Emerging Downdrift Erosion by Twin Long-Range Jetties on an Open Mesotidal Muddy Coast, China

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Abstract: Downdrift shoreline recession associated with the construction of a shore-crossing hard structure represents one of coastal erosional hotspots that must be addressed for an integrated, sustainable coastal zone management. To prevent siltation within the navigation channel, two rubber-mounted jetties were installed at the Sheyang River mouth on the open mesotidal muddy coast in Jiangsu province, China, in October 2013. The north jetty is 7.9 km long, while the south jetty is 7.8 km long. The net longshore sediment transport is from the north to the south due to flood-tide dominance. As disclosed by high-resolution satellite images, a 36-km-long downdrift shoreline stretch had experienced remarkable retreats at alongshore varying rates by March 2019. The eroding shoreline planform does not resemble a classic “S” shape, a crescentic shape, or a parabolic shape but an irregularly indented curved shape. Transect topographic survey also reveals an almost immediate response of the downdrift coast from the original accretionary scenery to an erosional regime, with the erosion front translocating downcoast at a much faster speed than a normal speed of 1–1.5 km/yr. Using FVCOM and SWAN, 2DH process-based numerical simulations are performed to simulate the flow, the sediment transport, and the yearly-magnitude accretion/erosion distribution in the jetty-affected area by a representative tidal force and an annual-magnitude wave force. The results demonstrate that the reciprocal tidal flow is predominantly responsible for the muddy sediment accretions at downdrift intertidal and surf zones shallower than a 4.0-m isobath, whereas big wind waves play a decisive role in triggering and developing the downdrift erosional process. The predicted spatial extent of the downdrift erosional segment matches closely the actual eroding front. The loss of the net annual longshore sediment transport volume, i.e., 3.08 million m 3 due to the blockage by the twin jetties is recovered from a much larger spatial extent than the 36-km-long retreating shoreline stretch. With regard to the Bruun model, the one-line model, the headland-bay model, and the 2DH numerical model, the potential maximum recession length and the planform shape of the downdrift erosional shoreline arc are further elaborated to gain new insights into the spatial and temporal impact of a hard structure on the adjacent shoreline and flat (beach).

Keywords: jetty; downdrift erosion; muddy coast; shoreline model; marsh scarp

1. Introduction

To provide shipping service, an inlet on a barrier coast or a river mouth on an open coast is often used as an entrance channel to a quay berth. To stabilize and prevent siltation within the entrance channel, a single jetty or twin jetties have to be installed. Such an anthropogenic interference will undoubtedly alter the local hydrodynamic field and cause morphological adjustment. Common observations show that shoreline progradation occurs on the updrift side of a jetty, while the shoreline on the downdrift side experiences recession due to the blockage of longshore sediment transport [1,2].

This type of downdrift erosion is one of the coastal erosional hotspots ascribed to hard structures [3,4]. It may be superimposed upon the chronic regional shoreline recession...
(e.g., [5]) or emerge on an original accretionary shore due to an abrupt sediment deficit. As the downdrift erosion causes damages to properties and land loss, it is critically important to understand the process and magnitude of the downdrift erosion for an efficient erosion-mitigation management strategy. Though considerable efforts are made, the temporal and spatial behaviors of the downdrift eroding shoreline remain not fully understood.

Pelnaro-Considère (cited in [1]) suggested that the erosion arc might expand downdrift infinitely. Bruun [1] proposed a conceptual model for the evolution of the downdrift erosional shoreline arc. It consists of a short component immediately adjacent to the jetty and a long-distance segment that is driven by sediment starvation and wave action. The two segments are separated by a no-erosion or a slightly accreting “bump” point, and the entire feature has a characteristic “S” shape, with the downcoast eroding front expanding over time.

Galgano [6] further elaborated that the erosion arc is non-stationary and expanding downdrift at a non-linear rate. The spatial magnitude of the erosion arc is also site specific, with maximum lengths ranging from 5.8 km to 18.6 km.

Meanwhile, many shoreline evolution models, such as varying-complexity one-line models (e.g., [7–11]), are often resorted to predict the downdrift shoreline receding process. The outputs by these models rarely resemble the “S”-shaped downdrift arc of Bruun [1] but manifest a crescentic shape. Headland-bay beach models, including the logarithmic spiral model, the hyperbolic tangent model, and the parabolic model [12], however, can predict the static equilibrium embayed shoreline planforms.

Nonetheless, the majority of existing research findings in the literature are applicable to sandy coasts where a jetty or groyne blocks the longshore coarse sediment transport dominantly driven by waves [13]. Investigation is rare into the evolution of a downdrift shoreline and the intertidal flat on a muddy open coast where stronger reciprocal tidal currents control the longshore transport of fine sediments and salt marsh vegetation is extensively present.

This study presents a unique example to gain insights into the spatial and temporal impact of a hard structure on an adjacent muddy coast. The morphological response of the downdrift shoreline and the tidal flat to the construction of the jetties of the Sheyang port in Jiangsu province, China, is firstly examined using high-resolution satellite images and transect topographic surveys, etc. And then the responsible hydrodynamic force in governing the downdrift erosion process is ascertained by performing processed-based 2DH simulations using FVCOM and SWAN models. The downdrift recovering process for the blocked net annual longshore muddy sediment volume is further investigated. The merits and limitations of existing empirical and numerical models in predicting the potential maximum recession length and the planform shape of the downdrift erosional shoreline arc are elaborated. Finally, concluding remarks is given.

2. Physical Background and the Construction of the Sheyang Port

Sheyang River is running into Yellow Sea through the middle part of the Jiangsu coast, which is a mesotidal open muddy coast (Figure 1). A giant radiant sandy ridge system (RSRs for short) with extensive intertidal flats is present at the southern coast, whereas an abandoned Yellow River delta (AYRD for short) has been continuously eroding at the northern coast since the Yellow River avulsed into Bohai Sea in 1855 [14].

The local tide is regular semi-diurnal. The average tidal range is about 2.56 m, and the spring tidal range is up to 3.4 m [15]. The deep water annual mean significant wave height is about 0.88 m, and the annual mean wind speed (U_{10}) is about 6.58 m/s.

To satisfy the increasing demand for shipping service, the Sheyang port has completed two-phase projects. The first phase aims at enabling 10,000-ton bulk cargo to get in and out. According to NNUESTRI [16], twin rubber-mounted jetties on both sides of the Sheyang River mouth began to be constructed in March 2011 and was completed in September 2012. The south jetty is 6.2 km long, while the north jetty is 6.3 km long. Except for the emerged shore-connected segments, the middle and tip parts of both jetties are
submerged at high tides. An 80–142-m-wide, 8.0–9.1 m-deep (LAT, Lowest Astronomical Tide) navigation channel was dredged from February 2012 to October 2013 and then the port began to operate.

![Figure 1. The Jiangsu muddy coast. The inserted satellite image shows the layout of the Sheyang port.](image)

The second phase was started in March 2015 and finished in October 2016. It enables 35,000-ton bulk cargo to enter and depart. The project consists of jetty-heightening and channel-dredging work. The crest levels of the original two jetties are all heightened to emerge over MHW (mean high-water level), and both jetties are extended by 1.6 km. The jetty tips sit at around 10 m isobath. The navigation channel is 10 km long, 160–360 m wide, and 10.5–11.3 m deep (LAT) [16].

3. Emerging Downdrift Shoreline Erosion

With a significant portion of eroded deposits from the AYRD being transported southeastward to RSRs, the mouth of the Sheyang River was once clarified as the transitional spot from the eroding coast in the north to the accreting coast in the south [17,18]. Before the construction of the Sheyang port, i.e., March 2011, both the salt marsh and the tidal flat to the north of the Sheyang River mouth experienced persistent erosion, whereas the salt marsh and the intertidal flat to the south of the Sheyang River mouth still advanced seaward, albeit the subaqueous bed was concomitantly retreating, demonstrating an “upper silting and lower scouring” mode (Figure 2). According to Wang et al. [19], the shoreline from the Sheyang River mouth northward to the Biandangang (a 41.2 km distance) receded at a spatially mean rate of 10.0 m/yr whereas the shoreline from the Sheyang River mouth southward to Xinyanggang (a 28.6 km distance) and from Xinyanggang to Doulonggang (a 23.1 km distance) prograded at spatially mean rates of 8.9 m/yr and 16.2 m/yr, respectively, during the period of 1988–2006.

The shoreline in this study is delineated by the Spartina alterniflora vegetation boundary of the salt marsh on high-resolution satellite images dated December 2013 and March 2019. This vegetation boundary is normally close to the MHW of neap tides [20]. As computed from the respective shoreline position, the downdrift shoreline evolution pattern immediately emerges (Figure 3). A 2.7-km-long shoreline segment adjoining the south jetty advances at a mean speed of about 31.2 m/yr. Further southward, the shoreline starts to recede at increasingly larger speeds of 12.7 m/yr, 20.5 m/yr, and 30.3 m/yr. However, at the northern flank of the Xinyanggang River mouth, the receding speed drops to 4.8 m/yr. At the southern flank of the Xinyang River mouth, the receding speed is about 8.2 m/yr. It firstly increases to 20.5 m/yr and then drops southward to 7.0 m/yr and 2.1 m/yr. At the final receding segment, the speed is slightly increased, to 4.8 m/yr. The spatially averaged receding rate is about 14.1 m/yr. Passing the eroding front, the shoreline to the Doulonggang River mouth advances at a speed of 1.3 m/yr. South to the Doulonggang River mouth, the cross-shore profile shape still maintains the characteristic “upper silting and lower scouring” mode [21].
Figure 2. Two topographic transects spanning 1984–2006 at the north and south sides of the Sheyang River mouth. The 0 m isobath is depicted for the year of 2006 (datum is above MSL; the same hereinafter).

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Figure 3. Downdrift shoreline evolution from December 2013 to March 2019. “SYG” denotes the Sheyang port, “XYG” denotes Xinyanggang, and “DLG” denotes Doulonggang. The inserted pseudocolor satellite images illustrate the variation in the salt marsh vegetation cover (green arrows) on the respective date.
The overall planform shape of the eroding shoreline arc does not resemble an “S” shape [1] or a crescentic or logarithmic-spiral planform. Its straight-longshore distance is 36 km or so. As constrained by the earth embankments used for the protection of shrimp ponds, the downdrift shoreline segment (i.e., at the receding rates of 12.7 m/yr and 20.5 m/yr) near the southern jetty cannot develop to its full capacity (Figure 4). As a result, the downdrift intense recession stretch does not appear close to the south jetty but expands over a large distance. The local shoreline, however, may retreat a considerable distance if the embankment is breached by wave attacks.

![Figure 4. The eroding shoreline bound by shrimp-pond earth embankments and remnant salt marsh. UAV photo is looking northward and was taken on 28 June 2019. The location is shown in Figure 3 (camera A).](image)

Vertical eroding scarps are developing at the saw-tooth-like salt marsh edges (Figure 5). As impacted by incident waves, the salt marsh vegetation firstly dies and then the salt marsh edge experiences a series of mass failures.

![Figure 5. The jagged eroding salt marsh boundary. Photo (A): vertical scarps; photo (B): mud blocks from toppling failure. Both photos were taken on 15 September 2020. The location is shown in Figure 3 (camera B).](image)

The whole eroding shoreline is irregularly indented. It is roughly intersected into different segments by river mouths and tidal creeks, where concave planform shapes appear with relatively lower receding rates. These rivers have all been installed with tidal gates, and the sediment volume that can be transported into the nearshore water is considerably reduced.

### 4. Process-Based Numerical Modeling

#### 4.1. Mode Setup and Driving Forces

The numerical model covers an area of about 5200 km². The domain is meshed into over 250,000 triangular elements with more than 120,000 nodes (Figure 6). The grid cell varies from 500 to 2000 m at the open sea boundary to 100 to 200 m shoreward and refined to around 10 m close to the salt marsh edge. The shoreline configuration is determined...
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The bathymetry within the computation domain is assembled and interpolated with multi-source data, including localized topographic survey, nautical charts, and inferred elevations of salt marsh vegetation, where no surveyed data are available. Non-uniform sediments of five classes, ranging from clay to fine sand ($D_{50} = 0.0389$ mm) are computed; the critical shear stress for erosion is 0.79 N/m$^2$.

Many field observations indicate that morphological adjustment in tidal flats is controlled by the balance between accumulative calm-weather tidal accretion and episodic erosion events by big wind waves (e.g., [22,23]). This understanding promotes the present study to obtain the net yearly-magnitude accretion/erosion distribution in the jetty-affected area by performing respective numerical simulations forced by representative tide only and representative wave force alone. In so doing, it is easy to discern the distinct role played by the reciprocal tidal current and incident wave in initiating and developing the downdrift erosion process, and considerable computation efforts can be saved by neglecting smaller waves from varying directions, which demand complex, time-consuming wave–tide interactions.

FVCOM is an unstructured grid, finite-volume, 3D, parallel-computed ocean model that is able to simulate both tidal motion and morphological evolution [24]. It is well calibrated with the observed water level, tidal current, and suspended sediment concentration at gauge sites in 2018 and the adjacent topographic evolution. The calibrations of the tidal current and bottom suspended sediment concentration at one of four gauges (see location in Figure 6) are shown for example (Figure 7). This study executes a quazi-2DH mode to simulate firstly the accretion/erosion distribution for the year of 2019 solely by tidal current. The simulation is forced by the hourly water level of the spring tide measured in the summer of 2018 at the open sea boundary nodes and is sped up by using an appropriate “morphological factor” [25].

A high water is critical for the wave-induced seabed and beach profile evolution, especially on the middle to upper flats [26], for reasons that the wave-induced bed shear stress within the wave-saturation zone reaches the highest value only at high tides [27]. More importantly, an eroding salt marsh scarp is predominantly produced by approaching wave attacks and the wave energy dissipation is maximized when the tidal level is just around the marsh platform elevation [28].
waves are reduced considerably and are negligible when they approach the shore in the wind sea wave [30]. The wind rose diagram and the wave rose diagram for the deep water (about 48 km from the shore) are presented in Figure 8. They are produced with 40-year prevalent wave comes from the ESE-SSE directions, and the SE wave has the relatively largest frequency. Nevertheless, due to the shielding and breaking effects of the extensive shallow sandy ridges (RSRs in Figure 1), the magnitude and frequency of the around-SE waves are reduced considerably and are negligible when they approach the shore in the study area, whereas the NNE waves propagate into the dominant N direction due to refraction, diffraction, and reflection processes, as demonstrated by the wave frequency rose diagram for significant waves larger than 1 m based on a 3.5 km to shore wave buoy observation north of DFG (Figure 8c).

This study thus chooses the representative annual wave force [29] to continuously act at a high water level for a period of time. The wave climate in the study area is dominantly wind sea wave [30]. The wind rose diagram and the wave rose diagram for the deep water (about 48 km from the shore) are presented in Figure 8. They are produced with 40-year (1979–2019) 1-h hindcast time series extracted from the 0.5-degree-resolution global wave model of Oceanweather Inc. In the offshore water, the prevalent big waves come from the NNE directions and the NNE wave is the biggest, with the largest frequency. The secondary prevalent wave comes from the ESE-SSE directions, and the SE wave has the relatively largest frequency. Nevertheless, due to the shielding and breaking effects of the extensive shallow sandy ridges (RSRs in Figure 1), the magnitude and frequency of the around-SE waves are reduced considerably and are negligible when they approach the shore in the study area, whereas the NNE waves propagate into the dominant N direction due to refraction, diffraction, and reflection processes, as demonstrated by the wave frequency rose diagram for significant waves larger than 1 m based on a 3.5 km to shore wave buoy observation north of DFG (Figure 8c).

Figure 7. Calibrations of the tidal current (a) and bottom suspended sediment concentration (b) at the gauge site.

Figure 8. The deep-water wind rose diagram (a) and wave rose diagram (b). Data for (c) is extracted from [31].
As a result, this study chooses the dominant NNE one-year-recurrence biggest significant deep-water wave as the representative annual wave force in the study area, i.e., $H_s$ is 2.92 m and the mean wave period is 6.8 s [21]. It propagates at the MHW for about 48 h.

The third-generation 2D wave model SWAN [32] is exercised to simulate the morphological adjustment solely under the action of the representative annual-magnitude wave force. The wave model uses the same numerical grid as FVCOM. It is driven by a constant NNE wind field, while the prescribed annual-magnitude wave propagates at the boundary nodes into the computation domain. The wave-driven longshore current is accounted for by coupling computations of SWAN and FVCOM, albeit the latter is prescribed only a static water level at the MHW.

### 4.2. Simulation Results

The southeastward flooding and northwestward ebbing tidal currents are reciprocal and parallel to the isobaths (Figure 9). The ebb duration is longer than the flood duration, whereas the flooding velocity is larger than the ebbing velocity. The tidal wave belongs to standing waves, with the maximum flow occurring at mid-tide. The spatially-mean depth-averaged current velocity for the respective flooding period and ebbing period varies by 0.17–0.57 m/s and 0.13–0.42 m/s. Correspondingly, the maximum depth-averaged velocity can reach up to 2.96 m/s and 1.96 m/s around the jetty head.

**Figure 9.** The flow at flood peak (a) and at ebb peak (b) and the depth-averaged suspended concentration at flood peak (c) and at ebb peak (d) of the spring tide. The coordinate axis unit is UTM (m).
The depth-averaged suspended sediment concentration can be up to 1.65 kg/m$^3$ in spring tides (Figure 9). Due to the flood-tide dominance, the net longshore transport of suspended fine sediments is also from the northwest to the southeast.

The simulated yearly accretion/erosion distribution for the 2019 year in the jetty-affected area is shown in Figure 10. It can be observed that the twin jetties are able to block longshore sediment transport driven by flooding/ebbing tidal currents at waters shallower than a 10-m isobath where the jetty tips are located (Figure 10A). As a result, sediment accretions occur on both the updrift side (northern) and the downdrift side (southern) of the twin jetties. Away from the two jetties, tidal currents are responsible for the accretions at both intertidal zones and subaqueous zones approximately shallower than a 4.0-m isobath. In other words, tidal currents do not induce downdrift intertidal flat erosion and shoreline recession. On the contrary, the ebb flow contributes to the accreting body adjoining the south jetty. Due to a flood-tide dominance, the updrift accretion volume is much larger than the downdrift counterpart. The seabed accretion in the deeper offshore area reflects the fact that a significant portion of eroded deposits from the AYRD is still transported by flooding tidal currents southeastward to RSRs on contemporary days.

**Figure 10.** Yearly morphological adjustment in 2019 by solely tide (A) and wave force alone (B) and the net output (C). The coordinate axis unit is UTM (m).
Wave force causes significant erosion at both intertidal flats and surf zone beds on the downdrift side of the jetties compared to the updrift side. The overall pattern does not show a classic response of “updrift progradation while downdrift recession” on a sandy beach (Figure 10B). The majority of erosion occurs within the active longshore transport zone roughly shallower than 9.1-m isobaths. In the deeper offshore area, the seabed experiences minor accretion as the wave bottom shear stress falls below the critical erosion stress of 0.79 N/m².

The net yearly accretion/erosion adjustment is obtained by adding the tidal output and the wave output together (Figure 10C). The downdrift shoreline erosion arc immediately emerges. It starts from the boundary of the 2.7-km-long accreting body bounding the south jetty (see Figure 3) and terminates at the northern flank of the DLG river mouth, very close to the eroding front identified on the satellite image. South to the DLG river mouth, the intertidal flat experiences accretion, with the shoreline concurrently advancing seaward (Figure 11), albeit net erosion occurs roughly below the 1.6-m isobath [21]. This indicates that the transitional spot from the eroding northern Jiangsu coast to the southern accreting Jiangsu coast has rapidly shifted a distance of about 39 km from the SYG mouth to the DLG river mouth in about eight years since the twin jetty began to construct.

Figure 11. The prograding salt marsh south to the “DLG” river mouth. UAV photo is looking southward and was taken on 29 June 2019.

Computed from the deposited body on both sides of the jetties in Figure 10C, the annual longshore sediment volume blocked by the north jetty is roughly 3.02 million m³, while the blocked annual longshore sediment volume by the south jetty is about 0.48 million m³. The annual siltation volume within the navigation channel to the Sheyang port (assuming no dredging activity since the port began operating in 2013) is 0.54 million m³. As the majority of the siltation volume within the navigation channel is contributed by flood currents, it should be added to the blocked sediment volume by the north jetty. So, the net southward longshore fine sediment transport volume is 3.08 million m³.

Meanwhile, the eroded volume by waves at the downdrift intertidal zone (bounded by MLWS line) from the boundary of the 2.7-km-long accreting body to DFG is roughly 1.69 million m³, whereas the accreted volume at the intertidal zone south to DLG is about 1.78 million m³, respectively. They are roughly equal to each other. Nevertheless, if the computed zone expands to the zone shallower than a 3.5-m isobath, the eroded volume and the accreted volume become 4.98 million m³ and 1.94 million m³, respectively, and the net erosion volume is thus 3.03 million m³, which is close to the blocked net longshore sediment transport volume of 3.08 million m³ by the twin jetties. This computed zone extends to a distance of about 67 km.

The above-outlined recovery process indicates that wave force is predominantly responsible for recovering the blocked net longshore sediment volume, playing a decisive role in controlling the downdrift erosion process.
5. Discussions

The assessment of the maximum shoreline recession length, its location, and the spatial scope of the downdrift eroding shoreline is of critical importance for an efficient, sustainable coastal erosion mitigation strategy. Considerable discrepancy exists if the planform shape of the downdrift eroding shoreline arc and its evolution with time are not reliably predicted. The short arc or the near arc advocated by Bruun [1] may provide a proximate tool to predict the maximum shoreline recession length. The magnitude or length of the maximum cut into the downdrift near arc could be estimated with reference to the angle $\alpha$ between the initial shoreline and the downstream arm of the near arc, which is $2-4^\circ$ (Figure 12). Alternatively, the maximum erosion site can be roughly estimated as the point where the wave-driven alongshore currents separate in opposite directions [6].

![Figure 12. Illustration of the downdrift shoreline erosion arc planform of Bruun [1]. Adapted with permission from 1995 The Coastal Education and Research Foundation, Inc., compared to a parabolic shoreline shape.](image)

Both Bruun [1] and Galgano [6] argued that at the point of maximum cut, sediments are transported in equal amounts to opposite directions and the reversal transport is attributed to the wave refraction process. It is this reversal sediment transport that causes the shoreline to prograde along the structure. In reality, the sediment accumulation at this shadow zone can be promoted by downcoast longshore sediment transport by waves on a sandy coast (e.g., [33]) or by the reversal tidal currents on a muddy coast, as the case in this study.

With the progressive sediment accumulation and concomitant shoreline progradation at the wave shadow zone, the near erosion arc will be pushed away from the jetty. This is probably the true reason for the downdrift translocation of the near erosion arc and the extension of the eroding front as well as the "sediment infilling" of the near erosion arc with time (Figures 5 and 7 in [6]).

The success of the "S"-shaped double-arc model of Bruun [1] relies on pinpointing a mildly advanced or no-recession "bump" point. The "bump" could be a result of a rock or other protruding outcrops in the nearshore water [1]. It could also be the site where a bypassing sand bar is attached to the shore, as in the case of the Shinnecock inlet, Long Island, NY, reported by Kraus and Galgano [3].

Nonetheless, such a mildly accreting, convex-curved bump zone is not always visible or identifiable on the downdrift shoreline when no sediment bypassing occurs or no longshore bar welds to the shoreline. The situation becomes even more complicated when a number of river mouths intersect the erosion arc with concave-curved mouth configurations, as in the case of this study.

However, if the reversed longshore sediment transport is lacking, the shadow zone will have no chance to accommodate the sediment to advance the shoreline. Instead, this zone will be eroded heavily by the relatively stronger reversal breaking-wave-driving current. The shoreline may eventually recede to a crescentic plan shape with the deepest cut point close to the jetty.

As to this type of downdrift shoreline erosion, one-line shoreline models can be resorted to predict the spatial extent and temporal behavior and thus pinpoint the location...
and magnitude of the maximum shoreline recession (e.g., [10,12]). Nevertheless, one of the key assumptions one-line models rely on is that the beach cross-shore profile maintains its constant equilibrium shape [34,35] during the shoreline advancing/retreating process. This concave type of equilibrium profile is normally applicable to sandy beaches where the bedload transport mode is dominant by the wave orbital movement. For a muddy coast where suspended fine sediment dominates the transport process, a suitable equilibrium profile for tidal flats is needed [26]. In addition, it is not easy to accurately determine the gross longshore sediment budget, and a small deviation of the incident wave angle from real wave propagation could cause significant computation discrepancy in the longshore sediment transport rate and consequent shoreline morphology [1].

As elaborated in Section 4, if only the intertidal zone is assessed, the sediment volume eroded within the 36-km-long downdrift arc is not enough to recover the blocked annual longshore sediment transport volume of 3.08 million m$^3$ by the twin jetties. The recovery needs additional sediments eroded from the inner surf zone (i.e., shallower than a 3.5-m isobath), which extends to a 67-km-long distance.

This may attest to the reasonability to doubt the premise used by existing shoreline models that the area of downdrift shoreline erosion equals the impoundment of sediment along the updrift jetty [6]. If the loss of the net annual longshore sediment transport volume is assumed to be recovered only within the downdrift shoreline recession area, the maximum shoreline recession length predicted by one-line models might be exaggerated, whereas the spatial extent of the receding shoreline might be underestimated.

Moreover, in contrast to what Bruun [1] reported, i.e., an eroding front normally moves downdrift at a rate of about 1–1.5 km/yr, it seems that the downdrift erosion stretch expands fast in response to the construction of the twin jetties of the Sheyang port. The jetties began to be built in March 2011. The bare intertidal flat to the south of the XYG mouth, an approximately 27 km distance from the south jetty, was subjected to remarkable scouring by July 2011 (Figure 13). This is probably due to the constraints to the shoreline recession imposed by the earth embankments of shrimp ponds north of the XYG mouth (Figure 4). As a result, the downdrift erosion arc expands downcoast at a much faster speed to recover the loss of longshore sediment transport volume. Nonetheless, once the recovery configuration is established, the eroding front will probably be stationary for a while or migrate downdrift slowly, depending on the accreting speed of sediments in the shadow zone of the jetty, and the receding speed of the erosional shoreline stretches. It can be envisaged that the erosion front will probably move southward faster due to the accelerated sediment deficit if the eroding shoreline and the intertidal flat are protected with revetment or groynes in the future.

![Figure 13. Rapid accretion-erosion transformation during October 2010–July 2011 (data are extracted from [15], with the datum shifting to MSL).](image-url)
Meanwhile, it is not an easy task to predict reliably the long-term downdrift shoreline evolution using a shoreline evolution model (e.g., [36]). The same is true for a process-based numerical model. The lower prediction reliability can be ascribed to both intrinsic and epistemic uncertainty [37].

One of uncertainties arises from the assumption that hydrodynamic forces, i.e., the tidal boundary condition and the incident wave climate, etc., remain unchanged for the long forecasting period, as those used for model calibration. Another uncertainty is induced by a lack of knowledge of the spatial distribution and evolution (horizontal and vertical) of non-uniform sediment properties of the beach and seabed, which can significantly influence the wave energy dissipation by bottom friction in the wave saturation zone [24]. For a muddy coast, the scenery becomes even more complicated due to the presence of salt marsh vegetation.

The presence of vegetation can greatly modify erosion and deposition processes on the marsh platform via attenuating wave energy and current velocity, increasing soil resistance to erosion by root mats, enhancing sediment trapping and belowground organic production by vegetation biomass, etc. [38]. In particular, salt marsh vegetation itself also experiences spreading, colonizing, and massive coverage when an intertidal flat progrades whereas it experiences dying and withering in the case of marsh recession.

One characteristic erosion feature at the salt marsh edge is the formation and retreating of nearly vertical scarps (Figure 5A). The scarp is produced by the combined boundary processes involving hydrodynamic, soil mechanics, plant dynamics, etc.

In particular, scarp erosion chiefly occurs through a series of mass failure processes, such as cantilever failure, toppling failure (Figure 5B), and rotational slip [39]. Such mass failure processes become non-uniform both in the horizontal and in the vertical direction under wave impact and thrust movement and consequently produce a jagged or saw-tooth-like marsh boundary intersected with tidal creeks.

Due to the complexities involving wave striking, mass failure, vegetation dynamics, and sediment movement, the state of the art for the process-based numerical simulation of the marsh edge recession, to the best of the authors’ knowledge, has remained at the 1D cross-shore model (e.g., [38,40,41]) that is incapable of describing the formation and retreating of the jagged marsh boundary. Thus, the accurate simulation of the long-term downdrift vegetated shoreline evolution using 2DH process-based numerical models remains a challenge and needs further intense study in the future.

An empirical equation (Equation (1)) that correlates the long-term recession rate of a marsh scarp to the averaged wave power [42] based on historic (often natural) shoreline data can be invoked in 2DH numerical modeling. The incident wave power or the wave energy flux is not distributed uniformly along the downcoast, as demonstrated by the remarkable variability in the shoreline receding rate (Figure 3). A mandatory boundary receding process may give rise to a considerable deviation from the engineering-intervened shoreline evolution in the long run.

\[ E = 0.35P^{1.1} \]  

where \( E \) is the recession rate (m/yr) and \( P \) is the wave power (kW/m) in front of the marsh edge.

Alternatively, a parabolic shape model may be exercised to estimate approximately the potential magnitude of shoreline recession. Such a model could predict a static equilibrium embay planform if a downcoast control point can be located with uncertainties examined [12,43,44]. It is normally applicable to the scenery in which the equilibrium planform is determined by the dominant incident wave and the concomitant net longshore sediment transport.

6. Concluding Remarks

A shore-crossing structure can trigger the updrift/downdrift shoreline adjustment due to the blockage of longshore sediment transport. This study presents a unique example
to discern the downdrift shoreline and the intertidal flat erosion pattern emerging on the open muddy coast in Jiangsu province, China, post the construction of two long-rang jetties at the Sheyang River mouth. The northern jetty is 7.9 km long, and the south jetty is 7.8 km long. The net longshore sediment transport is from the north to the south due to the flood-tide dominance.

As identified in high-resolution satellite images dated December 2013 and March 2019, a 2.7-km-long shoreline advances immediately downcoast of the south jetty; further southward, a 36-km-long shoreline experiences remarkable retreats at varying alongshore rates, ranging from 2.1 m/yr to 30.3 m/yr, with a spatially averaged rate of about 14.1 m/yr. The planform shape of the downdrift eroding shoreline arc does not resemble the classic “S” plan shape of Bruun [1], a crescentic plan shape, or a parabolic shape but an irregularly indented curved shape. Topographic transect of different periods also reveals that the downdrift coast responds almost immediately to the jetty construction, from the original accretionary scenery to the erosional regime.

FVCOM and SWAN, 2DH process-based numerical models, are used to simulate the flow, the non-uniform sediment transport, and the yearly-magnitude erosion/accretion distribution in the jetty-affected area. The distinct role played by the reciprocal tidal current and the incident wave in governing the downdrift erosion process is ascertained by performing separate simulations forced solely by a representative spring tide and a representative annual-magnitude wave force alone. The simulated downdrift shoreline erosion arc is close to the spatial extent identified on high-resolution satellite images. The net annual longshore sediment transport volume of 3.08 million m$^3$ blocked by the twin jetties is recovered by the downdrift eroded material, including both intertidal zone and inner surf zone, with a spatial extent of 76 km. In contrast to the counterparts on sandy coasts, the tidal flow is the dominant constructive force for the accretion of muddy sediment at the updrift/downdrift intertidal and surf zones approximately shallower than a 4.0-m isobath, whereas big wind waves play a decisive role in initiating and developing the downdrift erosional process.

The assessment of the maximum shoreline recession length, its location, and the spatial scope of the downdrift eroding shoreline is of critical importance for an efficient, sustainable coastal zone management strategy. These key parameters for quantitatively pinpointing the impact of a hard structure on the adjacent shoreline and flat (beach) depend on the reliable prediction of the planform shape and its evolution with time. In this regard, the Bruun model, the one-line model, the headland-bay model, and the process-based 2DH numerical models may be used for a muddy coast where marsh vegetation is extensively present, with the feasibility and limitation carefully examined.

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