Rapid formation of marsh-edge cliffs, Jiangsu coast, China

Yangyang Zhao a,b,1, Qian Yu a,⁎, Dandan Wang a, Ya Ping Wang a, Yunwei Wang c, Shu Gao d

a Ministry of Education Key Laboratory for Coast and Island Development, Nanjing University, Nanjing 210023, China
b State Key Laboratory of Marine Environmental Science, Xiamen University, Xiamen 361102, China
c College of Harbour, Coastal and Offshore Engineering, Huabei University, Nanjing 210098, China
d State Key Laboratory of Estuarine and Coastal Research, East China Normal University, Shanghai 200062, China

ARTICLE INFO

Article history:
Received 12 September 2016
Received in revised form 17 January 2017
Accepted 2 February 2017
Available online 4 February 2017

Keywords:
Marsh-edge cliffs
Tidal flat evolution
Spartina alterniflora community
Hysteresis effect
Jiangsu coast

ABSTRACT

The artificial introduction and rapid expansion of Spartina alterniflora Lois. (S. alterniflora) have greatly changed the natural evolution of tidal flats, especially along the Jiangsu coast, China. The occurrence of an elevation gap at the seaward margin (marsh-edge cliffs) manifests that tidal flats have experienced severe regional erosion and the state transformation from the smooth slope to the stepped profile. However, the mechanisms controlling the formation of marsh-edge cliffs and the state transformation of tidal flats remain elusive while the future evolution of a marsh-cliff-flat system is unclear. Here we show the rapid formation process of marsh-edge cliffs within less than five years in Yancheng Nature Reserve, Jiangsu coast, China, as a fantastic example for the first time, via examining the cross-shore topographic changes during 2008–2014 and investigating sediment characteristics, vegetation biomass, marsh deposition rate, currents and waves during 2013–2014. A one-dimension cross-shore model was used to estimate the relative predominance of waves and currents in response to the variations of tidal flat cross-shore profile and help interpret the formation of marsh-edge cliffs. The fast accretion of the marsh area after the introduction of S. alterniflora (approximately 4.3–5.3 cm/yr) accelerated the seaward shift of marsh margin, narrowing the width of bare flat; meanwhile, reduced sediment supply from offshore triggered the down-cutting of the lower bare flat (e.g., the maximum of about 0.5 m at the mean sea level from 2010 to 2011), strengthening wave actions to erode bare flat landward. These two processes jointly led to the predominance of waves (0.43–0.90 N/m²) over currents (mostly <0.40 N/m²) on the upper bare flat, which is consistent with the numerical modeling outcomes. As a consequence, the bed slope increased from 0.6‰ in 2008 to 1.0‰ in 2014, being vulnerable to wave attacks. The vegetation-root bonding of underlying sediments and the wave breaking at the marsh margin finally facilitated the formation of marsh-edge cliffs during 2011–2013. The cliffs were roughly estimated to retreat 20–110 m during 2013–2014. The future evolution of tidal flat accompanied by the continuous backward shifting of cliffs, the initiation and seaward spread of new marsh area, and the formation of new marsh-edge cliffs, resemble the hysteresis effect, as seaward advance and backward erosion will take almost equivalent time of nearly 9–50 years. This hysteresis effect of tidal flat evolution might further alter the local biological community structure and the pathway of coastal organic carbon flows.

© 2017 Elsevier B.V. All rights reserved.

1. Introduction

The subsistence of vegetation in the upper intertidal zone greatly complicates the geomorphological evolution of tidal flats, as it buffers currents and waves (Leonard and Luther, 1995; Möller et al., 1999, 2014; Leonard and Croft, 2006; Möller, 2006), inhibits resuspension of bottom surface sediments (Christiansen et al., 2000; van Proosdij et al., 2006a; Le Hir et al., 2007; Widdows et al., 2008) and thereafter prevents erosion and accelerates sediment accumulation (Mudd et al., 2010; Moskalski and Sommerfield, 2012; Fagherazzi et al., 2012; Wang and Temmerman, 2013) over the marsh area. The reduced flow velocity and increased sediment stability in turn facilitate plant growth, driving the salt marsh to accrete continuously and expand seaward (Wang et al., 2006; Mariotti and Fagherazzi, 2010). Such positive feedback between vegetation expansion, flow diminution and fast sedimentation narrows the lower bare flat and intensifies wave attacks due to increased steepness of bed slope (Tonelli et al., 2010), triggering the landward retreat of the intertidal flat. In this scenario, tidal flats with a smooth slope would not reach an equilibrium (Maan et al., 2015), but be constrained by the negative feedback of the wave erosion and the retreat of marsh margins (Marani et al., 2011; Leonard and Fagherazzi, 2014). This might result in the transformation of tidal flats into another state with a stepped profile (Marani et al., 2010; Wang and Temmerman, 2013), on which cliffs are formed at the boundary of the marsh platform and the lower bare flat (van de Koppel et al., 2005;
Rising sea level also can facilitate this process (Reed, 1995; Day et al., 1999; Kirwan and Murray, 2007; Mariotti and Fagherazzi, 2010). Tidal flats with marsh-edge cliffs have been widely reported along the mid-latitude coasts in the northern hemisphere since 1934 (Fig. 2). Other than sediment resuspension and settling that mainly

Fig. 1. Diagram showing different states of tidal flat geomorphology. (a) The smooth tidal flat of a gentle sloping bed; (b) The stepped tidal flat with a marsh-edge cliff at the boundary of the marsh platform and the unvegetated flat.

Fig. 2. Distribution of reported marsh-edge cliffs (Montmagny (Dionne and Bouchard, 2000), Sainte-Anne-de-Beaupré (Dionne, 2000), Bay of Fundy (Davidson-Arnett et al., 2002), Rehoboth Bay (Schwimmer, 2001), Southeast Essex (Greensmith and Tucker, 1965), Morecambe Bay (Allen, 1989; Pringle, 1995; Chauhan, 2009), Dovey Estuary (Richards, 1934), Severn Estuary (Allen, 1989), Wadden Sea Barrier Islands (Dijkema, 1997; Pedersen and Bartholdy, 2007; Van Loon-Steensma and Slim, 2013), Oosterschelde (Van Eerdt, 1985; Oenema and DeLaune, 1988; de Jong et al., 1994), Westerschelde Estuary (Gelisholt and Kristensen, 2003; van de Koppel et al., 2005; Van der Wal et al., 2008; Callaghan et al., 2010; Suzuki and Klaassen, 2011); Dengie Peninsula coast (Müller and Spencer, 2002), Venice Lagoon (Francalanci et al., 2013), Sado Estuary (Moureira, 1992), Jiangsu coast (Zhang et al., 2004; Zhao et al., 2014), Changjiang Estuary (Xie et al., 2013)).
control the shape of a smooth-slope tidal flat profile (Friedrichs and Aubrey, 1996; Yang et al., 2003; Pritchard and Hogg, 2003; Fan et al., 2006), the lateral erosion and retreat of marsh margins are at least of similar significance to the evolution of a marsh-cliff-flat system (Allen, 1989; McLaughlin et al., 2015; Leonardi et al., 2016). Generally, retreat rates of marsh-edge cliffs range from ~1 m/yr in semi-closed embayments (e.g., Allen, 1989; Schmimmer, 2001) to over 40 m/yr at large-scale estuaries and open coasts (e.g., Greensmith and Tucker, 1965; Pringle, 1995). Marsh margins retreat through toppling failure or rotational slips as waves erode the lower mud unit faster than the overlying root mat to create an overhang that eventually topples onto the bare flat (Allen, 1989, 2000; Schmimmer, 2001; Bendoni et al., 2014). The development of wave-cut gullies incising the marsh boundary further accelerates the landward shift of marsh-edge cliffs (Priestas and Fagherazzi, 2011; Zhao et al., 2014). Yet whether or not the retreat of cliffs will stop and the marsh-cliff-flat system will maintain stable are still uncertain. Allen (1989, 2000) proposed that marsh-edge cliffs were products of cyclic aggradation-erosion of marsh margins and finally left abandoned in salt marshes under the coupled impacts of relative sea level change, vertical accretion and sediment auto-compaction. On a shorter timescale, the disappearance of marsh-edge cliffs was observed after a rapid retreat of tens of meters during an extreme strong storm event (Xie et al., 2013), transforming tidal flats into the smooth-slope state.

As two states of tidal flats undergo distinct sediment dynamic processes and geomorphological behaviors, the evolution of tidal flats is closely related to their state transformation, that is, the formation and vanishment of marsh-edge cliffs. In northwest European, marsh-edge cliffs are considered to be inherent characteristics in the development of salt marshes (Pedersen and Bartholdy, 2007). Growing exposed salt marsh platforms terminate at the bare flat with an erosional cliff, where a small marsh-parallel depression is formed. van de Koppel et al. (2005) pointed out that self-organization within a tidal flat system was responsible for the formation of marsh-edge cliffs. As the increasing physical gradient characterized the marsh-flat boundary, salt marshes approached a critical state, susceptible to disturbance, for instance, a storm, leading to a cascade of vegetation collapse and severe erosion at the marsh margin. Marsh-edge cliffs were also found in an accretionary environment. Gao and Collins (1997) inferred that the formation of marsh-edge cliffs could be attributed to the deposition rate difference between salt marshes and bare flats. Contrast to the continuous upward growth of vegetated areas, the seasonal cyclic tidal flat elevation could cause a discontinuity at the bare/vegetated boundary, which might initiate the development of a small cliff once this discontinuity grows large enough for plant die-off to occur (Callaghan et al., 2010). The lateral shifts of tidal channels or creeks also could promote the occurrence of marsh-edge cliffs (Pringle, 1995; Shen et al., 2003).

In addition to these hypotheses proposed to illustrate the formation mechanisms of marsh-edge cliffs, numerical models have been established to reproduce and predict the evolution of tidal flats. A one-dimensional model by van de Koppel et al. (2005) stimulates the positive feedback between plant growth and sediment accumulation on the spatial development of salt marshes. As deposition is predicted to be highest at the exposed side of salt marshes (Temmerman et al., 2003; van Pootdij et al., 2006b; Pedersen and Bartholdy, 2007), the model develops toward an equilibrium characterized by strong sloping deposition at the bare/vegetated boundary. Once disturbances instigate local vegetation collapse, severe erosion of exposed sediments by focused wave action would produce a steep marsh-edge cliff (van de Koppel et al., 2005; Callaghan et al., 2010). This model also predicts the regrowth of vegetation in front of cliffs as cliffs erode simultaneously, which could be finally arrested when the newly established vegetation approaches cliffs. Mariotti and Fagherazzi (2010) further explored the evolution of the marsh boundary under different scenarios of sediment supply and sea level rise. Simulations indicate that vegetation determines the rate of marsh progradation and regression, and the scarp between salt marsh and tidal flat, resulting from redistribution of sediments within the intertidal area (Fagherazzi et al., 2012), is a distinctive feature of marsh retreat (Gao and Collins, 1997). A high rate of sea level rise leads to a deeper tidal flat and therefore higher waves that erode the marsh boundary (Kirwan and Murray, 2008; Mariotti and Fagherazzi, 2010). When the rate of sea level rise is too high the entire marsh would be drowned and transformed into a mud flat (Kirwan et al., 2010; Mariotti and Fagherazzi, 2010).

However, the detailed field observational study on the emergence of marsh-edge cliffs is few, which limits the understanding of the inherent physics. To further figure out the formation process of marsh-edge cliffs from a smooth-profile tidal flat and the controlling factors with respect to its “rapidness”, field investigations in Yancheng Nature Reserve, Jiangsu, China provide a fantastic example. Combined with geomorphological, hydrodynamic and sedimentological measurements, the development of marsh-edge cliffs and adjacent environmental characteristics were described in detail, and the controlling factors and potential interpretation for the rapid formation of salt marsh cliffs were further proposed. The fate of marsh-edge cliffs and the future evolution of this marsh-cliff-flat system were also predicted, which might serve as a bio-geomorphological foundation for coastal managements.

2. Study area

In situ observations and samplings were conducted in the intertidal zone in Yancheng Nature Reserve core region, middle Jiangsu coast, China (Fig. 3a). Since the sediment discharge into the coastal ocean from rivers is only 15% of sediment accumulation in the intertidal zone (Zhu et al., 1986), the evolution of tidal flats is governed mainly by Subei Coastal Current and nearshore residual currents with sediment supply from the erosion of abandoned Yellow River delta (Ren, 1986; Gao and Zhu, 1988). Due to the abundant sediment supply, nearshore suspended sediment concentration (SSC) is over 0.2 kg/m² in the surface layer and over 1.0 kg/m² in the bottom layer, especially in winter up to 1.5–6.0 kg/m³ (Xing et al., 2010; Wang et al., 2012; Yu et al., 2014, 2017). The high SSC led to the rapid progradation of the coast, at a rate around 150 m/yr in 1980s (Jiang, 1991; Wang et al., 2012). However, the reduced retreat rate of abandoned Yellow River delta has gradually resulted in accelerated erosion in the lower part of the intertidal zone but continuous accretion in the upper part (Zhang et al., 2006) due to the sediment trapping by marsh vegetation (Jiang, 1991). The nearshore tides are irregular semi-diurnal, with the tidal range of Xinyanggang, a gauging station close to the observation site (Fig. 3b), being 2.56 m on average and 3.40 m during spring tides. Tidal currents in the intertidal zone are characterized by remarkable diurnal inequality and time-velocity asymmetry, with the duration of ebb tides being 1.8 h longer than that of flood tides. The monsoon-driven winds are southeast in summer and north in winter (Ren, 1986; He et al., 2010). The multi-year averaged wind speed is 4–5 m/s on land areas and 5–7 m/s in the nearshore area (Ren, 1986). The mean and maximum 1/10 wave heights are recorded 1.0 m and 4.1 m, respectively, with minor interannual variability (Zhao et al., 2014). The recent sea level along the Jiangsu coast rises at a rate of 2–3 mm/yr (Wang, 1998).

Marsh vegetation Spartina anglica Hubbard (S. anglica) and S. alterniflora were consecutively introduced here in 1963 and –1989 to protect coasts and promote sediment deposition (Zhong and Zhu, 1985; Xu et al., 1993). Due to its strong adaptability and vitality, S. alterniflora community rapidly expanded to occupy the lower ecological niche than original vegetation communities (Zhang et al., 2004), not only colonizing seaward the adjacent bare flats, but also encroaching landward the neighboring marsh communities (i.e., S. anglica and Suaeda salsa) (Yao et al., 2009). Its expanding rate reached over 30% in 1990s and then gradually decreased to below 10% (Zhang et al., 2004). Until 2010, S. alterniflora marsh platform has enlarged to the width of 2–4 km and the area of 38 km², accounting for 28% of total salt marsh area (Fig. 3b).
3. Materials and methods

3.1. Surveys of bed-level changes

High accuracy elevation and positional data across the marsh platform and bare flats (Fig. 3b) were obtained by a Magellan Z-MAX GPS RTK (a differential GPS system). The GPS RTK has a static accuracy of 5 mm + 0.5 ppm for horizontal position and 10 mm + 0.5 ppm for elevation measurements. The dynamic accuracy decreased to 10 mm in the horizontal position and 20 mm for elevation. Bed-level surveying was conducted six times, in April 2008, October 2010, July 2011, November 2012, September 2013 and July 2014, respectively. As the topographic changes in the longshore direction are small here, these bed-level surveying points were all projected onto the projection line, which is perpendicular to the local shoreline, to normalize the distances from the Nature Reserve Gate for comparison (Fig. 3b).

3.2. Sediment sampling and analysis

A total of 10 bottom surface sediments were collected in July 2014 and five sediment cores were collected using PVC tubes with the diameter of 9 cm (Fig. 3c). Two cores (SA-C1 and SA-C2) were on S. alterniflora marsh platform (collected in July 2014) and three (BF-C1, BF-C2 and BF-C3) on the bare flat (collected in December 2012), with the depth of 1.91 m, 1.83 m and 0.65 m, 0.68 m, 0.73 m, respectively (Fig. 3d). Each core was sliced into 1-cm-thick contiguous samples. The grain size distributions of the sediment samples were measured by the Malvern Mastersizer 2000 laser granulometer (measuring range 0.02–2000 μm, with a duplicate measurement error of <3%). Moment statistics (McManus, 1988) was used to calculate the grain size parameters from the distribution curves. Sediment deposition rates of S. alterniflora marsh platform were estimated based on the expansion history of S. alterniflora here and the sediment characteristics in marsh sediment cores.

3.3. Investigations of micro-geomorphology and vegetation

Between the locations of sediment cores BF-C1 and BF-C3, wavelengths and wave heights of bottom surface sediment ripples were measured at the equal interval of about 20 m in December 2012 (Fig. 3c). Topography and geomorphological features at the marsh margin were also investigated and measured in December 2012, including the geometry of cliffs, the erosion/deposition bedforms and vegetation growth status. On September 30, 2013, 3 quadrats of 1 m × 1 m were sampled...
on the S. alterniflora marsh (Fig. 3c), and their stand density, wet mass density and stand height and diameter above the bed were also measured and calculated.

3.4. In-situ hydrodynamic observations and measurements

In-situ measurements were undertaken on the bare flat, about 800 m from the marsh-edge cliffs (Fig. 3c), from 8:00 on September 24 to 10:00 on September 29, 2013. An Acoustic Doppler Velocimeter (ADV), at 0.30 m above the bed, and an optical backscatter (OBS) sensors (D&A OBS-3A), at 0.30 m above the bed, was mounted on a stainless steel frame, with poles that penetrated the bare flat surface by > 1 m to provide stability. Currents and water depths were measured by ADV, with 16 Hz pressure measurement for 256 s per burst over a 300-second period during 8:00 on September 24 to 19:00 on September 25 and for 256 s per burst over a 600-second period during 10:00 on September 26 to 20:00 on September 29. ADV data with signal correlations > 70% and signal to noise ratios (SNR) > 20 were retained. In every burst, mean water depth was calculated, and wave period was determined as the average time that water depth takes two successive wave crests or troughs to pass the mean water depth. Subsequently, wave height was calculated as the average range of water depth in each wave period. Water samples were collected during the hydrodynamic measurements to calibrate the turbidity recorded by the OBS-3A sensors.

3.5. Calculation of skin-friction shear stresses

Skin-friction shear stress is calculated following the Soulsby (1997) method. The amplitude of the wave orbital velocity (u_w, m/s) at the top of the wave boundary layer can be expressed as follows:

\[ u_w = nH/|T \sinh(kh)| \]  

where \( H \) (m) is the wave height, \( T \) (s) is the wave period, \( k = 2\pi/L \) (/m) is the wave number, \( h \) (m) is water depth, and \( L = (gT^2/2\pi) \tanh(kh) \) (m) is the wave length.

The skin-friction shear stress due to waves (\( \tau_w \), N/m²) was calculated as follows:

\[ \tau_w = 0.5 \rho_w \bar{f}_w u_w^2 \]  

where \( \rho_w \) (kg/m³) is the seawater density, and \( \bar{f}_w \) is the wave friction factor, dependent on wave Reynolds number \( R_w \) for the smooth bed:

\[ R_w = u_w A / v \]  

\[ \bar{f}_w = \begin{cases} 2R_w^{-0.5} & \text{for } R_w \leq 5 \times 10^3 \\ 0.0521R_w^{-0.187} & \text{for } R_w > 5 \times 10^5 \end{cases} \]

where \( A = u_wT/2\pi \) (m) is semi-orbital excursion and \( v \) (m²/s) is the kinematic viscosity, which is the function of seawater temperature \( T \) (°C).

\[ v = \left[ 1.14 - 0.031 \times (T - 15) + 0.00068 \times (T - 15)^2 \right] \times 10^{-6} \]

Skin-friction shear stress due to currents (\( \tau_c \), N/m²) was calculated according to Eq. (6) as follows:

\[ \tau_c = \rho_o u_c^2 = \rho_o C_D u_c^2 \]  

where \( u_c \) (m/s) is the current-induced friction velocity, \( u_c(z) \) (m/s) is the current velocity at height \( z \) (m) above the bed, and \( C_D \) is the corresponding drag coefficient, expressed as Eq. (7) in case that the logarithmic velocity profile is assumed to hold throughout the water depth.

\[ C_D = k/\left[ \ln(z/z_0) \right]^2 \]

where \( k = 0.4 \) is Von Karman’s constant, \( z_0 = \kappa \delta_0 / (30 \text{ m}) \) is the bed roughness length, \( k_0 = 2.5\delta_0 \) (m) is the Nikuradse equivalent sand grain roughness, and \( \delta_0 \) (m) is the median grain size of the bulk sediment.

4. Results

4.1. Bed-level changes

The bed levels of P. australis and S. salsa marsh change little from 2008 to 2014, in contrary to that of S. alterniflora marsh and the bare flat, especially around the marsh margins, varying dramatically (Fig. 4a). The overall bed slope of salt marshes keeps around 0.2%, while the bed slope of the bare flat ranges from 0.6% in 2008 to 1.0% in 2014.

With the seaward expansion of S. alterniflora community, the bed level just in front of the marsh platform increases over 0.6 m during 2008–2014 (Fig. 4b). Meanwhile, the marsh margin shifts seaward about 500 m until 2013, but retreats landward nearly 20–110 m during 2013–2014 (Fig. 4b). The occurrence of cliffs at the marsh margin, which terminates at the bare flat, was observed in December 2012. The height of cliffs, i.e., the bed-level differences between the marsh platform and the bare flat, ranges from 0.73 to 1.22 m, with an average of 1.07 m. From 2013 later on, the bed level of the marsh platform is over 1.3 m higher than that of the bare flat (Fig. 4b).

The bed level of the bare flat keeps lowering during 2008 to 2014 (Fig. 4b). During 2008–2010, the bed level of the upper part of the bare flat changes little. Until 2011, the bare flat has been generally lowered, with the maximum eroded thickness of about 0.5 m at the MSL and decreasing eroded thickness to both sides. During 2011 to 2012, the upper part of the bare flat is continuously eroded, but the lower part accretes a little. From 2012 later on, the bed level of the bare flat changes in the range of 0.1 m.

4.2. Grain size distributions and deposition rates

4.2.1. Grain size distributions

The median grain sizes (\( d_{50} \)) of surficial sediments from the bare flat ranges from 3.61 to 3.91 \( \Phi \), with an average of 3.80 \( \Phi \), while the median grain sizes of surficial sediments from S. alterniflora marsh are averaged as 6.69 \( \Phi \) (Fig. 5a). The average contents of sand, silt and clay are 62.5%, 35.9% and 1.6%, respectively, from the bare flat and 5.8%, 85.1% and 9.1% respectively, from the S. alterniflora marsh (Fig. 5b).

The median grain sizes of core sediments on the bare flat are mainly in the range of 3.39–4.22 \( \Phi \), with the exception of some thin mud layers over 5.00 \( \Phi \) below the bed surface (Fig. 5a). Three cores all indicate that sandy sediment layers with the content of sand over 50% (Fig. 5b) cover the bare flat above the elevation of 0.5 m. For two cores in S. alterniflora marsh, almost in the upper 0.42 m, i.e., above the elevation of 1.5 m, the median grain sizes of core sediments average out to 6.57 \( \Phi \) (SA-C1) and 7.10 \( \Phi \) (SA-C2), respectively, and the sand contents are both below 17%. Below 0.42 m in depth, the median grain sizes gradually increase with the \( \Phi \) values decreasing to about 4.0 and the sand contents increase to over 50% until the depth of 0.64 m in core SA-C1 and 0.72 m in core SA-C2, respectively, below which sediment characteristics are similar to those from the bare flat. However, combined with the median grain size and the sand content of sediments, the characteristics of the sediments between 0.42–0.64 m in SA-C1 and 0.42–0.72 m in SA-C2 are quite different, as the former is similar to marsh sediments while the latter resembles the sediments from the bare flat.

4.2.2. Deposition rates

The introduction of S. alterniflora not only enhanced the sediment deposition, but also effectively sorted sediments by trapping finer-grain particles in the marsh platform to differ significantly from...
sediments of bare flats (Wang et al., 2006; Yang et al., 2008). Based on this, downcore grain size distributions, together with the introduction and expansion history of *S. alterniflora* in Jiangsu coast, could be used for a rough estimate of deposition rates in *S. alterniflora* marshes.

*S. alterniflora* community first introduced in this research area was around 1989 (Xu et al., 1993), but it expanded to the site of core SA-C1 until 1998 and core SA-C2 around 2005, on the basis of the expansion history of *S. alterniflora* by remote sensing imagery in Fig. 3. Hence, the sediments in these two cores have been influenced by *S. alterniflora* community as long as 15 years and 8 years, respectively. In Fig. 5, we could find above the depth of 0.64 m in core SA-C1 and 0.42 m core SA-C2, the sand contents decrease rapidly to below 10% and the median grain sizes are reduced to 6–7 Φ, which are the representative characteristics of sediments trapped by the marsh plants to distinguish from typical sediments of bare flats below the depth of 0.64 m in core SA-C1, 0.72 m in core SA-C2 and the exposed bare flat (Wang et al., 2006; Yang et al., 2008). The sediments above the depth of 0.64 m in core SA-C1 and 0.42 m core SA-C2 are therefore considered to be accumulated after *S. alterniflora* plants covering these two sites. So we can obtain the average deposition rates in these two cores are 4.3 cm/yr and 5.3 cm/yr, respectively.

4.3. Geomorphological features and salt marsh biomass

4.3.1. Geomorphological features

Sediment ripples are developed on the bare flat in front of the cliffs (Fig. 6a), with wave length ranging from 6.7 to 8.1 cm and wave height ranging from 1.5 to 2.4 cm (Fig. 6b). Mean wave length and wave height are 7.3 cm and 2.1 cm, respectively.

Marsh-edge cliffs are characterized by wave-cut gullies incising the marsh edges, resulting in indented geometry along the marsh margins (Fig. 6c), which has been described by Zhao et al. (2014) in detail and could be responsible for the different retreating distances along the coast. A marsh-parallel small depression is formed adjacent to the cliffs on the bare flat (Fig. 6a). At the head of wave-cut gullies, a fan-shaped ramp (Fig. 6d) is usually formed by sediments scoured from the lower part of cliffs, on which sometimes shell debris is deposited (Fig. 6e). From the cliffs backward, a 10 m-wide belt of *S. alterniflora* community

---

**Fig. 4.** Bed-level changes from 2008 to 2014. (a) Bed-level changes across salt marsh and bare flat. (b) Bed-level changes around the marsh margin, marked in (a) as red box. (c) Bed level of section 1 at the cliffs in 2014 (2014.07-1) with the same axis titles of (b). MHW: mean high water; MSL: mean sea level; MLW: mean low water. I: ponds; II: *P. australis* marsh; III: *S. salsa* marsh; IV: *S. alterniflora* marsh; V: bare flat. The origin refers to the gate of Yancheng Nature Reserve, marked in Fig. 3b. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

---

**Fig. 5.** From the cliffs backward, a 10 m-wide belt of *S. alterniflora* community
is die-off by wave attacks, left the exposed root mat and the eroded pits of 10–20 cm in depth (Fig. 6f). Far backward until around 30 m from the cliffs, a mass of S. alterniflora plants are lodged toward the wind direction (Fig. 6g). Intact shells of Meretrix are exposed on the surface and relatively packed around the plant roots (Fig. 6h). Noticeably, a marsh-parallel shell bank was observed in July 2014 about 10 m backward from the cliffs, with the width of 10 m and the height of 0.4–0.5 m (Fig. 4), which have not been observed before.

4.3.2. Salt marsh biomass
For three quadrats, their stand densities and wet mass densities of S. alterniflora community are as high as 146/m², 225/m², 122/m² and 3.94 kg/m², 3.33 kg/m², 5.28 kg/m², respectively. The mode of stand height reaches no less than 1.0 m and the mode of diameter over 0.5 cm. It is evident that high and dense S. alterniflora community occupies the marshes at the study area.

4.4. Hydrodynamics, bed shear stresses and suspended sediment concentration (SSC)

The hydrodynamic observation recorded 9 valid tides from the spring to the neap, showed in Fig. 7. The water depth of high level ranges from 2.5 m to 1.0 m, and the maximum current speed at flooding and ebbing decreases from over 0.4 m/s to about 0.2 m/s. Correspondingly, the significant wave height drops about 0.6 m from the spring (tide 1–3) to the neap (tide 7–9), while the mean wave period
maintains 4–5 s during the high water level. The current-induced skin-friction shear stress varies with the current velocity in a single tide, but it reaches the maximum 2.0 N/m² during the spring, far greater than that during the neap, fluctuating around 0–0.4 N/m². The wave-induced skin-friction shear stress keeps stable, ranging from nearly 0.9 N/m² during the spring to 0.43 N/m² during the neap. As a result, the SSC during the spring is as high as about 3.0 kg/m³, more than 3 times of the SSC during the neap.

5. Discussions

5.1. Rapid formation processes of marsh-edge cliffs

From the smooth-slope tidal flat developed into the presence of marsh-edge cliffs, it only took less than five years (within the period of 2008–2013), rather “rapid” for the state transformation of tidal flat. Based on the topographic changes displayed in Fig. 4b, where the missing parts of the tidal flat profiles have been conjectured in line with the trend of elevation variations in Fig. 4, this transformation process can be sketched into five different profiles (Fig. 8a) and attributed to two key subprocesses: the continuous accretion of salt marshes and the landward lateral erosion of the bare flat.

Before 2008, S. alterniflora community has been introduced here for over 20 years (Xu et al., 1993). Its strong adaptation to the adverse environment and high productivity (3.33–5.28 kg/m² in wet weight) made the small patches to rapidly expand longshore and outspread cross-shore (Xiao et al., 2010; Zhang et al., 2004). Until 1998–2005, S. alterniflora community patches were connected to cover the bare flat around the mean high water (Fig. 3b). Since then, the grain size of sediments got finer (from −4 Φ to 6–7 Φ upward in core SA-C1 and SA-C2,

![Fig. 6. Geomorphology at the marsh margins (photos taken in December 2012). (a) Wave ripples on bare flat in front of cliffs. (b) Scales of wave ripples. (c) Wave-cut gullies on the cliffs. (d) Fan-shaped ramp at the head of gullies. (e) Shell debris deposited on the ramp. (f) Remnant exposed root mat of S. alterniflora. (g) Lodged S. alterniflora plants. (h) Packed intact shells around the plant roots.](https://example.com/f6.jpg)
Fig. 7. Hydrodynamics and suspended sediment concentrations at height around 0.30 m above the bed on the bare flat, nearly 800 m in front of marsh-edge cliffs. (a) Water depth. (b) Current velocity. (c) Significant wave height and mean wave period. (d) Current- and wave-induced skin-friction shear stress. (e) Suspended sediment concentrations. The time “0” is 8:00 on September 24. Each valid tide is noted as tide number from 1 to 9.

Fig. 5a) as the high, dense S. alterniflora community (e.g., >1.0 m in height and 122–225 m² in standing density) trapped more mud to deposit (e.g., sand content decreases from over 50% to below 17% upward in core SA-C1 and SA-C2, Fig. 5b) (Wang et al., 2006). Benefited from the high SSC (e.g., as high as ~3.0 kg/m³ during the spring) and the strong trapping capacity of marsh plants, the deposition rate of S. alterniflora marsh (4.3–5.3 cm/yr) was much higher than the marshes colonized by S. salsa (1.9–2.1 cm/yr) (Liu et al., 2008; Liu et al., 2010), P. australis (1.2–1.7 cm/yr) (Liu et al., 2010) and unvegetated bare flat (1.5–2.3 cm/yr) (Liu et al., 2008; Li et al., 2011) along the Jiangsu coast. A similar rapid positive shift in elevation from bare to vegetated state was reported by Wang and Temmerman (2013).

The accelerated accretion of salt marshes (e.g. from 2008 to 2013) diminished the tidal prism and then reduced the current velocity on the bare flat (Le Hir et al., 2000), resulting in the accumulation of mud sediments in front of the marshes without great intensification of waves. The elevation of neighboring bare flat thus further created the advantageous environment for the seaward expansion of S. alterniflora community (Zhang et al., 2004; van de Koppel et al., 2005; Pedersen and Bartholdy, 2007). All above processes were coupled to push the seaward margin of S. alterniflora marsh reaching over 8.5 km from the origin until 2008 (Fig. 4b).

In contrast to the progradation of the supratidal zone above the mean high water, the subtidal zone below the mean low water has transferred from depositional state to erosional state since 1980 (Fig. 9) (Wang et al., 2003; Zhang et al., 2006). As the Yellow River had veered northward into the Bohai Sea and its sediment supply was cut off, the abandoned Yellow River delta was experiencing reduced retreat since 1855, leading to the decrease of sediments transported southward alongshore (Ren, 1986). The eroded area of the abandoned Yellow River delta was gradually extending to both sides (Zhang and Chen, 2009). In fact, under this factor, the lower intertidal zone had retreated at a rate of over 150 m/yr during 1980–1986, which decreased to about 30 m/yr in the subtidal zone during 1986–2001 (Zhang et al., 2006) under the impacts of the introduction of Sportina plants (Fig. 9).

These two subprocesses continued to be maintained before 2011, by the positive feedbacks between vegetation expansion and supratidal accretion, and hydrodynamic enhancement and subtidal retreat, respectively. As a result, the intertidal zone between the mean high water and the mean low water were further narrowed with a remarkably increasing gap of their elevations; that is, the bed slopes at the seaward margin of S. alterniflora and adjacent bare flat were steep to be about 2.0%. until 2011 (Fig. 8a). The increasing bed slope triggered the salt marsh to be a critical state and vulnerable to wave attacks (van de Koppel et al., 2005). When the waves broke at the marsh margin, the marsh sediments were bonded by the intricate network of S. alterniflora root mat while the unconsolidated sediments on the adjacent bare flat were scoured away (Allen, 1989, 2000; Schwimmer, 2001; Bendoni et al., 2014). As a consequence, the margin of the marsh platform was collapsed to the failure of the smooth slope during 2011–2013, forming a fracture of the bed elevations, i.e., the marsh–edge cliffs (Fig. 8a). The occurrence of cliffs further strengthened the wave attacks on the margin of the marsh platform (Tonelli et al., 2010), forming wave-cut gullies or troughs (Fig. 6c, d, e) perpendicularly cutting into the marsh boundary (Schwimmer, 2001; Prietas and Fagherazzi, 2011; Zhao et al., 2014). Meanwhile, without the bond of S. alterniflora root mat, underlying sediments at the margin were prior to be undercut by waves, leaving the overhang eventually toppled onto the front bare flat (Fig. 6d) when the sediment tensile strength was exceeded (Bendoni et al., 2014). Consequently, the down-cutting of the bare flat caused by the landward erosion (e.g., over 0.5 m near the MSL during 2010–2012) and the undercutting of the marsh boundary complementarily aggrandized the gap of the elevations of the marsh platform and the neighboring bare flat, which was up to over 1 m in height (Fig. 4b). So far the cliffs were formed to become the boundary of the marsh platform and the bare flat. Noticeably, comparing the tidal flat profiles of different years, we
could find that the erosion process of the bare flat tended to be stable and its bed slope maintained approximately 1.0‰ (Fig. 8a) accompanied by the retreat of marsh-edge cliffs after their formation.

5.2. Hydrodynamic-topography interactions

In the rapid formation process of marsh-edge cliffs, the accelerated vertical deposition resulted from the retardance of currents and the dampening of waves by the existence of vegetation on the upper tidal flat (Leonard and Luther, 1995; Möller et al., 1999; Davidson-Arnott et al., 2002; Möller, 2006), which trapped suspended sediments to be rapidly settled (Christiansen et al., 2000). Regarding the tidal flat as a system, the lateral erosion of the lower bare flat could be attributed to the decrease of sediment supply from the offshore. However, how the lateral erosion of the bare flat and the vertical deposition of the marsh area resulted in such evolution of tidal-flat profiles displayed in Fig. 8a remained unclear. Generally, tidal currents dominate over other hydrodynamic forcing on the tidal flat (Reineck and Singh, 1980), whereas wave action still acts as a stirring mechanism making sediments available for transport by tidal currents (Grant and Madsen, 1979). They both control sediment mobility and strongly affect the morphology of intertidal landscapes (Fagherazzi et al., 2007); as well they vary with the changes of intertidal topography. As observed in 2013, skin-friction shear-stress due to waves (varying from ~0.9 N/m² during the spring to...
0.43 N/m² during the neap) was much larger than that due to tidal currents (averaging ~0.4 N/m²) (Fig. 7d), which implies the stronger erosion capacity of waves. For this point, the dramatic topographical changes in the study area might have altered the flow regime across the tidal flat. Thus, the influence of bare-flat profile changes on the cross-shore distribution of currents and waves was discussed to help interpret the rapid formation of marsh-edge cliffs.

A one-dimension cross-shore model was established to probe into the cross-shore distribution of currents and waves over the different profiles of tidal flat during 2010–2014, based on the sketched profiles of the tidal flat (Fig. 8a). As waves propagate landward from offshore, the calculation of currents and waves started from the seaward opening boundary and the step pace was set as 1 m. The seaward opening boundary was set at the distance of 9.8 km from the origin and the hydodynamic calculation was based on the spring tide condition. Since the erosion was rather weak in the marsh area as marsh vegetation could strongly change the currents and waves (Leonard and Croft, 2006; Möller, 2006; Möller et al., 2014), here we only calculated the currents and waves over the bare flat, despite the marsh topography also post its influence on the currents and waves over the bare flat (Möller and Spencer, 2002).

Supposed that the water elevation keeps horizontal over the whole tidal flat no matter of the tidal phase, the magnitude of the cross-shore current can be estimated by considering the conservation of volume as the water level fluctuations (Friedrichs and Aubrey, 1996). Thus, the tide-averaged cross-shore current velocity $\bar{u}_c (\text{m/s})$ at the location of $x$ (km) from the origin on the bare flat was calculated as Eq. (8).

$$\bar{u}_c = 4P/(hT_{\text{age}})$$ (8)

where $P$ (m²), $h$ (m), $T_{\text{age}}$ (s) are tidal prism of unit area, the maximum water depth during spring tide, and the inundation duration in a tidal cycle at the location of $x$ (km) from the origin. Assuming the water level experienced a cosine variation during flooding and ebbing, the tide period at each location can be obtained as the function of water depth.

$$T_{\text{age}} = \frac{T_{\text{m}}}{\pi} \cos^{-1} \left(1 - \frac{2h}{R_{\text{M}}} \right)$$ (9)

where $T_{\text{m}}$ (s) is the period of M2 tide, valued as 12.42 h, and $R_{\text{M}}$ (m) is the maximum tidal range.

Following the linear theory and applying the conservation of energy for a monochromatic wave propagating on a uniform slope, a relation between the wave height $H$ (m) and the water depth $h$ (m) could be obtained according to the differential equation (Le Hir et al., 2000)

$$\frac{d}{dx} \left( \frac{H^2}{h^2} \right) = \frac{2f_w}{3\eta} H^3 \eta^{-3/2}$$ (10)

where $f_w$ is the wave friction factor, which is set as 0.05 based on the model of Soulsby (1997), considering the measured scale of bedform on the bare flat (Fig. 6b). Since the bed elevations changed little at the seaward opening boundary during 2010–2014, there the incident wave heights are set as 1.0 m and the wave period is set as 4.5 s on the basis of the observation. So the wave-orbital velocity could be obtained as Eq. (1).

Cross-shore tidal currents have been considered the driven force to shape the up-convex profile of the tidal flat and maintain the gentle bed slope, while wave actions would lead to the increased bed slope and the narrowed width, finally resulted in the up-convex profile of the tidal flat (Friedrichs and Aubrey, 1996; Kirby, 2000; Pritchard and Hogg, 2003). As for the small bed slopes, waves would not break but decay by the bottom friction (Le Hir et al., 2000), which still greatly affect sediment transport (Roberts et al., 2000; Fagherazzi et al., 2006).

The landward propagation of waves was strongly related to the topography. During 2010–2012, the decay rate of waves became slow with the down-cutting of the bare flat, the increased water depth and therefore the reduced bottom friction. In this case, the wave height at the boundary of the marsh area and the bare flat increased from 0.33 m to 0.51 m (Fig. 8c). Meanwhile, even though the erosion of the bare flat increased the tidal prism of lower bare flat, the accumulation on the marsh significantly diminished the tidal prism of the upper bare flat, which reduced the current velocity from ~0.13 m/s to 0.05 m/s (Fig. 8b). Comparing the cross-shore current velocity and wave-orbital velocity, their ratio decreased greatly, especially at the boundary of the marsh area and the bare flat, decreasing from 0.25 to 0.11 (Fig. 8d). This indicates that with the erosion of the bare flat and the accretion of the marsh area, the hydrodynamic frame over the upper bare flat have been inversed, that is, compared to waves, the power of tidal currents was weakened rapidly. With the decreasing distance from the boundary, waves became relatively stronger than currents (Fig. 8d). Thus, waves turned into the dominate hydrodynamic factor at the marsh margin. So the erosion by waves attacks at the boundary of the marsh area and the bare flat occurred, which triggered the formation of marsh-edge cliffs combined with the protection of marsh vegetation root mat on the upper layer of marsh sediments.

The profile of 2013 indicates that the cliffs had been formed at the boundary of the marsh area and the bare flat, and the marsh platform continuously accreted (Fig. 8a), which led to the further decrease of current velocity over the upper bare flat (Fig. 8b). The distribution of wave height was similar to that in 2012 over the bare flat, except for waves broke on the cliffs (Fig. 8c). Wave breaking would dissipate most power and gave rise to the intense erosion of cliffs (Gao and Collins, 1997; Marani et al., 2011; Bendoni et al., 2014). Thus, the tidal flat system has been transformed from the smooth tidal flat with a gentle sloping bed to the stepped tidal flat with a marsh-edge cliff at the boundary of the salt-marsh platform and the unvegetated flat (Fig. 1). The topographic evolution during 2012–2013 was a positive feedback process, that is, the increasing bed slope in front of the marsh promoted the stronger wave power relative to currents, which led to the further increase of the bare flat slope bordering the marsh area, until the formation of marsh-edge cliffs.

From 2013 to 2014, the cliffs were eroded to retreat but the marsh platform was able to keep on accreting (Fig. 8a). Sometimes the shell debris deposition in the front of the marsh platform (Fig. 4c) further cut down the tidal prism as its peak was higher than MHW and held up seawater on the bare flat. This resulted in decreased current velocity in front of the marsh platform (Fig. 8b). As a result of the cliff retreat, the bare flat turned to become wider, making waves to dissipate more energy on the bare flat before they attacked on the cliffs. So the wave height continued to decrease landward keeping the trend of 2013. In 2013, the wave height was 0.54 m at the cliffs, but after the cliffs retreated about 110 m, it decreased to 0.43 m (Fig. 8c). For this point, waves remained predominante at the marsh margin compared to currents, indicating that the marsh-edge cliffs will continue to retreat backward and the tidal flat will maintain its stepped state with marsh-edge cliffs.

5.3. Future evolution of marsh-edge cliffs and the tidal flat

The rapid formation of marsh-edge cliffs is the result of the combined interaction of hydrodynamics, marsh vegetation and sediment transport, which as well continue to control the future evolution of marsh-edge cliffs and tidal flats. Supposed the bed slope of the bare flat could be maintained about 1%, as displayed in Fig. 8a during 2012–2014, the greatest distance the cliffs could retreat is estimated to be nearly 1200 m, that is, the cliffs would be eroded back to the distance of 7.8 km from the origin, close to the landward margin of S. alterniflora marsh (Fig. 3c).

However, with the retreating of cliffs, the eroded marsh platform frees up space for added tidal prism (Fig. 10a), if without the
development of shell debris deposition, increasing the tidal currents but dissipation of more wave power before arriving at the cliffs. Following this trend, the predominance of waves over currents over the upper bare flat would be reversed before the vanishing of marsh-edge cliffs, further slowing down the backward shifting of the marsh margin. Moreover, the elevated bed level at the base of cliffs makes the elevation gap between the marsh platform and the adjacent bare flat smaller to a certain extent that the deep root of *S. alterniflora* could protect the underlying sediments on the cliffs from being eroded. At the stage of Fig. 10a, suspended sediments might start to accumulate in front of the cliffs, elevating the bed level to enhance the bottom friction for waves to decay. Though waves are greatly damped, the cliffs will still retreat (Fig. 10b) until the *S. alterniflora* community begins to colonize the elevated bare flat in front of the cliffs. As the cliffs retreated 20–110 m in 10 months during 2013–2014 (Fig. 4b), it will still take about 9–50 years for the cliffs to be abandoned if we suppose that the retreat rate of cliffs decreases linearly. Actually, this retreating time of cliffs may still vary with cliffs’ retreat rate, which greatly depends on the following interaction of hydrodynamics, marsh vegetation, and topography. After that, the positive feedback of sediments and vegetation will result in the vertical accretion of new marsh area (van de Koppel et al., 2005; Pedersen and Bartholdy, 2007) and the lateral expanding of *S. alterniflora* community.

With the restoration of new marsh community in front of the abandoned cliffs, slowed tidal currents over the lower bare flat offer the opportunity for the waves to under-cut under the background of the continuous expansion of erosional regions. The erosion of subtidal zone below the mean low water would extend upward to steepen the bed slope between the new marsh area and its adjacent bare flat, and increase the wave energy at the seaward boundary of tidal flats. Considering the predominance of waves over currents will be more significant, the new cliffs would be formed when the new marsh platform remains under the mean high water level (Fig. 10b). This might cut down the time for the formation of cliffs. Hence, the restoration of the new marsh platform and the formation of new cliffs are estimated to take at most the same time for cliffs retreating.

The above processes might be repeated again as Fig. 10c and Fig. 10d, similar to Fig. 10a and Fig. 10b. It is interesting to find out that these processes somewhat reflect the ‘hysteresis effect’, as they make up a round-trip starting from the vertical accretion of the marsh platform but ending with the lateral backward erosion of the marsh margin to be abandoned. Within these processes, the retreated marsh margins shrink the niche for macro-invertebrates in the marsh platform, but create more space for sandy-living burrowing benthos. This might result in the variations in their abundance and community structure, maybe the inverse succession just after the introduction of *S. alterniflora* community. In addition, organic matter buried in the marsh sediments from the photosynthesis of *S. alterniflora* community is considered to play a considerable role as carbon sink (Chmura et al., 2003). However, from a long-term perspective, most preserved organic carbon will be degraded by microbes and returned into the atmosphere directly (Benner et al., 1991; Kirwan and Blum, 2011). In fact, the backward erosion of marsh-edge cliffs releases the stored organic carbon to the coastal sea, where the remineralized carbon could be reused by phytoplankton and marine calcareous organism (Miller and Zepp, 1995; Hedges et al., 1997; Opsahl and Benner, 1997; Hermes and Benner, 2003). For this point, the processes of marsh accretion and cliff retreating greatly alter the pathway of carbon cycles in the coastal area.

This ‘hysteresis effect’ of tidal flat evolution might also occur in environment with lower suspended sediment concentrations, but with a much smaller rate. As in this study area the formation of marsh-edge cliffs is much faster than those ever reported, the complete round-trip might be observed here in the near future. In practice, our understanding of this rapid evolution of the marsh-cliff-flat system also could provide evidence for slow-process measurements.

Even the marsh platform accretes and the cliffs retreat in different pathway, both halves of the round-trip take the almost equivalent time. As the new marsh platform might be terminated to accrete under the mean high water level, the new cliffs would not reach the location of the cliffs already formed during 2012–2013 (Fig. 10). Yet the location of new cliffs and the time they take to complete a round-trip still cannot be predicted exactly, and whether the tidal flat would be ultimately inversely transformed to the smooth-slope state remains unknown. These need further exploration taking into consideration the sea level changes and sediment supply for the longer-term evolution (Kirwan and Megenigal, 2013).

### 6. Conclusions

The rapid formation processes of marsh-edge cliffs can attribute to two subprocesses: the vertical accretion of the marsh area and the landward erosion of lower bare flat. The introduction and expansion of high dense *S. alterniflora* community (>1.0 m in height and 122–225/m² of stand density) here accelerated the deposition rate (approximately

---

**Fig. 10.** Sketched future evolution of marsh-edge cliffs. The black dashed line indicates previous topography, and the red solid line indicates new-established topography. The four subgraphs show two cycles of marsh-edge retreating and new marsh area formed. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)
References


