Coastal morphodynamics and Chenier-Plain evolution in southwestern Louisiana, USA: A geomorphic model

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Abstract

Using 28 topographic profiles, air-photo interpretation, and historical shoreline-change data, coastal processes were evaluated along the Chenier Plain to explain the occurrence, distribution, and geomorphic hierarchy of primary landforms, and existing hypotheses regarding Chenier-Plain evolution were reconsidered. The Chenier Plain of SW Louisiana, classified as a low-profile, microtidal, storm-dominated coast, is located west and downdrift of the Mississippi River deltaic plain. This Late-Holocene, marginal-deltaic environment is 200 km long and up to 30 km wide, and is composed primarily of mud deposits capped by marsh interspersed with thin sand- and shell-rich ridges (“cheniers”) that have elevations of up to 4 m.

In this study, the term “ridge” is used as a morphologic term for a narrow, linear or curvilinear topographic high that consists of sand and shell material accumulated by waves and other physical coastal processes. Thus, most ridges in the Chenier Plain represent relict open-Gulf shorelines. On the basis of past movement trends of individual shorelines, ridges may be further classified as transgressive, regressive, or laterally accreted. Geomorphic zones that contain two or more regressive, transgressive, or laterally accreted ridges are termed complexes. Consequently, we further refine the Chenier-Plain definition by Otvos and Price [Otvos, E.G. and Price, W.A., 1979. Problems of chenier genesis and terminology—an overview. Marine Geology, 31: 251–263] and define Chenier Plain as containing at least two or more chenier complexes. Based on these definitions, a geomorphic hierarchy of landforms was refined relative to dominant process for the Louisiana Chenier Plain. The Chenier Plain is defined as a first-order feature (5000 km²) composed of three second-order features (30 to 300 km²): chenier complex, beach-ridge complex, and spit complex. Individual ridges of each complex type were further separated into third-order features: chenier, beach ridge, and spit.

To understand the long-term evolution of a coastal depositional system, primary process–response mechanisms and patterns found along the modern Chenier-Plain coast were first identified, especially tidal-inlet processes associated with the Sabine, Calcasieu, and Mermentau Rivers. Tidal prism (Ω) and quantity of littoral transport (Mtotal) are the most important factors controlling inlet stability. Greater discharge and/or tidal prism increase the ability of river and estuarine systems to interrupt longshore sediment transport, maintain and naturally stabilize tidal entrances, and promote updrift deposition. Thus, prior to human modification and stabilization efforts, the Mermentau River entrance would be classified as wave-dominated, Sabine Pass as tide-dominated, and Calcasieu Pass as tide-dominated to occasionally mixed.

Hoyt [Hoyt, J.H., 1969. Chenier versus barrier, genetic and stratigraphic distinction. Am. Assoc. Petrol. Geol. Bull., 53: 299–306] presented the first detailed depositional model for chenier genesis and mudflat progradation, which he attributed to changes in Mississippi River flow direction (i.e., delta switching) caused by upstream channel avulsion. However, Hoyt’s model oversimplifies Chenier-Plain evolution because it omits ridges created by other means. Thus, the geologic evolution of the Chenier Plain is more
complicated than channel avulsions of the Mississippi River, and it involved not only chenier ridges (i.e., transgressive), but also ridges that are genetically tied to regression (beach ridges) and lateral accretion (recurved spits).

A six-stage geomorphic process-response model was developed to describe Chenier-Plain evolution primarily as a function of: (i) the balance between sediment supply and energy dissipation associated with Mississippi River channel avulsions, (ii) local sediment reworking and lateral transport, (iii) tidal-entrance dynamics, and (iv) possibly higher-than-present stands of Holocene sea level. Consequently, the geneses of three different ridge types (transgressive, regressive, and laterally accreted) typically occur contemporaneously along the same shoreline at different locations.

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1. Introduction

The Chenier Plain of southwestern (SW) Louisiana extends from Sabine Pass at the Texas/Louisiana border, eastward almost 200 km to Southwest Pass at Vermilion Bay (Fig. 1). This Late-Holocene, marginal-deltaic environment is up to 30 km wide and is composed primarily of mud deposits that are capped by marsh and interspersed with thin sand- and shell-rich ridges known as cheniers. The term “chenier” comes from the French Cajun word chêne, which means “oak tree.” In the Chenier Plain, oak trees line these ridges, which are better drained and topographically higher than the surrounding marsh.

Shifts in the course of the Mississippi River and the geomorphic development of the Chenier Plain were discussed first by Russell and Howe (1935) and Howe et al. (1935). Fisk (1948) discussed the geology of Cameron and Vermilion Parishes and provided a general description of Chenier-Plain morphology and geology. Furthermore, Fisk emphasized the importance of shoreline orientation and relict river courses for interpreting shifts in shoreline position. Using geomorphology, stratigraphy, and extensive radiocarbon dating, Gould and McFarlan (1959) and Byrne et al. (1959) documented the geologic framework of the Chenier Plain and described seven major shorelines.

Hoyt (1969) presented the first detailed depositional model for chenier genesis and mudflat progradation (Fig. 2a). According to Hoyt’s (1969) model, Chenier-Plain development reflects changes in Mississippi River flow direction that are caused by channel avulsion (see Aslan et al., 2005), which results in downstream delta switching along the coast, as illustrated in Fig. 2b and c (see Scruton, 1960; Coleman, 1988; Roberts, 1997, 1998 for details regarding the delta-switching process). The Hoyt model correlates mudflat progradation with periods of Mississippi River discharge to the west (Fig. 2a, b). When the river becomes hydraulically unstable and avulses to a shorter, more eastward course to achieve hydraulic efficiency, fine-grained sediment supply to the Chenier Plain is reduced significantly, enabling wave reworking and erosion of mudflats, thus forming cheniers (Fig. 2a, b). Hoyt’s hypothesis that the changing position of the Mississippi River is important to overall Chenier-Plain development is valid, but because it incorporates only regressive mudflats and transgressive cheniers, omitting ridges created by other means, his model oversimplifies Chenier-Plain evolution in SW Louisiana. This oversimplification has been challenged by subsequent researchers (Gosselink et al., 1979; Kaczorowski, 1979, 1980; Kaczorowski and Germant, 1980), who recognized that the geologic evolution of the Chenier Plain was more complicated than channel avulsions of the Mississippi River, involving not only chenier ridges (i.e., transgressive), but also ridges that are genetically tied to beach ridges, recurved spits, eolian deposits, storm berms, natural levees, and ancient oyster reefs. Subsequent reviews (e.g., Walker and Coleman, 1987; Penland and Suter, 1989) supported the original scientific findings of Gosselink et al. (1979) and Kaczorowski and Germant (1980) that the Chenier Plain contains more than true cheniers (i.e., transgressive ridges); however, a comprehensive understanding of Chenier-Plain evolution and a more-advanced geomorphic model that explains the genesis and distribution of different ridge types still are lacking.

The goal of this paper is to explain the geomorphic evolution of the SW Louisiana Chenier Plain. Specifically, the objectives are to (i) refine the definition of the term Chenier Plain; (ii) clarify what types of ridges are found in the Chenier Plain and map their spatial distribution; (iii) use the combination of shoreline-change data and tidal-inlet morphodynamics along the present Chenier-Plain coast as a modern analog by which to understand coastal processes and landforms along relict shorelines (ridges); (iv) identify global, regional, and local processes that are responsible for
Fig. 1. Geomorphology of the SW Louisiana Chenier Plain illustrating the orientation and distribution of ridges from Sabine Pass, Texas/Louisiana to Southwest Pass at Vermilion Bay, Louisiana (modified from Taylor et al., 1996).
ridge formation; and (v) present a six-stage geomorphic process-response model that synthesizes the geologic evolution of the Chenier Plain.

2. Regional setting

The Louisiana coast, which consists entirely of Holocene sediments deposited directly or indirectly by the Mississippi River, is naturally divided into two primary geomorphic zones: the deltaic plain of southeastern Louisiana and the Chenier Plain of SW Louisiana (Fisk, 1944; Gould and McFarlan, 1959; Roberts, 1997). Except for local progradational areas around the active mouths of the Mississippi and Atchafalaya Rivers, as well as short segments east and west of Sabine and Calcasieu Passes and along the easternmost portion of the Chenier Plain (i.e., the Freshwater Bayou area), the outer shorelines of the deltaic plain and Chenier Plain have been dominated by rapid shoreline retreat, land loss, and deterioration for more than 100 years (Morgan and Larimore, 1957; Morgan and Morgan, 1983; Byrnes et al., 1995b; McBride and Byrnes, 1995, 1997).

The modern beach, Holocene ridges, and estuaries, as well as saline, brackish, and fresh marshes, form the 5000-km² marginal-deltaic deposits and environments of SW Louisiana known as the Mississippi River Chenier Plain. During the Middle to Late Holocene, wind-driven currents transported fine-grained fluvial sediment westward from distributary plumes of the Mississippi River to the SW Louisiana coast, creating a 30-km-wide, low-lying coastal plain. Currently, the sediment plume of the Atchafalaya River is advected in a net-westerly direction and nourishes the easternmost shelf and shoreline of the Chenier Plain with clay and silt (Morgan et al., 1953; Wells and Kemp, 1981; Wells and Roberts, 1981; Kemp, 1986; Wells, 1986; Huh et al., 1991; Kineke et al., 2006).

Distinct ridges, most of which represent relict shoreline positions, are interspersed in the mud-dominated coastal depositional system. Ridges typically are oriented shore-parallel to subparallel, are 10 to 90 km long and <1 to 5 m thick, and are ∼1 km wide.

The oldest, most-landward ridge is ∼15 km from the modern Gulf shoreline. Ridge relief ranges from 0 to 4 m above mean sea level (MSL), and ridges generally are linear-to-concave with an asymmetrical profile and a steep seaward face. Some ridges have multiple crests and swales, but most ridges have smooth southern (seaward) sides and lobate to irregular northern (landward) boundaries. Ridges are most easily distinguished by the presence of live oak trees (Quercus virginiana) that are rooted in their well-drained, shelly sand deposits. Sediment that comprises the ridges and modern beaches ranges from well-sorted medium sand to coarse, shelly deposits with a fine-grained sand matrix (see Taylor et al., 1995; Byrnes et al., 1995a; Anderson et al., 1995; Taylor, 1996; McBride et al., 1997; Anderson et al., 1998 for discussions on ridge sedimentology and macrofauna).

Most ridges are bounded abruptly on the north and south by marsh deposits that range from 1 to 10 km in width. The northernmost marsh of the Holocene Chenier Plain is fresh and becomes brackish southward, in the region where ridges are found. Salt marsh only exists within a thin band along the modern shoreline (O’Neil, 1949; Chabreck and Linscombe, 1978; Visser et al., 2000).

The study area encompasses a low-energy, microtidal, storm-dominated environment that experiences episodic sediment supply (Wells and Roberts, 1981; Byrnes et al., 1995b). Mean spring tide is mainly diurnal, ranging from 0.6 to 0.8 m (Harris, 1981; NOAA, 1985a; Forbes, 1988). Dominant nearshore currents are to the west (Beall, 1968; Becker, 1972; USACE, 1974; Crout and Hamiter, 1981; NOAA, 1985b) and are controlled by winds and waves that are predominantly from the southeast (USACE, 1974; Hubbertz and Brooks, 1989). Gammill (2002) provides a detailed review of historical water levels and their impact on marsh evolution throughout the Chenier Plain. According to recent tide gage data, the average rate of relative sea-level rise for the Chenier Plain is 0.57 cm/year (Penland and Ramsey, 1990). Most of this change can be attributed to compactional subsidence of Holocene sediment.

3. Methods, landform definitions, and geomorphic hierarchy

A detailed geomorphic base map for the entire Louisiana Chenier Plain (Fig. 1) was needed because regional, geologic field and mapping studies of the area have been lacking since the Byrne et al. (1959) study. Also, high-quality, color infrared imagery datasets were collected in subsequent years, thus providing additional resolution and contrast for air-photo interpretation of form/process relationships. Specific datasets used to construct the geomorphic map include (i) NASA color infrared National High Altitude Aerial Photography (December 1990, scale 1:58,000) for mapping ridge morphology, distribution, and cross-cutting relationships; (ii) U.S. Geological Survey 7.5-min (1980) quadrangles for photograph rectification; and (iii) the Geologic Map of Louisiana (Snead and McCulloh, 1984, 1:500,000) for identifying the approximate subaerial location of the Pleistocene/Holocene boundary.
Fig. 2. (a) Depositional model explaining Chenier-Plain development through mudflat progradation, wave reworking and ridge development, followed by mudflat progradation, thus creating a chenier (from Hoyt, 1969). (b) Geologic model showing Chenier-Plain evolution in response to channel avulsions of the Mississippi River and the resultant, downstream delta-switching process along the coast (modified from McBride et al., 1997). (c) Holocene delta complexes of the Mississippi River illustrated chronologically from oldest (Maringouin/Sale-Cypremort) to youngest (Atchafalaya/Wax Lake) as based on Frazier (1967), Coleman (1988), and Roberts (1997). In order to gain a shorter, steeper course to the ocean (base level) and thus achieving hydraulic efficiency, the Mississippi River has experienced channel avulsions and attendant delta switching naturally about once every 1000 to 1500 years.
Various coastal geomorphic features and coastal river deflects often are used to infer processes and direction of net longshore drift (Davies, 1958; Allen, 1965; Johnson, 1965; Curay et al., 1967; Staton, 1975; Jacobsen and Schwartz, 1981). Using 1990 NASA photography, different coastal landforms were interpreted based on the presence or absence of truncations (i.e., cross-cutting relationships) and the configuration of ridges. This information was used together with the shoreline-change data quantified by Byrnes et al. (1995b) to infer process along paleoshorelines of the Chenier Plain.

In addition, 28 topographic profiles were constructed to delineate the morphology of relict ridges and modern beaches in more detail. (Twenty three of the 28 profiles are presented in this paper, as shown in Fig. 3.) Profiles were located wherever distinct changes in ridge and beach topography were identified. Using standard surveying equipment (theodolite, level, stadia rods) and methods, individual elevation measurements were collected along a line oriented perpendicular to the trend of a ridge or modern beach, thus creating a topographic profile. For each profile, elevation measurements were taken at regularly spaced intervals on flat terrain, but at shorter intervals when a break in slope occurred in order to define the topographic details of the geomorphic feature. All profiles were tied to the 1929 National Geodetic Vertical Datum (NGVD 1929), using known elevations in the field. Moreover, historical shoreline-change results from Byrnes et al. (1995b) were further evaluated to document past and present trends in shoreline movement and coastal dynamics along the Chenier-Plain coastline.

In this paper, we use the term “ridge” simply as a morphologic term defined as a narrow, linear or curvilinear topographic high (≤4.25 m in elevation) that consists of sand and shelly material accumulated by waves and other physical coastal processes. Under this definition, most ridges in the Chenier Plain represent relict shorelines. On the basis of past movement trends of individual shorelines, ridges can be further classified as transgressive, regressive, or laterally accreted. Curay (1964) defined a transgression as the landward migration of a shoreline, regardless of the forcing mechanism (e.g., sediment supply variations due to delta switching, relative sea-level changes, and storm impacts), whereas a regression is the seaward movement of a shoreline, regardless of the forcing mechanism. Consequently, and being consistent with Curay (1964), transgressive, regressive, and laterally accreted ridges are topographic features that were created through landward, seaward, and lateral movement of the shoreline, respectively. A transgressive ridge surrounded by littoral mud deposits also is known as a chenier, a regressive ridge is known as a beach ridge, and a laterally accreted ridge is known as a spit or a recurved spit.

In this paper, geomorphic zones that contain two or more regressive, transgressive, or laterally accreted ridges are termed complexes. A beach-ridge complex contains two or more parallel to subparallel, regressive ridges, and is net progradational, without extensive fine-grained interridge deposits. A chenier complex contains two or more parallel to subparallel, transgressive ridges that are separated by regressive, littoral mudflats (muddy units), and it is net progradational. A spit complex contains two or more laterally accreted ridges that primarily have prograded laterally along the downdrift end of ridges. Previously, a Chenier Plain was defined as containing two or more parallel to subparallel ridges (cheniers), separated by progradational littoral mudflats (Price, 1955; Otvos and Price, 1979). In this paper, however, we refine the definition by Otvos and Price (1979), defining a Chenier Plain as containing two or more chenier complexes. (In addition to the required chenier complexes, a Chenier Plain also may contain beach-ridge complexes and/or spit complexes.)

Based on these definitions, we developed a geomorphic hierarchy of landforms relative to dominant process for the Louisiana Chenier Plain (Table 1). The Chenier Plain is defined as a first-order feature (5000 km²), and it is a marginal-deltaic environment because of its genetic link to the Mississippi River. The Chenier Plain is composed of three types of smaller, second-order features (30 to 300 km²): chenier complex, beach-ridge complex, and spit complex. Individual ridges of each complex type are further separated into third-order features: chenier, beach ridge, and spit (Table 1).

4. Results: ridge chronology, origin, and distribution

4.1. Relative chronology of ridge trends (relict shorelines)

Fifteen shoreline trends (including the modern shoreline) are identified and presented in chronological order in Table 2. Although shorelines S1, S2, and S3 are the oldest ridges in the Chenier Plain, they are limited spatially and are considered minor shorelines (Fig. 1; Table 2). Because of its regional extent, the Little Chenier–Little Pecan Island trend (S4) is the oldest prominent shoreline trend in the area. Cypress Point and Fire Island may be eastward extensions of S4, as they are the most-landward ridges in
the Pecan Island area. High Island (S5) is part of a major shoreline trend whose possible extensions include the prongs off the western end of North Island, as well as those off the Back Ridge in the Pecan Island area (Sweet Bay Ridge, Cane Ridge, Lambert Ridge, and Coupe Ridge). Shoreline S6, which is composed of Back Ridge (east of Calcasieu River), Chenier Perdue, North Island, and Back Ridge (Pecan Island area), was the next major shoreline to form, and it may extend westward to include Smith Ridge and Buck Ridge near Johnsons Bayou. Creole Ridge, Pumpkin Ridge, Tiger Island, and possibly Kochs Ridge compose shoreline S7. Pumpkin Ridge truncates and coalesces with S6 as interchenier mudflat width decreases to the west. Shoreline S7 possibly continues to the west of the Calcasieu River as Sanders Ridge in the Sabine Pass area.

Happy Ridge and Hackberry Ridge (east of the Mermentau River) form shoreline S8. As shown in Fig. 4.1, ridges at this point are spaced more tightly, which could suggest a decrease in the supply of fine-grained sediment to the area. Eugene Island and Cow Island/Indian Point form shoreline S9, which appears to continue west as Cameron Ridge near the Calcasieu River. Dan’s Ridge (S10) is a long but thin chenier that extends between the Mermentau and Calcasieu Rivers. The Grand Chenier trend (S11) is by far the longest semicontinuous shoreline in the Chenier Plain, stretching more than 150 km in length and encompassing Front Ridge (west), Oak Grove Ridge, Grand Chenier, Long Island, Pecan Island, and Front Ridge East. Holly Beach, Peveto Beach, and Blue Buck Ridge possibly are the westward extension of S11. Ridges that formed just before the modern shoreline in the Freshwater Bayou area include Cheniere au Tigre (S12), Beef Ridge (S13), Sand Ridge (S14), and Mulberry Island (S14). Hackberry Beach and Mesquite Ridge (S14) form a relict shoreline closest to the modern beach (S15) at the old mouth of the Mermentau River. Shorelines S12 and S13 are considered minor shorelines because of their limited lateral extent.

4.2. Ridge origin and distribution

Ridges and modern beaches of the SW Louisiana Chenier Plain are classified as transgressive, regressive, or laterally accreted (Fig. 4). Evolution of the modern beach is similar over time; that is, local areas of transgression provide sediment for downdrift beach regression and lateral accretion at tidal entrances.

Ridge distribution and morphology are interpreted for two geomorphic zones. The western zone encompasses ridges between the Sabine and Calcasieu Rivers, and includes the ridge complex immediately west of Sabine Pass in Texas. The eastern zone includes regressive, transgressive, and laterally accreted ridge complexes that are between the Calcasieu River and Southwest Pass at Vermilion Bay.

4.2.1. Sabine River to Calcasieu River

The abandoned Calcasieu deltaic headland is responsible for the general lack of ridges between Constance Beach and Calcasieu Pass, as Fig. 1 illustrates (see Fisk, 1948; Penland and Suter, 1989). Ridges in the western geomorphic zone are concentrated on either side of the Sabine River in two regressive-ridge and laterally accreted-ridge complexes and two updift, transgressive segments (Constance/Peveto Beach area and east of Holly Beach) that are past and present sources of sediment for downdrift accretion at Sabine Pass and other smaller entrances (e.g., Hamilton River and Johnsons Bayou; Figs. 4 and 5). The local pattern of updift erosion and downdrift accretion along the modern shoreline between the Sabine and Calcasieu Rivers (Byrnes et al., 1995b) also is evident landward along relict shorelines (Fig. 4).

Fig. 5 illustrates that the ridge complex east of Sabine Pass contains 20 to 25 closely spaced beach ridges (2 m above NGVD 1929) and spits that diverge from Blue Buck Ridge. Ridges curve toward the NW at Hamilton Lake (a relict Calcasieu River channel) and Johnsons Bayou (a relict Sabine River channel), as explained by Russell and Howe (1935). Lateral migration of Blue Buck Ridge forced westward migration of Hamilton Lake (Fisk, 1948); however, the southernmost ridge of the Blue Buck trend, immediately north of Salt Work Ridge, clearly is transgressive because it truncates ridge complexes to the north (Figs. 4 and 5). South of Salt Work and Hackberry Ridges, another 25 topographically low, closely spaced ridges (0.6 to 1.4 m above NGVD 1929) without truncations trend toward the SW, a result of net-seaward sediment deposition and interpreted as beach ridges by Kaczorowski and Gernant (1980). Buck and Smith Ridges form a distinct ridge trend north of the Blue Buck Ridge, and are separated by Johnsons Bayou (Figs. 1 and 5). A profile across Smith Ridge (Figs. 5 and 6, profile 18) and aerial photography show a transgressive, single-crested ridge that is unlike the regressive and laterally accreted ridge complex immediately to its south.

For the period 1883 to 1994, the modern shoreline between the Sabine and Calcasieu Rivers experienced net accretion in the west, but retreated or remained stable in the east (Byrnes et al., 1995a; McBride and Byrnes, 1995). The shoreline between Holly Beach and Ocean View Beach retreated at an average rate of 1.2 m/year (Fig. 7a) and truncated the eastern ends of Blue
Fig. 3. Map showing the location of topographic profiles (numbered arrows) that were collected at various sites across paleoshorelines and modern beaches throughout the Louisiana Chenier Plain.
West of Ocean View Beach, the outer shoreline advanced at 3.5 m/year due to updrift sediment erosion, lateral sediment transport, and deposition (Byrnes et al., 1995b), as shown in Fig. 7a. Westward divergence and curvature of ridges formed east of Sabine Pass indicate that sediment was transported predominantly westward during their construction.

The formation of regressive ridges immediately west of the Sabine and Calcasieu Rivers (Figs. 1 and 4)
indicates local reversals in net westward sediment transport (Byrnes et al., 1995b; Taylor, 1996). Ridges west of Sabine Pass (in Texas) diverge rapidly to the east and form three distinctly regressive and laterally accreted ridge complexes. The two northerly complexes, separated by 2 km of marsh, each contain 20 to 25 ridges with elevations of ~1.5 m. Ridges to the south are topographically lower (up to 0.9 m) and likely represent rapid sediment deposition as beach ridges. The modern shoreline west of Sabine Pass, contrary to the overall trend of regression, has retreated >2 m/year over the past 100 years, except for a short, 2-km accretionary stretch immediately west of the western jetty at Texas Point (see Morton, 1975; Westphal et al., 1996).

4.2.2. Calcasieu River to Southwest Pass

The eastern geomorphic zone between the Calcasieu River and Southwest Pass extends for 120 km (Figs. 1 and 4). Overall, modern shoreline-change trends mimic relict shoreline (ridge) configurations in this zone. Individual ridges are up to 250 m wide and reach elevations of 3.4 m NGVD 1929 (Taylor et al., 1996). Interridge mudflats in the eastern zone are up to 10 km wide, but narrow to the west. Cheniers separated by wide mudflat deposits are common; however, complexes with laterally accreted and regressive ridges also exist, indicating widespread local variability in depositional processes (Fig. 4). East of the Calcasieu River, ridges diverge and are separated by mudflats that are several kilometers wide. The Mermentau River flows through the center of the eastern zone and has been deflected 25 km to the west by the lateral growth of ridges (Fig. 4). The Little Chenier–Little Pecan Island trend (S4) constitutes one of the oldest paleoshorelines in the area and may continue east as Cypress Point and Fire Island (Figs. 1 and 4, Table 2). This ridge trend has similar morphology to the rapidly transgressing modern shoreline. Little Chenier is dominated by a narrow, single-crested transgressive deposit (Figs. 8 and 9, profiles 19 and 20), as well as minor areas and deposits of lateral accretion and regression (see Zenero et al., 1995; Taylor et al., 1996). Multiple, curved ridges typical of environments that have experienced lateral growth (Figs. 8 and 9, profiles 21 and 22) are found at the eastern end of Little Chenier, near the Mermentau River. To the east, transgressive and laterally accreted ridges characterize Little Pecan Island (Figs. 1 and 4). The central section of the ridge is transgressive and truncates older ridges diverging to the west that have a more northwesterly orientation (Figs. 4, 8, and 10, profile 25). West and east (e.g., Long Island, High Island, and Little Islands) of its central section, Little Pecan Island is extremely low-lying and narrow (Figs. 8 and 10, profiles 23 and 27). These single-crested, narrow, washover-dominated ridges were formed during transgression (Fig. 4).

Several kilometers of marsh separate the Little Chenier–Little Pecan Island trend from Chenier Perdue to its south. Near the Mermentau River, Chenier Perdue is a single, narrow, well-defined transgressive ridge (Figs. 1, 4, and 11, profile 17). About 5 km from its eastern end, the ridge begins to bifurcate to the west into a series of wider, lower, and less distinct ridges (Figs. 1 and 11, profile 15). The bifurcations indicate net westward longshore sediment transport.

Pumpkin Ridge, a low (<1 m above NGVD 1929) transgressive ridge with lobate washover fans along its northern (landward) margin, coalesces with Chenier Perdue a few kilometers east of Creole Ridge (Figs. 1 and 4). Interridge mudflat width in the eastern geomorphic zone decreases to the west, and ridges near the Calcasieu River coalesce, forming a 5-km-wide area of regression (Fig. 4). Tiger Island, Hackberry Ridge, and Cow Island/Indian Point (all laterally accreted ridges truncated at their eastern ends by the Grand Chenier trend) are eastward extensions of Pumpkin Ridge, Happy Ridge, and Eugene Island, respectively (transgressive ridges; see Zenero et al., 1995), as shown on Figs. 1 and 4.

Ridges in the Pecan Island area may be eastward extensions of ridges to their west, but direct physiographic correlation is difficult without precise dates (see Gould and McFarlan, 1959, for radiocarbon dates). For instance, North Island and Back Ridge (to the west and east of the Pecan Island area, respectively) appear to be a continuation of the Chenier Perdue shoreline trend. North Island and Back Ridge (Figs. 3 and 11, profile 28) are both transgressive; each shoreline truncates at least six closely spaced laterally accreted ridges directly to the north (Figs. 1 and 4).

The Grand Chenier trend [which includes Front Ridge (west), Oak Grove Ridge, Grand Chenier, Long Island, Pecan Island, and Front Ridge East] is the most laterally continuous ridge on the Chenier Plain, traceable for more than 80 km, although it may extend farther west as Blue Buck Ridge (Fig. 1). This regional ridge trend contains regressive, transgressive, and laterally accreted deposits, making it analogous to the modern shoreline (Fig. 7a, b). Front Ridge East, the easternmost ridge section of the Grand Chenier trend, is transgressive and truncates approximately 20 paleoshorelines (Figs. 12a and 13, profile 13). Farther west, Front Ridge East becomes a very narrow transgressive ridge (Figs. 12a and 13, profile 12). Between Pecan Island and the eastern end of Grand Chenier, the ridge is indistinct and unidentifiable on
Fig. 4. Distribution of transgressive, regressive, and laterally accreted ridges along paleoshorelines and modern coast of the SW Louisiana Chenier Plain.
Fig. 5. Chenier-Plain geomorphology between Holly Beach, Louisiana, and a location about 10 km west of Sabine Pass, Texas–Louisiana. Profile 18 is located on Smith Ridge, Louisiana.
aerial photographs for about a 5-km stretch (Fig. 1); however, 1930s black-and-white aerial photographs reveal that the western end of Pecan Island curved to the NW at Floating Turf Bayou, indicating that this bayou was an inlet during ridge formation (Fig. 1).

A prominent laterally accreted ridge complex on the Grand Chenier trend exists east of the Mermentau River (Figs. 8 and 12a, profile 10), for which sediment was provided by easterly updrift areas of retreat and ridge truncation (e.g., Cow, Tiger, and North Islands) (Figs. 1 and 4). A similar scenario occurs along the modern shoreline in the Hackberry/Rutherford Beach area, where truncation of relict laterally accreted ridge deposits at Hackberry Beach (beach scarps) (Figs. 8 and 14, profiles 3 and 4) and long-term shoreline erosion (Byrnes et al., 1995b) provide sediment for downdrift shoreline advance at Rutherford Beach, indicated by multiple regressive ridges (Fig. 14, profiles 1 and 2).

Fig. 6. Topographic profile 18 across Smith Ridge, Louisiana, showing a low, single-crested ridge indicating transgression through overwash processes (from Taylor, 1996). See Figs. 3 and 5 for profile location.

Fig. 7. Net shoreline change along the entire modern shoreline of the Louisiana Chenier Plain (from Byrnes et al., 1995a). (a) Sabine Pass to Calcasieu Pass, from 1883 to 1994. (b) Calcasieu Pass to Southwest Pass, from 1883/86 to 1994. The change rates shown on the plot below each map correspond to the graphic representations of advance and retreat. Values along the bottom inside of the plot indicate average rates of change in m/year for designated local sections of coast.
Topographic profiles across the Grand Chenier trend west of the Mermentau River reveal a simple, single-crested, transgressive ridge (Figs. 8 and 12a, profile 8). Evident at the western end of the Grand Chenier trend [Front Ridge (west) area] is a 5-km-wide regressive and laterally accreted ridge complex (Figs. 1 and 4). This regressive deposit represents the terminus of longshore sediment transport, where the Calcasieu River has interrupted littoral transport and caused long-term deposition. Accumulation here of 32 closely spaced ridges indicates long-term advance, which is consistent with historical shoreline-change data presented by Byrnes et al. (1995b), as shown in Fig. 7b. Regressive morphology (e.g., lack of truncations) in the Front Ridge (west) area (Figs. 8 and 12b, profile 5) is comparable to that observed along the modern shoreline (Figs. 8 and 14, profile 1). Sediment for continued advance of the modern ridge complex comes from local updrift areas of erosion (Fig. 4). The well-preserved, shoreface-derived shell assemblage and the high concentration of quartz sand also support the interpretation of Front Ridge (west) as a regressive setting (Anderson et al., 1998).

Cheniere au Tigre and Mulberry Island/Beef Ridge/Sand Ridge represent shorelines that were formed after significant mudflat progradation south of the Grand Chenier trend. The shore-normal orientation of Cheniere au Tigre and Belle Isle indicates that they may be relict Vermilion Bay shorelines or former oyster reefs that were oriented along drowned natural levees of abandoned distributary channels, like those observed now in Vermilion Bay and along the southern shoreline of Marsh Island (see Coleman, 1966; Taylor, 1996).

5. Coastal morphodynamics and Chenier-Plain evolution

To understand the long-term evolution of a coastal depositional system, primary process–response mechanisms and patterns found along the modern coast must first be identified. As such, present-day coastal processes operating along the outer shoreline of the Chenier Plain are synthesized. These processes play a central role in providing new insight in the construction of a geomorphic model for Chenier-Plain evolution.
Fig. 8. Chenier-Plain geomorphology between Calcasieu Pass and Little Constance Bayou, Louisiana. Specific topographic profile locations (numbered arrows) are shown along paleoshorelines and the modern beach for subsequent figures.
5.1. Tidal-entrance dynamics and ridge occurrence

Inlet processes associated with the Sabine, Calcasieu, and Mermentau Rivers are important factors in the development, morphology, and distribution of ridges on the Chenier Plain. The inlet stability criterion of Bruun (1967) and Bruun et al. (1978) is used in this study to examine the stability and dynamic diversion (Todd, 1968) potential of three primary tidal entrances along the Chenier Plain, and to explain the distribution of different ridge types.

Tidal prism ($\Omega$) and the quantity of littoral sediment transport ($M_{total}$) are the most important factors that control inlet stability (Bruun et al., 1978). Greater discharge and/or tidal prism increase the potential for river and estuarine systems to (i) interrupt longshore sediment transport; (ii) maintain and naturally stabilize tidal entrances; and (iii) promote updrift deposition (Bruun, 1967; Todd, 1968). In characterizing inlet stability, Bruun (1967) assigned a stability ratio number (ratio of tidal prism to longshore sediment transport quantity, or $\Omega/M_{total}$) to tidal entrances around the world. According to Bruun et al. (1978), for stable inlets, $\Omega/M_{total}$ is >150; for moderately stable inlets it ranges between 100 and 150; and for unstable inlets it is <100. As such, unstable and moderately stable inlets will migrate laterally, whereas stable inlets are much more stationary and are classified as tide-dominated (see Nummedal et al., 1977; Nummedal and Fischer, 1978; Davis and Hayes, 1984; McBride, 1999).

Table 3 summarizes river, estuary, and inlet characteristics along the Chenier Plain and reports the $\Omega/M_{total}$ for each tidal entrance. Along the Chenier Plain, maximum river discharge ($Q_{max}$) from each of the three primary rivers increases westward, from 32 to 234 m$^3$/s. Suspended sediment discharge (SSQ) also increases, from 0.19 to $0.29 \times 10^9$ kg/year, as does tidal prism ($\Omega$) of Chenier-Plain estuaries, from 0.94 to $4.10 \times 10^7$ m$^3$. Longshore sediment transport ($M_{total}$) is about the same along the entire modern Chenier-Plain shoreline, at 0.62 to $1.0 \times 10^5$ m$^3$/year (Bruun, 1967; USACE, 1971), as is average significant wave height ($H_s = 1.0$ m), wave approach (157°), and tidal range (0.61 to 0.76 m).

Sabine Pass, which has good inlet stability ($\Omega/M_{total} = 410$), has effectively interrupted longshore sediment transport, causing long-term shoreline advance immediately west and east of the inlet (Figs. 1 and 4; Table 3). Calcasieu Pass has not always been as stable as the present channel. To the west, relict deflected channels of the Calcasieu River, such as Hamilton Lake and spits curving into this channel, indicate past periods of inlet instability that allowed longshore sediment transport to dominate fluvial and tidal processes. A regressive ridge complex adjacent to the present Calcasieu Pass denotes...
increased stability. Although present discharge of the Calcasieu River is much less than that of the Sabine River (73 vs. 234 m^3/s), tidal prism is of a similar volume and accounts for good inlet stability (Ω/M_total=390) over the past 1 kyear or so, judging from the accumulation of regressive ridges south of the Grand Chenier ridge trend (Figs. 1 and 4; Table 3). The Grand Chenier trend has been radiocarbon dated at about 1.1 kyears according to Gould and McFarlan, 1959.

Historically, the Mermentau River has been characterized as relatively unstable (Ω/M_total =94). Although longshore sediment transport was not great enough to fill the river mouth naturally, it was sufficient to dominate fluvial discharge (32 m^3/s) and tidal exchange (9.4×10^6 m^3), forcing westward migration of the entrance and lateral shoreline accretion (Figs. 1 and 4; Table 3). Long-term instability of the Mermentau River during the development of most of the Chenier Plain is indicated by westward movement of the inlet of up to 25 km and associated depositional features (e.g., Indian Point, Cow Island, and Hackberry Beach). On the other hand, the mouth of the Mermentau River appears to have had a period of greater stability during an earlier time interval, between the development of Little Chenier (~2.8 kyears ago) and that of Pumpkin Ridge (~2.1 kyears ago), when its orientation was more shore-normal (Fig. 1). Thus, prior to human modification and stabilization efforts, the Mermentau River entrance would be classified as wave-dominated, the Sabine River entrance as tide-dominated, and the Calcasieu River entrance as tide-dominated to occasionally mixed.

5.2. Geomorphic model of Chenier-Plain evolution

Chenier-Plain evolution is described in terms of shoreline response to the interaction between longshore sediment transport and tidal prism of local estuaries, delivery of fine-grained sediment from a distant source, and local reworking of sediment (Fig. 15). Delta
switching of the Mississippi River, caused by upstream channel avulsion, is a well-documented and important province-scale process for Chenier-Plain evolution (i.e., fine-grained sediment delivery and wave reworking) (Fig. 2b, c). However, the concentration of coarser-grained deposits and the overall response of ridges involve lateral variations in processes and landforms that are not explained fully by simple delta switching as discussed by Russell and Howe (1935), Fisk (1948), Gould and McFarlan (1959), Hoyt (1969), and Penland and Suter (1989).

As such, a six-stage geomorphic process-response model is presented that explains Chenier-Plain evolution primarily as a function of: (i) the balance between sediment supply ($Q_s$) and energy dissipation ($E$), (ii) local sediment reworking and lateral transport, and (iii) tidal-entrance dynamics (Fig. 15). When the Mississippi River (primary sediment source) occupies a westerly position near the Chenier Plain (Fig. 2b), westward-transported, fine-grained sediment is deposited along the Chenier-Plain shoreline, resulting in mudflat advance where $Q_s > E$ (Fig. 15, stage 1). When the Mississippi River channel avulses to the east, fine-grained sediment delivery to the Chenier Plain is reduced, enabling wave energy to dominate ($Q_s < E$) and transgression to occur, thus concentrating coarser material into ridges (Figs. 2b and 15, stage 2). Suppose that two inlets exist in this model, where the eastern inlet is wave-dominated and unstable ($\Omega/M_{total} < 150$) while the western inlet is tide-dominated and stable ($\Omega/M_{total} > 150$). As the shoreline recedes in some localities, creating cheniers, sediment is supplied downdrift and is trapped by way of lateral accretion (recurved spits) at the wave-dominated inlet and by net shoreline advance (beach ridges) farther west, where inlet processes dominate longshore sediment transport at the tide-dominated inlet. Thus, the geneses of three different ridge types can occur contemporaneously along the same shoreline, but at different locations.

![Fig. 13. Chenier-Plain geomorphology between Pigeon Bayou and Cheniere au Tigre, Louisiana. Locations of profiles 12, 13, and 28 are shown.](image)

As noted by Taylor (1996), Chenier-Plain environments are characterized by an array of sedimentary features that are unique to the region. These features include mudflats, spits, and ridges, all of which are the result of complex interactions between the Mississippi River and the coastal environment. The sediment is delivered to the Chenier Plain by the river, and the resulting sedimentary deposits are shaped by the dynamic coastal processes.

![Fig. 14. Topographic profiles across the modern Chenier-Plain outer shoreline (modified from Taylor, 1996). Profile 1 is of a regressive ridge west of Rutherford Beach. Sediment is supplied from updrift erosion at Hackberry Beach and ridges farther east (profiles 3 and 4). Profile 2 crosses a laterally accreted ridge complex at Rutherford Beach (Old Mermentau River mouth). Profile 4 is a perched transgressive ridge east of Hackberry Beach. See Figs. 3 and 8 for profile locations.](image)
During stage 3, the Mississippi River avulses back to the west ($Q_s > E$), and mudflats advance seaward of the ridges that were formed during stage 2 (Fig. 15). When the Mississippi River switches back to the east (Fig. 15, stage 4), fine-grained sediment supply is reduced, and transgressive, regressive, and laterally accreted ridges again form contemporaneously. The unstable, wave-dominated inlet migrates farther west, while the ridge accretes laterally with sediment from local updrift erosion and ridge truncation (Fig. 15, stage 4). The area around the stable, tide-dominated inlet advances farther seaward.

During stage 5, the Mississippi River occupies a westerly position, and mudflat growth seaward of the ridges is renewed; however, fine-grained deposition does not occur along the entire Chenier-Plain shoreline. This permits further redistribution of ridge sediment and continued seaward advance of the shoreline that is adjacent to the stable inlet. During stage 6, fine-grained sediment supply to the Chenier Plain is reduced again when the Mississippi River switches to the east (Fig. 15, stage 6). Local transgression is more extensive, which allows the active shoreline to truncate older ridge deposits, thus releasing trapped sediment into the littoral transport system for lateral accretion to the spit complex updrift of the unstable inlet. The regressive ridge complex updrift of the stable inlet continues to advance seaward with sediment from local updrift erosion.

Regressive and laterally accreted ridge complexes, and transgressive ridges separated by mudflats, are evident through time, although previous researchers who investigated Chenier-Plain development only incorporated prograding mudflats and transgressive ridges in their discussions and depositional models (e.g., Hoyt, 1969). Several laterally accreted ridge complexes form over time at unstable inlets ($\Omega/M_{total} < 150$), whereas regressive ridge complexes form at stable inlets ($\Omega/M_{total} > 150$).

6. Discussion

The geomorphic process-response model presented in this paper explains Chenier-Plain evolution at province scale and clarifies several important processes that are responsible for the genesis of different ridge types, such as local updrift erosion and downdrift sediment trapping at tidal entrances. Remaining critical issues entail the exact sources of sand and shell that compose ridge deposits and the forcing mechanisms responsible for Chenier-Plain evolution.

6.1. Littoral drift direction and sources of sand and shell

Overall, net direction of longshore sediment transport during evolution of the Chenier Plain has been from east to west, although local reversals exist today and have occurred in the past. The presence of recurved spits on the eastern side of the Mermentau River and of regressive ridge complexes updrift (east) of the Calcasieu and Sabine entrances indicates that transported sediment is predominantly from easterly sources. An overall decrease in width of interridge marsh deposits from east to west also suggests an easterly sediment source and westward transport for fine-grained deposits. Furthermore, westward deflection of the Mermentau River by 25 km is evidence of net long-term westward sediment drift (Fisk, 1948; Taylor et al., 1996). However, ridge complexes west of Sabine and Calcasieu Passes indicate local longshore sediment-
Fig. 15. Six-stage geomorphic process-response model illustrating evolution of the southwestern Louisiana Chenier Plain, where the genuses of three different ridge types (i.e., transgressive, regressive, and laterally-accreted) typically occur contemporaneously along the same shoreline at different locations (modified and expanded from Taylor, 1996). The tidal-prism/longshore-sediment-transport ratio ($\Omega/M_{\text{total}}$) determines inlet stability and influences ridge response. Regressive ridge complexes form adjacent to stable inlets ($\Omega/M_{\text{total}} > 150$) and laterally accreted ridge complexes form updrift of unstable inlets ($\Omega/M_{\text{total}} < 150$), whereas transgressive ridges form updrift of both types of ridge complexes through local shoreline erosion. Sediment for regressive and laterally accreted ridges and complexes comes from truncation of updrift relict ridges and from sand winnowed from mudflat deposits. Fine-grained sediment for mudflat advance is provided when the Mississippi River delta switches to the west in response to upstream channel avulsion. The four dots in each stage represent the same position through time.
transport reversals. Also, NE-oriented ridges along the eastern end of Little Chenier near the Mermentau River and associated eastward deflection of the Mermentau River suggest periods of east-directed longshore sediment transport.

Although net westward movement of sediment along the Chenier Plain and supply of mud to the Chenier Plain are well-documented, the exact source of sand and shell composing isolated ridges within a mud-dominated coastal depositional system needs further discussion. Most authors report that coarser material is winnowed from eroding mudflats and is concentrated to form sandy/shelly beaches (Hoyt, 1969; Otvos and Price, 1979), but presently, only clay and fine silt (2 to 6 μm) are transported from the Atchafalaya River mouth and Wax Lake Outlet to the eastern end of the Chenier Plain, shelf and shoreline near Cheniere au Tigre (Roberts et al., 1980; Wells and Kemp, 1981; Wells and Roberts, 1981; Kemp, 1986; Wells, 1986; Kineke et al., 2006). Some fine sand could have been reworked from mudflat deposits; vibracores taken from interridge mudflats do contain small amounts (∼4%) of fine sand (McBride et al., 1997); however, the ultimate source of sand and how it initially was incorporated into the Chenier Plain have yet to be fully explained.

The majority of sand likely was supplied to the Chenier Plain via three mechanisms. The first mechanism involved shoreface erosion of sandy Pleistocene deposits (fluvio-deltaic sediments of the Prairie Complex) during the last postglacial rise in eustatic sea level, when the open-Gulf shoreline transgressed across an exposed shelf from ∼16 to ∼4 kya (Anderson et al., 2004) before Chenier-Plain development. The open-Gulf shoreline was characterized by some type of barrier-beach/barrier-island system that scoured (shoreface ravinement) the upper portion of underlying Pleistocene deposits through erosional shoreface retreat, and incorporated some of the released sand into retreating coastal deposits (e.g., washover fans). As this transgressive barrier shoreline (i.e., transgressive ridge) experienced the apex of the Holocene highstand in sea level and/or an increase in sediment supply, and thus ceased its landward migration, the Little Chenier/Little Pecan Island shoreline trend (S4) was created (Table 2).

For the second mechanism, distributary sand and coarser delta-front sediment possibly were delivered directly to the eastern Chenier Plain when the Mississippi River deposited the Teche–Maringouin delta complex (Frazier, 1967), as indicated by the location and sandy texture of Trinity and Tiger shoals on the Louisiana inner continental shelf (see Fig. 2c), just south of Cheniere au Tigre (see Beall, 1968; Penland et al., 1989; Pope et al., 1991). However, in light of Gould and Morgan’s (1962) work regarding the Maringouin delta complex being deposited at a lower stand of Holocene sea level (i.e., stacked delta-plain concept (see Table 4)), it is more likely that coarser sediment was delivered directly to the Chenier Plain when the Mississippi River had switched to the early Lafourche delta lobe of the modern delta plain. Coarse material that was deposited and transported westward subsequently may have been reworked many times, with the admixture of locally supplied shells (Anderson et al., 1998) and sand that were winnowed from mudflat deposits to form distinct ridge trends interspersed in this mud-dominated system (Fig. 1; Table 2). The third mechanism involves small amounts of coarser sediment on the inner continental shelf that are transported onshore during storms (cold fronts, tropical storms, and hurricanes) and are preserved as part of storm surge overwash deposits and mudflats along the coast (see Huh et al., 1991; Draut et al., 2005).

Local rivers generally are not considered a primary source of sand-sized sediment. Most sediment carried by the Sabine, Calcasieu, and Mermentau Rivers is deposited in the upper reaches of the estuaries (see Barrett, 1971). Nichol et al. (1992, 1996) suggested that the Calcasieu River is not capable of transporting sand-sized sediment the length of the estuary and into the nearshore zone. For example, Holocene fluviodeltaic deposits (i.e., bayhead deltas) terminate 31 km from the modern outer shoreline in the middle of Calcasieu Lake.

6.2. Forcing mechanisms that control ridge and Chenier-Plain formation and timing

Clastic terrigenous shorelines respond primarily to sediment supply, intensity of coastal processes, and relative changes in sea level (Curray, 1964). Although the geomorphic model presented here explains different ridge types and their distribution within ridge complexes, the autycyclicity of the Mississippi River remains the sole forcing mechanism, driving ridge formation and Chenier-Plain development through fluctuating sediment supply and attendant incident processes. However, other forcing mechanisms have not been addressed in detail. Thus, to put the geomorphic model into context, the discussion below focuses on the progression of key concepts regarding Chenier-Plain development and on the possible roles played by Holocene climatic oscillations, storm impacts, and sea-level changes.

6.2.1. Climatic oscillations and storm impacts

Minor climatic oscillations and storm frequency have been postulated as possible forcing mechanisms for
Research results provided by different authors over the years regarding Holocene sea-level position during Mississippi River delta switching:

<table>
<thead>
<tr>
<th>Author(s)</th>
<th>Year(s)</th>
<th>Key Findings</th>
</tr>
</thead>
<tbody>
<tr>
<td>Fisk (1948)</td>
<td></td>
<td>Present stand of Holocene coast was eroding due to Mississippi River switching.</td>
</tr>
<tr>
<td>Gould and McFarlan</td>
<td></td>
<td>Same as Gould and McFarlan (1959)</td>
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<tr>
<td>Gould and Morgan</td>
<td></td>
<td>Same as Gould and McFarlan (1959)</td>
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<tr>
<td>Penland and Suter</td>
<td></td>
<td>Same as Gould and McFarlan (1959)</td>
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<tr>
<td>McFarlan (1959)</td>
<td></td>
<td>Abandonment of Teche delta complex and eastward switch to Lafourche delta complex.</td>
</tr>
<tr>
<td>Kaczorowski and Gernant</td>
<td></td>
<td>Same as Gould and McFarlan (1959)</td>
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<tr>
<td>McFarlan (1959)</td>
<td></td>
<td>Abandonment of Lafourche delta complex and westward switch to Pecan Island trend.</td>
</tr>
<tr>
<td>Gould and Morgan</td>
<td></td>
<td>Same as Gould and McFarlan (1959)</td>
</tr>
<tr>
<td>McFarlan (1959)</td>
<td></td>
<td>Abandonment of Lafourche delta complex and eastward switch to Pecan Island trend.</td>
</tr>
<tr>
<td>McBride et al. (this study)</td>
<td></td>
<td>Same as Gould and McFarlan (1959)</td>
</tr>
<tr>
<td>McBride et al. (this study)</td>
<td></td>
<td>Elaborates tidal inlet morphology, but does not address tidal inlet morphodynamics.</td>
</tr>
</tbody>
</table>

Note: The table above summarizes key findings from various researchers over the past 70 years. The inlet stability ratio (Ω/M) is a critical indicator of tidal inlet morphodynamics. Higher values of Ω/M indicate more stable tidal inlets. The table also highlights the presence of different ridge types, such as lateral-accreted ridges and transgressive-regressive ridges, and their role in shaping the coastline.
6.2.2. Sea-level changes and progression of key concepts regarding Chenier-Plain formation

Regarding global sea level, the literature currently contains two primary scenarios that describe eustatic sea-level changes during the Middle-to-Late Holocene. Today, cold fronts and tropical storms produce elevated local water levels of up to 2 m, 20 to 30 times a year, and hurricane surges cause 2- to 4-m elevations approximately every seven years (Morgan et al., 1958; Chamberlain, 1959; Nummedal, 1982). Along sandy beaches of the Chenier Plain, storms play a significant role in shoreline erosion, cross-shore transport of sand and shell, and back-marsh deposition through overwash processes, whereas muddy beaches tend to prograde during storms because fluid mud is accreted onto the shoreface and foreshore and/or may be deposited as muddy washover fans (Morgan et al., 1958; Kemp, 1986; Huh et al., 1991). However, variation in storm frequency or intensity caused by minor climatic changes plays no significant role in ridge construction because topographic highs along the beach (e.g., berm crests and colian dunes) are flattened, not built, by storm processes (Morgan et al., 1958; Stone et al., 1999; Morton et al., 2000). Hence, elevated water levels associated with storm impacts during warmer periods of the Holocene are not responsible for primary ridge genesis in the Chenier Plain. Shore-parallel ridges along the active beach (e.g., beach ridges) are constructed slowly, by multiple depositional events and by wave run-up from swell, rather than from storm waves (see Taylor and Stone, 1996; Blum et al., 2001).

6.2.2.1. Fisk (1948). Fisk (1948) postulated that the delta-switching process of the Mississippi River was responsible for Chenier-Plain development and occurred entirely during the past 5000 years, when Holocene sea level reached a stillstand at present levels (see Table 4). He identified three primary transgressive shorelines (initial, second transgression, and present shoreline) and two general periods of mudflat progradation (i.e., outgrowth of marshlands). His initial shoreline delineated the most landward position of the last postglacial rise in eustatic sea level and was recognized by Fisk (1948) as the Little Chenier–Little Pecan Island trend (Fisk also included Lost Island north of Pecan Island and Belle Isle). Sediment and marsh that are found landward of the Little Chenier–Little Pecan Island trend aggraded vertically during and after the last stages of the postglacial rise in sea level.

Fisk (1948) identified the western position of the Teche delta complex as being responsible for discharging sediment directly to the Chenier Plain (Fig. 2c), thus causing the first period of mudflat progradation seaward of the Little Chenier–Little Pecan Island trend. When the Mississippi River abandoned the Teche delta complex and shifted eastward to the Lafourche delta complex, diminishing sediment supply to the Chenier Plain, the second major transgression occurred, which created the Grand Chenier trend [Note: In 1948, the chronological details regarding the Mississippi Delta were still emerging, and Fisk incorrectly stated that the Mississippi River shifted to the Lafourche delta position, instead of to the St. Bernard delta position]. Although Fisk recognized the second period of mudflat progradation seaward of the Grand Chenier trend, he had difficulty explaining it. Instead of employing the delta-switching process directly, he suggested that fine-grained sediments may have been sourced from deterioration of pre-existing deltas (e.g., Teche) and the enlargement of Vermilion and West Cote Blanche Bays, as well as from the erosion of Marsh Island, and transported westward. As the Mississippi River shifted to its present course (Birdfoot delta), erosion of the Chenier Plain outer shoreline was renewed, as Fisk observed in the 1940s.

6.2.2.2. Gould and McFarlan (1959). Although Gould and McFarlan (1959) agreed that the shifting course of the Mississippi River at the present stand of sea level was responsible for Chenier-Plain development, they proposed a model that differed from Fisk (1948) in certain details: (i) on the basis of radiocarbon dating, they placed the
beginning of the Holocene standstill in sea level at 3 kyepm, instead of at 5 kyepm, and (ii) they proposed a different origin for the marsh-covered mudflats that are landward of the Little Chenier–Little Pecan Island trend. Additionally, the chronology of the Mississippi River delta complexes had been further refined by the late 1950s (e.g., McIntire, 1954; Van Lopik, 1955), which influenced Chenier-Plain investigations.

According to Gould and McFarlan (1959), marsh-capped sediment landward of the Little Chenier–Little Pecan Island trend was deposited in response to the rapid influx of mud from the Teche delta complex that is located along the western margin of the deltaic plain (Fig. 2c). Consequently, they interpreted the contact between the Pleistocene Prairie Complex and the Holocene marshes as representing the Holocene highstand open-Gulf shoreline (Table 4), rather than the Holocene bay shoreline at the leading edge of the postglacial transgression, as Fisk (1948) presented.

When the Mississippi River subsequently abandoned the Teche delta complex and avulsed to the St. Bernard delta complex on the far eastern margin of the deltaic plain (Fig. 2c), the Chenier Plain was transgressed and the outer shoreline retreated landward and created the Little Chenier–Little Pecan Island trend. Although deposits between the Little Chenier–Little Pecan Island trend and the Grand Chenier trend are characterized by a series of alternating progradational mudflats and ridges, Gould and McFarlan (1959) were unable to correlate specific deltaic processes with geomorphic features on the Chenier Plain, except to mention that moderate amounts of sediment may have been discharged at certain times through Bayou Barataria (i.e., the western margin of the St. Bernard delta complex), which may have allowed sediment to reach the Chenier Plain episodically. The Grand Chenier trend was created when the course of the Mississippi River was on the extreme eastern margin of the St. Bernard delta complex. Extensive marsh-covered mudflats that are seaward of the Grand Chenier trend were deposited when the Mississippi River abandoned the St. Bernard delta complex and shifted westward to the Lafourche delta complex. Gould and McFarlan (1959) further explained that the general stability of the modern shoreline along the Chenier Plain occurred when the Mississippi River abandoned the Lafourche delta complex and shifted eastward to the modern Birdfoot delta, whereas the more recent rapid progradation along the eastern Chenier Plain was in response to the increased discharge down the Atchafalaya River as documented by the work of Morgan et al. (1953).

6.2.2.3. Gould and Morgan (1962). Although Gould and Morgan (1962) differed little with Gould and McFarlan (1959) regarding the timing and formation of the Chenier Plain, they documented for the first time that the Maringouin delta complex was part of an older Mississippi River delta plain (see Fig. 2c) that was deposited when Holocene sea level was lower than present. By laterally tracing eastward a widespread peat layer at a depth of ∼3.6 m in the eastern portion of the Chenier Plain that is dated at ∼4 kyepm, Gould and Morgan (1962) found that it stratigraphically overlapped buried Maringouin natural-levee deposits. This stratigraphic relationship signifies that the Maringouin delta complex was abandoned before sea level reached present-day levels and that it was not active during Chenier-Plain development. Previous work by Fisk (1948) and Gould and McFarlan (1959) had assumed that Chenier-Plain development was directly related to the delta-switching process among all the Mississippi River delta complexes (Maringouin/Sale-Cypremort, Teche, St. Bernard, Lafourche, Plaquemines-Balize [modern Birdfoot], and Atchafalaya/Wax Lake [see Fig. 2c]) at a common highstand of Holocene sea level (i.e., single Holocene delta-plain concept), whereas Gould and Morgan (1962) documented that delta switching occurred during at least two different stands of Holocene sea level (stacked Holocene delta-plain concept), and that Chenier-Plain development occurred during the latter stand (Table 4). Specifically, the St. Bernard, Lafourche, and Plaquemines-Balize [modern Birdfoot] delta complexes built the most recent Mississippi River delta plain, as presented in Fig. 1 of Gould and Morgan (1962).

6.2.2.4. Penland and Suter (1989). Penland and Suter (1989) continued with the stacked delta-plain concept as it was presented originally by Gould and Morgan (1962), and they labeled the two delta plains as Late Holocene and modern. The Late-Holocene delta plain included the Maringouin/Sale-Cypremort and Teche delta complexes (which possibly are the same delta complex), whereas the modern delta plain consisted of the St. Bernard, Lafourche, Plaquemines-Balize [modern Birdfoot], and Atchafalaya/Wax Lake delta complexes (see Fig. 2c). Penland and Suter (1989) agreed with Gould and McFarlan (1959) that the Chenier Plain is ≤3 kyepm old and that it developed at the present stand of Holocene sea level. Although Penland and Suter (1989) differed with Gould and McFarlan’s (1959) interpretation regarding (i) the Pleistocene–Holocene contact, (ii) the Holocene marsh that is landward of the Little Chenier–Little Pecan Island trend, and (iii) the Little Chenier–Little Pecan Island trend itself, Penland and Suter’s (1989) claim of
reinterpretation of these abovementioned features actually is identical to the concepts that Fisk (1948) originally presented more than 40 years earlier (Table 4). Most importantly, Penland and Suter (1989) linked the development of the Chenier Plain to delta-lobe switching in the Lafourche delta complex within the modern Mississippi River delta plain, which greatly narrowed the scope of deltaic processes that were operating updip while the Chenier Plain was being built.

6.2.2.5. Higher-than-present Holocene sea levels: a new perspective. Until Morton et al. (2000) and Blum et al. (2001, 2002) published their findings from the adjacent Texas and nearby Alabama coasts, overall development and geomorphic evolution of the Chenier Plain were commonly accepted to have been driven entirely by the autocyclicity of the Mississippi River (i.e., the delta-switching process and excess sediment supply) during a slowly rising or stable Holocene sea level. Their higher-than-present sea-level scenario provides a mechanism for reconsidering the primary factors in ridge development and net progradation of the Chenier Plain. The underlying question is whether ridge formation and overall net progradation were driven primarily by (i) fluctuating sediment supply from the Mississippi River during a stable or slowly rising Holocene eustatic sea level, or (ii) fluctuating sediment supply from the Mississippi River during Middle-to-Late Holocene highstands and a net drop in eustatic sea level?

The key to understanding the impact of rising or falling Holocene sea level on Chenier-Plain evolution lies in the development of various ridges on the Chenier Plain that have maximum elevations that exceed 2.5 m NGVD 1929. These ridges include Little Chenier (+2.7 m NGVD 1929) and Little Pecan Island (+2.7 m NGVD 1929) along the Little Chenier–Little Pecan Island trend (Table 2, S4), Grand Chenier (+3.1 m NGVD 1929) and Pecan Island (+3.4 m NGVD 1929) along the Grand Chenier trend (Table 2, S11), and Hackberry Beach (+2.8 m NGVD 1929) close to the modern shoreline (Table 2, S14). Maximum elevations tend to occur along two prominent shoreline trends, the Little Chenier–Little Pecan Island trend and the Grand Chenier trend, which are the two most important paleoshorelines on the Chenier Plain because of their lateral extent and relief (see Table 2).

Regarding scenario 1, traditional Gulf of Mexico sea-level curves indicate that the postglacial rise in sea level stabilized approximately 3 kyears ago. The Little Chenier–Little Pecan Island trend (Table 2, S4) is the most landward and the oldest major shoreline in the Chenier Plain and is dated at around 2.6 to 2.8 kyears (Gould and McFarlan, 1959). This shoreline trend correlates with the onset of the sea-level stabilization phase (scenario 1) and marks the point in time where net shoreline movement changed from recession to advance. In other words, prior to the formation of the Little Chenier–Little Pecan Island trend, relative sea-level rise overwhelmed sediment supply, causing net retrogradation.

Since the development of the Little Chenier–Little Pecan Island trend, sediment supply to the Chenier Plain has overwhelmed relative sea-level rise, causing net progradation. During this period of relatively stable sea level and net progradation (at least 2.8 kyears ago to present), delta switching of the Mississippi River greatly influenced ridge formation and mudflat accretion. During ridge formation, regional and local processes operated concurrently to create transgressive, regressive, and laterally accreted ridges. During mudflat accretion, regional processes dominated.

In their sea-level scenario (scenario 2), Morton et al. (2000) and Blum et al. (2001) postulated that sea level during the Middle-to-Late Holocene rose to ∼1 to 2 m higher than present in the western Gulf of Mexico and subsequently has fallen to present levels, with possible higher-than-present sea-level oscillations during its fall [However, a consensus has not been reached regarding the existence of higher-than-present Holocene sea-level oscillations as discussed by Otvos (2001), Tornqvist et al. (2004), and Rodriguez and Meyer (2006)]. Under this second sea-level scenario, two possible explanations for net progradation of the Chenier Plain are (i) that a highstand and subsequent net fall in sea level were the primary factors, and variations in sediment supply from the Mississippi River played a secondary role; or (ii) that both factors in (i) exerted nearly equal influence. The second explanation is more likely, where the Little Chenier–Little Pecan Island trend (S4) would represent the +1- to +2-m highstand shoreline, as presented by Morton et al. (2000) and Blum et al. (2001). Subsequent net sea-level fall to present levels, along with sediment input from westerly positions of the Mississippi River, would drive net progradation of the Chenier Plain. The origin of subsequent paleoshoreline trends (e.g., Table 2, S5 through S10) would represent decreases in sediment supply as a result of more easterly positions of the Mississippi River mouth (i.e., delta switching) that enabled wave processes to erode the Chenier-Plain shoreline and to concentrate coarser sediment into ridges. The origin of the Grand Chenier
trend may be related to a major decrease in sediment supply from the Mississippi River as a result of delta-lobe switching within the Lafourche delta complex, or it may be related to a higher-than-present sea-level oscillation that occurred 1.1 to 1.2 kyears ago. Modern berm-crest heights along the outer Chenier-Plain shoreline average +1.5 m NGVD 1929 (e.g., Fig. 14, profiles 2 and 4). Maximum berm-crest heights along the Little Chenier–Little Pecan Island trend are +2.7 m NGVD 1929, and they range from +3.1 to +3.4 m NGVD 1929 along the Grand Chenier trend (e.g., Fig. 9, profile 20; Fig. 12a, profile 10; Fig. 12b, profile 6). Thus, to reconstruct the height of mean sea level along the two most continuous paleoshorelines of the Chenier Plain, 1.5 m must be subtracted from the ridge elevations. Subtracting 1.5 m from the maximum elevations along the Little Chenier–Little Pecan Island and the Grand Chenier trends places Middle-to-Late Holocene sea levels at +1.2 and about +1.8 m NGVD 1929, respectively, which correlates with the findings of Morton et al. (2000) and Blum et al. (2001, 2002).

7. Summary and conclusions

(i) The SW Louisiana Chenier Plain is a complex geomorphic feature that contains not only transgressive ridges (traditional cheniers), but also regressive ridges (beach ridges) and laterally accreted ridges (spits or recurved spits). Thus, a geomorphic hierarchy of primary landforms relative to dominant coastal process has been devised. The term Chenier Plain is defined as a first-order feature and is composed of three types of smaller, second-order features: chenier complex, beach-ridge complex, and spit complex. Individual ridges of each complex type are further separated into third-order features: chenier, beach ridge, and spit. Hoyt’s (1969) model only incorporates regressive mudflats and transgressive cheniers, however, and thus, he oversimplifies Chenier-Plain development by omitting ridges created by other means. Consequently, Chenier-Plain evolution involves more than just channel avulsion of the Mississippi River and attendant delta switching.

(ii) We agree with Hoyt (1969) and Otvos and Price (1979) on the definition of a chenier, but we present a new definition for the term Chenier Plain. Previously, a Chenier Plain was defined as containing two or more parallel to subparallel ridges (cheniers) separated by progradational littoral muddy units or mudflats (Price, 1955; Otvos and Price, 1979); however, in this paper, we further refine this definition by defining Chenier Plain as containing two or more chenier complexes. As such, we propose that this new, refined definition of Chenier Plain and the geomorphic hierarchy presented above be used whenever the term Chenier Plain is applied to a coastal depositional system.

(iii) Throughout the Chenier Plain, net long-term westward sediment transport is indicated geomorphologically (a) by westward deflection of rivers; (b) by the presence of beach ridges and beach-ridge complexes on the eastern sides of inlet entrances; and (c) by an overall decrease in interridge marsh width to the west. Local reversals in net westward sediment transport occur, such as those at the eastern end of Little Chenier and on the western side of Sabine and Calcasieu Passes.

(iv) Modern shoreline response to dominant coastal processes offers an approach to understanding the spatial distribution of landward relict ridges and ridge complexes. Paleoshoreline configurations that are similar to modern shorelines are evident throughout Chenier-Plain formation. Along the modern shoreline, two major erosion zones (sediment sources) exist: (a) the Holly Beach area from Ocean View Beach to approximately 1 km west of the Calcasieu River, and (b) the 63-km segment between the mouth of the Old Mermentau River and western Mulberry Island (dominated by the southern boundary of Rockefeller Refuge). These two local erosion zones are characterized by transgressive ridges that grade downdrift (westward) into regressive and laterally accreted ridge complexes (i.e., sediment sinks) near the termini of longshore sediment transport east of Sabine and Calcasieu Passes and into a laterally accreted spit complex (sediment sink) at the Mermentau River mouth. As such, transgressive, regressive, and laterally accreted ridges are produced concurrently along the present outer shoreline. Similar shoreline response trends are found along a paleoshoreline (Grand Chenier trend [S11]) landward of the modern shoreline in the southern White Lake area, where a large erosion zone with transgressive ridges exists along the Long Island–Pecan Island trend between North Island and Cypress Point Ridge. Sediment transported westward by littoral drift created the recurved spit complex along the eastern margin of the Mermentau River (Grand Chenier area) and coalesced east of Calcasieu Pass, forming a large beach-ridge/recurved-spit complex (Front Ridge West).

(v) Although the Mississippi River plays a primary role in Chenier-Plain development, Chenier-Plain
evolution and ridge response cannot be attributed solely to channel avulsion and the resultant delta-switching process. Other critical depositional and erosional processes include local reworking and downdrift transport of sediment, and the stability and sediment-trapping capability of tidal entrances. The location and stability of tidal entrances determine the genesis and distribution of beach ridges and recurved spits and their associated complexes. In the SW Louisiana Chenier Plain, regressive beach-ridge complexes develop at stable inlets ($\Omega/M_{\text{total}} > 150$), whereas laterally accreted spit complexes form at unstable inlets ($\Omega/M_{\text{total}} < 150$).

(vi) A six-stage geomorphic model is presented that synthesizes Chenier-Plain evolution primarily as a function of (a) the balance between sediment supply and energy dissipation, (b) local sediment reworking and lateral transport, and (c) tidal-entrance dynamics. Our geomorphic model presented in this study differs from previous models because it incorporates for the first time the important role of tidal-entrance morphodynamics, where the relationship between tidal prism and net littoral drift is considered.

(vii) In addition to the impact of updrift channel avulsions of the Mississippi River (delta switching), local sediment reworking and longshore sediment transport, and tidal-entrance dynamics along the Chenier Plain, overall Chenier-Plain evolution must take into consideration Holocene sea-level changes. Two paleoshorelines in the Chenier Plain have maximum elevations that exceed +2.5 m NGVD 1929: the Little Chenier–Little Pecan Island trend and the Grand Chenier trend. Although controversy remains regarding the exact height and behavior of past stands of Holocene sea level, geomorphic evidence from this study suggests that, to construct the two abovementioned paleoshoreline trends, sea-level oscillations most likely included higher-than-present stands during the Middle-to-Late-Holocene epoch. This evidence is consistent with the findings of Morton et al. (2000) and Blum et al. (2001, 2002) for the adjacent Texas and nearby Alabama coasts.

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