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Morphology and hydrodynamics of wave-cut gullies

A.M. Priestas *, S. Fagherazzi

Department of Earth Sciences and Marine Program, Boston University, Boston, MA, USA

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ABSTRACT

Wave-cut gullies are sub-triangular incisions common along deteriorating marsh scarps. Wave gullies may be equispaced to quasi-equispaced and enlarge in time, incising the marsh boundary. A high resolution survey is provided for ten wave gullies formed along the chenier plain of the Rockefeller Wildlife Refuge, Louisiana, USA. The measurements capture the morphologic character, evolution, and erosion rates of wave-cut gullies over a two month period. The data relate changes in morphology to geometric factors and shoreline retreat. Finally, the first analysis of wave data measured by acoustic Doppler velocity profilers is presented to show how propagating waves are transformed inside a wave-cut gully in order to describe the processes leading to their formation. Results show that waves of intermediate period (4–6 s) yield very strong swash currents that hit the gully head, detaching marsh substrate and triggering headward erosion. A conceptual model of wave gully evolution is presented as an explanation for this non-uniform, episodic shoreline erosion.

1. Introduction

Along with the well-known functions of providing unique habitats to fish and waterfowl, salt marshes are now known to filter natural and human waste products from water, buffer storms, and play a critical role in the cycling of chemical and biological compounds (Mitsch and Gosselink, 2000). Thus, understanding the rates and processes responsible for marsh erosion is key for the preservation and restoration of these environments.

The two primary mechanisms responsible for the deterioration of coastal salt marshes are platform submergence and marsh edge retreat. Deterioration by submergence depends on the competing influences between erosion and deposition processes. Deposition of both inorganic and organic sediments is influenced by sediment supply, vegetation productivity (through organic production) and vegetation effects on sediment transport (e.g., sediment trapping and increase in the threshold shear stress for erosion), and the oscillations of relative sea levels (DeLaune et al., 1983, 1994; Baumann et al., 1984; Orson et al., 1985; Finkelstein and Hardaway, 1988; Reed, 1988; Kearney et al., 1988; Day et al., 1994; Cahoon et al. 2006; D'Alpaos et al. 2007; Marani et al. 2007). Drowning kills the halophytic vegetation that stabilizes the marsh platform, so that the substrate can be easily eroded and dispersed in the ocean (Morris et al., 2002; Kirwan and Murray, 2008).

In contrast, salt marsh deterioration via wave attack is arguably the primary reason for lateral retreat of the seaward edge, and is thus a function of the local wave climate (Schwimmer, 2001; van de Koppel et al., 2005). Salt marshes undergoing lateral retreat tend to have a prominent scarp at its seaward edge; this erosive feature, and the mechanism responsible for its migration, were explained by 2-D modeling efforts by van de Koppel et al. (2005) and more recently by Mariotti and Fagherazzi (2010), even though Pye and French (1993) considered that the formation of a scarp was not necessarily indicative of edge retreat if the rate of sedimentation on the mudflat is lower than that of the adjacent marsh. Mariotti and Fagherazzi (2010) showed that, for a given sediment supply, the marsh progrades or erodes as a function of sea-level rise. A high rate of sea-level rise leads to a deeper tidal flat and, therefore, higher waves that erode the marsh boundary. However, their model is only two-dimensional, and thus might neglect important three-dimensional processes affecting marsh erosion. The bathymetry of the adjacent tidal flats is also critical for the erosion of the marsh boundary, since water depth controls wave formation and propagation, and ultimately the energy with which the waves impact the scarp (Fagherazzi and Wiberg 2009). Recent results by Mariotti et al. (2010) indicate that wave energy at the marsh boundaries is sensitive to wind direction, and increases remarkably with higher sea-level elevations and storm surges. The process of marsh erosion by wave impact is complex and modulated by tides. Tonelli et al. (2010) show with a high resolution Boussinesq model that wave thrust on the marsh scarp strongly depends on tidal level, increasing with tidal elevation until the marsh is submerged and then rapidly decreasing. Similarly, wave energy dissipation reaches the maximum just above the marsh platform elevation when breaking occurs at the marsh edge, beyond which, wave energy is further dissipated through the marsh vegetation canopy (Möller et al., 1999, Möller, 2006).

* Corresponding author.
E-mail address: priestas@bu.edu (A.M. Priestas).

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Edge retreat is increasingly recognized in the literature as a significant cause of marsh erosion, and though the relative contribution of salt marsh loss due to edge retreat or drowning is still unclear, the dominance of one of the two processes likely depends on the geographic setting. For example, investigating Louisiana salt marshes, Turner et al. (2004) speculated that shoreline erosion may not be as significant as that from increased flooding. Kearney et al. (1988) reported interior ponding was the primary mechanism for wetland loss in the upper reaches of the Nanticoke Estuary, Chesapeake Bay. Finally, Ravens et al. (2009) argued that decreased sediment supply from damming operations was the principle cause of marsh loss in West Galveston Bay, Texas. In contrast, Schwimmer (2001) found that salt marshes in Rehoboth Bay, Delaware, are mostly eroded via wave attack, rather than platform drowning, as supported by Pb^{210} sediment accumulation rates. He correlated the short-term erosion rates of the marsh edge to a function of wave power, which was hindcast from local wind, fetch, and bathymetric data. For marshes in East Galveston Bay, Texas Hall et al. (1986) reported that prolonged exposure to wind-generated waves could be responsible for more erosion than the effects of submergence during hurricanes.

Van der Wal and Pye (2004) indicated that changes in the wind/wave climate near estuaries in the Greater Thames area of the UK corresponded to rapid and recent erosion of the marsh edge, while Chauhan (2009) reported that edge retreat at a study site in northwest England is one stage of autocyclic erosion and progradation — though the evidence for such cycles (e.g., visibly abandoned clifflets from local sedimentation of the adjacent mudflat) does not appear to be present in microtidal salt marshes in the United States.

Our research involving sediment and tidal fluxes along the Louisiana chenier plain (Fagherazzi and Wiberg, 2009) led to the recognition of a unique mechanism for salt marsh boundary erosion by waves in gullies. These peculiar geomorphic features incise the shoreward edge of the marsh, and appear to be created by persistent and direct wave impact. Herein termed wave-cut gullies, they are morphologically similar to features described by previous researchers. For example, Hall et al. (1986) noted “points” and “cuts” along the eroding edge. Likewise, Schwimmer (2001) described a “cleft” and “neck” formation as a series of v-shaped notches cut along the marsh shoreline. Wave-cut gullies were also observed in Plum Island Sound, MA, and Hog Island Bay, Virginia, though they differ somewhat in size, extent, and spacing. Gully formation gives rise to an undulating pattern along the marsh boundary, though this is not to be confused with a “wave-etched shoreline” which is defined an initially straight shoreline made irregular by differential wave erosion of materials of varying resistance (Gary et al., 1974).

The origin of wave gullies is still unknown, but they might form as a self-organized process by which small perturbations of the scarp morphology facilitate wave erosion. These initial indentations then develop into gullies due to compression of wave crests and wave shoaling in a positive feedback by which erosion increases gully length, thus enhancing erosive wave processes within the gully.

In this paper, a high resolution survey is used to capture the morphologic character, evolution, and erosion rates of wave-cut gullies over a two month period. We seek to relate changes in morphology to geometric factors and shoreline retreat. Furthermore, the first analysis of wave data captured by Acoustic Doppler Velocity Profilers (ADCPs) is presented to show how propagating waves are transformed and concentrated (by changes in velocity and wave height) inside a wave-cut gully as an explanation for non-uniform shoreline erosion. Finally, a simple conceptual model is presented in order to describe their origin and evolution in time.

2. Study site

The study site is located in Little Constance Bayou within the Rockefeller Wildlife Refuge, along the central chenier coastal plain of southwest Louisiana, USA (Fig. 1). The Louisiana chenier plain is composed of Late Holocene deltaic muds interspersed by a series of widely separated sand and shell ridges termed “cheniers” (Otvos and Price, 1979). During the last 3000 years, net long-term westward currents transported fine sediments from the Mississippi delta to this location forming a 30 km-wide chenier plain (McBride et al., 2007).

In recent years, these plains have been affected by rapid erosion. Shoreline retreat in this area is one of the highest in the United States, averaging rates greater than 10 m/year between 1884 and 1994 (Byrnes et al., 1995) probably due to a decrease of sediment inputs from the Mississippi (Draut et al., 2005b), to the high rate of relative sea-level rise of 0.57 m/year (Penland and Ramsey, 1990), and to an offshore slope of about 1° coupled with wind waves propagating from offshore (Elgar and Raubenheimer, 2008).

An exception to this occurs along the eastern chenier plain, 30 km east of the study site, where sediments from the Atchfalaya river combined with resuspended shelf sediments in energetic conditions promote chenier accretion (Draut et al., 2005). The primary source of offshore sediments to the study site is likely the Atchafalaya subaqueous delta (Draut et al., 2005a) which terminates 10 km east of the study site (Draut et al., 2005a). Fagherazzi and Priestas (2010) found that despite the large volumes of sediment carried into the marsh at the study site, most of the material is returned to the ocean during the subsequent tidal cycle, which implies that the marsh can capture only a small fraction of the available mineral sediment. Many Louisiana marshes accrete via below-ground root production (organogenic sedimentation) (Nyman et al., 2006) and the combination of organic and inorganic sedimentation allows the marsh surface to keep pace with sea-level rise despite high subsidence rates (Reed, 2002).

The regression of the shoreline has considerably reduced the length of the tidal channel indicated in Fig. 1 in the last 50 years, so that the embayment in front of the channel mouth is in reality a vestige of a channel meander.

While storm waves periodically rework and deposit shell fragments along the beach (white areas in Fig. 1C), the entire system is mud-dominated. The region is microtidal with a maximum diurnal tidal range of 60 cm measured at Calcasieu Pass (~70 km west of the study site). The 15-year average significant offshore wave height is approximately 1 m with a dominant wave period of about 6 s, as measured from National Data Buoy Center station 42035 located about 30 km offshore of Galveston, Texas, and approximately 170 km from the study site (the closest available data). However, significant wave heights as measured 30 m offshore of the embayment (Fig. 1) were only 0.50–0.75 m, and were typically produced by southerly winds of 10–15 m/s (Fagherazzi and Priestas, 2010). Large storm events, however, can produce offshore (30 km) significant wave heights of 4–5 m. The offshore bathymetry is very shallow and gently sloping, reaching 5 m in depth at a distance of 2 km from the shoreline. The muddy seafloor causes considerable wave energy dissipation, reducing the height of the waves impacting the shoreline (Elgar and Raubenheimer, 2008).

3. Methods

3.1. Short-term erosion measurements

Ten gullies were identified at the study site, and their short-term headward and lateral erosion rates were monitored using erosion pins inserted horizontally (and vertically where noted) into the head-, side-walls, and shore-facing ends of each gully (see Fig. 2). The gully head represents the landward point, while the gully end is the location within the gully entrance estimated at the approximate shoreline. The sides represent the gully surface and floor at the approximate midpoint of each (see Fig. 2). Erosion pins were inserted on 01/15/2008 and recorded erosion measurements on 01/16/2008 (1 day), 01/17/2008 (1 day), 03/12/2008 (2 months), and also on 03/