environments, Inc. (1977).

Although much less well-studied than other coastal plain surfaces, interfluves (uplands separating valleys) of Pleistocene coast-parallel terraces have distinct landform-associated cultural deposits. When sea level was much lower than present during the last glacial-interglacial cycle, these uplands extended out onto the continental shelf and exhibited a collection of relict natural levees of channel belts and deltaic distributaries and relict beach ridges. Better drained because of their coarser texture and greater elevation, the crests of these features are likely the preferred location of prehistoric occupation and accumulation of cultural deposits, as suggested by limited surveys conducted for cultural resource management studies, such as Weinstein et al. (1979a). Within the coast-parallel terraces, cultural deposits also have been found associated with the summit of innumerable pimple mounds that cover the surface of the Pleistocene coast-parallel terraces (Aten and Bollich 1981, Jones and Schuman 1988). As discussed further below, these uplands on the exposed shelf have little to no preservation potential during sea-level rise.

5.2. Paleolandscape and Cultural Resources Preservation

From the previous discussion, it is obvious that large areas of the Gulf of Mexico continental shelf were exposed as land during the last glacial-interglacial cycle, but have become submerged and reworked as a result of sea level change. It is also well documented that human populations were present within the Gulf of Mexico Coastal Plain during at least the last 13,000 years and possibly for the last 20,000 years. This means that the Louisiana and East Texas continental shelves were potentially the sites of pre-historic human occupation. The main limiting factor in terms of preservation potential is whether the paleosurfaces on which cultural deposits might have accumulated survived the Late Wisconsin–Holocene transgression.

5.2.1 Transgression: coastline flooding surfaces related to paleolandscape preservation

Within the southeastern Texas coast in the region of the McFaddin Beach Site, Site 41 JF 50, is an example of the relatively gentle submergence of Pleistocene coast-parallel terraces by rising sea level. Behind the modern shoreline, the terraces are covered as sea level rises by a blanket of marsh that allows for little, if any disturbance of the surface, except by bioturbation. Elsewhere, where a narrow lagoon exists between a barrier island and landward edge of the lagoon, there often exists a marsh fringe that also protects the terrace as it is submerged. Similar conditions can be observed along the inland edge of the Louisiana Chenier Plain, where the Pleistocene coast-parallel terraces, along with intact pimple mounds, are disappearing beneath coastal marsh.

Within estuaries, the effects of the transgression of the bayline flooding surface can be highly variable, depending on factors such as relief and steepness of slope of geomorphic surfaces being submerged and fetch of the bay. In case of fluvial terraces, bayhead deltas, and floodplains, found within fluvial valleys, their generally flat nature allows large areas of them to be rapidly flooded in response to relatively small rises in relative sea level (Rodriguez et al. 2005). This would allow for rapid flooding, submergence, and subsequent preservation of various geomorphic surfaces and associated cultural deposits within valleys during rises in relative sea level. Similarly, the episodic periods of rapid, relative sea level rise noted by Milliken et al. (2008b and 2008c), Rodriguez et al. (2005), and others should have the same effect of rapidly submerging relatively flat to gently sloped geomorphic surfaces and ultimately abetting their preservation with paleovalley fills. Finally, the aggradation of a prism of fluvial sediment upriver from bayline should have the effect of burying and preserving the surfaces of crevasse splays, natural levees, and geomorphic surfaces well in advance of the flooding of a river valley.

On the other extreme, cultural resource studies, e.g., Lowery (2001) and Lowery and Stanford (2013), of large estuaries, such as Chesapeake Bay, Maryland, have found that strong wave action can be generated where the over-water distances across which the wind blows or “fetch” is large enough and the orientation to prevailing winds is optimum. Such wave action can result in severe erosion of the shoreline and destruction of cultural resources during coastline flooding. In addition, where wave action is parallel to the eroding shoreline, littoral drift has the potential to move artifacts great distances along the shorelines.

In the case of the Mississippi River Delta, the effect of soundline or bayline flooding and associated land loss has long been noted by numerous studies. However, except for briefly noting the reduction of shell and other middens into erosional beach lags, precise effects of soundline or bayline flooding on coastal sites has been noted by only a relative few studies, such as Gagliano (1984), Gagliano et al. (1982), Perrault et al. (1994), Weinstein et al. (2012), and Saltus et al. (2003). Although only as little as the upper 0.5 to 1 meter of the deltaic surface might be eroded, it is enough for waves and currents to winnow out finer-grained material from either shell middens or earth mounds, earth middens, and other cultural deposits. This process destroys the integrity of a site in terms of stratigraphy and geometry, and reduces it to a reworked, erosional lag of the coarsest components, such as shell and artifacts. These components are often thrown back on the marshy shore as a perched beach composed of a thin veneer of shell and other site debris. Because of subsidence within the delta plain, additional intact cultural deposits often occur on the level of shoreline erosion and survive either soundline or bayline flooding (Gagliano 1984, Gagliano et al. 1982, Perrault et al. 1994, Saltus et al. 2003).

Judging from what is currently known about coastline flooding, it quite likely that significant preservation of geomorphic surfaces and associated cultural resources occurs during this early stage of a transgression as the bay-shore processes encroach on paleolandscapes. Even if some disturbance of geomorphic surfaces occurs, it appears that a significant proportion of extant cultural resources survive. However, as discussed later, the ultimate preservation of geomorphic surfaces and associated cultural resources depends how much they are impacted by subsequent shoreface erosion during marine transgression.

5.2.2 Transgression–shoreface ravinement

As pointed out by Pearson et al. (1986) and Waters (1992), the most destructive part of a transgression, in terms of geomorphic surfaces and cultural resources, is the result of shoreface erosion and corresponding formation of a ravinement surface. As previously noted, shoreface erosion along the gulfward edge of barrier islands, chenier plains, and barrier island arcs will result in the removal by erosion of a substantial thickness of the uppermost sedimentary deposits and the cultural deposits that they contain forming these landforms (Figure. 4.1). Finer-grained material would be removed in suspension and coarser material would be transported offshore by geostrophic currents, and along the coast by longshore currents, and deposited along the shore as a transgressive beach. Some of the denser and larger material, e.g., shells, artifacts, and pedogenic nodules, tend to remain in place and concentrate as an erosional lag associated with and immediately overlying the ravinement surface. Some of the less dense and smaller artifacts would also be thrown onshore and incorporated into transgressive beaches and beach ridges.

Saltus et al. (2003) and Murphy (1990) report the presence of submerged, in situ cultural deposits gulfward of the eroding shoreline of coastal barrier islands that are associated with either a submerging delta plain or coast-parallel terrace. From these observations, both concluded that cultural deposits can survive shoreface erosion. However, both studies are looking at the beachface, which is only in the initial phase of erosion, and uppermost segment of the entire shoreface. Because shoreface erosion is as much as 6 m to 15 m along the Texas coast (Rodriguez et al. 2001, Wallace et al. 2010); 8 m to 12 m along the front of the Chandeleur Islands (Miner et al. 2009), and 3 m to 5 m along the Isles Dernieres (Penland et al. 1985), it appears that shoreface erosion effectively destroys most geomorphic surfaces and associated cultural deposits. The main preservation of paleosurfaces occurs in paleotopographic features, e.g., paleovalleys and in the northeastern Gulf of Mexico offshore Florida, sinkholes, and paleokarst valleys, that lie below the level of shoreface erosion (Figure 5.4) (Pearson et al. 1986, Faught 2004).

The McFaddin Beach Site (41JF50) is a well-documented example of a mixed assemblage containing Clovis, Eden, San Patrice, and other projectile points, along with other artifacts and Late Pleistocene mammoth, camel, and horse bones concentrated on a transgressive beach by shoreface erosion of Holocene and Late Pleistocene deposits (Coastal Environments, Inc. 1977, Long, 1977, Stright et al. 1990).

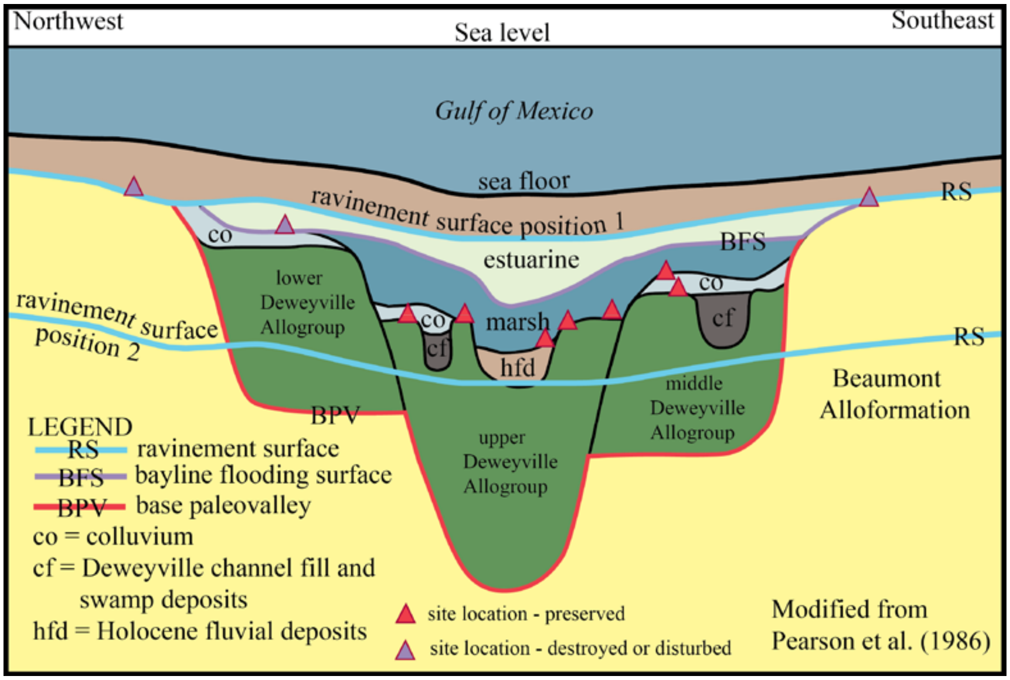

Figure 5.4. Preservation potential of cultural deposits within a major paleovalley fill (e.g., Sabine) in respect to depth of erosion of the ravinement surface.

Modified from Pearson et al. (1986).

Although the bulk of the barrier island and lagoonal sedimentary sequences underlying the barrier island and associated lagoon are preserved from shoreface erosion, the beach ridges and other landforms underlying surficial deposits are eroded and reworked. It is highly unlikely that any cultural deposits associated with beach ridges and other landforms can survive a shoreface retreat intact.

5.2.3 Tidal ravinement

The formation of a tidal ravinement also has the potential to disturb and destroy paleosurfaces and associated cultural deposits. Anchored tidal ravinements, which are associated with the inlets of barrier arcs of subsiding sublobes of the Mississippi River Delta, widen and deepen the paleochannels of their former deltaic distributaries. As previously noted, the result of overall enlargement of such paleochannels is to erode and disperse with time the natural levees of and any cultural deposits associated with the paleochannels. Similarly, the creation of a sweeping tidal ravinement by the lateral migration of a tidal inlet will erode the sequence of sediments that lie above the depth of the thalweg of the tidal channel as it migrates. This depth of tidal ravinement can be greater than that of shoreface ravinement at deep inlets, if the inlet is deeper than the depth of shoreface erosion (Miner et al. 2007). At the least, any beach ridges and typically submerged or buried natural levees of a deltaic distributary paleochannel will be destroyed, along with any associated cultural deposits as a tidal inlet migrates through them and a laterally extensive tidal ravinement surface is left behind. At worst, the tidal channel erodes deeply enough to cut entirely through the coastline flooding surface and into either channel belt sands of a paleovalley fill or sediments from a previous sea level cycle.

As reported by Saltus et al. (2003), an example of the ongoing destruction of a paleosurface and associated cultural resources is the Fort Livingston Site (16JE49) that was constructed at the mouth of Barataria Bay and the western end of Grand Terre Island. On the westernmost end of this island, numerous ancillary buildings and a rail line were built on a sand dune spit during the construction of this fort. By the 1930s, enlargement of the tidal inlet had completely eroded this spit and converted this area to open water that later refilled with sand as the spit was rebuilt. Magnetometer investigations by Saltus et al. (2003) in this spit and adjacent tidal channel found magnetic expressions that are likely associated with the heavier remnants of these historic structures. These cultural materials were likely dropped down by erosion to the bottom of the tidal channel when these structures were destroyed by enlargement of the tidal channel and later reburied when the spit reformed. McNinch et al. (2006) and Quinn (2006) describe the scour and settling processes that would cause the downward movement of individual artifacts by removal of sand and smaller-sized sediment.

5.2.4 Normal regression–progradation

As discussed by Waters (1992), normal regression, as the result of progradation of strand plains, chenier plains, and barrier islands is most conducive to the creation and preservation of geomorphic surfaces and associated cultural resources. As the shoreline builds gulfward, their associated landforms are temporarily removed from destructive coastal processes as former coastal sites are left further and further inland as the shoreline progrades gulfward. If progradation is periodically interrupted by a cessation in sediment supply, they will be subjected to wave erosion and shoreline retreat and shoreface erosion. This process can produce transgressive beach ridges and unconformities in beach ridge sequences. The transgressive beach ridges can contain a mixture of older cultural material that has been reworked from the older deposits (Coastal Environments, Inc. 1977, Gagliano 1984, Gagliano et al. 1982, Waters 1992).

These prograding shorelines create an ordered sequence of beach ridges that become progressively younger toward the shoreline on which lie archaeological sites that overall also become progressively younger toward the shoreline (Figure. 5.2.3). The basic cause of this chronological ordering of archaeological sites is that prehistoric people found it advantageous to locate their sites near the ocean. Such chronological ordering of archaeological sites has been documented in the chenier plain of western Louisiana. Within it, the distribution of initial occupation sites corresponds to the sequential development of major chenier ridges (McIntire 1971, Coastal Environments, Inc. 1977, Gagliano 1984, Gagliano et al. 1982, Waters 1992). However, as previously noted, regardless of whether a set of beach ridges are gradually or rapidly submerged, it is quite unlikely that they and associated cultural resources will survive intact.

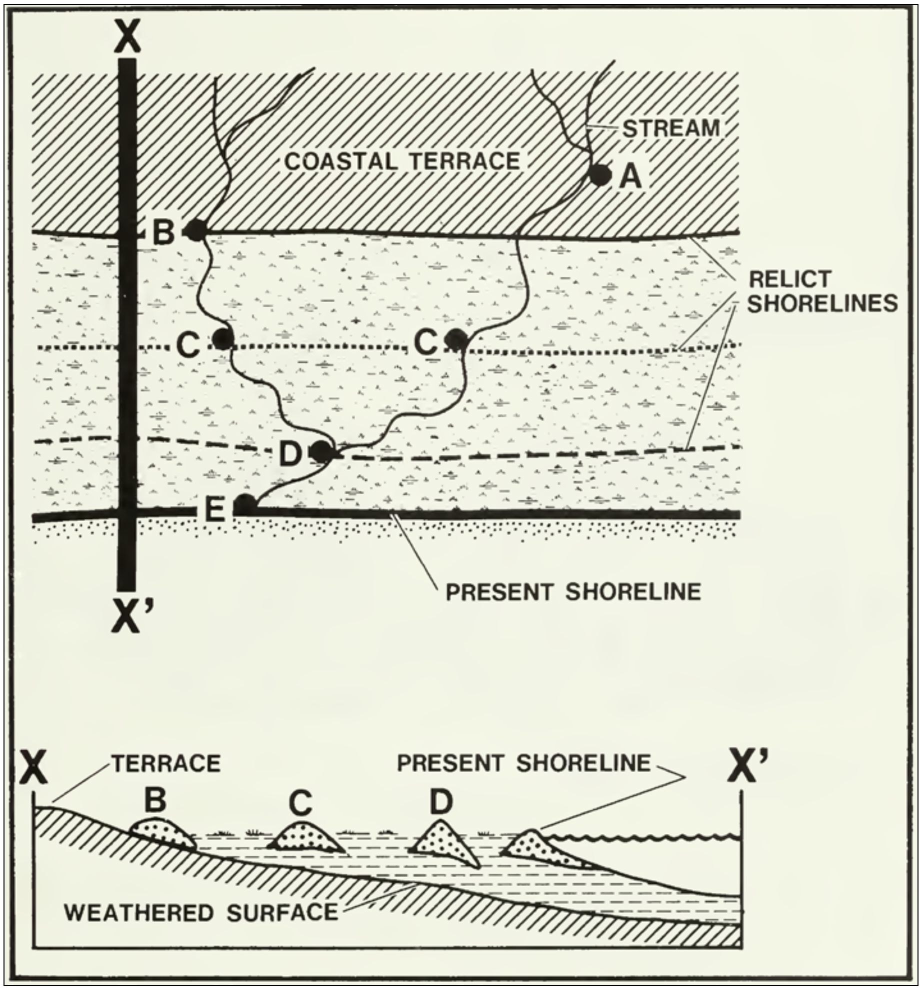

Figure 5.5. Distribution of initial occupation sites in a prograding beach sequence. From A-oldest; to E-youngest.

Reprinted from Gagliano et al. (1982).

Similarly, the Holocene progradation of the Mississippi River Delta created an environment conducive to the creation and preservation of geomorphic surfaces and associated cultural resources (McIntire 1971, Coastal Environments, Inc. 1977, Gagliano et al. 1982, Waters 1992). After establishment of a prodelta platform, deltaic distributaries prograde and bifurcate, which results in the active construction of the plain of a delta lobe. At this time, permanent settlements are lacking from this delta plain because it is too prone to flooding to be comfortably settled. The active construction of a delta plain continues until the distributary course is no longer hydraulically efficient. Then it is abandoned for a new trunk channel that starts construction of a new delta lobe. After a delta lobe is abandoned, , its plain is actively occupied because of the reduced flooding and greater productivity and diversity in fauna and flora. In addition, the abandonment of a delta lobe initiates the transgressive phase of the delta cycle. Initially, as the delta plain subsides, fine-grained sediments accumulate and slowly bury natural levees and mouth bar accretionary ridges and their associated cultural deposits in a protective mantle (Figure 5.6). An example of such a site is the Bayou Jasmine Site (16SJB2), which consists of *Rangia* shell and earth middens, is situated on natural levee ridges of Mississippi River distributaries, and is now subsided as much as 5.5 m and lies buried beneath fine-grained deltaic sediments. However, as the abandoned delta lobe subsides, coastal processes rework the gulfward margin of the delta plains and create a sandy barrier shoreline backed by bays and sounds. At first, the surficial deposits and cultural resources are disturbed and reworked by coastline flooding (Figure 5.7). Later, as the barrier islands move inland, the deltaic and associated cultural deposits are deeply eroded by shoreface processes, as previously described (Figure 4.1).

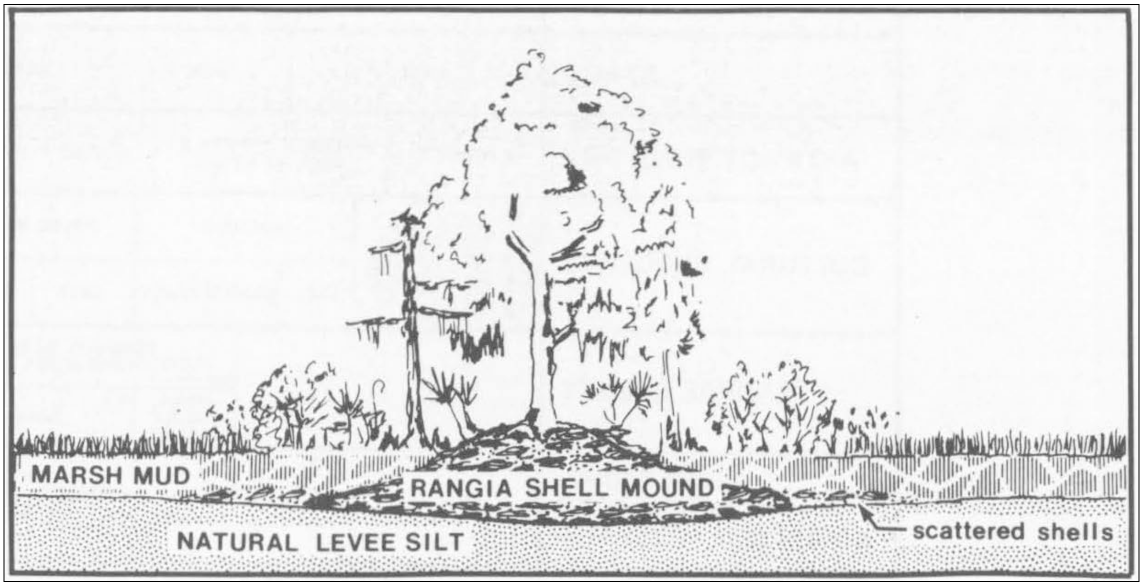

Figure 5.6. Shell mound on a subsided natural levee ridge enveloped by marsh mud. Reprinted from Gagliano et al. (1982).

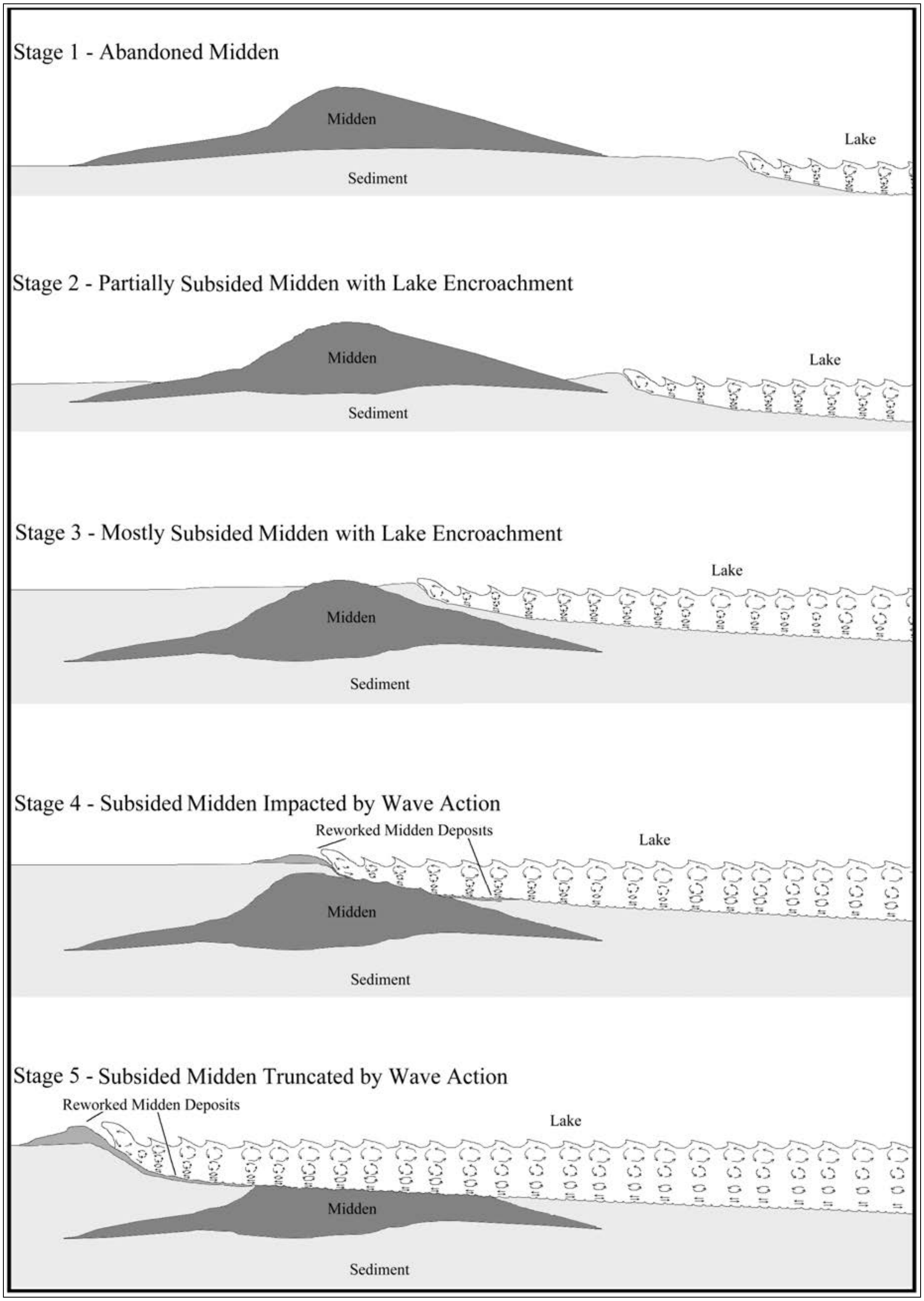

Figure 5.7. Schematic dip cross section, central coastal plain, Texas.
Reprinted from Saltus et al. (2003).

5.2.5 Forced regression

Forced regressions resulting from the fall of relative sea level would have been very conducive for the creation and preservation of geomorphic surfaces and associated cultural resources (Figure 4.4). However, they would have been created during the last glacial cycle at times before prehistoric Native American occupation of the northwestern Gulf of Mexico Coastal Plain. Although the relict landforms created by forced regressions might have later been ideal locations for Native American occupations, many of the relict landforms created by forced regression would have largely occupied the relatively flat, coast-parallel terraces between major river valleys. As a result, they would be destroyed along with associated cultural resources by shoreface erosion when relative sea level rose. Only the landforms associated with strips of coastal plain preserved from shoreface erosion by back-stepping of barrier islands and fluvial terraces within paleovalleys might have survived relatively intact and undisturbed.

6. Stratigraphy of Study Area

As discussed by Young et al. (2010 2012), considerable effort has been made over the last hundred years in understanding the stratigraphy and sedimentology of the sedimentary strata underlying the northwestern coastal plain of the Gulf of Mexico. The earliest geologic studies, e.g., Deussen (1914 and 1924), Barton (1930), Trowbridge (1932), Plummer (1932), Price (1934), Weeks (1933 and 1945), and Fisk (1940 and1948), focused on outcrop description and correlation. Later studies, e.g., Doering (1935 and 1956), Bernard (1950), Holland et al. (1952), Fisk and McFarlan (1955), Bernard and LeBlanc (1965), and Otvos and Howat (1997), incorporated the concept by Fisk (1940 and 1944) of a direct one-to-one correlation between sea level fluctuations, coastal-parallel terrace formation, and only four major continental glaciations, the Nebraskan, Kansan, Illinoian, and Wisconsinan stages of Kay and Apfel (1929) and Reed and Dreeszen (1965). The combination of these concepts resulted in the arbitrary and ultimately temporary subdivision of the Lissie Terrace into Bentley and Montgomery terraces within southwest Louisiana and southeast Texas (Heinrich 2009). Some studies, e.g., Akers and Holck (1957) and McFarlan and LeRoy (1998a and 1998b), attempted to incorporate subsurface data, primarily geophysical logs of oil and gas well and proprietary lithologic and paleontologic data, into their mapping and subsurface cross-sections that extended offshore. Although discredited by Boellstorff (1973 and 1978), these studies based their correlation on the incorrect presumption that there had been only four major continental glaciations during the Pleistocene. Most recently, only a set of three coast-parallel terraces has been consistently recognized and mapped along the northwestern Gulf of Mexico Coastal Plain (Winker 1979 and 1990, Snead and McCulloh 1984, DuBar et al. 1991, Heinrich 2009).

In Texas, this research was evaluated and is summarized by the Bureau of Economic Geology in its Geologic Atlas of Texas series (Aronow 1975, Shelby et al. 1968, Proctor et al. 1974, Aronow and Barnes 1975, Aronow et al. 1975, Brewton et al. 1976a, Brewton et al. 1976b). This mapping was revised in their Environmental Geologic Atlas of the Texas Coastal Zone series (Brown et al. 1977 and 1980, Fisher et al. 1972, McGowen et al. 1976a and 1976b.

In Louisiana, this research was evaluated and summarized in large-scale geologic maps. First, Snead and McCulloh (1984) remapped units and revised stratigraphic nomenclature for their 1:500,000 scale state geologic map. In their mapping, terraces were inferred to represent unconformity-bounded “complexes” composed of sedimentary strata. In addition, the Bentley and Montgomery terraces were collapsed into a single Intermediate complex in the absence of recognizable distinctions in terrace surfaces, terrace scarps, bounding unconformities, and other features. Later, Snead et al. (1999) and McCulloh et al. (2003) reclassified the informal complexes nomenclature into more formal allostratigraphic units. Their classification formed the basis for the nomenclature used in the compilation of 1:100,000 geological maps (Heinrich et al. 2002 and 2003, Heinrich 2005b and 2006).

The overall conclusion of these studies is that multiple unconformity-bounded, seaward dipping, clastic wedges, ranging in age from Pliocene to Late Pleistocene, underlie the coast-parallel terraces that compose the northwest Gulf of Mexico Coastal Plain (Figure 6.1). These clastic wedges consist of nonmarine sediments that grade gulfward into coastal and eventually shallow marine sediments. In updip areas, each of these clastic wedges erosionally truncates and onlaps an underlying clastic wedge. Thin erosional remnants, isolated terraces, onlapping veneers, and Holocene alluvial cover make it difficult to establish regional correlations between outcropping and subsurface stratigraphic intervals (Winker 1979, Winker 1990, DuBar et al. 1991).

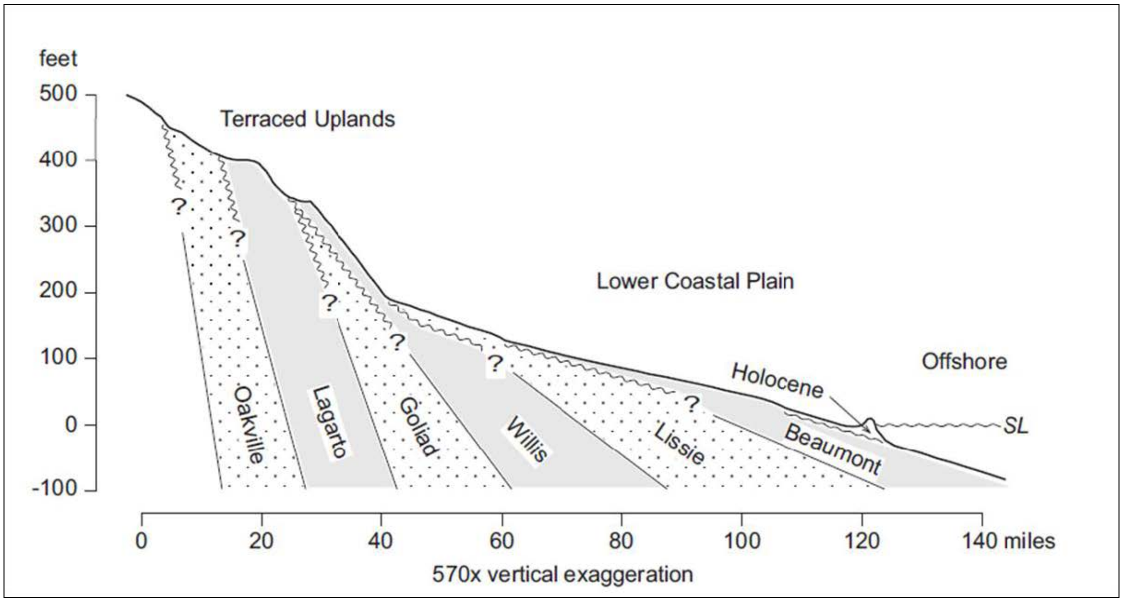

Figure 6.1. Regional dip cross section showing relationships between outcropping formations and subsurface stratigraphy, northwestern Gulf of Mexico Coastal Plain.

Reprinted from Young et al. (2012).

6.1 Allostratigraphic Approach to Pleistocene Unit Definitions

Traditional sedimentary units in geology are defined by their stratigraphic position and their lithology. This is effective in older hard-rock geology with extensive beds of consistent rock types. However, the largely unconsolidated, heterogeneous, and nondiagnostic nature of the siliciclastic fluvial, deltaic, coastal, and shallow marine sediments of the northwestern Gulf of Mexico Coastal Plain that were deposited within dynamic and shifting fluvial, deltaic, and coastal environments created widely varying packages of sediment juxtaposed among each other. All of these deposits consist of complex interfingering depositional facies that are composed of varying proportions of sand, silt, clay, peat, and minor amounts of gravel (McCulloh et al. 2003, Heinrich et al. 2015).

As a result, traditional lithologic formation categories are of less value here and Quaternary geologists must use other distinguishing criteria. In the subsurface, it becomes useful to redefine them as alloformations, where the package of sediment is defined by its bounding unconformities. An allogroup is several related alloformations. Also, sometimes it is useful to subdivide these alloformations into allomembers. Where exposed at the surface, the upper bounding surface of an alloformation typically forms a distinctive coast-parallel or fluvial terrace surface that can be recognized and mapped using topographic and pedologic data (Snead et al. 1999, Young et al. 2012, Heinrich et al. 2015).

By the late 1980s, the Louisiana Geological Survey, e.g., Autin (1989), started examining the application of allostratigraphic concepts and nomenclature to the definition and mapping of surface Plio–Pleistocene units. It was eventually concluded that allostratigraphy offers an effective, if not essential, approach to their delineation and classification. It was also noted that Plio–Pleistocene units of northern Gulf of Mexico Basin subsidence exhibit a clear spectrum of preservation of surface morphology varying from pristine relict and active landforms of the Holocene coastal plain to deeply dissected terrace surfaces of the oldest Pleistocene allostratigraphic units. In case of older Pliocene strata, the original constructional surfaces have been destroyed by erosion. As a result, allostratigraphic nomenclature has been heavily used in the STATEMAP-funded geologic mapping projects of the past two decades to map approximately three-fourths of the surface of Louisiana that is covered by Quaternary strata (McCulloh et al. 2003).

6.1.2 Beaumont Alloformation (Prairie Allogroup)

The Prairie Allogroup, formerly called the “Prairie complex” or “Prairie Formation,” is a collection of Pleistocene depositional sequences of alloformation rank (Autin et al. 1991, McCulloh et al. 2003, Heinrich et al. 2002, Heinrich et al. 2003, Heinrich 2005b, Heinrich 2006). The sediments of the Prairie Allogroup accumulated within a diverse variety of coastal plain settings, i.e., fluvial (meander-belt, backswamp, and braided-stream), colluvial, estuarine, deltaic, and shallow marine environments. It had been argued that these sediments largely accumulated over a considerable part of the late Pleistocene (Autin et al. 1991, Otvos and Howat 1997, Otvos 2005, McCulloh et al. 2003, Heinrich 2009). More recently, correlation of the unconformity at the base of the Beaumont Alloformation by Young et al. (2012) indicates that it accumulated during a large part of the middle Pleistocene and the earliest part of the late Pleistocene.

The surface of the Prairie Allogroup is a coast-parallel terrace that extends along the northwest coast of the Gulf of Mexico from a point about 110 km (67 mi) south of the Rio Grande in Mexico over to at least Mobile Bay, Alabama. This coast-parallel terrace is the lowest continuous terrace and lies directly above Holocene coastal and flood plains (Figures 3.3 and 6.2.1.1). Unlike older coast-parallel terraces, the surface of the Prairie Allogroup exhibits recognizable constructional topography. However, the relict landforms that it exhibits are distinctly more poorly preserved than those seen on the fluvial terraces of the Deweyville. It comprises multiple stratigraphic units of alloformation rank (McCulloh et al. 2003, Heinrich et al. 2002 and 2003, Heinrich 2005b, Heinrich 2006).

The Beaumont Alloformation is an unconformity bounded stratigraphic unit. It is separated from the underlying Lissie Alloformation by a regional unconformity that has been defined and mapped by Young et al. (2010 and 2012). Except where cut by valleys formed during sea level lowstands of the last glacial cycle, the upper boundary of the Beaumont Alloformation consists of the surface of the coastal plain.

It was originally named the “Beaumont Clay” by Hayes and Kennedy (1903) for the City of Beaumont, Jefferson County, Texas. There is a lack of any type section because the original description of the Beaumont Clay is a composite compiled from subsurface information from numerous wells.

Since it was named by Hayes and Kennedy (1903), the Beaumont Alloformation was intensively studied as either the Beaumont Clay or Beaumont Formation in the southeast Texas coastal plain by various geologists (e.g., Deussen 1914, Deussen1924, Barton 1930, Aronow 1971, Aronow 1986, Aronow and Barnes 1975, Fisher et al. 1972, Shelby et al. 1968, Van Siclen 1985). It was first mapped across most of the Texas Coastal Plain by Darton et al. (1937) and later in Southwestern Louisiana as the either the Prairie Formation or Prairie Terrace(s) (e.g., Fisk 1944, Bernard 1950). Doering (1956) subdivided it into the Eunice and Oberlin formations. More recently, Snead and McCulloh (1984), Mossa and Autin (1989), and Autin et al. (1991) recognized it as an informal allostratigraphic unit called the “Prairie complex.” In recognition of the persistence of the name “Beaumont” for more than 80 years and rules of priority, Snead et al. (1999) and McCulloh et al. (2003) redesignated the “Prairie complex” as the Beaumont Alloformation in southwestern Louisiana. This is the nomenclature that has been used in Louisiana Geological Survey geologic mapping. Most recently, the Texas Water Development Board, e.g., Young et al. (2010 and 2012), redefined it as an unconformity bounded stratigraphic unit.

As indicated by its original name, the Beaumont Alloformation is a predominately fine-grained stratigraphic unit that consists largely of varicolored, laminated to massive, calcareous silty clays that often contain calcareous nodules (Figure 6.2). The clays contain organic matter, partly decomposed tree remains (predominantly cypress), and shell beds (*Rangia cuneata* and *Crassostrea spp.*). Other animal fossils include mammoths, horses, and an indigenous foraminiferal fauna. These clays enclose brown, yellow, and blue sands and silty sand that occur as dip-oriented, narrow, ribbon-like channel belts and deltaic-distributary channels that are 6 to 15 m thick and are locally stacked as thick as 45 m. Sandy sediments also comprise a coast parallel sandy beach ridge complex known as the Ingleside barrier and/or strandplain system. The interbedded muddy units are generally of similar thickness to the sands (Hayes and Kennedy 1903, Price 1934, Knox et al. 2006, Young et al. 2012).

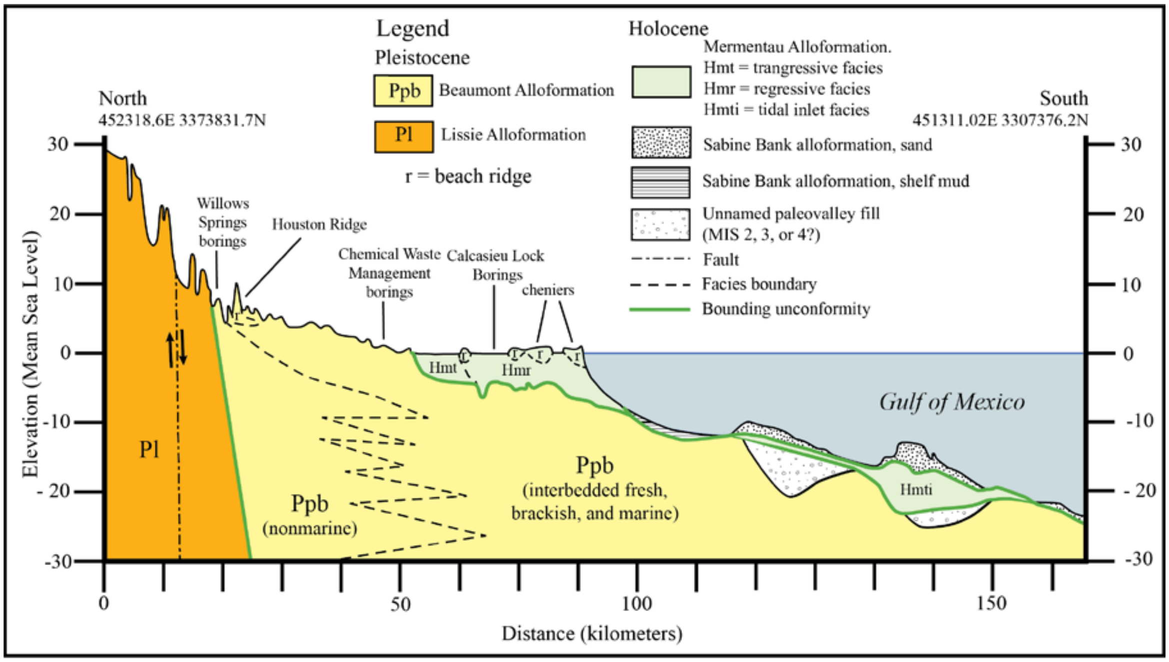

Figure 6.2. Regional dip cross section showing relationships between stratigraphic units within the study area and onshore Southwest Louisiana.

Except where cut by river valleys that formed during sea level lowstands of MIS 4, 3, and 2 and the Lafayette meander belt, the upper contact of the Beaumont Alloformation is known informally as the “Beaumont terrace” or “Beaumont surface” (Figure 3.2). This surface is an extensive coast-wise terrace that extends from the western valley wall of the Mississippi River Alluvial Valley past the Rio Grande to the Tamaulipas Range in northeastern Mexico (Bernard 1950, Doering 1956, Winker 1990). The surface of the Beaumont Alloformation exhibits moderately well preserved relict depositional topography of regional rivers and small streams (Fisk 1948, Fisk and McFarlan 1955, Aronow 1986, Winker 1990, Saucier and Snead 1989). It also includes extensively modified, coast-parallel, sandy beach ridges often known as the Ingleside Shoreline, Ingleside barrier system, or Ingleside strandplain system. Within Southwest Louisiana, the crest of this ridge ranges in elevation from 9 m to 11 m above mean sea level and rises 2 m to 4.5 m above the adjacent surface of the Beaumont terrace (Heinrich 2007). The Beaumont terrace is generally flat, little-dissected, and slopes gently. Within Southwest Louisiana, its slope is about 0.2 m to 0.3 m / km and 20 km to 28 km wide in the Lake Charles area. This surface and associated relict landforms have been extensively modified by pedogenesis and pedoturbation, resulting in its upper part being significantly weathered and overconsolidated (Fisk 1948, Fisk and McFarlan 1955, Johnson et al. 2008). The alteration of the surface of the Beaumont Alloformation has also resulted in the formation of a variable combination of innumerable pimple mounds and “blow-out” depressions along with extensive gilgai topography (Plummer 1932, Johnson et al. 2008). This surface extends beneath terminal Pleistocene to Holocene coastal-plain deposits and marine deposits and extends offshore beneath the Texas and Louisiana continental shelves. Winker (1990) and DuBar et al. (1989) correlate it with his R1 reflector. Within interfluves and uplands between Holocene and Wisconsinan paleovalleys and channel belts, it forms part of the Holocene-Pleistocene surface.

Until recently, there has been a lack of any agreed-upon definition of the base of the Beaumont Alloformation, as noted by Aronow (1971). Winker (1979) has projected its position to a seismic reflector offshore that he regards as correlative to the top of the Lissie Formation. Although this inferred base may prove to be valid, the documentation does not constitute a satisfactory definition. More recently, Young et al. (2010 and 2012) define the lower contact of the Beaumont Alloformation as a regionally extensive and laterally extensive flooding surface that is correlated with the Trimosina A micropaleontological zone offshore at about 0.6 Ma (Figure 6.3). Updip, this flooding surface is correlated northward along the bases of major channel sands to where it crops out and the Beaumont Alloformation onlaps on the surface of the Lissie Alloformation (Young et al. 2012).

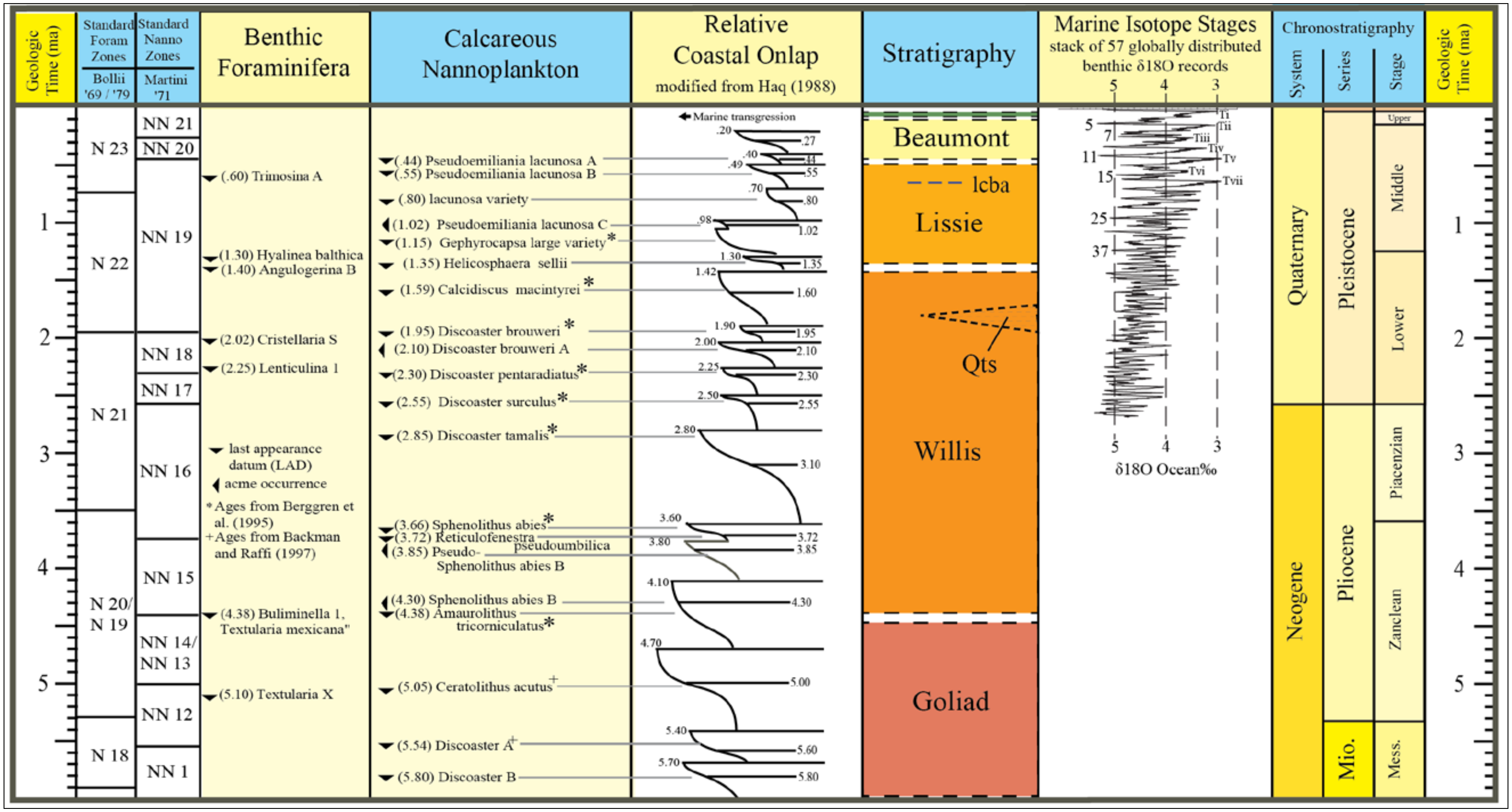

Figure 6.3. Correlation chart of regional Late Neogene strata showing the relationship between biostratigraphy, coastal onlap curve, and marine oxygen isotope stages.

Icha = Lava Creek Ash and Qts = Terrebonne Shale. Because of scale, Deweyville Allogroup (green) and Holocene units are shown in Figure 6.4.

Contrary to the assumptions of previous studies, the Beaumont Alloformation consists of sediments that accumulated during multiple high-frequency glacio-eustatic sea-level fluctuations (Figure 6. 4). Onshore, the bulk of its deposition occurred during interglacial highstands of sea level in fluvial, deltaic, and marginal-marine systems (Blum et al. 2013). Within Texas, the highstand positions of the Beaumont shoreline almost coincide with those of the modern shoreline in Texas and as far as 60 km inland in Southwest Louisiana (Solis 1981, Knox et al. 2006, Heinrich 2007, Young et al. 2012).

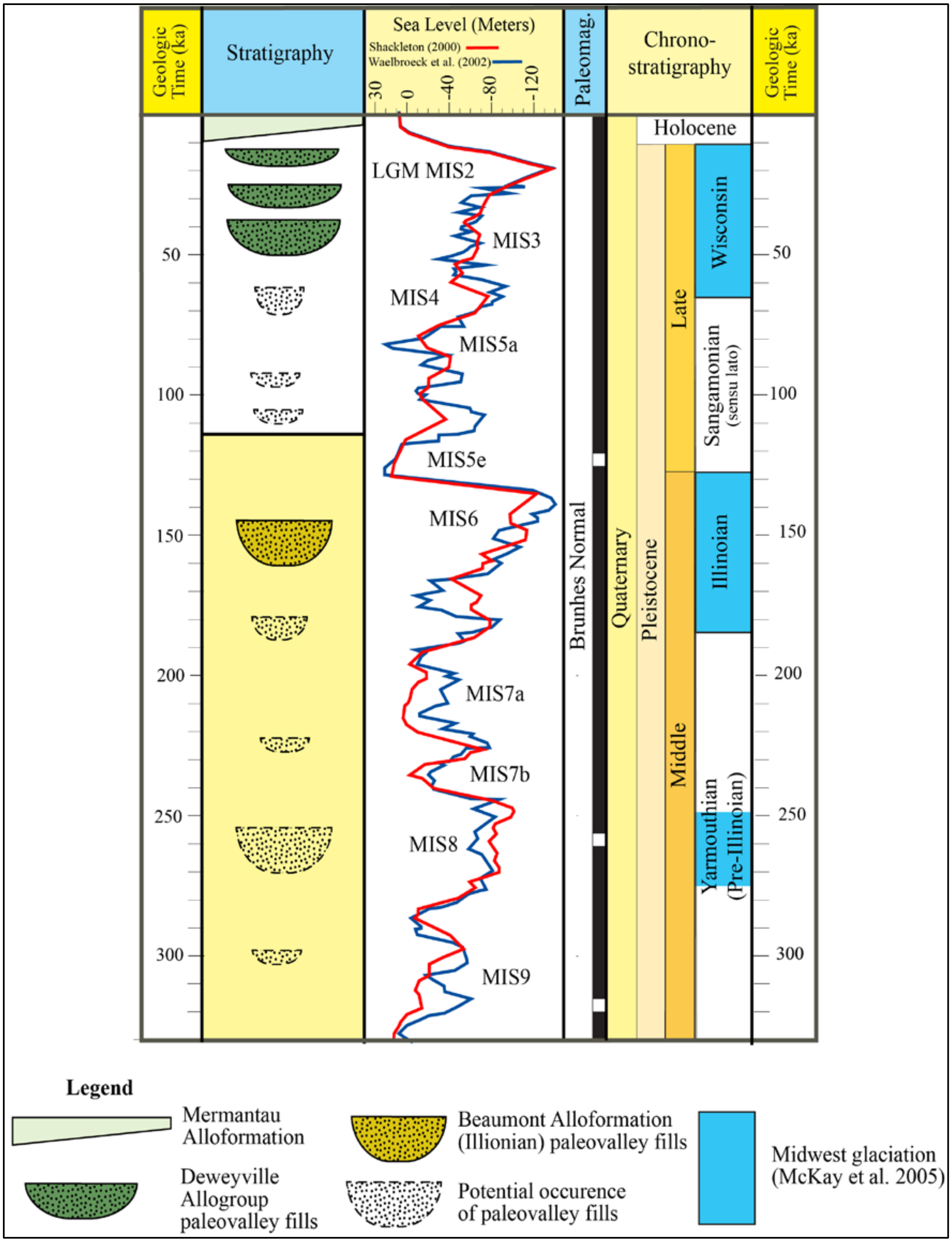

Figure 6.4. Late Pleistocene stratigraphy of Southwest Louisiana and offshore study region showing the relationship between major coastal plain stratigraphic units, global sea level, paleomagnetic chronology, and chronostratigraphy.

The Beaumont Alloformation has been long considered to be late Pleistocene in age (DuBar et al. 1991). Pleistocene-age fossils have been found in the Beaumont Alloformation at numerous locations on the

Texas Coastal Plain (Maury 1920, Maury1922, Plummer 1932, Price 1934). Initially, geologists H.N. Fisk and his associates (e.g., Fisk 1938, Fisk 1944, Fisk and McFarlan 1955, Bernard and LeBlanc 1965) assigned a mid-Wisconsinan age to the Beaumont Alloformation that corresponds approximately to MIS 3. Later, Mange and Otvos (2005) and Otvos (2005) obtained optically-stimulated luminescence ages from the Beaumont Alloformation that ranged between Eowisconsin (MIS 5d–5a) and Wisconsinan (MISs 4 and 3) ages. However, optically-stimulated luminescence ages from the Beaumont Alloformation and correlative Hammond Alloformation indicate that the Beaumont Alloformation is older than Wisconsinan (MIS 4 and 3) and its surface sediments accumulated during marine isotope stages 5e and 5a. Optically-stimulated luminescence ages from younger Deweyville Allogroup and Mississippi River valley train deposits substantiated this interpretation (Shen et al. 2012 and 2013).

Furthermore, Simms et al. (2013) obtained optically-stimulated luminescence ages of 119 to 127 ka from samples from depths between 1.5 m and 4.0 m beneath the surface of the Ingleside strandplain. Samples between depths of 1.0 m and 1.5 m beneath the surface of the Ingleside strandplain yielded ages of 1.32 to 57 ka. These ages were interpreted as representing later eolian reworking of the surface of the Ingleside strandplain.

Within Southwest Louisiana, the position of relict channel belts gulfward of the Ingleside Shoreline and the presence of incised fluvial ridges and beheaded deltaic distributary ridges as seen in LiDAR DEMs, topographic maps, and aerial photographs indicates that the final accumulation of sediments comprising the Beaumont Alloformation occurred during the falling stages of marine isotope stage 5. The Beaumont Alloformation clearly predates the terraces of the Deweyville Allogroup that lie within the valleys of the Sabine, Trinity, Calcasieu, and other rivers and the adjacent Lafayette meander belt of the Avoyelles alloformation. Downdip correlations by Young et al. (2012) argue that the Beaumont Alloformation began accumulating 600,000 years ago.

Within the Beaumont Alloformation, its significant cultural and sand resources will be restricted to its upper contact.

Braid belts and paleovalleys associated with ancient paleochannels of either the Red, Calcasieu, Mississippi, and other rivers contain large quantities of sand. However, only those fluvial sand bodies at the upper contact of the Beaumont Alloformation and underlying a thin blanket of marine sediments would be shallow enough to be economic to use.

Similarly, because of the age of the Beaumont Alloformation, its accumulation predates the accepted range of time during which this part of North America was first populated (Stright 1990, Anderson et al. 2013). As a result, cultural materials can be expected to be restricted to the the surface of this stratigraphic unit and within paleosols, and have been buried by younger deposits, e.g., cover sands. Preferred locations for the occurrence of archaeological sites on the surface of the Beaumont Alloformation include the tops of pimple mounds and within the sandy soils associated with them (Aten and Bollich 1981; Jones and Shuman 1988), the relict natural levees of Pleistocene paleochannels, and terrace edges overlooking valleys cut into the Beaumont surface by modern streams and bayous (Weinstein et al. 1979b, Coastal Environments, Inc. 1977).

6.1.3 Avoyelles Alloformation

The Avoyelles alloformation is an informal unconformity-bounded stratigraphic unit. At its base, it is separated from the underlying Beaumont Alloformation by an extensive unconformity that represents the basal scour surface of a MIS 3 Mississippi River meander belt. The upper contact of the Avoyelles alloformation is a surface that includes the channel belt of the Lafayette Meander Belt and associated flood basin (Figure 3.2). This surface is covered by a westward thinning blanket of Late Wisconsinan (MIS 2) Peoria Loess that ranges from about 7 m in thickness at the western valley wall of the Mississippi

Alluvial Valley to about 1 m in thickness within a distance of about 8 km (Miller et al. 1985, Saxton 1986, Callihan 1988, Rouly 1989, Mateo 2005).

The Avoyelles alloformation was informally named by by Snead et al. (1999) and McCulloh et al. (2003) for Prairie des Avoyelles and Avoyelles Hills within Avoyelles Parish, Louisiana. These are the northernmost occurrence of the alloformation in Louisiana and where Fisk (1940) described and recognized the significance of the surface morphology of the Avoyelles alloformation.

Currently, there is no formal type section for the Avoyelles alloformation. However, two cores that partially penetrated it within the Avoyelles Prairie are described in detail by Autin and Aslan (2001). Also, Mateo (2005) illustrates four cores, nos. 0394-006, 0494-022, 0494-023, and 0494-024, from a point bar of the Lafayette Meander Belt near Carencro, Louisiana that penetrated the Avoyelles alloformation. Lithologic logs of geotechnical borings that penetrate the Avoyelles can be found in Appendix 4 of Rouly (1989) and in the Type III Solid Waste Permit Application for the ANGCO facility near Carenco, Louisiana (KourCo 2011).

The most detailed information about the lithology of Avoyelles alloformation comes from the borings of Mateo (2005) taken from a point bar of the Lafayette Meander Belt north of Opelousas and geotechnical logs of the ANGCO facility near Carenco, Louisiana. They show that beneath the Peoria loess, the channel belt deposits of the Avoyelles alloformation consist of sand that fines upward and contains occasional ripple and parallel laminations and interbeds of silty loam. The uppermost unit of the Avoyelles alloformation consists of a 1.5 m to 6 m-thick, massive, gray to brown silty clay loam, which occasionally includes a weakly developed paleosol. It is well indurated and contains iron and manganese mottling. This unit might correlate with the “Yellow Clay” unit of Autin and Aslan (2001). So far, no fossils have been reported from the Avoyelles alloformation.

The surface of the Avoyelles alloformation is a loess-mantled fluvial terrace that forms the southernmost segment of the western valley wall of the Mississippi Alluvial Valley (Figure 3.2). This terrace occupies a triangular 0 to 22 mi (0 to 35 km) wide area, which extends eastward from the western valley wall of the Mississippi Alluvial Valley from New Iberia, Louisiana to Abbeville, Louisiana, and just west of Intracoastal City, Louisiana. This fluvial terrace consists of a Pleistocene Mississippi River channel belt that truncates older, loess-covered fluvial-deltaic plains of the Red River, which are underlain by the Beaumont Alloformation (Saucier 1994, Mateo 2005, Shen et al. 2012). Its slope is difficult to determine because of faulting and the variable thickness of its loess cover. However, the analysis of available LiDAR DEMs where the loess is thinnest suggests that it slopes gulfward at about 0.07 m per kilometer. Just before it disappears beneath the Holocene sediments of the Mermentau Alloformation, this channel belt bifurcates into distinct channel belts. Both channel belts exhibit moderately to poorly preserved relict constructional landforms, relict river courses, meander loops, and ridge and swale topography that are comparable in size to those of the modern Mississippi River (Fisk 1948, Jones et al. 1954, Saucier 1994, Shen et al. 2012). South and southeast of Jefferson Island and within Iberia Parish, a featureless and flat clay plain, a relict flood basin, forms the surface of the Avoyelles alloformation between the Lafayette Meander Belt and the adjacent Holocene delta and Chenier Plains (Autin 1984). In places, the surface of this relict meander belt and overlying loess blanket have been displaced by faulting to form east-west trending fault-line scarps (Heinrich 2006, Heinrich et al. 2003).

The base of this allostratigraphic unit is likely a composite and diachronous erosion surface created by the slight incision and lateral migration of an ancient Mississippi River. The sediments underlying it are Red River deposits of the Beaumont Alloformation.

The Avoyelles alloformation consists largely of thick point bar, crevasse, overbank, and other fluvial sediments. The channel deposits are overlain by 2 m to4 m of Mississippi River overbank deposits. In turn, the overbank sediments are veneered by 2 m to5 m of Peoria Loess (Saucier 1994, Mateo 2005, Shen et al. 2012). The sediments that comprise the Avoyelles alloformation accumulated within the Late Pleistocene Lafayette meander belt and associated flood basin (Heinrich 2005b). The ancient delta associated with this meander belt now lies underwater offshore of Cameron Parish where it forms part of the southern edge of the Louisiana Continental Shelf. Likely, one of the meander belts terminates with the West Louisiana Delta of Wellner et al. (2004).

The age of the Avoyelles alloformation is constrained by stratigraphy, optically stimulated luminescence ages, and thermoluminescence ages. it is incised into and younger than Red River deposits of the Beaumont Alloformation that are best interpreted to date to MIS 5a (Shen et al. 2012, Shen et al. 2013). They are overlain by MIS 2 loess of the Peoria Loess (Bettis et al. 2003, Mateo 2005). Excluding a 25 ± 3 ka age that is anomalous relative to the age of the overlying Peoria Loess, using optically stimulated luminescence, Mateo (2005) dated the Avoyelles alloformation at about 55 ± 6 ka. This age is consistent with the cross-cutting relationship between the Beaumont and Avoyelles alloformations and ages obtained from the Marksville segment of the Avoyelles alloformation. Optically-stimulated luminescence ages of 85 ± 10 and 86 ± 14 ka and a thermoluminescence age of 54 ± 6 ka were reported by Otvos (2005) from borrow pits in the northern and southwestern sector of the Lafayette Meander Belt. However, it is hard to interpret them in light of the stacking of the deltaic, meandering river and loess deposits in this area because the stratigraphic context for the dated samples is lacking.

Offshore, the Avoyelles alloformation would be a significant source of high quality sand for coastal restoration projects. As indicated by the core descriptions of Mateo (2005) and KourCo (2011) the channel belt sand deposits of it consist of a thick layer of well-sorted (poorly-graded) sand that because of shoreface erosion should lie relatively close to the surface of the gulf bottom.

As in case of the Beaumont Alloformation, the deposition of Avoyelles alloformation predates the accepted range of time during which this part of North America was first populated (Stright 1990, Anderson et al. 2016). As a result, cultural materials can be expected to be restricted to the surface of this stratigraphic unit and within paleosols, and to have been buried by younger deposits, e.g., possibly loess. Preferred locations for the occurrence of archaeological sites would be the the relict natural levees of Pleistocene paleochannels, and terrace edges overlooking valleys cut into the Lafayette Meander Belt by modern streams and bayous (Weinstein et al. 1979a, Weinstein 1981, Coastal Environments, Inc. 1977).

6.2 Trinity-Sabine Unconformity

Both onshore and in the study area, the Deweyville Allogroup is separated from the older deposits of the Prairie Allogroup that form the base and sides of paleovalleys cut into these older deposits (Figure 6.5). The base of the valley is a diachronous, composite surface formed by at least three cut-and-fill events and subsequent lateral migration of the associated fluvial system (Sylvia and Galloway 2006). Each of these events deposited a recognizable allostratigraphic package of fluvial sediments bounded on the sides and base by erosional surfaces and on their top by a terrace surface (Blum and Valastro 1994, Blum et al. 1995). Within the study area, the basal unconformity of these paleovalleys is known informally as the “Trinity-Sabine unconformity.”

Contrary to the reconstructions of the prehistory of the Mississippi River Alluvial Valley, this surface was never subaerially exposed. Thus, it was never available for habitation regardless of the age of formation of a specific surface.

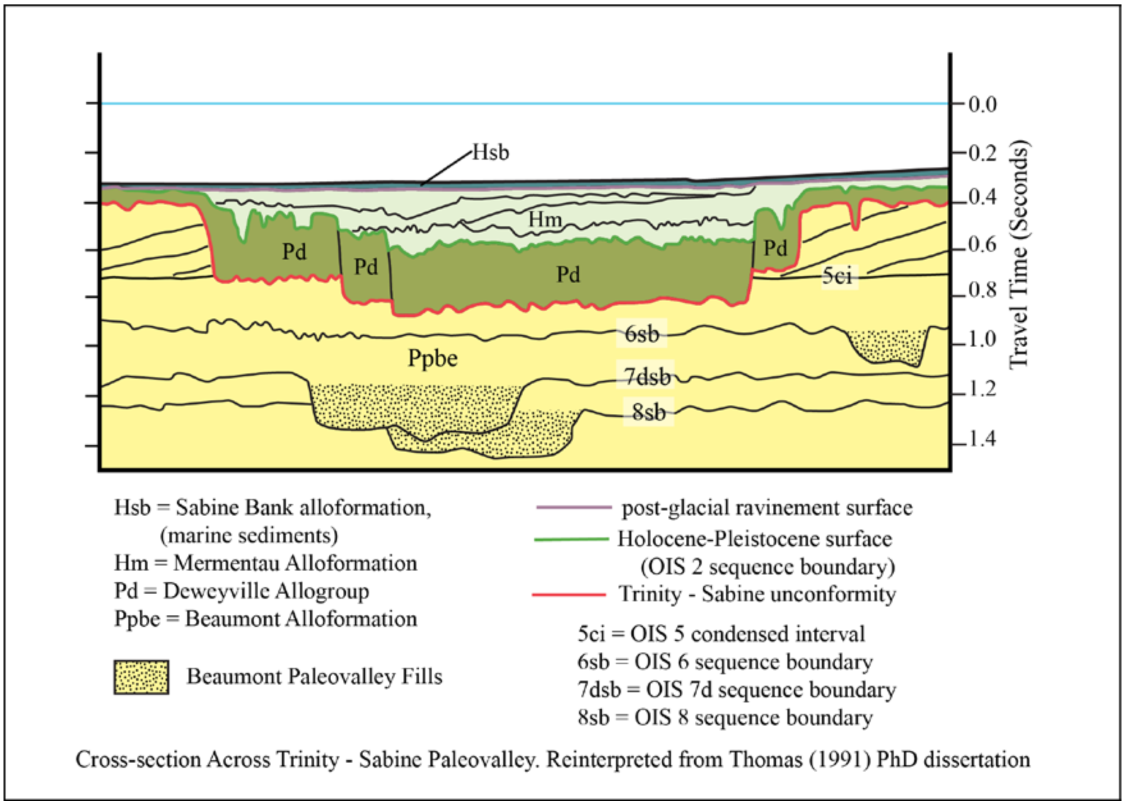

Figure 6.5. Cross-section across the Trinity-Sabine Paleovalley.

Modified and reinterpreted from Thomas (1991) showing major bounding unconformities and allostratigraphic units.

6.3 Deweyville Allogroup

The Deweyville Allogroup is an unconformity-bounded sedimentary package that underlies sub-regionally extensive fluvial terraces that lie topographically below the position of the surface of the Beaumont Allogroup, the Prairie terrace or Prairie surface, and geomorphically above the modern flood plain along streams and rivers of the northern Gulf coastal plain (Bernard 1950) (Figure 6.6).

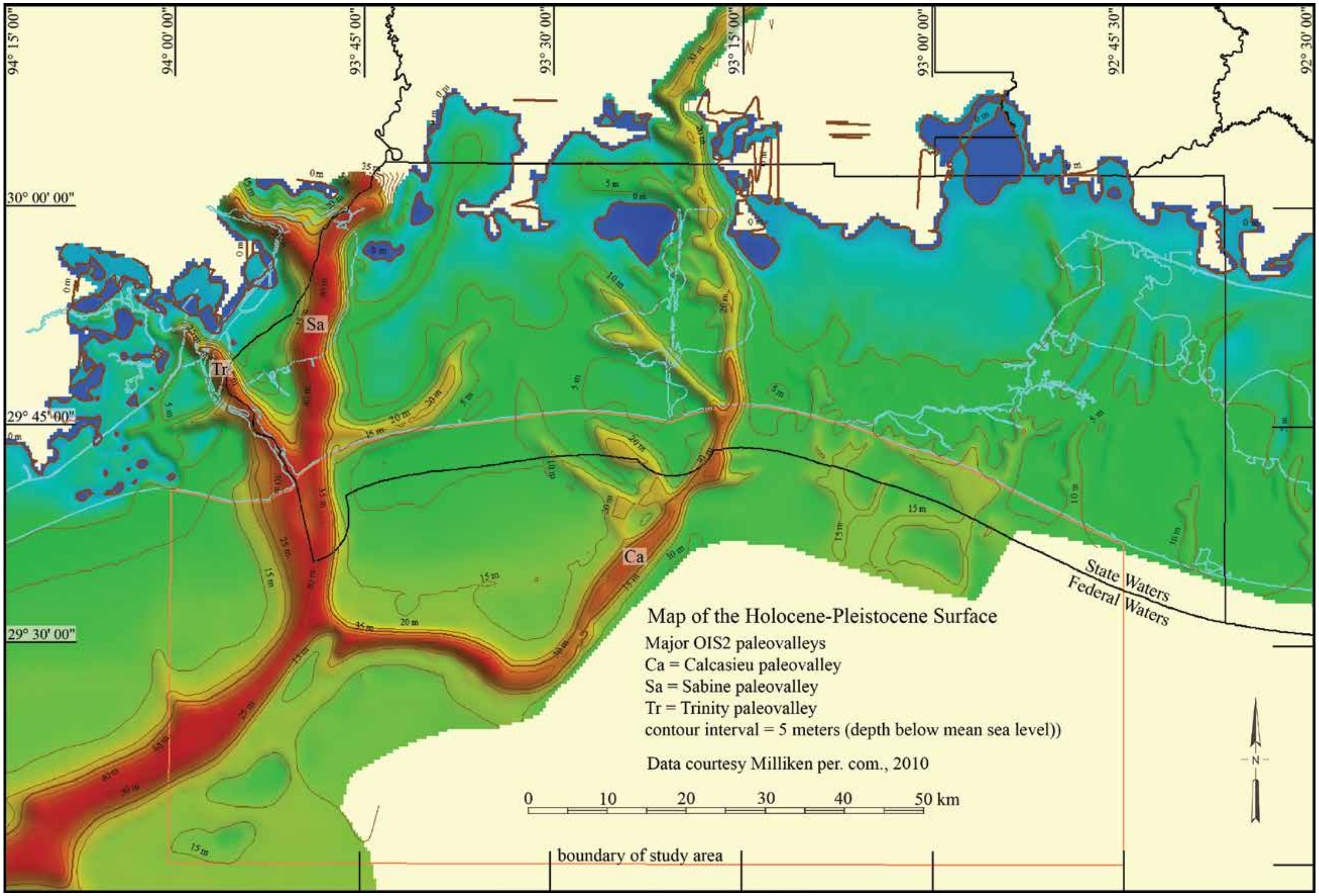

Figure 6.6. Holocene–Pleistocene Surface, 5-m Contour Interval.

The deep trenches are the paleovalleys of the Trinity (Tr), Sabine (Sa), and Calcasieu (Ca) rivers. Figure created using Global mapper from data supplied by Milliken (2010).

Bernard (1950) focused on the uppermost of three fluvial terraces in the area of Deweyville, Texas and did not designate a type section. Gagliano and Thom (1967) later identified fluvial terraces along much of the Gulf and Atlantic coastal plains as the Deweyville terrace. However, terraces and associated fluvial sediments compatible with the original definition of Bernard, (1950) are best restricted to streams of intermediate size in Texas, Louisiana, Arkansas, and Mississippi, e.g., Fisher et al. (1972), Kesel (1980), Snead and McCulloh (1984), Winker (1990), Blum et al. (1995), Garvin (2008), and Blum et al. (2013). Pearson et al. (1986) recognized and mapped the Deweyville terrace for a distance of up to 48 km south of the current shoreline along the offshore Sabine River paleovalley beneath the submerged Louisiana-Texas continental shelf. Thomas (1991) also recognized the presence of three buried terraces and associated fluvial sediments within offshore Sabine and Trinity River paleovalleys.

The base of this allogroup is a composite and diachronous erosion surface created by the periodic incision, lateral migration, and channel belt formation of streams of intermediate size in Texas, Louisiana, Arkansas, and Mississippi (Figures 3.3 and 6.5). The scoured base of each of the, typically three, Deweyville channel belts progressively truncate the scoured base of an older channel belt to form the base of a paleovalley that consists of a composite of erosional surfaces that are younger with depth. The exposed surface of the Deweyville typically exhibits well-preserved oversized fluvial features (Sylvia and Galloway 2006, Garvin 2008, and Blum et al. 2013).

The top of the Deweyville Allogroup is a flight of typically three fluvial terraces that lie topographically below the position of the surface of the Beaumont Allogroup and geomorphically above the modern flood plain along streams and rivers of the northern Gulf coastal plain (Figures 3.3). Gulfward along these streams, these terrace surfaces are progressively buried by Holocene flood plain deposits that, in turn, are buried within paleovalleys by estuarine sediments adjacent to the coastline and shallow marine sediments offshore (Gould and McFarlan 1959, Byrne et al. 1959, Milliken et al. 2008b, Milliken et al. 2008c). In southeast and central Texas, the terraces of the Deweyville Allogroup are sometimes blanketed with a few-meter thick layer of windblown sand, known as the “Holocene sandy mantle,” that has accumulated over the Early to Late Holocene. Although thin, these windblown sand deposits have been found to contain in situ and stratified archaeological deposits. Where covered by these younger deposits, either overbank or eolian, paleosols are commonly found developed within the Deweyville Allogroup (Abbott 2001, Aiuvalasit 2006). In general, the sediments of the Deweyville Allogroup are separated from younger undifferentiated Holocene fluvial and estuarine sediments by a surface created by the accumulation of sediment during postglacial sea level rise and local erosion surfaces created by the postglacial lateral migration of streams and rivers.

The Deweyville Allogroup can be further subdivided into unnamed bodies of fluvial sediments of alloformation rank (Blum and Vastaldo 1994). The tops of each of these allostratigraphic units are characterized by a single terrace, associated channel belts, sand bodies, and lateral scour surfaces. It is interpreted to represent a single episode of fluvial incision, stability, and lateral migration (Blum and Vastaldo 1994, Blum et al. 2013).

The Deweyville Allogroup consists of largely coarse-grained fluvial sediments that underlie the terraces that lie intermediate in position between the Prairie surface and modern flood plain along the Trinity River. Detailed descriptions of these sediments for the Trinity River in Southeast Texas can be found in Garvin (2008) and Abbott (2001); for the Sabine River within the Deweyville Texas area by Bernard (1950); and the Brazos River by Sylvia and Galloway (2006). The Deweyville channel belt sands typically are coarser than Beaumont or Holocene channel belt sands. Individual subdivisions of the Deweyville Allogroup consist of multiple cross-cutting and therefore laterally amalgamated sandy and gravelly channel belts that contain isolated muddy paleochannel fills. Deweyville channel belt deposits fine upwards as is typical of point–bar deposits described from the literature, e.g., Bridge 2003. Typically, they are considerably coarser than corresponding point bar sediments than younger and older deposits. Finally, vertical accretion facies are limited to non-existent vertical accretion facies within Deweyville deposits (Bernard 1950, Blum et al. 1995, Abbott 2001, Blum and Aslan 2006, Garvin 2008, Blum et al. 2013).

Little is known about systematic variations in thickness of the Deweyville Allogroup. The most detailed study of variations in its thickness is Garvin (2008). He found that, along the Trinity River in southeast Texas, the three recognized subdivisions of the Deweyville within its valley all increase in thickness gulfward. His "High Deweyville," the oldest and stratigraphically lowest of these subdivisions of alloformation rank, increases in thickness from about 5.5 m updip to ~11 m downdip. His "Middle Deweyville" increases in thickness about 6 m updip to about 9–11 m downdip. His "Low Deweyville," the youngest and stratigraphically highest of these subdivisions, increases from about 7 m updip to about 11 m to12 m downdip Garvin (2008).

The Deweyville allostratigraphic units have proven difficult to date. They are currently interpreted to have been deposited during MIS 4, 3, and 2 (Blum et al. 1995, Morton et al. 1996, Blum and Törnqvist 2000, Blum and Aslan 2006, Blum et al. 2013). For example, of 11 samples subject to optically stimulated luminescence dating by Garvin (2008), eight yielded age estimates that agree with cross–cutting relationships, whereas three yielded dates that violate cross–cutting relationships. One sample taken from his High Deweyville unit yielded an age representing the timing of channel belt activity from 35 to 31 ka. Four samples from his Middle Deweyville unit yielded age ranges from 34 to 23 ka. Three samples from his Low Deweyville unit yielded age ranges from 23.2 to 18.8 ka. The coarse-grained valley fills of the Deweyville Allogroup represent the abandonment and entrenchment of valleys within the Beaumont alluvial plains by river systems ca. 100 ka, and multiple episodes of lateral migration, aggradation, and/or degradation within those valleys during the MIS 4, 3, and 2. These fluvial systems were graded to shorelines at midshelf or farther south (Blum et al. 1995 and 2013).

The Deweyville Allogroup is a significant stratigraphic unit in terms of both sand and cultural resources. The thick, laterally continuous, and sandy nature of the Deweyville channel belts make it a significant potential source of sand for beach restoration where they occur near the gulf bottom. As previously discussed, the edges of Deweyville terraces and oxbow lakes and swamps have been documented to be preferential locations of settlement and other activities in prehistoric times. Some of these cultural deposits would have been buried by eolian sediments to form stratified sites. Others would have been buried initially by overbank sedimentation and later by estuarine sediments as sea level rose during postglacial sea level rise within the valley in which they lie. Such sites depending on other factors could have been sheltered within a paleovalley below the level of erosion by shoreface erosion and resulting ravinement formation.

The youngest of the Deweyville alloformations would have been young enough to have been contemporaneous, in part, with the human occupation of North America. As a result, the natural levees of its paleochannels and oxbow lakes are young enough to have been occupied while a particular paleocourse was active; this would indicate the potential for the presence of cultural resources being found buried within the sediments of this part of the Deweyville Allogroup.

6.4 The Holocene-Pleistocene Surface

Separating Prairie and Deweyville allogroups from younger terminal Pleistocene and younger Holocene deposits is an unconformity that has been informally referred to either as “base of the Holocene,” “top of the Pleistocene,” or “Holocene-Pleistocene surface.” (Figures 6.2.1.5, 6.2.1.6). The “Holocene-Pleistocene surface” is used in this study because in the gulfward extent, it is as old as Late Wisconsin (MIS 2) and is in places described as the “Late Wisconsin unconformity.” This surface is a regional unconformity that extends along the Gulf of Mexico coast from at least the Central Texas continental shelf, where it is known as the “Holocene-Pleistocene exposure surface,” across the Louisiana Coastal Plain and continental shelf, and eastward of the Chandeleur Islands, where it is known as the “Late Wisconsin unconformity.” This surface represents the former Late Pleistocene coastal plain during the last lowstand of sea level that occurred during the Last Glacial Maximum of MIS 2 about 29 to 11 ka (Heinrich et al. 2015).

Between Late MIS 2 though MIS 1, 18 to 4 ka, this low relief coastal plain was submerged as deglaciation led to a rapid rise in sea level. As sea level rose, the valleys of the Calcasieu, Mermentau, Sabine, and Trinity rivers were flooded. The lowstand flood plain soils that had developed in the terrace surfaces of the Deweyville Allogroup initially were inundated by rising sea level, causing the aggradation of a prism of fluvial and deltaic sediments that buried the terrace surfaces and formed the Holocene-Pleistocene surface. Later the flooding of valleys further buried this surface under estuarine deposits. Within paleovalleys, these fluvial deltaic deposits were in some places either partially or completely eroded by the transgressing shoreline within an estuary or lagoon to create a bayline or lagoonal ravinement surface that merged with and became the Holocene-Pleistocene surface. Within the uplands lying between paleovalleys and shallow paleovalleys, the shoreface of the main Gulf shoreline eroded deep enough to completely remove pre-existing transgressive deposits down and into either the underlying Beaumont Alloformation, Avoyelles alloformation, or Deweyville Allogroup. In such a case, the marine ravinement surface merged with and now comprises the Holocene-Pleistocene surface (Kane 1959, Smyth et al. 1988, Thomas and Anderson 1989, Smyth 1991, Thomas 1991, Milliken et al. 2008b, Milliken et al. 2008c, Blum et al. 2013). The nature of the Holocene-Pleistocene surface is important because it constrains the potential for the preservation of offshore buried and submerged cultural resources.

Beneath the Louisiana Chenier Plain, the Holocene-Pleistocene surface separates the Holocene Mermentau Alloformation from the underlying Deweyville and Prairie allogroups of the Pleistocene (Milliken et al. 2008b and 2008c). Gould and McFarlan (1959) inferred that all of the Holocene-Pleistocene surface underlying the Mermentau Alloformation from its northern pinch out at the juncture of the Chenier Plain with the subaerially exposed Prairie surface south to the shoreface of the modern shoreline is a marine ravinement surface. In contrast, Penland and Suter (1989) argue that north of Little Cheniere, the Holocene-Pleistocene surface beneath the Mermentau Alloformation consists of a bayline ravinement-flooding surface that developed behind a shoreline parallel trend of sand shoals between Ship Shoal and Sabine Bank. As discussed previously, it is possible that intact archaeological sites might have survived formation of a bayline ravinement-flooding surface and lie buried beneath the Holocene Mermentau Alloformation.

6.5 Mermentau Alloformation

The Mermentau Alloformation is an unconformity-bounded sedimentary package that fills paleovalleys and valleys cut by the Mermentau, Calcasieu, Sabine, Trinity, and other rivers, and their tributaries during MIS 3 and 4 and underlies the Louisiana Chenier Plain. The lower boundary of the Mermentau Alloformation is the Holocene-Pleistocene surface, which separates it from either the Deweyville Allogroup or Beaumont Alloformation (Figures 3.3, 6.2, 6.5, 6.7). Landward of the main gulf shoreline the upper boundary is either a bay bottom or the surface of the Louisiana Chenier Plain. Offshore, the upper boundary is an unnamed ravinement surface, which in many places separates the Mermentau from the overlying Sabine Bank alloformation (unnamed shelf sand in Figure 6.2). The Mermentau was first named as the "Mermentau Formation" by Jones et al. (1954) for water wells located near the Mermentau River in Calcasieu River.

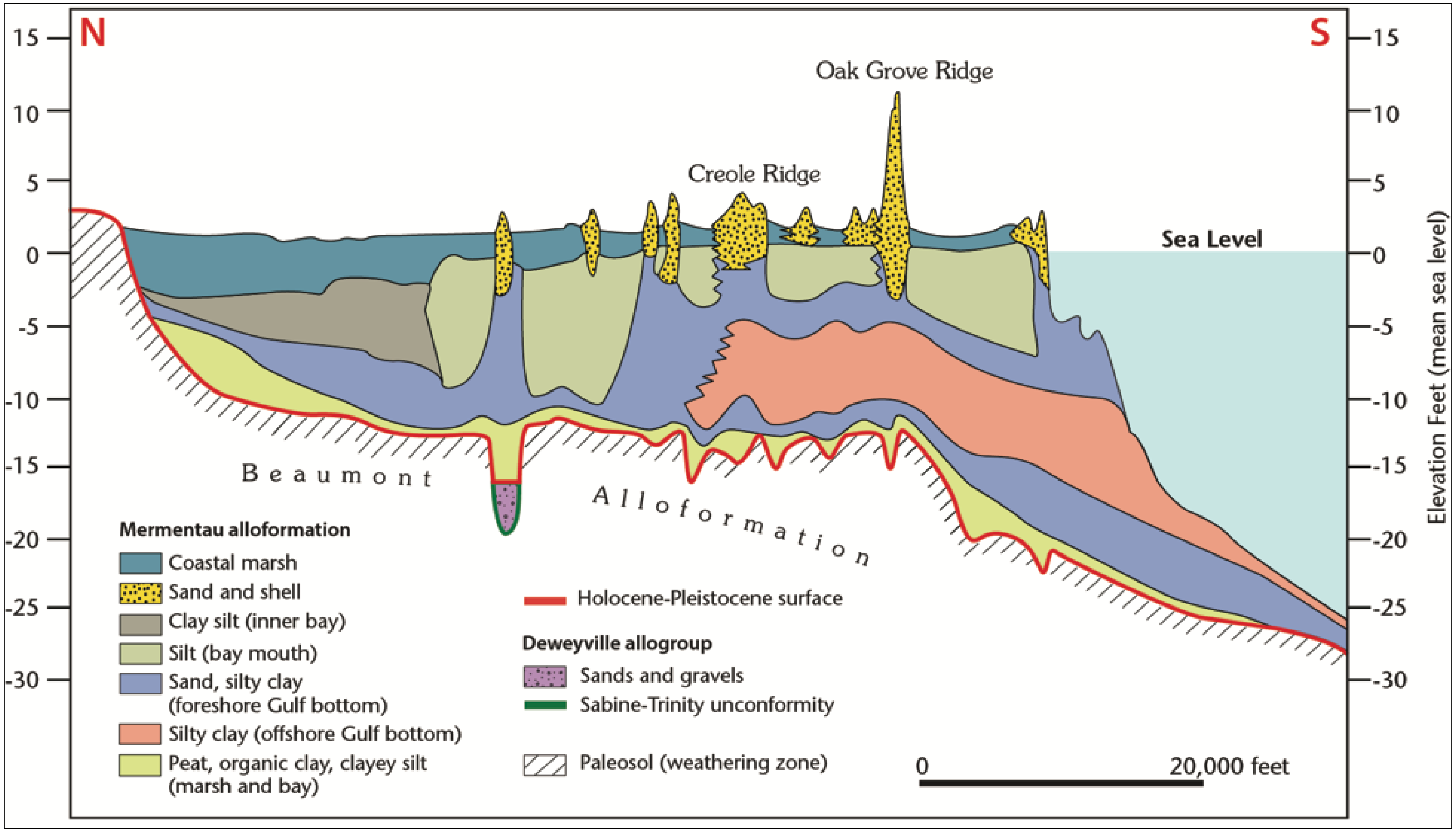

Figure 6.7 North-South cross-section across the Louisiana Chenier Plain showing sedimentary facies and bounding unconformities of the Mermentau Alloformation.

The Holocene Pleistocene surface is the bottom bounding unconformity of the Mermentau Alloformation. Adapted from Gould and McFarlan (1959). Reprinted from Heinrich et al. (2015).

The base of the Mermentau Alloformation is the Holocene-Pleistocene surface. As previously discussed, it is, in part, the surface of the MIS 2 lowstand coastal plain that has been submerged by postglacial sea level rise. It is a composite and diachronous flooding surface that has experienced varying degrees of subsequent modification by ravinement. It separates the Mermentau Alloformation from underlying sediments and terraces of the Deweyville Allogroup and unnamed Holocene fluvial deposits, which fill the bottom of paleovalleys and valleys of the Calcasieu, Sabine, other coastal rivers and streams. In cores, the Holocene-Pleistocene surface exhibits a sharp increase in shear strength–a change from overlying soft gray sediments to underlying hard, orange and tan mottled mud and sand. These changes represent the deposition of soft gray sediments of the Mermentau Alloformation on top of sediments that have been altered by subaerial exposure and pedogenesis (Fisk 1948, Byrne et al. 1959, Gould and McFarlan 1959, Milliken et al. 2008b, Milliken et al. 2008c, Heinrich 2015).

Offshore of the modern shoreline within the study area, the upper contact of the Mermentau Alloformation is a transgressive ravinement surface, known in this report as the "unnamed ravinement surface." The unnamed ravinement surface is an erosional unconformity created by the landward movement of the shoreface as the result of postglacial sea level rise. This surface completely truncates the package of sediments of the Mermentau Alloformation underlying the Louisiana Chenier Plain along the modern shoreline and partially truncates paleovalley fills offshore of the modern shoreline. Outside of paleovalleys, the unnamed ravinement surface has completely removed the Mermentau Alloformation and eroded down into the Beaumont Alloformation (Siringan 1993, Siringan and Anderson 1994, Rodriguez et al. 2001).

As described from Texas and Louisiana estuaries, the Mermentau Alloformation includes a heterogeneous mixture of fine, coarse, and organic sediments. Both the onshore and offshore paleovalley fills beneath the Louisiana Chenier Plain consist of a variety of 1) sandy bayhead delta deposits; 2) central-basin muds; 3) tidal inlet, tidal delta, and/or back-barrier estuarine sands; and 4) fine-grained shoreface muds. Within chenier plains, transgressive and regressive beach and shoreface sands are also present and form distinct beach ridges (Smyth et al. 1988; Thomas and Anderson 1989; Smyth 1991; Thomas 1991; Siringan 1993; Rodriguez et al. 2001, 2004; Milliken et al. 2008b, 2008c). In the valleys of Calcasieu and other coastal rivers, the Mermentau Alloformation also contains a blanket of organic-rich mud that accumulated within cypress swamps (Nichols et al. 1996).

Sirigan and Anderson (1994) and Rodriguez et al. 2001 and 2004) argue that the Sabine, Heald, and Shepard banks represent submerged paleoshorelines. They subdivided sandy sediments of these banks, from bottom to top, into 1) a back-barrier estuarine facies that is characterized by landward-dipping seismic reflectors and consists of interbedded sand and mud; 2) a fore-barrier, lower shoreface and/or ebb-tidal delta facies that is characterized by seaward prograding to chaotic seismic reflectors and consists of muddy sand; and 3) a storm reworked facies characterized by a chaotic to acoustically reverberating seismic reflection pattern that consists of interbedded shell hash and sand. According to their interpretations, the "lower back-barrier estuarine" and "fore-barrier, lower shoreface/ebb-tidal delta" facies would belong to the Mermentau Alloformation as lying between the unnamed ravinement surface and Holocene-Pleistocene surface (Figures 3.3 and 6.7). They interpreted these facies to be erosional remnants of lower shoreface and tidal inlet sands that represent the former locations of stillstand barrier island trends. More recent studies, e.g., Swartz (2016) and Swartz et al. (2016), dispute this interpretation and argue that the unnamed ravinement surface lies directly upon Pleistocene fluvial sediments and the sand shoals have been reworked from them.

The Mermentau Alloformation is a significant unit in terms of prehistoric cultural resources. First, given its age, its accumulation spans the human occupation of the Northeastern Gulf Coast. Its bayhead delta deposits have the potential to contain in situ and intact cultural deposits. Beach ridges that are associated with the Louisiana Chenier Plain are associated with numerous archaeological sites. However, whether such sites can survive shoreface transgression remains to be seen. Second, the sediments of the Mermentau Alloformation in places are thick enough to protect paleosurfaces and associated prehistoric cultural resources from shoreface erosion where paleovalleys are deep enough. Finally, the nature of its lower bounding unconformity, the Holocene-Pleistocene surface, determines whether significant paleosurfaces are preserved well and the nature of these paleosurfaces. For example, studies of the stratigraphy and sedimentology of the sediments of the Louisiana Chenier Plain clearly indicate that the Holocene-Pleistocene surface at their base is mantled by a layer of thin, but likely intact, marsh sediment. This layer suggests that the paleosurface of the underlying coastal plain has suffered little, if any, erosion during the transgression of it by marine or bay waters. If so, then associated prehistoric cultural resources associated with it should be relatively intact.

6.6 Unnamed Ravinement Surface

The unnamed ravinement surface is a transgressive ravinement surface and an erosional unconformity created by the landward movement of the shoreface as the result postglacial sea level rise. Offshore of the modern shoreline within the study area, the unnamed ravinement surface forms the upper boundary of either the Mermentau Alloformation, Deweyville Allogroup, or Beaumont Alloformation (Figures 6.2 and 6.5). Along the lower shoreface, it forms the pinchout of the Mermentau Alloformation that underlies the Louisiana Chenier Plain and truncates paleovalleys that are filled with it and Deweyville Allogroup. For large parts of the study area between major paleovalleys, the unnamed ravinement surface is deep enough to have completely removed the Mermentau Alloformation, down into underlying Pleistocene sediments, and has merged with the Holocene-Pleistocene surface.

6.7 Sabine Bank Alloformation

The Sabine Bank alloformation is an informal allostratigraphic unit that, within the study area, overlies the unnamed ravinement surface, which forms its basal unconformity (Figures 6.2 and 6.5). The surface of the Sabine Bank alloformation (Ha01m in Figure 6.5) is the Gulf bottom. It is informally named from Sabine Bank, where it forms a large offshore shoal and has been studied in detail using seismic surveys and cores. Within the interfluvial areas of the inner shelf, the Sabine Bank alloformation, consists of offshore marine sediments that are less than 1 m thick and composed dominantly of mud. These muds lack storm sand beds (Siringan and Anderson 1994, Rodriguez et al. 2001 and 2004).

As argued by Siringan and Anderson (1994) and Rodriguez et al. (2001, 2004), only the upper part of the Sabine, Heald, and Shepard banks lies above the unnamed ravinement surface. They interpret it as an upper seismic facies that is characterized by a chaotic to acoustically reverberating seismic reflection pattern and consists of interbedded shell hash and sand consisting of storm-reworked barrier island sediments. As noted above, more recent studies, e.g., Swartz (2016) and Swartz et al. (2016), dispute this interpretation and argue that the unnamed ravinement surface lies directly upon fluvial sediments and that the sand shoals consist entirely of offshore sands of uncertain provenance.

7. Paleogeography

7.1 Paleochannels and Their Fills

Various previous studies, including hazard surveys, in the southwest Louisiana continental shelf study area, have often conflated paleovalleys, paleochannels, and channel belts under a single category such as "paleochannels." At various times, the same features, occasionally even in the same paper, have been given different designations to denote the same landform. The landforms that were found most often conflated are fluvial valleys, channel belts, and paleochannels. Because each of these features comes with its own expectations as to occurrence and preservation of cultural resources, they need to be strictly defined and differentiated.

A channel can be either a subaerial or subaqueous landform. It is a geometric entity in the form of an open conduit through which a natural body of fluid and the sediment, which it transports, flows. On Earth, the fluid that occupies a channel from bank to bank is either water in a fluvial channel or an active turbidity flow in a submarine channel. The shape of a fluvial channel in cross section varies from a flattened to trapezoidal half-cylinder (Erkeling et al. 2014, NSSH 2008, Sharp and Malin 1975). A lengthy segment of a channel is a course, e.g., a river or stream course. A paleochannel is an abandoned, either historic or prehistoric, segment of river or stream that has been either partially or completely filled or buried by younger sediment (Neuendorf et al. 2005) (Figure 7.1 and 7.2). A continuous segment of significant length of an inactive historic or prehistoric segment of river or stream channel is a paleocourse.

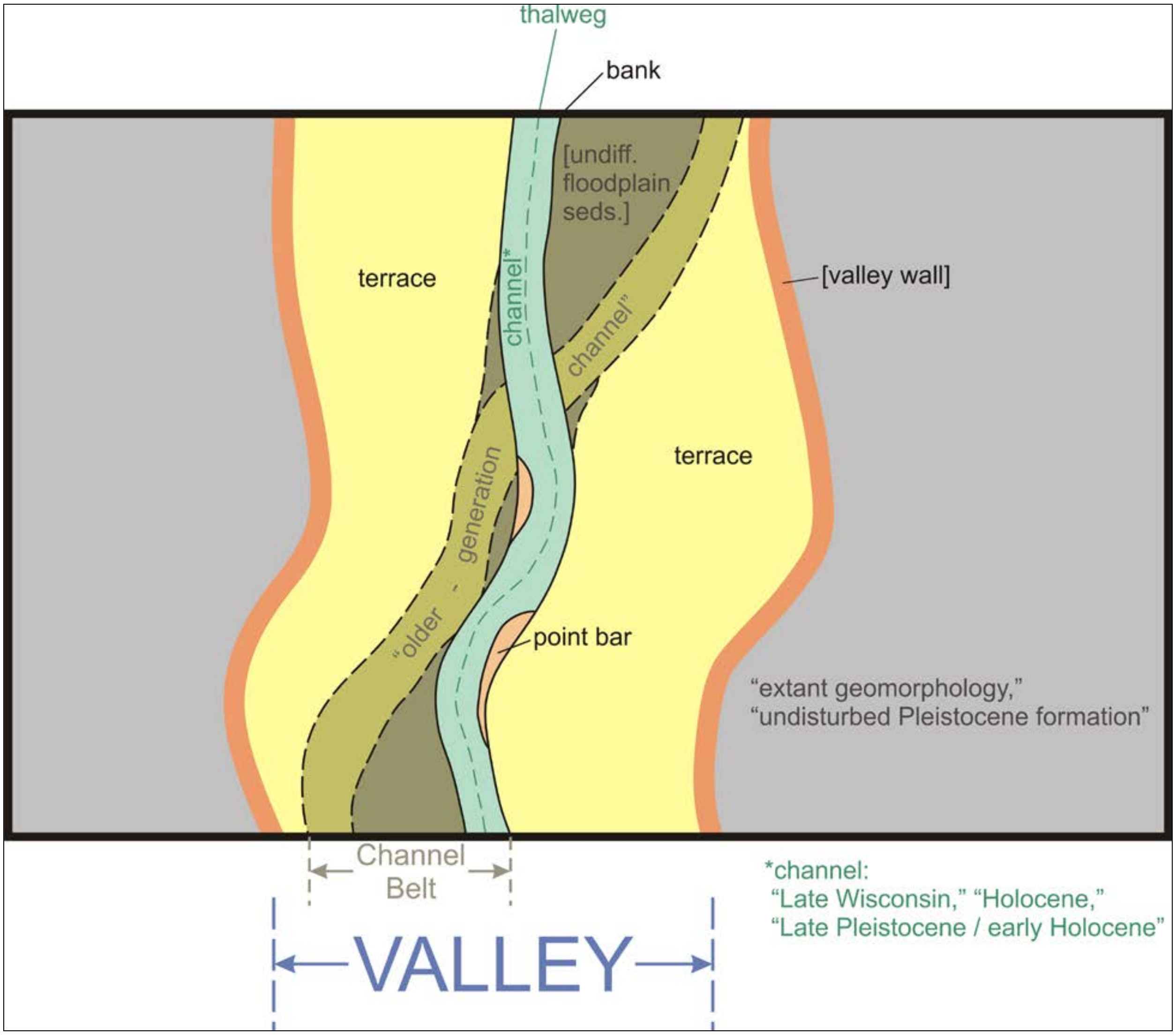

Figure 7.1. Nomenclature for Fluvial Valley Landforms.

Many of the terms used as labels were copied directly from block-hazard survey maps prepared over a period of 40 years. Some multiple listings for features reflect the usage of different investigators working in different areas and/or times. Other terms, though not copied directly from hazard surveys, clearly were implicit in them. Terms enclosed in brackets were neither used nor implied on the hazard-survey maps reviewed for this investigation.

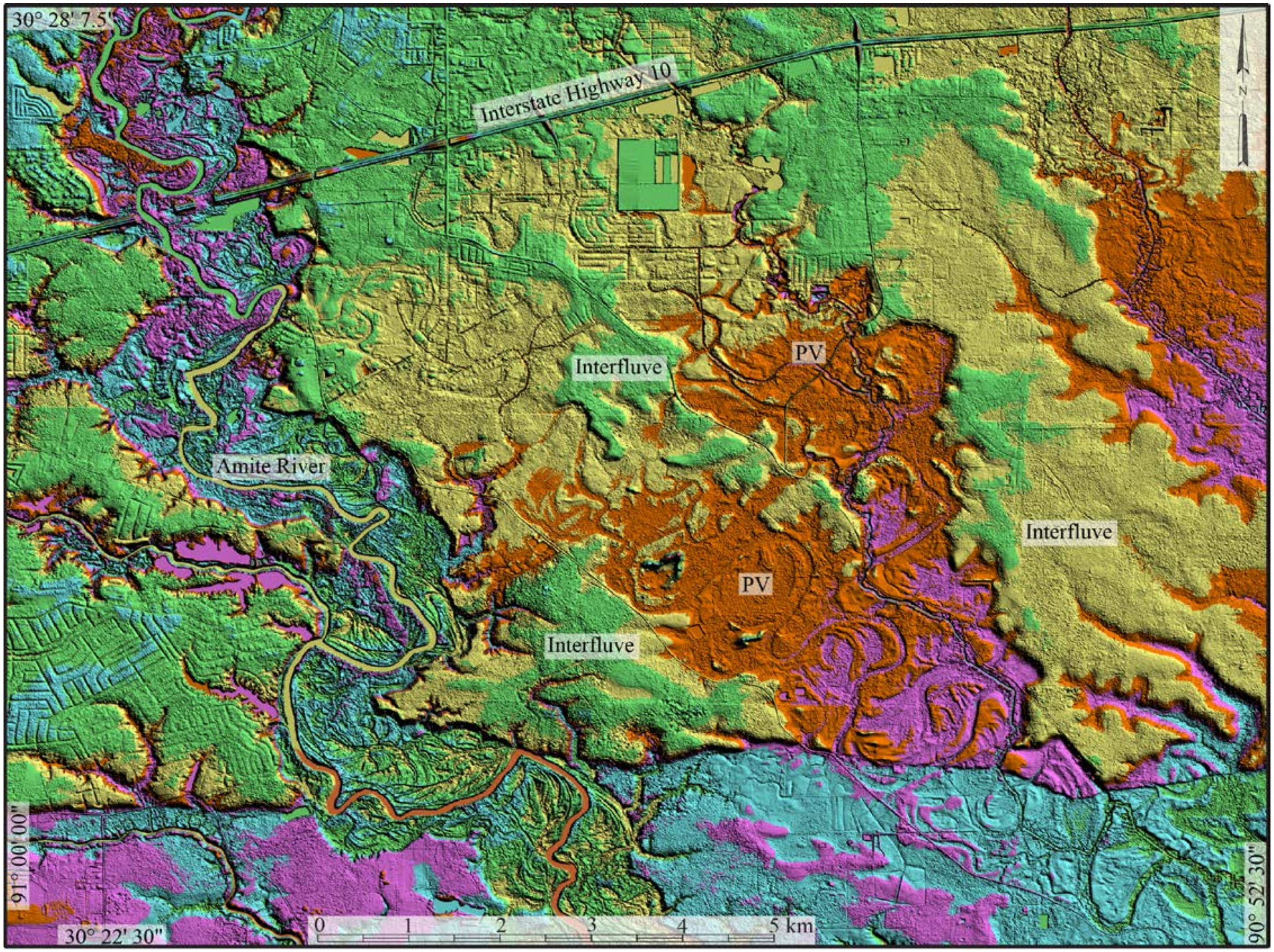

Figure 7.2. LiDAR DEM of Amite River Valley and Paleovalley (PV) showing channels, paleochannel, and intervening interfluves. Image constructed using Global Mapper and LiDAR DEMs from Atlas: The Louisiana Statewide GIS.

As noted by Blum et al. (2013), a number of papers, including Leopold and Maddock (1953), Wolman and Miller (1960), and Williams (1978), have demonstrated that the dimensions, depth and width of channels scale to discharge. The vertical dimension between the depth of scour and the bar top is a proxy for bankfull depth. In addition, the bankfull cross-sectional areas have been shown to adjust to flood discharges with recurrence intervals of 1 to5 years, the so-called bankfull discharge (Blum et al. 2013).

In the Gulf of Mexico region, modern fluvial channels range from hundreds of meters in width and a few meters in depth to the Mississippi River. For example, the Brazos River in central Texas is approximately 146 m wide and 17 m deep near Rosharon, Texas (USGS 08116650) and 150 m wide and 12 m deep at Richmond, Texas (USGS 08114000) (Heitmuller 2014). In southeast Texas, the Trinity River is about 200 m to 400 m wide and 5 m to 14 m deep near Moss Hills, Texas; near Liberty, Texas, it is about 100 m to 200 m wide and 6 m to 14 m deep (Phillips et al. 2005). Along the Texas-Louisiana border, the Sabine River is about 125 m wide and 10 m deep near Bon Weir, Texas, (USGS 08028500). At Ruliff, Texas (USGS 08030500), it is about 80 m wide and 14 m deep (Heitmuller 2014). Within Louisiana, the lower segment of the Mermentau River had a width that varied between 75 m and 300 m and a low-water depth of 3 m or more. Before channelization, the lower segment of the Calcasieu River was about 40 m wide and 1.5 m deep (Chief of Engineers 1919). Finally, at Baton Rouge, Louisiana, the Mississippi River varies between 600 m and 900 m wide and between 15 m and 30 m deep.

As reviewed extensively by Ethridge and Davidson (2011), reviewed briefly by Blum et al. (2013), and classified by Church (2006), modern fluvial channels exhibit a spectrum of plan-view channel patterns. They range between end-member patterns such as braided (rapidly shifting multi-channel systems), meandering (sinuous single channel), anabranching (stable multichannel systems), and straight (single channel) (Church 2006, Ethridge and Davidson 2011). These variations in stream pattern are interpreted to represent differences in stream power (proportional to the product of discharge and slope), bed material grain size, and the presence or absence of bank-stabilizing vegetation and/or muds (Church 2006).

The classical model of braided “lowstand” rivers and streams was developed following the observations of the glacial-period lower Mississippi River by Fisk (1944) and others. However, as discussed by Blum et al. (2013), the concept of lowstand braided rivers and streams is only applicable in general to the glacial-period Mississippi River and other meltwater channels for a number of reasons. First, unlike the glacial-period Mississippi River, other lowstand Gulf of Mexico rivers lacked glaciation in their headwaters and, as a result, were not braided (Saucier 1994, Blum et al. 1995, Blum and Aslan 2006, Blum 2007). Second, in modern, low-gradient, sedimentary basins that lack glacial influence in their contributing drainage areas, the braided-meandering transitions occur within uplifting to stable continental interiors, along rifted margins, or within a few tens of kilometers of orogenic fronts along active margins. As a result, in modern subsiding basins the majority of low-gradient river systems are dominated by either meandering or anabranching channels (Bridge 2003, Latrubesse 2008, Blum et al. 2013). Third, both Bridge (2003) and Blum et al. (2013) argue that it is unlikely that reliable interpretations of paleochannel patterns can actually be made solely from 2-D seismic data, well logs, or outcrops. This argument is supported by interpretations of braided fluvial systems being common in 2-D seismic, well-log, and outcrop studies; interpretations of braided channels in 3-D seismic data are rare (Ethridge and Davidson 2011, Blum et al. 2013). Finally, Blum et al. (2013) speculate that prehistoric braided rivers were common in certain tectonic settings, but are uncommon in the overall stratigraphic record, especially after the Early to Middle Paleozoic advent of primitive land plants, which introduced rooting systems, and promoted the production of meandering and anabranching fluvial systems and the preservation of their paleochannels. Therefore, the concept of lowstand braided fluvial systems likely does not apply to the Texas-Louisiana continental shelf. It is likely that the majority of the paleochannels within this region are meandering and anabranching in nature.

The assemblage of sediments that was created by the filling of a paleochannel is known as a "paleochannel fill" (Figures 7.1 and 7.2) It accumulates as the result of either the abandonment of a channel by neck cutoff, a local event that affects a single meander loop, or an avulsion, a regional event that affects an entire river course. Typically, channel fills are heterolithic. Normally, the upstream reach of a paleochannel is sand-rich because it is proximal to and remains connected with the newly active channel. A typical channel fill becomes increasingly mud-rich and eventually organic-rich along a paleochannel with distance away from where it connects to an active channel (Blum et al. 2013). Bridge (2003, 2006) provides a variety of detailed facies models for channel fills. Reijenstein et al. (2011) provide examples of paleochannels in high-resolution 2-D and 3-D seismic data.

Because of the overall heterolithic and generally fine-grained natures of channel fills, they are typically unsuitable as sources of sand needed for beach restoration and similar projects. Generally, they will occur as sinuous ribbons of muddy sediment within a fluvial channel belt sand that is composed of sand suitable for restoration purposes (Figures 7.3 and 7.4).

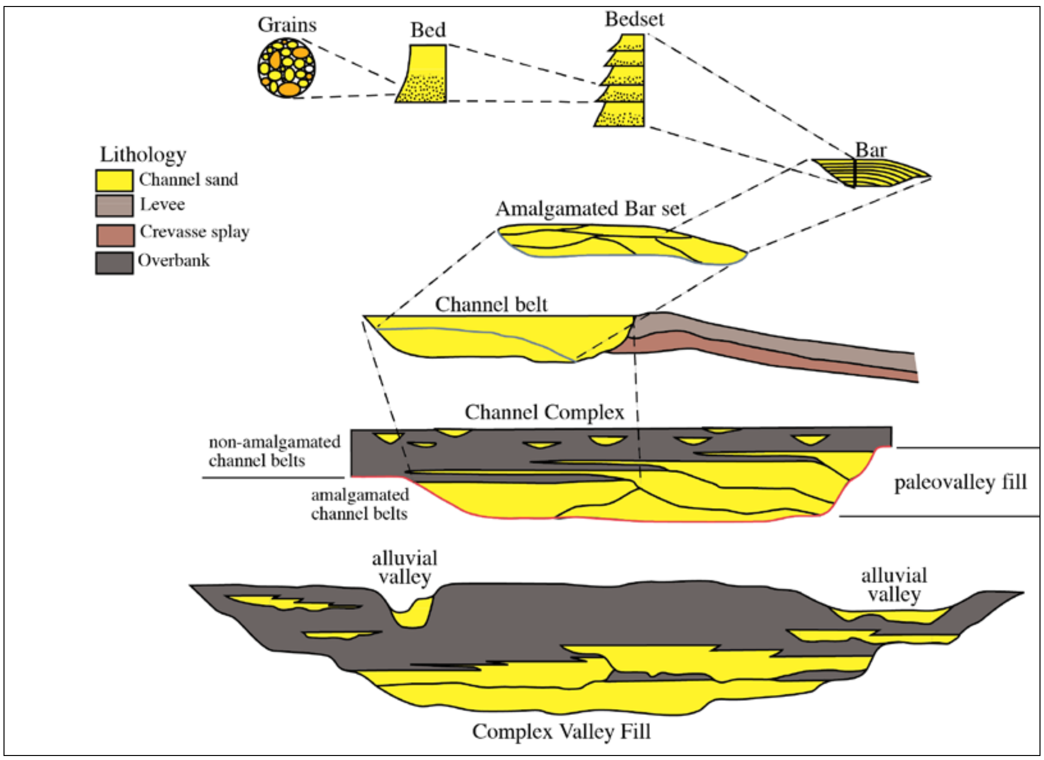

Figure 7.3. Hierarchy of fluvial architectural elements.
Modified from SEPM Stratigraphy Web (2019).

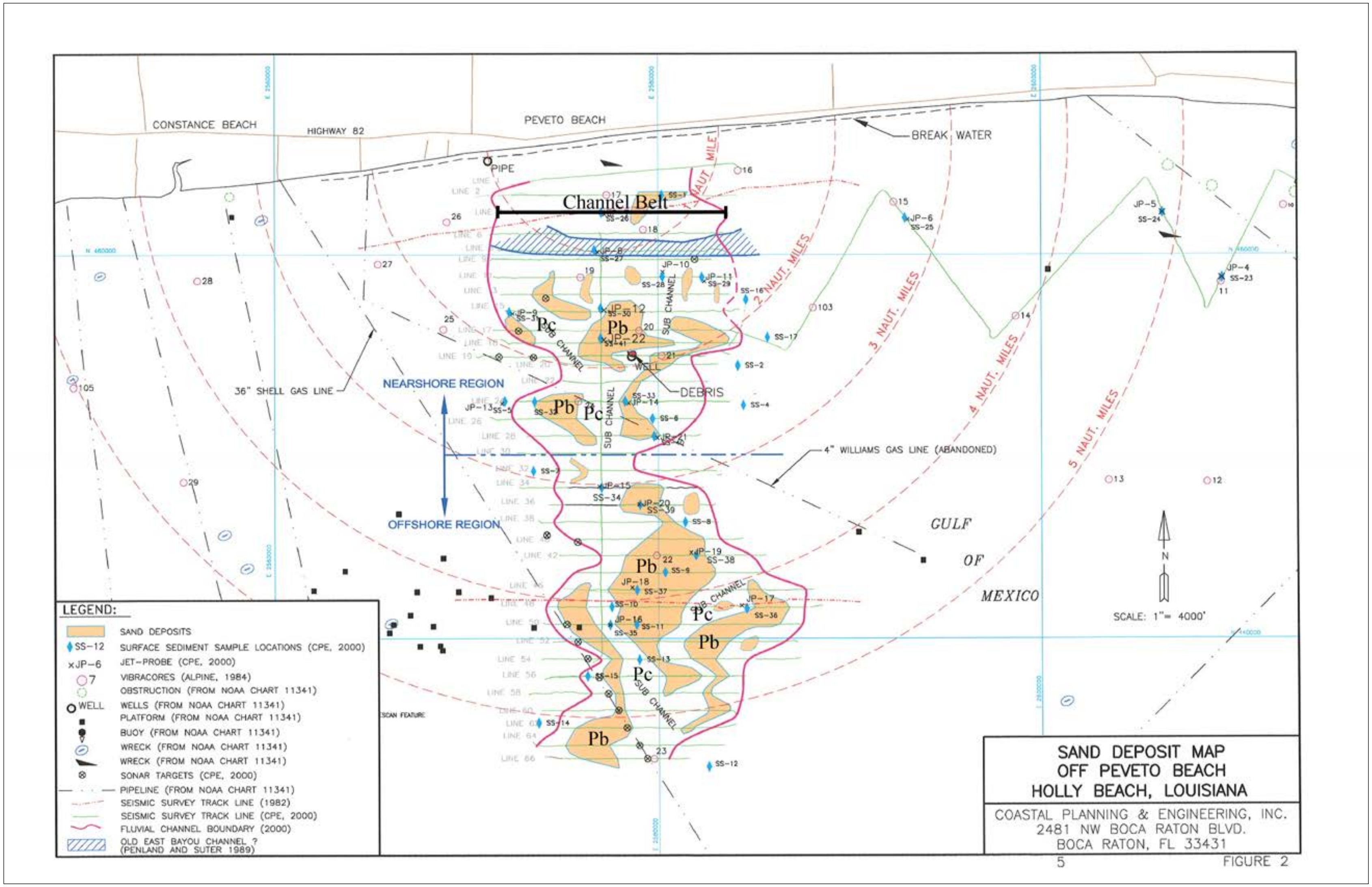

Figure 7.4. Peveto Channel sand body on the OCS used for Holly Beach Restoration interpreted as Late Pleistocene channel belt. Pb = point bar and Pc = paleochannel. Map from Coastal Planning and Engineering (2001).

7.2 Channel Belts and Deposits

As defined by Blum et al. (2013), a channel belt is a larger scale fluvial landform that consists of the area of floodplain over which a fluvial channel has migrated (Figures 7.1 and 7.2). This area includes individual paleochannels. Similarly, according to Blum et al. (2013), a channel-belt sand body, the channel body of Gibling (2006), is a single story ("storey") of sandy to gravelly sediments created by the lateral migration of a channel with no net aggradation (Figure 7.3). As defined and illustrated by Friend et al. (1979), a story is typically a flat-topped, and convex-downward body of sand, gravelly sand, to gravel with a major scour surface forming its base. A fluvial story typically fines upward and is overlain by overbank fines. The upper contact can also consist of the major scour surface at the base of an overlying channel belt sand body (story). Together they would comprise a multistory sand body (Potter 1963, Friend et al. 1979, Blum et al. 2013). The thickness of a single story of a channel-belt sand body generally approximates the vertical distance between the maximum depth of scour and the water surface at bankfull stage. This thickness may be significantly greater than the observed channel depth at low flow. In seismic records, a single set of large-scale inclined-strata sets will represent what is observable and diagnostic at the scale of seismic data (Reijenstein et al. 2011, Blum et al. 2013).

The thickness of a fluvial story and associated channel belt sand body will vary greatly with the scale of either the stream or river that created them. For example, in the case of the Oyster Creek course of the Brazos River near Dewalt, Texas, the channel belt sand body is about 15 m to 18 m thick and 2 km to 2.5 km wide (Bernard et al. 1970). The possible range of story thicknesses is from 2 m to 3 m for small streams up to 30 m to 60 m, for a fluvial system the size of the Mississippi River (Blum et al. 2013).

The channel belt and associated paleochannels are significant features in terms of both sand resources, paleosurfaces, and buried cultural resources. The channel belt sand body that underlies a channel belt is a prime target for finding economic offshore sources of sand suitable for the restoration of beaches and similar coastal projects. Although the bulk of the channel belt sand body may consist of suitably sandy sediments, any associated paleochannels will largely consist of ribbon-like bodies of muddy sediments, which are unsuitable for restoration projects, lying within the sandy channel belt sediments. In addition, if buried and preserved, the surface of a channel belt is a paleosurface and part of a relict floodplain that would have been in prehistoric times a favored location for prehistoric settlement. The natural levees that flank these paleochannels would be a preferred location for the location of prehistoric settlements and the accumulation of archaeological deposits (Weinstein et al. 1979a, Weinstein 1981, Gagliano et al. 1982). In many places, the floodplain–channel belt paleosurfaces have been destroyed by either ravinement processes or channel scour during the aggradation of a coastal plain or within a paleovalley during a marine transgression (Pearson et al. 1986).

7.3 Paleovalleys and Their Fills

The next level of fluvial landforms found within the study area are paleovalleys. Typically, the term "incised valley" is used in the literature instead of "paleovalley." However, Van Wagoner et al. (1990), Zaitlin et al. (1994), and Boyd et al. (2006) define "incised valleys" to include: 1) significant erosional relief with truncation of older strata, 2) juxtaposition of fluvial or estuarine strata on marine deposits, and 3) demarcation of a significant basinward shift in facies, with subaerial exposure on interfluves. As noted by Blum et al. (2013), many fluvial valleys are incised, but do not meet criteria 2 and 3 of how an incised valley is defined. In addition, the modern analogues for "incised valleys" are typically buried beneath Holocene delta plains, or submerged on the shelves, and lack morphological expression. Therefore, following Blum and Törnqvist (2000) and Blum et al. (2013), valley or paleovalley are preferred over incised valley because, by definition, all valleys are incised.

Like Blum and Törnqvist (2000) and Blum et al. (2013), it is assumed here that relative sea-level fall produces incision and an upstream-propagating wave of stream rejuvenation. Step-wise sea level fall results in step-wise incision and the formation of terraces and their associated allostratigraphic units. During sea level lowstand, a basal channel-belt sand body forms by the subaqueous scouring of migrating meandering or anabranching channels. Contrary to the often-reproduced illustration of the Mississippi River model of Fisk (1944), the bottom of a valley is never subaerially exposed. Later, a paleovalley eventually becomes filled by a predictable succession of sedimentary facies as sea level rises, e.g., Blum (1993), Boyd et al. (2006), and many other publications too numerous to mention (Figure 7.3).

As discussed by Blum and Törnqvist (2000), Boyd et al. (2006), Blum et al. (2013), and many other publications, the changing fluvial and estuarine processes during a sea level cycle create a predictable succession of three sets of fluvial and estuarine deposits. Initially, the forced regression of the shoreline causes entrenchment of the river into its fluvial ridges and deltas as the river course extends seaward. If the river course is abandoned during this process, entrenched meanders cut into its fluvial ridge and beheaded distributaries of its delta are left behind as exhibited by the surfaces of the Prairie Allogroup units in southwest Louisiana.

More commonly, relative sea-level not only falls to the point that deep incision of the coastal plain and valley formation occurs, but incision occurs in a step-wise fashion. As happened during MIS 4 through 2 within the northwestern Gulf of Mexico coastal plain, each period of incision was separated by periods of lateral migration and extensive channel-belt deposition. This results in the creation of a valley where a series of terraces overlying channel belt sand bodies flanking either side of the valley step downward along strike. In this situation, the oldest channel belt sand body lies topographically higher than, and is cross-cut by, successively younger channel belt sand bodies. This type of degradational stacking of channel belt sand bodies and associated terraces is typical of the northwestern Gulf of Mexico coastal plain where subsidence rates are relatively low and shelf gradients are similar to the river longitudinal profile. As a result, major slope adjustments are not necessary as the river course extends across the shelf as relative sea level falls (Blum and Valastro 1994, Blum and Price 1998, Blum et al. 2013).

Second, during a relative sea level lowstand of a glacial cycle, e.g., MIS 2, coastal rivers lay confined within valleys that they entrenched into the coastal plain and former highstand continental shelf. As a result, the shifting of either meandering or anabranching channels deposited an extensive single-story, channel belt sand body that overlies a scoured surface cut by lateral migration of the course within a coalesced channel belt. Furthermore, when relative sea level initially rose in step-wise fashion, the migration of a river course within the confined valley typically eroded and truncated the upper part of the underlying story as it deposited another channel belt sand body (Figure 7.3). This action created a multistoried, amalgamated, lower valley-fill consisting of superimposed channel belt sand bodies that are separated by laterally persistent scour surfaces. This amalgamated, lower valley-fill, following the model of Fisk (1944) for the Mississippi River Valley, has been commonly interpreted to mean that many lowstand Quaternary coastal plain rivers were “braided” during sea-level lowstands and that the amalgamated lower valley-fill of a paleovalley represents the braided-stream deposits (Blum 2007, Blum et al. 2013). The basal fluvial fill and terrace deposits represent potential sand resources if they occur close enough to the gulf bottom. If a paleovalleys is sufficiently deep, the terrace surfaces and basal channel belt within a paleovalley represent paleosurfaces that might have cultural resources preserved below the unnamed ravinement surface (Pearson et al. 1986).

Finally, during rising sea level, the valley fills with a mud-rich non-amalgamated upper valley-fill sequence that represents aggradation in response to relative sea-level rise. In the coastal plain away from the transgressing shoreline, the mud-rich sediments typically consist of separate fluvial channel belt, backswamp, and lacustrine sediments that are fundamentally avulsive and distributive (Saucier 1994, Törnqvist et al. 2000, Blum et al. 2013) (Figure 7.3). Adjacent to the transgressing shoreline, the upper valley fill of mud-rich sediments is dominated from bottom to top, by tidal inlet and/or flood tidal delta; central bay mud; and upper bay and/or upper bay delta sediments. These depositional environments often abruptly back-step along flooding surfaces. They overlie channel belt sands at the base of river and stream valleys. In the case of the smaller, intrabasinal rivers, these basal fluvial sands often consist of multistoried, amalgamated channel belt sand bodies (Anderson et al. 2008, Milliken et al. 2008b, Milliken et al. 2008c).

The transgressive part of these valley fills accumulated within a zone of backwater effects associated with the shoreline. As a result, it accumulated as a landward-tapering wedge that is proportional to backwater length. At their landward end, a trangressive valley fill onlaps onto amalgamated fluvial channel belt and floodplain sediments that accumulated during the MIS 4–2 period of relative sea-level fall and lowstand. The landward termination and onlap of trangressive valley fills migrates upstream through time as the valley fills (Törnqvist et al. 2001, Milliken et al. 2008b, Milliken et al. 2008c, Blum et al. 2013).

7.4 Paleochannels Compared to Channel Celts and Paleovalleys

As previously noted, the landforms mapped in the hazard and archaeological surveys are identified by a bewildering multitude of names. The most common components of this nomenclature are “paleochannel” and “channel.” However, judging from their plan view and cross-section where they cross the USGS seismic data, most of these paleochannels and channels are either paleovalleys or channel belts.

Coastal plain paleovalleys differ from paleochannels in plan view, internal structure, scale, and relief. In plan view, paleovalleys exhibit a strong tendency to widen downstream independent of relative sea-level in contrast to the relatively parallel-sided and sinuous form of an observed segment of channels and paleochannels, e.g., Martin et al. (2011) and Reijenstein et al. (2011). Similarly, paleovalleys also differ noticeably in plan view from channel belts in that channel belts, of which paleochannels are a part and inset into, are typically of variable width. They also have roughly subparallel and often arcuate boundaries that differ from the edges of both paleovalleys and paleochannels. In the case of scale relative to that of contemporaneous channel belts, coastal-plain paleovalleys range from comparably wide to wider. However, both paleovalleys and channel belts are significantly wider and more variable in relative width than the paleochannels inset into them (Miall 2002, Ethridge and Schumm 2007, Reijenstein et al. 2011, Blum et al. 2013).

Paleovalleys differ from paleochannels and channel belts in cross section, as seen in seismic sections. Basically, the valley walls of a paleovalley extend above the height of a single-story channel belt sand. Thus, the fill of a paleovalley will consist of either basal single-story channel belt sand or multistory channel belt sand bodies overlain by a mud-rich non-amalgamated upper valley-fill of either fluvial or estuarine origin. The depth of incision of a channel belt will be limited to the thickness of its channel belt sand body. Also, the fluvial channel bar deposits of a channel belt sand body often exhibit distinctive inclined seismic facies, e.g., Reijenstein et al. (2011) and Blum et al. (2013). However, a paleovalley that has been truncated by erosion down to its basal channel belt sands would be indistinguishable from a channel belt sand body. Paleochannels will appear as depressions that have relief equivalent to or less than the thickness of a single channel belt sand body. The fill of such depressions has been observed as a continuous and small-scale sag in seismic reflections as discussed by Reijenstein et al. (2011).

7.5 Southwest Outer Continental Shelf Study Area

The application of the above criteria to the southwest Outer Continental Shelf (OCS) study area indicates that the majority of features mapped as either channels or paleochannels by the hazard surveys are largely paleovalleys that in some cases are mixed in with channel belts and paleochannels. Many of the features exhibit the noticeably divergent, irregular plan view with distinct tributaries that are more consistent with paleovalleys than either channel belts or paleochannels as in West Cameron Blocks WC151, WC165, and WC166 (Figure 7.5 In West Cameron Blocks WC115, WC128, WC129, and WC130, the largest of these paleovalleys can be identified as the MIS 2 paleovalley of the Calcasieu River (Figure 7.5). A few of these features have the parallel sides and meandering plan view that is typical of true paleochannels as seen in East Cameron Blocks, EC30, EC31, and EC44. In West Cameron Blocks, WC132, WC149, WC150, and WC167, one large sinuous feature is large enough to be a possible channel belt (Figure 7.5). Unfortunately, the data are too fragmentary and inconsistent to determine the interrelationships between and complete extent of many of these fluvial features. Also, the lack of a representative seismic record for each of the specific fluvial features mapped prevented their reinterpretation. Because of this, it was impossible to correlate and reconstruct a relative chronology of paleovalleys, channel belts, and paleochannels and to confidently identify the nature of many of the mapped fluvial features. Unlike the Calcasieu, Sabine, and Trinity paleovalleys, it could not be determined which of these fluvial features either belong to the Beaumont, Deweyville, or Mermentau alloformations, because many of the fluvial features are truncated by the unnamed ravinement surface. As a result, they potentially vary in age from Late MIS 5 to either late MIS 2 or early MIS 1.

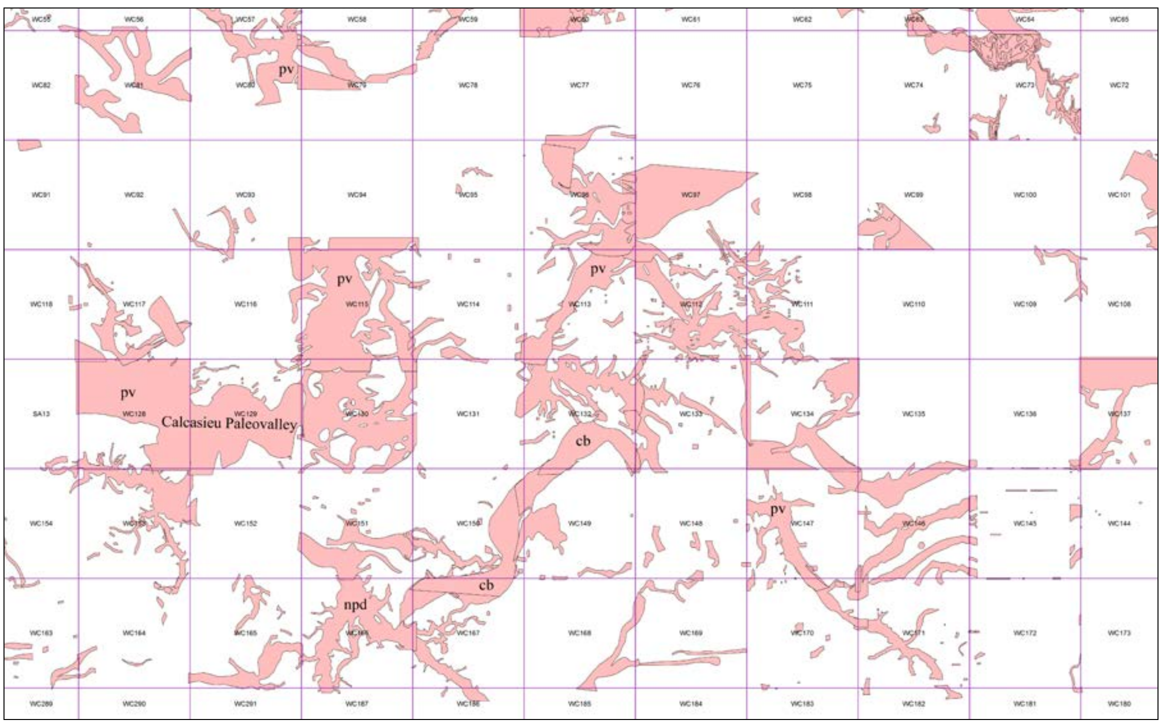

Figure 7.5. Excerpt from plot of fluvial features mapped by offshore hazard surveys.
Excerpt shows features interpreted to be paleovalleys (pv) and channel belts (cb). Note northward-oriented paleodrainage (npd).

However, within the hazard-survey data, there are noticeable drainage anomalies. The first is the well-known coast-parallel paleovalleys of the Calcasieu, Mermentau, and Sabine rivers that once flowed westward into the Trinity River's paleovalley and then to the shelf edge. Second, possibly associated with this anomaly is a partial, northward-draining paleovalley system in West Cameron Blocks WC115, WC128, WC129, and WC130 (Figure 7.5). This orientation is the opposite of what one would expect, given the gulfward slope of the present continental shelf. One possible explanation for both anomalies is that the crest of the forebulge of the Laurentide Ice Sheet was at one time during the Late Pleistocene south of the study area. As a result, a topographic east-west trough existed within the study area when it was an exposed coastal plain that diverted the drainage of the Mermentau and Calcasieu rivers westward and parallel to the coast. During the MIS 2 lowstand, this drainage pattern was superimposed on the exposed shelf when the Mermentau and Calcasieu rivers further entrenched into it along with northward draining tributaries of it. Finally, the apparent channel belt in West Cameron Blocks WC132, WC149, WC150, and WC167 is at right angles to and abruptly terminated by the northward-draining paleovalley (Figure 7.5). This suggests that the paleovalleys and other fluvial features mapped by hazard surveys represent an amalgamation of multiple periods of paleovalley and channel belt formation.

With few exceptions, the relationship of the paleovalleys, channel belts, and paleochannels of the hazard surveys to the Pleistocene fluvial features (referred to as stream “courses,” “fluvial channels,” and “buried fluvial channels”) of Suter and Berryhill (1985) and Suter (1987) is problematic. From comparison of the fluvial features mapped by Suter and Berryhill (1985) and Suter (1987) and the hazard surveys in the project GIS it is clear that the Calcasieu River paleovalley closely matches one of Suter's fluvial channels (Figure 7.6). Nevertheless, an examination of the distribution of both sets of mapped fluvial features suggests that some segments of his fluvial channels likely represent highly generalized groupings of separate paleovalleys. However, for other of his fluvial features there is an overall lack of any definable correlation between the features mapped by Suter and Berryhill (1985) and Suter (1987) and those mapped by the hazard surveys (Figures 7.6 and 7.7). Also, from their plan and seismic characteristics alone, they are regarded to be neither “channels” nor “paleochannels” because they are too large, too deep, and often multistoried. To effectively evaluate the relationships among Suter's fluvial features and those mapped by the hazard surveys, Suter's analog seismic data need to scanned and reexamined in light of the hazard-survey mapping and new subsurface data collected.

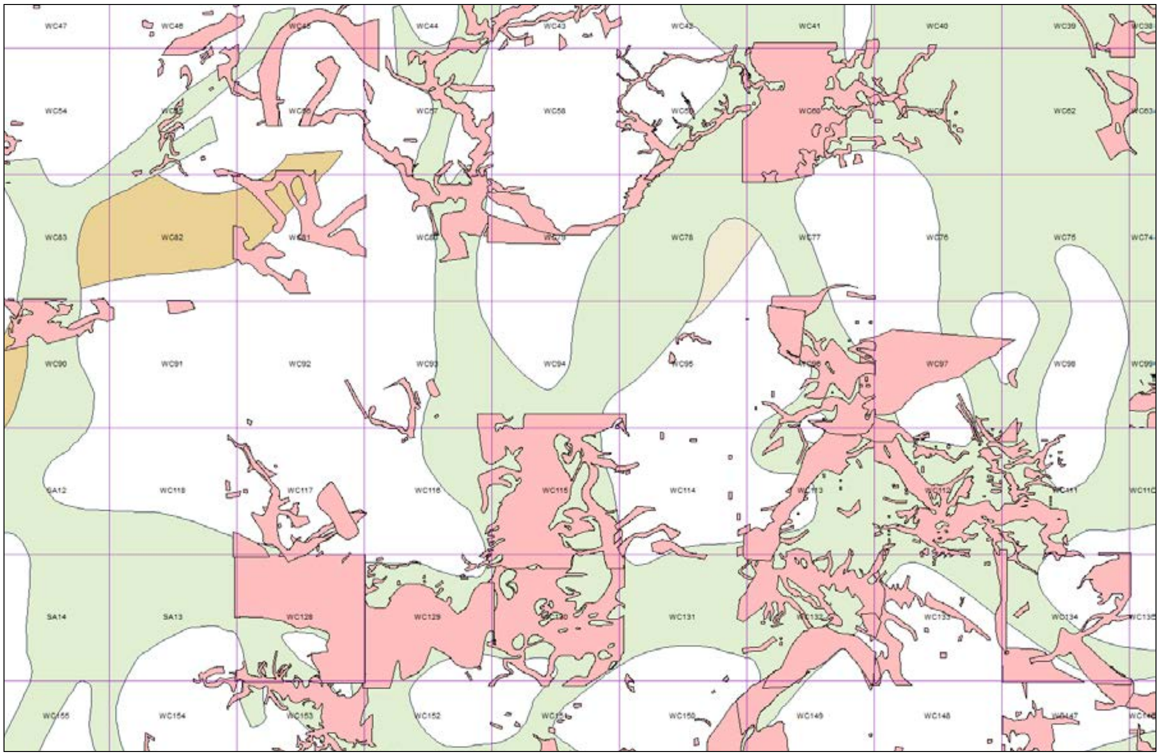

Figure 7.6. Excerpt from plot of fluvial features mapped by offshore hazard surveys and Suter (1987).

Excerpt shows lack of correspondence with fluvial “channels” and related features mapped by Suter (1987), and the disconnected nature of Hazard survey mapping.

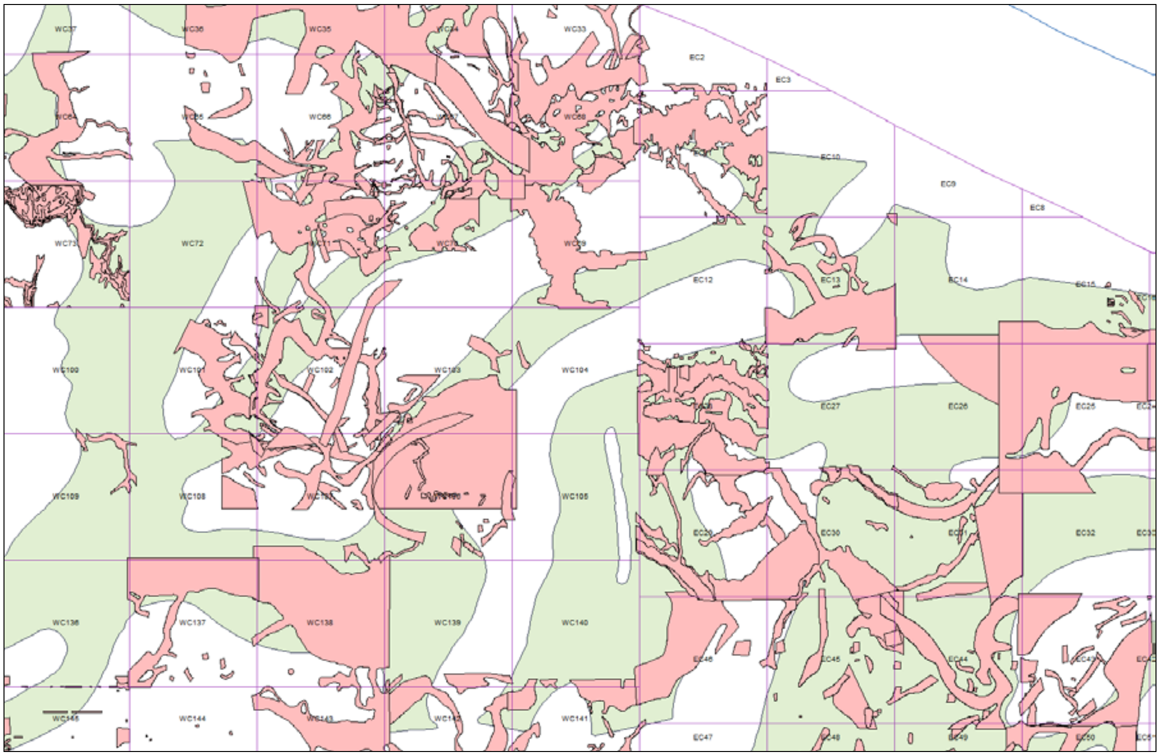

Figure 7.7. Excerpt from plot of fluvial features mapped by offshore hazard surveys and Suter (1987).

Maps shows the lack of correspondence with fluvial “channels” and related features mapped by Suter (1987), and the disconnected nature of Hazard survey mapping.

Another problematic aspect of the fluvial features mapped by Suter and Berryhill (1985) and Suter (1987) is their relationship to onshore channel belts and paleochannels. Although clearly degraded, the paleochannels and channel belts exhibited by the Prairie surface can be readily reconstructed using historic aerial photography, USD soil surveys, and LiDAR DEMs. As a part of the preparation of Heinrich et al. (2002, 2003) and Heinrich (2005b, 2006), these channel belts and paleochannels were mapped and compiled (Figure 7.7). Comparing the onshore MIS 5 paleochannels with the fluvial features of Suter and Berryhill (1985) and Suter (1987) revealed that many of his fluvial features “terminate” abruptly at the coastline without any onshore continuation, with only two observable exceptions. Given that his fluvial features should be falling stage, MIS 5a to MIS 3, or slightly younger, they should have onshore continuity with either mappable paleochannels or channel belts. One exception is the MIS 3 Lafayette Meander Belt of the Avoyelles alloformation that should continue offshore as at least one of these fluvial features (channel belt or belts) offshore to the West Louisiana delta of Anderson et al. (1996) and Wellner et al. (2004). The other exception is the previously mentioned Calcasieu paleovalley, which continues onshore as Lake Calcasieu and the associated estuary.

7.6 Paleoshorelines and Inner Shelf Shoals

Rodriguez (1999) and Rodriguez et al. (1999, 2004) collected and analyzed sediment cores, high-resolution seismic data, and high-resolution side-scan sonar data from Sabine, Heald, Shepard, and Thomas banks on the east Texas inner continental shelf and within the current study area. They observed that all four shoals have similar sediment facies, structure, and presumably genesis. On the basis of the location of Heald, Sabine, Shepard, and Thomas banks and mapping of tidal-inlet valley fill within the Trinity River paleovalley, Rodriguez (1999) and Rodriguez et al. (1999, 2004) argued that the sandy sediments of the Sabine alloformation that compose these shoals are remnants of a barrier island chain that has been eroded and reworked by storms and wind-driven currents into offshore sand shoals. They argued that what they interpreted to be shoreface/ebb-tidal delta deposits are the only portion of the barrier island preserved and that the overlying dune, beach, and shoreface sediments were eroded and reworked into the upper storm-influenced facies that comprises the sand shoals.

More recently, Swartz (2016) and Swartz et al. (2016) collected and analyzed about 90 km of high resolution CHIRP data that were collected by the University of Texas marine geophysics field course over Heald Bank and the adjacent continental shelf. They interpreted these seismic data as showing that a large and complex "channel" network underlies Heald Bank and the surrounding sea floor. These authors reported a lack of any distinct shoreface sequences associated with the bank, which they argued sits on a ravinement surface, the "unnamed ravinement surface" of this report, that truncates underlying layered channel fills of undetermined age and correlation. They propose that Heald Bank is not a drowned barrier island.

8. Synthesis and Summary

As discussed in the preceding section, the current data collections involving offshore hazard and cultural resources surveys have significant shortcomings in terms of data collection. One major concern is that they focus on what can be said to be a "postage stamp size" view that lacks integration into a regional framework. As a result, the absolute age, if determined at all, of fluvial and coastal landforms that are being mapped is determined by what is best described as the "count from the top" method. This is a method that has led to considerable confusion and misinterpretations in Quaternary studies elsewhere, as in case of the five-stage Midwest glacial chronology. Accurate dating of individual paleovalleys, channel belts, and paleochannels mapped by hazard and cultural resources surveys is important, because a large number of these fluvial landforms likely predate the human occupation and, as a result, lack any predictable and significant cultural resources associated with them. If their age could be established with any confidence, such fluvial features could be considered as potential sources of sand for coastal restoration.

Another concern is the confusion in the interpretation of seismic data in the identification of and nomenclature used to designate paleovalleys, channel belts, and paleochannels. For example, the examination of the available seismic data found that many paleovalleys and channel belts are being misidentified as paleochannels or indiscriminately designated as "paleochannels." This is an important distinction, because the expected distribution of cultural resources associated within and around a fluvial landform will vary greatly as whether it is either a paleovalley, channel belt, or paleochannel. As a result, the accurate determination of the type of fluvial landform being mapped is an essential step in evaluating the presence of the type cultural resources present within a lease block. In case of a paleovalley that has been truncated below the top of its usually amalgamated, basal channel belt, it might be impossible to differentiate it from a channel belt. However, for cultural resource purposes, the difference between the two is moot because this degree of erosion of a paleovalley has destroyed any existing paleosurfaces within it and the potential for the existence of significant cultural resources within it.

The limited size of offshore hazard and cultural resources surveys and lack of an accessible regional seismic survey also presented limitations. The lack of a usable regional overview creates large data gaps that prevent correlation of mapping from one group of lease blocks to another and the confident and detailed reconstruction of drainage patterns. Also, focusing on a single lease block or group of lease blocks makes it easy to miss the geomorphic context of the area being evaluated. For example, it would be hard to tell that a lease block is lying in a large paleovalley if the valley walls lie outside the immediate study area. A large seismic dataset was acquired for the research by Suter and Berryhill (1985) and Suter

(1987). However, it exists only in hard copy format that is impractical for the vast majority of researchers to access and use.

Also, an overall lack of consistency was found in the interpretation of data; the manner in which they were mapped; and the stratigraphic and geomorphic and nomenclature used to designate features mapped. In the hazard and cultural resources survey report, rarely were the specific criteria used to map the boundaries of fluvial landforms stated. As a result, it is hard to know how the boundaries were drawn from the seismic data and how these differences might have influenced the mapping of these landforms. In rare cases, the thalweg of a fluvial landform was mapped instead of its boundaries, which made comparison of mapping quite difficult if not impractical. However, it does appear that in some cases effort was made to edge match their data with the data of older, adjacent surveys.

Finally, in the course of this study, it was found to be very difficult to access and replicate previous interpretations of offshore hazard and cultural resources surveys. This could not be done because many of the offshore hazard and cultural resources surveys lacked a type seismic line that illustrated the seismic signature of each of the landforms mapped (or access to the seismic data). Similarly, the mapping compiled from the offshore hazard and cultural resources surveys could not be compared with the research by Suter and Berryhill (1985) and Suter (1987) because the bulky hard-copy format of their seismic lines made it impractical to reexamine critical segments of their seismic lines.

8.1 Site Preservation and Stratigraphy

Although much remains to be learned about the geoarchaeology of the Louisiana-Texas offshore continental shelf, some preliminary generalizations can be made about where paleosurfaces, which might be associated with cultural resources, can be found. First, the important surface that is associated with preserved cultural resources is the Holocene-Pleistocene surface. It was the Late Pleistocene, MIS 2, coastal plain when the northwestern Gulf of Mexico was occupied by Native Americans. The extent to which it is preserved and unmodified by relative sea level rise will be a major control on the preservation of cultural resources during the last and ongoing transgression. As previously noted by numerous other researchers, Gagliano et al. (1982), Pearson et al. (1986), and Waters (1992), the primary location where this surface will be preserved is within deeply entrenched paleovalleys. The preservation of paleosurfaces within paleovalleys will depend on the rate and corresponding degree of modification that they experience from the transgressing bayline. Also, if a paleovalley was not entrenched deeply enough into the surrounding coastal plain, shoreface erosion potentially could have cut deeply into the basal channel belt within its valley fill and largely could have destroyed any paleosurfaces that existed within it. Within the Pleistocene coast-parallel surfaces and associated Pleistocene and Holocene channel belts on these surfaces, shoreface erosion and ravinement surface formation will likely have extensively modified the Holocene-Pleistocene surface and any cultural deposits associated with it.

It is theoretically possible that periods of very rapid sea level rise might have preserved strips of the Holocene-Pleistocene surface by the in situ drowning of existing barrier islands or chenier plains by a process known as step-wise retreat. In such cases, erosion of the beaches ridges during the initial drowning and by later tidal action would have likely heavily reworked these beach ridges and destroyed any associated cultural deposits. However, as the shoreface jumps to its new location, lagoonal deposits and a strip of the Holocene-Pleistocene surface and any associated cultural resources underlying them might survive the transgression.

In case of delta and chenier plains, there is a good chance that some cultural resources might survive the initial submergence along its bayline or soundline. However, the advancing shoreface and ravinement formation will likely extensively have eroded the surface of the delta or chenier plain and have destroyed any associated cultural resources. More research needs to be done as detailed studies and documentation about how the submergence of cultural resources actually affects both surface and buried cultural resources is lacking.

8.2 Recommendations for Management and Protection

This research has developed a number of recommendations for management and protection.

1. In terms of the mapping of submarine paleolandforms and potential paleolandscapes:
   - A. Regional terminology should be standardized.
   - B. There needs to be clear definition of, recognition of, and differentiation among paleovalley, paleochannels, and channel belts.
   - C. A type-seismic section for specific paleovalleys, paleochannels, and channel belts mapped for that block should be included in the report.
   - D. Geophysical data should be submitted with block survey reports for BOEM analysis and development of regional geologic models from multiple block surveys, and archiving for future or alternate uses such as sand resources identification.
   - E. Geologic sampling to ground-truth geophysical data and absolute dating of potential paleolandscapes within fluvial valleys should be conducted using appropriate techniques.

2. There needs to be a more regional approach to mapping submerged landforms and attendant bounding surfaces. This would include:
   - A. Block survey data (not just interpreted map products as done in this study) should be applied to develop a regional geologic model and conceptual model for shelf evolution.
   - B. When new surveys are conducted, provide regional models to operators for edge matching with previous studies and context locally for their study area.
   - C. Digitize and make readily available online the USGS seismic lines of Suter and Berryhill (1985) and other regional surveys conducted by USGS and LGS that are preserved only as analog or paper forms.
   - D. Develop uniform classification schemes, nomenclature, and recognition criteria for submerged landforms and potential paleolandscapes.

3. There needs to be a better understanding of the depositional and/or erosional response during transgression as ravinement surfaces are produced at the shoreface and within valley estuarine systems to determine preservation potential of prehistoric landscapes within valley fill packages.

4. Allow selected drilling on paleovalleys with a mandate that boring logs and “soil” samples from them be made available for scientific research by Quaternary geologists and geoarchaeologists. Any potential damage to sites will be offset by additional data that, due to funding constraints, would not otherwise be collected.

5. Avoidance criteria should be developed based on a strong understanding of shelf and/or valley-fill evolution in response to sea-level rise. The results of this study demonstrate that shelf preservation of paleolandscapes with potential to host prehistoric cultural resources would be extremely rare due to

significant erosion that occurs during sea level rise. Regional models that identify geologic units and bounding surfaces based on absolute age-verification should be applied because, based on the synthesis presented herein, it is possible that many of the paleo-fluvial systems being avoided in the vicinity of the study area are far too old for potential human occupation.

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