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Hydro-sediment dynamics in an abandoned estuarine pond under artificial mangrove restoration

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ABSTRACT

Mangrove wetlands are highly productive and provide essential services to coastal communities. However, global mangrove extent has declined, underscoring the need for restoration, particularly in abandoned aquaculture ponds. In this study, we conducted in-situ vegetation, hydro-sediment, and morphological observations, coupled with an idealized numerical model, in a former aquaculture pond in an estuary. Measurements revealed that tidal asymmetry shifted from ebb dominance to flood dominance along the inflow path, with the flood-to-ebb water flux ratio increasing from 0.4 in the outer channel to 1.5 in the inner pond. Currents exhibited a vertical structure characterized by opposite directions between upper and lower layers in the pond region. From outside in, bed shear stress decreased sharply from a peak of 1.4 Pa to 0.2 Pa, remaining below the critical threshold for deposition throughout the entire tidal cycle within the pond. Additionally, near-bed SSC during the flood phase (0.1 kg/m^3) was nearly 1.5 times that during the ebb phase (0.07 kg/m^3) , with suspended sediment flux showing a clear decline from the inlet to the pond's interior. Planted mangrove heights and canopy widths increased by 17 % and 40 %, respectively, with an overall mineral accretion of 5 cm observed in the pond over a nine-month period. The numerical simulation results show that both the increase in pond dike closure and the early growth of mangroves exhibited a non-linear effect (initially increasing and then decreasing) on flood-dominated asymmetry and bed shear stress ratio. Moreover, increasing planting density reinforced these effects from the scenario analysis. Our study highlights a shift in hydro-sediment dynamics within a mangrove-restored pond, offering insights into optimizing topographic closure and planting density in abandoned ponds to promote sustainable mangrove restoration.

1. Introduction

Mangrove forests are vital ecosystems in estuarine wetlands of subtropical and tropical regions throughout the world, providing various services such as wind and storm protection (Temmerman et al., 2023), carbon sequestration (Donato et al., 2011), air and water purification (Morrison et al., 2023), and ecological habitat (Loiola et al., 2023). However, despite covering an area surpassing 140,000 km² (FAO, 2023), global mangrove area has been declining over the past two decades, likely due to both human activities and climate change (Murray et al., 2022). Undoubtedly, the degradation of mangroves will lead to increased flooding and resource scarcity, affecting coastal communities (Gijsman et al., 2021).

The conversion of mangrove to aquaculture land is a major contributor to the decline of mangrove wetlands in many regions (Richards and Friess, 2016; Long et al., 2021; Hagger, 2022; Prakoso et al., 2023; Jia et al., 2023; Xiong et al., 2024). This loss has been recognized and reversed through various restoration strategies in countries such as Indonesia, Vietnam, Brazil, India and China (Joffre et al., 2015; Djamaluddin et al., 2018; van Bijsterveldt et al., 2020; De Lacerda et al., 2021; Gou et al., 2024). Several studies have suggested that integrated mangrove aquaculture (i.e. combining estuarine pond ecosystems with mangrove forests) could be a viable option for mangrove rehabilitation, productivity maintenance, and ensuring

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sustainable management (e.g., Kautsky et al., 2000; Krishnapriya et al., 2023; Ahmed et al., 2023). However, due to historical overexploitation and increased contamination, many aquaculture ponds that incipiently replaced native mangrove populations have been abandoned, making damaged ecological functions difficult to recover immediately (Primavera, 2005; Djohan et al., 2015; Amelia et al., 2023). To address this issue, afforestation has been reported as an effective strategy for ecosystem restoration in abandoned pond areas, following the modification of hydrological and elevation conditions (Matsui et al., 2010; Ouyang & Guo, 2016; Song et al., 2023). This approach stresses the critical role of human involvement in the initial restoration of severely damaged mangrove-pond ecosystems (Bliege Bird and Nimmo, 2018).

Despite the significant role of artificial reconstruction in the early stages of restoration, the key factor for the subsequent success is whether the integrated mangrove-pond habitat can sustain natural development (Alongi et al., 2002; Makowski and Finkl, 2018). Some researchers have indicated that significant biogeochemical differences may arise as restored vegetation coverage expands, such as variations in carbon storage, nitrogen fluxes, and microbial communities that support the ecological cycle within the regenerating habitat (e.g., Sidik and Friess, 2020; Lovelock and Reef, 2020; He et al., 2022). However, the primary concern for afforestation and recruitment is establishing an appropriate bed level to regulate tidal inundation and saline intrusion (van Bijsterveldt et al., 2020; Xiong et al., 2021; Huang et al., 2023). This requires stable hydrological conditions and sufficient sediment availability to provide a solid foundation for mangrove colonization (Lovelock et al., 2015; van Hespen et al., 2022). Using an ecomorphological model and satellite-derived sediment concentration in a natural mangrove wetland, Breda et al. (2022) demonstrated that soil elevation and wetland evolution should be maintained over the long term under periodic tidal inundation, even in the face of storm events. In addition, considering restoration along a living shoreline, Kibler et al. (2019) conducted a hydro-morphodynamic comparison between nonvegetated, mature vegetated, and restored vegetated regions, suggesting significant flow energy dissipation and a delayed contribution to land gains and plants recruitment. In other words, appropriately elevating the mineral surface is beneficial for sapling growth and restoration, with hydro-sediment dynamics playing a key role during the near-natural evolution of mangrove ponds (Xie et al., 2022; Glover et al., 2023).

Clearly, the presence of vegetation reshapes the hydrological regime, adding complexity to sediment dynamics in mangrove ponds (Capo et al., 2006; Asp et al., 2018; Yoshikai et al., 2022; Fanous et al., 2023). Field studies have indicated that mangrove plants-with their pneumatophores, rigid stems and canopy structures-coupled with vegetation density and coverage, play a significant role in attenuating fluid motion and dissipating wave energy (Hu et al., 2014; Yoshikai et al., 2022; Zhou et al., 2022). Analogous field observations have been conducted on small scales in afforested mangrove areas (Kibler et al., 2019; Best et al., 2022), one of which described a single velocity profile correlated to vertical plant structure, from root to canopy (Chang et al., 2020). However, most studies on horizontal variation of vertical flow profiles through vegetation were carried out indoors using physical models or numerical simulations (e.g., Nepf, 2012; Le Minor et al., 2019). These studies primarily focus on changes in unidirectional currents while often overlooking the potential effects of flood and ebb tides. Moreover, as afforested vegetation grows, they start to affect hydrodynamics on regional scales and exert influence on long-term evolution (Xie et al., 2022). This last aspect is still rarely considered in restoration scenarios. In general, although the mechanistic theory of how mangrove plants influence hydrology and, consequently, sediment dynamics has been extensively discussed (e.g., Willemsen et al., 2016; Bryan et al., 2017; Schettini et al., 2020), there remains a lack of research on hydromorphodynamics in field-scale restoration settings within tidal environments.

Given that most estuarine aquaculture ponds are located along semi-

enclosed shorelines near human settlements (Herbeck et al., 2020), the topographic factors influencing the hydro-morphodynamics of pond-tomangrove restoration are crucial. Furthermore, hydrological reestablishment tends to create a small-scale inlet-basin system within abandoned ponds (Djamaluddin et al., 2018), altering sediment transport pattern and substrate elevation. Some researchers have suggested that the combination of constricted topography and entrance channels can induce flood-directed shifts in tidal current asymmetry, based on 2D modeling and observations of tidally dominated basins at scale of 10-100 km (e.g., Moore et al., 2009; De Ruiter et al., 2019). Additionally, by applying different topographic combination variables in a 1D model, external tidal asymmetry has been shown to strongly influence the magnitude of net sediment transport, determining whether it leads to export or import (Ridderinkhof et al., 2014). For inlet-basin systems on a kilometer scale, flood dominance is more prevalent and often results in sediment infilling in the basin (Golbuu et al., 2003; Lehman et al., 2015; Hanegan et al., 2023). At the meter scale, Fleri et al. (2019) has examined the morphological and vegetative effects on accretion in a restored tidal marsh by combining field observations with numerical models. However, despite the fact that over 25 % of mangroves have been cleared for aquaculture worldwide (Richards and Friess, 2016; Duan et al., 2020), there remains a significant gap in understanding the role of topography in hydro-sediment dynamics at relatively small scales. This gap calls for detailed studies to ensure successful restoration of mangrove in abandoned ponds.

In this study, we investigated the hydro-sediment dynamics and its morphological implications at a newly afforested mangrove pilot site, where a monoculture of Kandelia obovata was introduced through artificial planting in a previously abandoned pond. To reveal the hydromorphological processes and examine human and vegetation influences, we employed a combination of field measurements and numerical simulations. The Results section presents detailed observations of hydrological conditions, sediment fluxes, vegetative morphology, and bed level changes. To further explore the influences of three critical factors that are difficult to measure in practical mangrove restoration within a pond-namely topographic modification, planting density, and plant growth-we developed a physics-based Delft3D model, calibrated and validated using field observations. Through this model, we investigated how these factors influence horizontal and vertical hydrodynamics, which are critical for sediment transport and deposition, respectively. A quantitative assessment of these interactions is provided in the Discussion Sections 4.2-4.3, followed by the patterns of vegetation-hydro-morphodynamic feedbacks within an estuarine pond undergoing early-stage mangrove restoration in Section 4.4.

2. Materials and Methods

2.1. Study area

The pond-to-mangrove pilot site, known as the Kongquewan (KQW) Mangrove Reserve, is located along the Jingu River at the northern end of Qinzhou Bay in Beibu Gulf, Guangxi, China (Fig. 1A–C). The Jingu River has an average depth of less than 4 m (Yang et al., 2019) and an annual average runoff of approximately 1.5×10^8 m³ (Chen et al., 2016). Characterized by irregular diurnal tides with higher velocities during ebb than flood tides, the mean tidal range at the river mouth is about 1 m during neap tides and 4 m during spring tides (Zhang et al., 2010; Lin et al., 2024). Many mangroves grow along the riverbank and play a crucial role in the ecological construction of the modern industrial development area, China-Malaysia Qinzhou Industrial Park (Chen et al., 2016).

The restoration pilot site, formerly used for aquaculture, has undergone artificial hydrological and geomorphological reconstruction since 2021 (Fig. 1D, E, F). Saplings of *K. obovata* were cultivated and colonized onto the reconstructed substrate within the 30,000 m² restoration zone, which was divided into two blocks featuring distinct dike



Fig. 1. Study area. (A) Location of Beibu Gulf (BBG); (B) Location of Qinzhou Bay (QZB); (C) Location of Jingu River and Kongquewan Mangrove Reserve (KQW). KQW (D) before and (E) after topographic modification. (F) Excavation for topographic modification; (G) Cultivation and colonization of *K. obovata*. Field scene (H) before afforestation in August 2021, and (I) after afforestation in June 2023.

closures and bordered by native mangrove stands (Fig. 1E, G). In the western part of the area, an enclosed pond was breached with a 2-meterwide water gate, while a more open platform was constructed at a higher elevation on the eastern side. A tidal inlet from the Jingu River traverses the entire restoration area, maintaining hydrological connectivity, and forming a 30° angle with the gate of western pond (Fig. 1E).

The K. obovate is one of the mangrove species found along the sandy, muddy, and mixed swamps of southern Vietnam, southeast China, Japan and Indonesia (FAO, 2023). It has a moderate growth rate, reaching heights of 2–3 m when mature (Duke et al., 2010). The plant features a grayish, smooth trunk, dark green, elliptic-obovate leaves, and lacks pneumatophores protruding from the substrate (Sheue et al., 2003). New plants can be cultivated and colonized via separate propagules from the parent tree, which need abundant water but minimal maintenance (Van Hespen et al., 2022). K. obovata has been planted in suitable areas to protect coastal communities from flood and storm impacts due to its strong prop trunk and dense canopy (Duke et al., 2010). Additionally, the recolonization of this species has been shown to restore the soil physicochemical properties and microbial community structure (Lin et al., 2021). Here in the pilot afforested area, individuals were initially planted at approximately 35 cm intervals, thus the saplings now exhibit a relatively uniform distribution. The mangrove plants have been

growing for more than two years in the pilot site and continue to show visually healthy growth (Fig. 1H, I).

2.2. In-situ observation setup

In the pilot area, we conducted integrated surveys focusing on biogeomorphodynamics. These surveys encompassed vegetation parameters, hydrodynamic processes (including currents and waves), turbidity, and elevation changes. Eight stations were set up in the pilot site for hydrological data acquisition, along with eight 2 m \times 2 m quadrats for plant morphology measurements. These stations were labeled as L/R1-4 and L/R A-D, arranged in ascending order from the interior to the exterior, respectively (Fig. 2A). Notably, hydrological stations labeled L4 and R4 were positioned near the outer channel, drained during low water levels (Fig. 2D), while the remaining stations were situated within the afforested pond (Fig. 2E). Vegetation quadrats were also placed within the afforested zones (Fig. 2C). A joint instrument frame, equipped with a 1 or 2 MHz Aquadopp Acoustic Doppler Current Profiler (HR-ADCP), T-wave 101 Gauge and RBRconcerto³ CTD + Tu logger or OS 304Plus CTD (Fig. 2B), was deployed to measure vertical velocity profile, high-frequency pressure and turbidity. The setup of the joint frame varied between stations due to the instrument configuration but the



Fig. 2. Field observation setup. (A) Overview of all observation locations: The blue pentagons represent the locations of hydrological stations, labeled L1-4 and R1-4 in blue on the western and eastern sides, respectively, with the numbers increasing from the shore to the river; the green squares indicate the locations of 2 m \times 2 m quadrats for vegetation morphological parameters measurement, labeled LA-D and RA-D in green on the western and eastern sides, respectively, with numbers ordered alphabetically from the shore to the river; the purple lines represent fixed elevation measurement transects, labeled T1-4 in purple, with numbers increasing from western side to eastern side. Two cross-sections are present in each pond on the western and eastern sides. The background contour map shows the surface elevation model obtained by drone in June 2023, calibrated to the local mean sea level, with positive values above mean sea level and negative values below mean sea level. (B) Diagram of the joint instrument frame for hydrological observation. (C) Measurement of vegetation morphological parameters. (D) Field observation environment at the station L4 and (E) L3. Note that all the stations are exposed to air during low tide. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

functions were the same (Fig. 2D, E). Wave gauges were deployed only at *L3*, *L4* and *R3*, *R4* to measure wave attenuation from outer channel. Initially installed in the eastern partition, the frame was relocated to the western partition one day later, covering a representative spring tidal cycle from June 18th to 20th in 2023 (See Table 1 for configuration parameters and Table S1 for complete overview of setup information).

In terms of vegetation survey, due to the uniformity of the afforestation layout, with approximately 35 cm between each pair of plants, the plant density remained consistent during the early sapling stage. Therefore, we selected 20 individuals for measurement. The following

Table 1

Instruments configuration and measurement parameters.

Instrument / Materials	Observation Object	Setting Parameters	Time
HR-ADCP (Nortek Inc., Norway)	Velocity Profile; Pressure	Burst Int.: 60 s; Duration: 30 s; Cell size: 30 mm	2023/6/ 18 ~ 2023/6/20
T-Wave (HydroSurvey Ltd., China)	High-frequency pressure	Burst Int.: 1200 s; Duration: 1024 s; Sample Freq.: 4 Hz	
CTD (RBR Ltd., Canada) / (Idronaut s.r.l, Italy)	Turbidity; Pressure	Continuous Mode: 1 Hz	
Quadrat Survey	Vegetation morphology	2 m × 2 m; Distance: 7 m	2023/06/ 18 2023/11/ 14 2024/03/ 30
RTK GNSS	Elevation	Distance: 3.5 m	2023/06/ 18 2023/ 11/14 2024/03/ 30
UAV images	Geomorphology		2023/6/18

parameters were recorded: plant height ($H = H_t + H_c$, where H_t and H_c refer to the trunk height and canopy height, respectively), canopy width ($W_c = \sqrt{ab}$, where a, b represent the lengths of the major and minor axes of the canopy), diameter at basal stem (D_s , measured at 5 cm above the ground) and breast height (D_{bh} , measured at the first branch junction) (Fig. 2C). Vegetation morphologies were measured three times, in June and November 2023 and in March 2024. The maximum and minimum values were excluded from each set of data prior to statistical analysis.

Drone aerial imagery and Real-Time Kinematic (RTK) GNSS surveys were utilized to obtain bed level data. A drone flight was conducted in June 2023, accompanied by a complete RTK elevation calibration of the entire restoration area. Two additional RTK elevation surveys were conducted along four transects (labeled *T1-4*) in November 2023 and March 2024 (Fig. 2A). The elevation data were calibrated to the local sea level datum, with the differences indicating deposition or erosion between surveys. In addition, bed surface sediment samples were collected across transects *T1-4* on June 20th 2023, and sediment grain sizes were analyzed using an LS-13–320 Particle Size Analyzer in the laboratory. The results of the cumulative percentage distribution curve of grain size, are shown in Fig. S2.

2.3. Hydro-sediment data processing

HR-ADCPs were positioned with an upward looking orientation and slightly elevated from the substrate to prevent sediment deposition. Velocity data were collected every minute and averaged over the first 30 s of each burst, with a cell size of 3 cm and a fixed blanking distance from the ground (Table S1). Raw velocity data with beam correlation lower than 75 % were considered unreliable and discarded (Nortek AS, 2008; Horstman et al., 2011). To validate the water column measurements, velocity data were filtered based on the water level derived from the collected pressure data. Outliers within the valid data were masked and replaced with the rolling average of nearby values. Subsequently, data from the two horizontal orthorhombic averaged velocity beams per minute (i.e., \bar{u} and \bar{v}) were processed to derive the dominant velocity magnitude U(z) in the main flood-ebb direction Φ . For each observation station, Φ was determined by maximizing the root-mean-square velocity \bar{u}_{rms} (You, 2006),

$$\overline{u_{rms}} = \left[\frac{1}{N} \sum_{i=1}^{N} (\overline{u_i} \sin\Phi + \overline{\nu_i} \cos\Phi)^2\right]^{0.5}$$
(1)

where *i* is the index of bursts, *N* is the number of bursts. Although the

streamwise directions of all cells of the profile varied, the mean value was treated as the dominant direction (Φ) for each station (Fig. 4). The velocity in each cell, U(z), as a function of cell position z, was then calculated as

$$U(z) = \overline{u}(z)\sin\Phi + \overline{v}(z)\cos\Phi$$
(2)

Subsequently, two types of characteristic velocity were obtained. The first is the vertically averaged velocity, which represents the instantaneous flow state of the water column, formulated as

$$\overline{U_{\nu}}(t) = \frac{1}{\eta} \int_0^{\eta} U(z,t) \, dz \tag{3}$$

where η is the water level at moment *t*, and *z* is the vertical position. The other is the averaged velocity of each layer in the flood or ebb direction over the tidal cycle, calculated as

$$\overline{U_{p,f}} = \frac{1}{A_f} \iint_{D_f} U(z,t) \, dA \tag{4}$$

where A_f is the area of the flood domain D_f , which is the integral of inundation depth over time during the flood phase (The word *domain* here refers to combined effects of tides, considering both the temporal and spatial influences of flood or ebb tide). Similarly, $\overline{U_{p,e}}$ represents the averaged velocity during the ebb phase. By calculating this averaged velocity, the contribution of reverse flow to the velocity profile at the same time can be separated, allowing the water flux, $\overline{U_p} \cdot A_f$, to be derived for different tidal phases. Additionally, the durations of flood and ebb phases were extracted at both the water bottom and surface to explore the vertical differences in tidal phases.

Pressure data were collected at a frequency of 4 Hz, with a 20-minute interval and a sampling duration of 1024 s. After removing the lower envelope of the time series, the raw pressure data were converted to water levels in meters at high frequency. Wave parameters, such as significant wave heights and periods, were then computed using the zero-crossing method (Tucker & Pitt, 2001).

Using the Nephelometric Turbidity Unit (NTU) probes, CTDs were primarily employed to measure suspended sediment concentration, while also recording the pressure data (Table S1). Date were acquired at a frequency of 1 Hz and averaged every minute. Additionally, surface sediment samples were collected in the field and brought back to the laboratory, for calibration against the optical turbidity values. The calibration curves for all the probes are shown in Fig. S1, demonstrating a strong linear correlation with high confidence ($R^2 > 0.98$, p < 0.001). These curves were then used as a basis for converting turbidity data into suspended sediment concentration (SSC).

Since all probes were located 10–30 cm above the bed, two dimensionless indices of near-bottom sediment dynamics during a spring-tide day can be derived. The first is the suspended sediment retention (*SSR*), which assesses the sediment availability influenced by external tidal dynamics and potential for sediment capture under tides. It was calculated as follows (Adame et al., 2010; Schettini et al., 2020):

$$SSR = \frac{\overline{SSC_{flood}} - \overline{SSC_{ebb}}}{\overline{SSC_{flood}}}$$
(5)

Another parameter is the suspended sediment flux (SSF_T), defined as the integral of the product of point velocity and SSC:

$$SSF_T = \frac{1}{\rho_s} \int_0^T SSC(t) \cdot A_w \cdot U(t) dt$$
(6)

where *T* represents the time duration of one ebb and flood tide at the sediment surface, U(t) is the post-processed velocity at the nearest cell to the probe position, A_w is the unit flow area, and *t* is the unit time of averaged variables. *SSF*_T directly reflects the volume of suspended sediment transport over a tide cycle.

2.4. Bed shear stress estimation

Bed shear stress is a key parameter for evaluating the potential for sediment movement induced by the hydrodynamics, with higher values indicating intensified erosion (Hu et al., 2015). Currents and waves are the two primary drivers of bed shear stress, which can be derived from observed velocity profiles and wave parameters, respectively.

The combined bed shear stress τ_m induced by both current and wave, is calculated as (Soulsby & Clarke, 2005):

$$\tau_m = \tau_c \left[1 + 1.2 \left(\frac{\tau_w}{\tau_c + \tau_w} \right)^{3.2} \right] \tag{7}$$

Since the bed shear stress induced by waves is small, both at the outer and inner stations, being $3 \sim 4$ orders of magnitude lower than that induced by current (i.e., $\tau_w \ll \tau_c$, see Fig.S8 for detailed information), the total bed shear stress τ_m reduces to $\tau_m \approx \tau_c$.

Current-induced bed shear stress τ_c is calculated using the formula:

$$\tau_c = \rho_w u_*^2 \tag{8}$$

where ρ_w is the water density, and u_{*} represents the friction velocity, which is determined by the logarithmic law:

$$\frac{u(z)}{u_*} = \frac{1}{\kappa} ln \left(\frac{z}{z_0}\right) \tag{9}$$

where κ is the von Kármán constant and equal to 0.4, u(z) is the velocity at a given height z of the profile. Using the least-squares method, the slope of the linear fitfor $u(z) \ln(z)$ is taken as $\frac{u_*}{\kappa}$, allowing u_* to be derived. Notably, u(z) must adhere to the logarithmic profile for the friction velocity u_* to be reliably obtained, which imposes certain limitations on the applicable range of water depths (Wilcock, 1996).

To assess the intensity of shear stress against bed erodibility, critical bed shear stress is a key indicator (van Rijn, 2020). For sand-mud bed mixtures, the critical bed shear stress τ_{ce} can be derived from:

$$\tau_{ce} = \phi_{cohesive} \tau_{ce,o} = (1 + p_{mud})^{\beta} \tau_{ce,o}$$
⁽¹⁰⁾

where the cohesive coefficient $\phi_{cohesive}$ is a function of the mud fraction percentage and the coefficient β , equated as

$$\beta = \left[1 + \left(p_{clay}/p_{mud}\right)^{a1} + \left(\rho_{dry,mix}/\rho_{dry,max}\right)\right]^{a2} \tag{11}$$

where p_{clay} is the percentage of particles with a diameter $< 8 \mu m$; p_{mud} is the percentage of particles with a diameter $< 63\mu m$; and $\rho_{dry,max}$ is the maximum dry bulk density of the mixture, which is $1,600 \text{ kg/m}^3$ here. $\alpha 1$ and $\alpha 2$ are empirical coefficients, here $\alpha 1 \approx 2$, $\alpha 2 \approx 1.5$ for sand fraction and ≈ 2 for mud fraction. $\rho_{dry,mix}$ is the dry bulk density of mud fraction, estimated using content of organic materials, clay, silt and sand of the sediment (van Rijn, 2020):

$$\rho_{dry,mix} = \left(1 - p_{org}/100\right) \left\lfloor 400 \left(p_{clay}/100\right) + 800 (p_{silt}/100) + 1,600 (p_{sand}/100) \right\rfloor$$

(12)

 $\tau_{ce,o}$ is the critical bed shear stress of cohesionless particle, which can be derived from the Shields Curve using the following formula:

$$\theta_{cr} = \frac{0.3}{1+D_*} + 0.055 \left(1 - e^{-0.02D_*}\right) \tag{13}$$

 $\tau_{ce,o} = \theta_{cr}[(\rho_s - \rho_w)gD]$ (14)

where D_* is the dimensionless grain size, calculated as $D_* = D\left[\frac{(\rho_s/\rho_w-1)g}{\nu^2}\right]^{1/3}$. The sand density $\rho_s = 2,650 kg/m^3$, and the kinematic viscosity coefficient $\nu \simeq 10^{-6}m^2/s$. The critical shear stress for sand fraction ($\tau_{ce,s}$) and mud fraction ($\tau_{ce,m}$) were computed, with *D* assigned as the median grain size for sand fraction and 63µm for mud fraction, respectively.

Additionally, the critical shear stresses for deposition of suspended particle, τ_{cd} , was calculated using the empirical formula (Berlamont et al., 1993; Carrasco et al., 2023):

$$\sqrt{\frac{\tau_{cd}}{\rho_w}} = \begin{cases} 0.008, w_s \le 5 \cdot 10^{-5} m/s \\ 0.094 + 0.02 \cdot log_{10}(w_s), 5 \cdot 10^{-5} < w_s \le 3 \cdot 10^{-4} m/s \\ 0.023, w_s > 3 \cdot 10^{-4} m/s \end{cases}$$
(15)

where w_s is the settlement rate of the cohesive sediment aggregates, calculated using the modified Stokes' velocity:

$$w_s = \frac{\alpha}{\beta} \frac{(\rho_s - \rho_w)gD_{50}^2}{18\rho_w \nu} \tag{16}$$

with shape factors $\alpha = \beta = 1$ for ideal spherical particles.

Finally, the bed shear stress induced by dynamics were compared with the critical shears to analyze the intensity of vertical sediment dynamics in both space and time within the restored area.

2.5. Numerical model analysis

In addition to the field data, a numerical analysis was conducted to investigate the human and vegetation factors on hydro-sediment dynamics of the study region utilizing the widely used, process-based model Delft3D (van Rijn, 1993; Lesser et al., 2004). A simplified grid with a 2-meter horizontal resolution was constructed to represent the shape of the *L*-labeled region, which consists of a rectangular pond zone and a waterway parallel to the outer edge (Fig. 3). The bathymetry assigned for the grids was based on the hypsometry derived from the UAV images and calibrated by the RTK-GNSS system. The fusion of the in-situ pressure data from Station L4 and tidal level data from the LMG

station in Qinzhou Bay (available at https://www.chaoxibiao.net/) were used to define the boundary tidal level propagating into the pond (Fig. S6, Fig. 3). The vertical plant structure was considered using the (Rigid) 3D Vegetation model, with cylindrical stem density and vertical width as the primary input parameters (Fleri et al., 2019). Plant density was set to 0, $4/m^2$ and $9/m^2$, representing no vegetation, sparse vegetation and dense vegetation, respectively, notably with 9/m² reflecting the actual planting density in the field. The vertical width distributions are derived from the vegetation surveys mentioned above. The cylindrical drag coefficient C_D is set to 1.0 in the model to mimic the effects attributed solely to plant shapes. Due to minimal impact of wave (see Fig. S8), only the FLOW module was employed for 3D hydrodynamic simulations, utilizing the σ coordinate system with 10 uniformly distributed vertical layers (Huang and Spaulding, 1996). Time step was set to 3 s to satisfy the Courant-Friedrichs-Lewy criterion (Deltares, 2020). Through calibration and validation, the model performed excellently in reproducing observed water levels and depth-averaged velocities (Fig. S3). Subsequently, two key bio-geomorphological factors-dike closure and plant growth-were examined by developing multiple scenarios based on reasonable assumptions derived from in-situ observations, totaling 30 and 18, respectively (Fig. S4, Table S2). All three plant densities were considered in the analysis of closure effects, and two densities— $4/m^2$ and $9/m^2$ —were used in the plant growth simulations, amounting to a total of 126 simulated scenarios.

Based on earlier work, we extract tidal asymmetry and bed shear stress ratio as two quantities characterizing horizontal sediment transport and vertical bed level change, respectively (Ridderinkhof et al., 2014; van Rijn, 2020). Tidal asymmetry was expressed as the skewness of the one-dimension flood-ebb speed at each grid point, formulated as:

$$Sk_U = \frac{T}{(T-1)(T-2)} \sum_{i=1}^{T} \left(\frac{U_i - \overline{U}}{s}\right)^3$$
(17)

where *T* is the timespan of the entire tidal cycle, U_i is the dominant velocity at moment *i*, \overline{U} is the average velocity, and *s* is the standard deviation of U_i , derived from equation $s = \sqrt{\frac{1}{T-1}\sum_{i=1}^{T} (U_i - \overline{U})^2}$. Since velocities in the flood (ebb) direction were processed as the positive (negative) values, flood-dominant asymmetry was inferred when $Sk_U > 0$, and vice versa (Guo et al., 2019). Notably, since there is a nearly proportional relationship between the modeled and observed depth-averaged velocities (Fig. S3 G-I), it can be easily inferred that the modeled Sk_U exhibits similar performance to the in-situ observations, despite the underestimation of velocity magnitude.

The bed shear stress ratio was defined as the ratio of bed shear stress to the estimated critical shear stress, τ_m/τ_{cr} . This is physically correlated to the simplified mass conservation for the bed level change derived by



Fig. 3. Idealized model grid and bathymetry of the mangrove afforested pond. The bathymetry was integrated into an idealized rectangle based on the surface elevation model obtained by drone. Tidal boundary condition is determined by the water level measured at station *L4*, which was also derived from data of the external tidal gauge through a calibrated conversion. Green dots represent the calibration and validation points, corresponding to the location and elevation of the field hydrological stations. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

Tao et al. (2024):

For no erosion scenarios,

$$\frac{\partial z_b}{\partial t} = \frac{cw_c \left(1 - \frac{\tau_m}{\tau_{cd}}\right)}{1 - p} \#$$
(18)

For no deposition scenarios,

$$\frac{\partial z_b}{\partial t} = -\frac{M\left(\frac{\tau_m}{\tau_{ce}} - 1\right)}{1 - p} \#$$
(19)

The two equations indicate a positive linear correlation between bed shear ratio and bed level increase under the condition that factors p, c, w_c or M are fixed. In addition, the estimation errors of bed shear stress can be corrected by multiplying by the square of the error ratio between observed and modeled velocities (Fig. S3G–I). Hence, both the two dimensionless values can be accurately estimated through numerical simulations. In particular, a focus region densely populated with K. obovata saplings was framed from the model for statistical analysis of the computational results, which aligned with the field observation stations (Fig. 3).

3. Results

3.1. Morphological features of vegetation

Over a period of more than nine months, changes in height and crown width of the afforested *K. obovata* saplings were documented through three observations (Fig. 4). In general, the temporal and spatial

Table 2		
Subregional median values of median	n vegetation parameters of each quadra	t.

Date	ate 2023/6/18		2023/11/	2023/11/14		2024/3/30	
Section	L	R	L	R	L	R	
Н	105.9	75.0	117.6	84.6	118.9	93.1	
H_t	*	_	65.9	52.1	59.6	53.2	
H_c		_	54.6	30.6	59.1	37.2	
W_c	26.9	22.7	37.3	29.8	34.8	34.4	
D_s	_	_	_	_	2.7	2.3	
D_{bh}	—	—	—	—	1.5	1.2	

*Note: The horizontal line indicates that the term was not measured at the time. All values in the table are in centimeters.

growth of the plants varied significant within the pond. Temporally, both heights and canopy widths increased across observations, with mean values rising from 90.5 cm to 106 cm and from 24.8 cm to 34.6 cm, respectively (Fig. 4). The averaged growth rates were approximately 17 % and 40 %. Spatially, although afforested at nearly the same time in 2021, the *K. obovata* saplings exhibited distinct morphology differences between western and eastern subregions. Mangroves in the *L*-labeled area, characterized by a higher degree of closure, are larger than those in the *R*-labeled region (Figs. 1, 4). Moreover, the initially high growth rates, reaching 4 cm per month at the beginning, gradually decreased. This trend exhibited a time in at the *R*-labeled region, maintaining a high growth rate of 1.8 cm per month during the later stage of observation, while the values in the *L*-labeled area were lower than 0.3 cm per month (Table 2).

Similarly, spatial variations were observed in diameters at basal and breast height, with a 1.2-fold difference between plants in the *L*-labeled and *R*-labeled regions (Table 2). Furthermore, H_t decreased in the *L*-



Fig. 4. Height and crown width of vegetation at different quadrats on (A) 2023/06/18, (B) 2023/11/14, and (C) 2024/03/30. Grey boxes denote the plant heights and green ones denote canopy widths. Subfigures (D-F) show the photos of the plant growth taken at these three stages, respectively. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)



Fig. 5. Hydrological conditions of the observation stations. Panels (A-D) correspond to stations L1-4 and (E-H) to R1-4. Arrows in circle of each box represent the maximum mean-root-square velocities of each layer, with the vertical color bar showing the corresponding position. $U_{rms}(z)$ is the maximum of mean-root-square velocity in the mainstream direction Φ of each layer: $U_{rms}(z) = [\overline{u_{rms}}(\Phi, z)]_{max}$. Φ_{flood} is the mean direction of all valid observation layers of each station, considered the primary transport direction and used to calculate the one-dimensional flood-ebb current velocity. Positive values in red color indicate velocities in flood direction and negative values in blue indicate velocities in ebb direction, as shown in the contour surface. The black dashed lines represent the turning interface between the red and blue colors, i.e., between flood and ebb directions. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

labeled area but slightly increased in the *R*-labeled region, while H_c rose at a higher rate in both from November 2023 to March 2024. This corresponded to the reduced growth process mentioned earlier and suggested a potential transition in growth strategy from elevation to lateral extension (Fig. 4, Table 2).

3.2. Spatial variations in hydrodynamics during spring tides

3.2.1. Hydrodynamic characteristics

Spatial variations in the vertical velocity structure during flood and ebb tides were significant in both at *L*- and *R*-labeled regions. In terms of direction, vertical diversion was slight near the outer channel, with the sheering angle derived from the root-mean-square circle remaining below 30° (Fig. 5D, H). In contrast, changes in direction were more complicated in the afforested regions. Specifically, *R1-3* showed vertical maximum discrepancies above 75° , while *L1-3* exhibited discrepancies above 45° , with sheering angles undergoing irregular oscillations (Fig. 5A-C, E-G). During the spring tidal cycle, the entire water column moved in and out in sync at *L4* and *R4* (Fig. 5D, H). However, opposite streams occurred vertically at the afforested regions *L1-3* and *R1-3* simultaneously, with the lower clusters of positive values wedged into upper negative ones (Fig. 5A–C, E–G).

Additionally, changes in flow direction, indicating flood or ebb tide, were counted cumulatively at both the bottom and surface to determine the durations of flood and ebb phases (Fig. 6A, B). It was observed that the durations of the flood and ebb phases were approximately symmetric between bottom and surface in the pond. For instance, at *L*3, the flood tide duration was 8.1 h at the bottom and 2.8 h at surface, while the ebb tide duration was 3.4 h at the bottom and 8.7 h at the surface (Fig. 6A, B). In contrast, at the outer stations *R4* and *L4*, the durations of the flood and ebb phases were consistent between the bottom and surface, with flood phases lasting 4.7 h and 5.2 h, respectively, and ebb phases lasting 8.7 h and 8.2 h at *L4* (Fig. 6A, B).

As for the magnitude of velocity, the maximum root-mean-square *U* decreased as the vertical position rose, meaning tidal flow generally ran faster in the mainstream direction at the bottom of each layer than at the surface (Fig. 5). However, it should be pointed out that velocities at *R4* and *L4* were greater at the surface water during the flood and ebb peaks



Fig. 6. Velocity and duration at flood and ebb phases. Duration at (A) bottom waters; (B) surface waters throughout flood or ebb tide. (C) Depth-averaged velocity at flood or ebb peak within the tidal cycle. (D) Continuous domain-averaged velocity in the flood (+) or ebb (-) phase. The green arrows indicate a decrease from flood tide to ebb tide, while the gray arrows indicate an increase from flood tide to ebb tide. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

Table 3

Ratio of velocity over flood and ebb tides.

Site	Ratio of duration of flood to ebb at bottom	Ratio of duration of flood to ebb at surface	Ratio of tidal domains (A_f/A_e)	Water flux of flood to ebb $(\overline{\frac{U_{pf}}{U_{p,e}}}, \frac{A_f}{A_e})$
R1	3.76	0.17	0.48	0.96
*R2	1.84	0.23	0.73	0.72
R3	2.29	0.32	1.11	3.00
R4	0.78	0.64	0.66	0.46
L1	3.12	0.54	1.08	1.51
*L2	2.65	0.62	1.79	3.55
L3	2.38	0.32	1.09	1.83
L4	0.54	0.63	0.54	0.37

*Note: Incomplete data were used in the second hydrological station after removing inaccurate values.

(Fig. 5D, H), whereas the darker color *iso*-surface, which represents higher velocity, was present at the bottom or middle water at other stations (Fig. 5A–C, E-G). Furthermore, the vertically averaged velocity increased by 8.2 and 9.4 cm/s from flood peak to ebb peak at R4 and L4, respectively, indicating ebb dominance in the outer channel (Fig. 6C). Conversely, at *L*1-3 and *R*1-3, the vertically averaged velocities were higher at flood peak than at the ebb peak, indicating flood dominance (Fig. 6C).

Following the aforementioned, unit-width water flux was obtained by calculating the product of averaged velocity, duration and inundation depth, with $[L^3/L]$ as the physical dimension, and the ratio of flood to ebb tide is provided in Table 3. An intense dominance of ebb flux was present at the outer stations with the ratio less than 0.5, whereas the component of flood flux was enhanced, especially with a ratio of 3.0 at *R3* and 1.83 at *L3*. To sum up, remarkable changes in vertical velocity structure and water flux over the flood and ebb phases were found along the path of tidal inflow, from the outer station to the pond.



Fig. 7. Bed shear stress induced by tidal current. Bed shear stress induced by current of (A) the *L*-labeled and (B) *R*-labeled region, respectively. Blue lines correspond to the outer observation station and red lines correspond to the entrance of the afforested area. Dashed lines denote the critical shear stress, with upper ones for erosion and lower ones for deposition. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

3.2.2. Bed shear stress

By examining τ_m in relation to the critical shear stress for erosion (τ_{ce}) and deposition (τ_{cd}) , we determined whether the bed shear stress is sufficient to suspend sediment form the substrate or allow it settle from entrainment, respectively (Roberts et al., 2000). Since the grain size across the entire region was fairly homologous (Fig. S2), τ_{ce} and τ_{cd} were estimated to be 0.52 Pa and 0.21 Pa, respectively, based on equations (10–16) (Fig. 7). During different tidal phases, the relationship between τ_c and the two critical shear stresses qualitatively illustrated various sediment motion states. At R4, τ_c initially increased to approximately 1.4 Pa until the flood peak, then decreased during the slack stage before rising again during the ebb tide. This suggests that particles became suspended during the flood and ebb peak ($\tau_c > \tau_{ce}$), remained entrained with minimal erosion or sedimentation when $\tau_{cd} < \tau_c < \tau_{ce}$, and settled during slack water when $\tau_c < \tau_{cd}$ (Fig. 7B). to R4, but with greater erosion (Fig. 7A). However, τ_c at L3, the entrance of the western semiclosed pond, stayed below the τ_{cd} dashed line throughout the tidal cycle, indicating continuous sedimentation (Fig. 7A).

3.3. Suspended sediment transport

From the outer channel to the inner pond, near-bottom SSC exhibited asymmetry between the flood and ebb phases (Fig. 8A, B). During the flood phase, mean SSCs reached approximately 0.09 kg/m³ at *R2-4* and exceeded 0.1 kg/m³ at *L2-4*, while at the innermost stations, *R1* and *L1*, about it decreased to 0.07 kg/m³, representing reductions of about 25 % and 35 %, respectively (Table 4). In contrast, mean SSCs during the ebb phase were notably lower across all stations except the outermost location. For example, in the *R*-labeled region, mean SSC was nearly equal between flood and ebb phase decreased to 0.055 kg/m³ at *R2-3* and 0.043 kg/m³ at *R1*, with an SSR of 37 % (Table 4). A similar pattern was observed in the *L*-labeled region, where SSC was consistently higher

Table 4

	<i>SSC_{Flood}</i> (kg/ m ³)	<i>SSC_{Ebb}</i> (kg/ m ³)	SSR (%)	SSF_{Flood} (\cdot 10 ⁻³ m ³ / cycle)	SSF_{Ebb} (\cdot 10 ⁻³ m ³ / cycle)	$SSF_T(\cdot 10^{-3} m^3/cycle)$
R1	0.068	0.043	36.37	4.94	0.38	5.33
R2	0.090	0.056	37.58	5.66	-0.01	5.65
R3	0.088	0.054	38.72	53.83	8.50	62.43
R4	0.097	0.096	0.40	50.89	-18.77	32.06
L1	0.069	0.038	45.04	6.77	-0.32	6.46
*L2	0.102	_	_	13.00	_	_
L3	0.110	0.048	56.57	41.88	-10.55	31.36
*L4	0.105	_	_	23.42	_	_

*Note: Data were loss during ebb tide on the second survey day resulting in blanks here. Positive values indicate net import toward the flood direction, while negative values indicate net export in the opposite direction.

during the flood phase compared to the ebb phase, and SSR decreased with inflow progression through the afforested region (Fig. 8C). Notably, the higher SSR of 56.6 % at L3 and 45.0 % at L1 indicated greater potential of sediment import and deposition driven by flood-dominant asymmetry (Fig. 8C).

The total suspended sediment flux (*SSF_T*) in the mainstream direction over the diurnal tidal cycle was positive, indicating a net inward import (Table 4). Within the afforested region, a significant reduction in *SSF_T* was observed, decreasing from 62.0 m³/cycle at *R3* to 5.3 m³/cycle at *R1*, and from 31.4 m³/cycle at *L3* to 6.4 m³/cycle at *L1* (Fig. 8C). During the flood phase, net sediment import into the afforested region was evident, with values of 41.9 m³/cycle at *L3* and 53.83 m³/cycle at *R3* (Table 4). during the flood phase, net sediment flux exhibited a declining trend from the entrances *R3* and *L3* to the innermost stations during the flood phase, indicating declining sediment mass along the



Fig. 8. SSC, SSR and SSF_T of observation stations. (A-B) SSC at different stations during a diurnal spring tidal cycle. (C) Suspended sediment retention (SSR) and sediment flux (SSF_T) over the entire tidal cycle.



Fig. 9. Bed Levels and changes along the four designated transects over three measurements in nine months. (A-D) *T1-4* represent the four designated transects in the restored area, shown from left to right in Fig. 2A. Relative distance refers the ratio of the distance between each point on the transect and the innermost point to the total length of the transect. Bed level differences along of the transects within time interval of (E) 2023/06/18–2023/11/14, (F) 2023/11/14–2024/03/30, and (G) 2023/06/18–2024/03/30. In E-G, the box spans 25% and 75% (the interquartile range) of the data, the red central line marks the median, whiskers extend to values within 1.5 times the interquartile range, and points beyond the whiskers indicate outliers. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

flood tide pathway (Table 4).

3.4. Evolution of bed level over time

The three elevation surveys indicated an overall accretion along the four designated transects over a nine-month period from June 2023 to March 2024 (Fig. 9A–D). Box plots of temporal elevation changes (ΔE) along each transect revealed variations in sedimentation intensity across different locations during two intervals of approximately 4.5 months (Fig. 9E, F). From June to November 2023, the median bed thickness increments (ΔE_1) were about 1.8, 4.0, 1.8, and 5.0 cm for *T1-4*, respectively (Fig. 9E). Over the subsequent interval, the medians (ΔE_2) were 3.8, 1.2, 3.6, 1.2 cm (Fig. 9F). Notably, the transects within the *L*-or *R*-labeled regions exhibited complementary patterns, resulting in

similar total median increments (*Total* ΔE) of approximately 5.0 cm for all transects. This corresponds to an annualized accretion rate of approximately 6.4 cm (Fig. 9G).

4. Discussion

4.1. Impact of hydrodynamics on sediment transport

Previous studies have demonstrated that various hydrodynamic processes play a crucial role in governing sediment transport within estuaries (e.g., Sidik and Friess, 2020). Remarkably, spring tidal cycle plays a significant role in sedimentation at higher elevations within intertidal areas, particularly in mangrove systems (Swales et al., 2019). In this mangrove system, spring tides account for over 65 % of the

inundation events and 75 % of the total inundation duration annually. Increasing tidal ranges also contribute to increasing water flux at the entrance of the pond (L3). This underscores the key role of spring tides in regulating water inflow and sediment transport in the restored pond. (See Fig. S7 for detailed analysis of local tidal cycles based on tide gauge data and simulations).

Focusing to this key role, two primary processes can be distinguished in sediment transport: the first is the vertical disturbance of the bed surface (Soulsby and Clarke, 2005; MacVean and Lacy, 2014). Given the weak forces induced by waves (Fig. S8), shear stress induced by current was regarded as the predominant driver initiating the motion of bed sediment. In the study area, the shift of τ_c from the outer channel to the inner pond indicated a general trend along the flood tidal route, contrasted with an increasing trend along the ebb route (Fig. 7). This suggests that the outer channel primarily served as the sediment source for the inner pond.

Although τ_c values were missing at stations closer to the interior (i.e., *R1-2*, *L1-2*) due to deviations from the logarithmic law, lower τ_c values can be inferred based on the reduced velocity magnitudes observed in Fig. 5 and Fig. 6. Currents exerted minimal erosion in the afforested pond, resulting in negligible damage to the colonized K. obovata saplings. Additionally, the ratio of the near-bed suspended sediment retention (SSR) over the entire diurnal tidal cycle, showed highly positive percentiles, suggesting that net sediment import and substantial sedimentation likely occurred within the pond due to flood-dominant asymmetry (Fig. 8C) (Schettini et al., 2020). Furthermore, the observation that larger SSR values occurred at the entrance rather than at the innermost station indicated sediment deposition throughout the flood phase (Table 4). On the other hand, the near-equilibrium SSC between the flood and ebb phases at R4 suggested minimal sedimentation during the tidal cycle. This observation aligned with the moderate τ_c values, which predominantly ranged between τ_{cd} and τ_{ce} , and underscored the critical role of the outer channel as a sediment delivery pathway (Table 4, Fig. 7). In summary, along the pathway from outer channel to the inner pond, the rapid decline in erosion due to the diminishing bed shear stress, combined with significantly higher SSC import during the flood tide compared to the ebb, created a favorable condition for net sediment import into the afforested pond.

The second key process involves horizontal sediment transport via flow entrainment (van Maren and Winterwerp, 2013). Tidal asymmetry is widely recognized as a key factor in shaping sediment transport pattern within estuarine systems, which could be characterized by variations in tidal constituents or imbalances in the water budget (Wu et al., 2014; Guo et al., 2019). Throughout the near-bed process, the pronounced asymmetry of currents-manifesting in both the flood and ebb phases and in averaged velocities (Table 3, Fig. 6)-undoubtably impacted regional sediment transport, a phenomenon that has been extensively documented (e.g., Horstman et al., 2011; Pang et al., 2019; Leuven et al., 2021). It's clear that the tidal regime undergone a dramatic shift as it propagates into the planted pond, transitioning from an ebb-dominant regime in the outer station to a flood-dominant regime (Figs. 5-6). Within the pond, the longer duration and higher velocity during the flood tide, combined with a greater volume of entrained sediment, resulted in a substantial horizontal sediment inflow at the entrance over the diurnal tidal cycle (Table 4). Therefore, the differences in the flood-ebb tidal regimes between the outer channel and the inner pond drove the near-bed sediment transport into the pond. Further, the decline in SSF from the entrance to inner station in the mainstream direction over the tidal cycle indicated concurrent sedimentation as sediment transport continued (Fig. 8C). This observed decline in sediment flux along the length of the pond aligns with the vertical sedimentation process discussed in the preceding paragraphs (Table 4). Overall, the concurrent vertical deposition and horizontal sediment transport under increasing tidal levels contributed to the development the afforested pond as a sediment sink within this dynamic system.

eater volume of entrained

side, the closed entrance on the western side exhibited relatively low τ_c values, indicating ongoing sedimentation. In addition, compared to the L-labeled region, the SSF_T did not accurately reflect the sediment import direction into R-labeled planted area, as the mainstream flow was aligned with the tidal channel rather than entering perpendicularly (Fig. 1A, H). As a result, the significantly higher values at R3 may be overestimated. Both the western and eastern sides generally maintained a similar spatial pattern of sediment transport from outer channel to the inner pond as mentioned in the two processes above.

Lastly yet importantly, the distortion of vertical structure observed in the velocity profile within the pond, similar to that seen in estuarine circulation or submarine channel profiles (Geyer and MacCready, 2014; Bolla and Imran, 2014), probably exerted a reinforcing effect on total sediment import in this shallow tidal environment. The source of this exertion lies in the combination of the wedge-shape velocity profile over time—in which flood dominates the bottom while ebb dominates the surface for most of the time in the planted pond (Figs. 5-6)—and the well-explored vertical SSC profile, which typically follows the parabolic Rouse formula (Rouse, 1937; Sumer and Muller, 1983):

$$C_z = C_a \left[\frac{a(h-z)}{z(h-a)} \right]^R \tag{20}$$

where *h* is the total water depth; C_z , C_a represent the SSC at the variable height *z* and the specified height *a*, respectively. The Rouse Number *R*, which determines the distribution pattern of SSC, is given by $\frac{W_z}{\kappa t u}$, which refers to the ratio of the settling velocity and shear turbulence (Tseng and Tinoco, 2021). The concentration profile (C_z) is approximately a reciprocal function of z^R , indicating that the SSC increase vertically downward. This theoretical deduction is further supported by numerous in-situ observations in shallow-water tidal environments (e. g., Williams et al., 1999; Venditti et al., 2016; Xing et al., 2022). Thus, despite the lack of SSC profile data, a hypothetical framework can be reasonably proposed to illustrate suspended sediment transport throughout the entire water column over tidal cycles, incorporating two collaborative configurations: the distorted vertical current profiles and the constant Rouse SSC profile concluded here.

Due to the two distinct vertical velocity profiles exhibited between channel and pond, the sediment transport throughout the entire water column during the tidal cycle can be summarized in space (Fig. 10). In the first pattern, the entire water column moves in the same flood-or-ebb direction temporally, with the turning interface of current's heading nearly perpendicular to the bed. Hence, the net motion of suspended sediment depends solely on the asymmetry of depth-averaged velocity (Fig. 10A), which, in the outer channel, represents the general form of suspended sediment transport on tidal flats (van Maren and Winterwerp, 2013; Xing et al., 2022). In contrast, the less frequently discussed wedgeshape velocity profile over time exerts a more complex influence on sediment transport in the inner pattern (Fig. 10B). Specifically, the flood domain (defined here as the entire spatial and temporal extent of the flood current direction) is confined to the lower part of the water column and persists during the falling water level. This configuration, coupled with the higher sediment concentration near the bottom, results in greater sediment import compared to that observed in a vertically unidirectional distribution (Fig. 10B). This appears to be analogous to sediment transport in a meandering channel, where more sediment is transported near the bottom from the concave bank to the convex bank, while sediment in the upper water layer is transported in the opposite direction (Blanckaert et al., 2013), suggesting the critical role of the unique geomorphology of ponds in this process. Additionally, the water budget is maintained through the offset by the ebb domain in the upper water layer, ensuring the replenishment of dissolved nutrients in the ecosystem, which are less subject to gravity (Valiela et al., 1978; Grunwald et al., 2010). In summary, the distortion in the vertical velocity profile transports more sediment particles into the inner pond without affecting the tidal water volume exchange.

Unlike the open entrance in the secondary channel on the eastern



Fig. 10. Patterns of collaboration between vertical current profiles and SSC profiles influencing suspended sediment flux throughout the water column. Panel (A) represents the pattern outside the pond, and panel (B) the pattern inside the pond. The blue horizontal arrow line is the time-axis, showing the progression of the tidal cycle; the gray vertical arrow line is the depth-axis, representing spatial changes over the tide. The green arrow line indicates an increase in SSC values along the arrow. The green curve, as a function of depth rather than time, describes the vertical variation of SSC transport. The gray arrows within the panels indicate domain in the flood or ebb directions, and are determined by both the spatial and temporal axes. Solid blue curves mark the turning interface between flood and ebb domains. Thus, differences between the two sediment transport patterns can be reflected in the product of the green curve and the integration within different tidal domains. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

4.2. Hydrodynamic effects of the semi-enclosed topography

The semi-enclosed topography of the pond could be a critical factor for hydrodynamics, as it modifies tidal asymmetry and bed shear patterns, consequently influencing sediment transport and accretion processes (Ridderinkhof et al., 2014 de Ruiter et al., 2019). But, determining the optimal degree of closure appears to be a key issue for the success of restoration efforts, which is fundamentally dictated by the selection and intensity of human reconstructions. To examine the hydrodynamic effects of varying closure degrees within the pond, thirty scenarios were simulated using the model described in Section 2.5, with increasing dike closure to higher scenario numbers (Fig. S4). The degree of closure is defined by a coefficient (C_c) calculated as follows:

$$C_c = \sum_{i=1}^{n} \gamma \cdot \frac{W_A}{W_A + W_B} \tag{21}$$

where *i* is the index of the closed edge; and W_A and W_B correspond to the widths of the edge above or below the reference level, respectively. The reference level is approximated by the high tidal level, set at 2.0 m here. The coefficient γ is the weight factor, with values of 0.5, 0.3, 0.2 assigned to the upper, right and bottom edges, respectively. Different weight factors are applied to distinguish variations in tidal exposure at different edges (Fig. 11). Notably, to emphasize the optimal selection of



Fig. 11. Illustrated diagram of components of the closure coefficient.

closure at the outer edge, high closure coefficient values were amplified for clearer visualization by transforming C_c into $\frac{-1}{ln(C_c)}$, which is positively correlated with C_c (Fig. S5). By correlating the closure coefficient with the tidal velocity skewness of the focus vegetated area shown in Fig. 3 under actual dense planting, a significant positive correlation was observed (R = 0.55, p < 0.01) within the range of closure degrees less than the actual closure degree. Additionally, average values were generally greater than 0 (Fig. 12C), indicating that flood dominance in the pond was enhanced due to the retention of the outer wall, which facilitates sediment import into the pond (Hanegan et al., 2023). An opposite trend was observed as the closure degree increased beyond the actual closure (R = -0.72, p < 0.05), forming a nonlinear relationship between closure degree and flood dominance within the pond (Fig. 12C). This break indicates that the maximum water import could occur at a closure degree near the value associated with actual restoration.

Within the semi-enclosed pond, bed shear stress was observed to be relatively low compared to the critical shear stress for deposition derived from the sediment characteristics (Fig. 7A). In addition, the simulated bed shear stress ratios were consistently much less than 1, indicating a general deposition condition (Fig. 12F). Therefore, despite the uncertainty in evaluating critical shear stress, the ratio τ_c/τ_{cd} still maintained a positive linear relationship with vertical deposition magnitude, according to the simplified formula (18) for non-erosive conditions proposed by Tao et al. (2024). The bed shear stress ratio exhibited a negative correlation with the closure coefficient (R = -0.82, p < 0.001) (Fig. 12F), indicating a high potential for sedimentation. Notably, the declining trend was not as pronounced for closure coefficients in the interval between 0.5 (the critical values of full openness at the outer edge) and 0.9 (Fig. 12F). This highlights the significant effect of the incremental increase at closure degrees above 0.9 exerted on shear stress. Thus, the optimal construction is likely associated with a high closure degree, integrating considerations of both asymmetric and shearing forces. However, total water inflow probably decreases as closure increases, resulting in reduced sediment import. Moreover, a deep, narrow channel is considered a key factor in triggering sediment inflow, contributing to morphological resilience within a restored marsh (Fleri et al., 2019). This effect may be analogous in the pond-tomangrove system at a similar scale. A precise evaluation of the optimal closure necessitates a more comprehensive analysis that incorporates various factors.

The impact of closure differences was also evident in the field



Fig. 12. Influence of closure degree on tidal velocity asymmetry and bed shear strength in the pond based on simulations. The relationship between different simulated closure degrees and tidal asymmetry under (A) no vegetation, (B) sparse planting, and (C) dense planting. The black horizontal line at 0 represents no asymmetry, values greater than 0 indicate flood dominance, and values less than 0 indicate ebb dominance. The blue vertical dashed line represents specific closure degrees, with 0.5 corresponding to the fully openness at the outermost edge. In panels A-C, the turning points of the two red fitted curves were determined by the closure degree applied in the actual planting, as shown in the Fig. S4, with Geo_scenario.0 as the index. The relationship between different simulated closure degrees and bed shear strength under (D) no vegetation, (E) sparse planting, and (F) dense planting. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

between the L- and R-labeled regions. Although both regions transformed tidal asymmetry into flood dominance, with enhanced effects in the bottom waters (Fig. 5, Fig. 6), bed shear stress ratio was significantly greater in the R-labeled region, which has a lower degree of closure compared to the L-labeled region (Fig. 7). This indicates an intensified sedimentation effect due to edge retention, consistent with the model results (Fig. 12D-F). Moreover, suspended sediment retention was higher in the L-labeled region, reflecting a greater potential for sediment settling in the more closed pond (Fig. 8C). The R-labeled region appeared to have a higher suspended sediment flux (Fig. 8C). However, unlike the L-labeled region, the main transporting direction at the entrance was not parallel to the inlet of the R- labeled region, but formed an angle of approximately 60° (Fig. 1E, 5G). This angular transport resulted in nearly equivalent sediment fluxes into both sides (Fig. 8C), which could explain the similar magnitudes of accretion observed in both regions (Fig. 9G).

Beyond tidal asymmetry and bed shear stress, wave and storminduced effects may be mitigated by the pond's closure, which was evident in the abrupt decline of the wave-induced shear stress from L4 to L3 and R4 to R3 (Fig. S8). Thus, a potential shelter effect of the closure on the planted mangrove was suggested, although wave forces were minor during the observations.

4.3. Effects of plant morphological growth and density on hydrodynamics

Under the tidal regime, the presence of vegetation can exert multifaceted influences on hydrodynamics in estuarine environments (Horstman et al., 2013; Lera et al., 2019). As demonstrated in the topographic influence analysis, the presence and densification of vegetation could significantly enhance the effect of closure. Notably, a nonlinear correlation between velocity skewness and vegetation density was observed, which was positive and then negative (Fig. 12A-C: for non-vegetated scenarios, R = 0.45, p < 0.05, and then R = -0.60, p =0.07; for sparse vegetation, R = 0.53, p < 0.05, and then R = -0.55, p =0.10; and for dense vegetation, R = 0.55, p < 0.01, and then R = -0.72, p < 0.05). Similarly, vegetation presence increased the negative correlation for bed shear stress ratio in these three vegetation scenarios, with decreasing *R* and *p* values as vegetation density increased (Fig. 12D-F). These divergent effects, brought by presence or absence of vegetation, were also observed in the deformation of river mouth bars (Lera et al., 2019). Denser vegetation appears to have a greater damping effect on tidal flows and waves, which could trap considerable amounts of sediment within the afforested pond, in combination with the flooddominant asymmetry and weak bed erosion (Horstman et al., 2013). Furthermore, the vertical structure of vegetation is known to distort the flow velocity profile, often generating a shear layer above the canopy, while the velocity magnitude varies with depth depending on the vegetation density below the canopy (Nepf, 2012; Wang et al., 2019). Dense vegetation clusters could elevate the maximum SSC vertically, shifting it from near the bottom to the canopy in inundated shallowwater flows (Li et al., 2020). Here in the afforested pond, K. obovata saplings were planted at regular spacing, with ratios of spacing to trunk diameter ranging from 20:1 to 40:3 (Fig. 4). This configuration may exert minimal influence on the development of a logarithmic velocity profile upward from the bed (Nepf, 2012). Consequently, two superimposed log-like profiles bounded by the canopy can be likely to form when the vegetation is fully submerged (Huai et al., 2009). Similarly, with the sparse, low-positioned and tapered trunks, vertical plant structure probably has little significance for the SSC profile of the planted mangrove. This behavior contrasts with the effects observed in dense herbaceous coverage, as described in the experimental setup of Li et al. (2020). Remarkably, the periodic tidal submergence, accompanied by the shifts in velocity profiles of both upper and lower layers (Fig. 5), adds substantial complexity to the sediment transport patterns, which can even display opposite behavior across different tidal stages (Temmerman et al., 2005). Hence, sediment transport processes within the afforested pond could be more intricate than the generalized pattern depicted in Fig. 10.

In evaluating the success of incipient restoration efforts, the potential early co-occurrence of plant growth emerges as a crucial factor. Herein, the growth process of colonized *K. obovata* saplings was tracked through in-situ observations of individual plants (Fig. 4). The saplings exhibited an initial phase of vertical elongation, followed by lateral expansion. Eighteen scenarios were simulated to represent the early stages of plant colonization, illustrating a hypothetical spiral growth process characterized by both vertical elevation and lateral spread (Table S2). For the actual planting density $(9/m^2)$, flood-dominant velocity asymmetry



Fig. 13. Progressive alteration of hypothetical vegetation morphological growth on hydrodynamics at the early afforestation stage. Impact of plant growth on (A) Sk_U , and (B) τ_c/τ_{cd} . SD in the figure denotes standard deviation, and CI denotes confidence interval. Red blocks represent the dense planting scenarios and blue blocks represent the sparse planting scenarios. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

showed a strong positive correlation with sapling growth during the very early stages (R = 0.89, p < 0.01). However, this correlation shifted to a significant negative value as the plants continued to grow (R = -0.69, p < 0.05), indicating that net sediment import increased shortly after initial afforestation but subsequently declined as the growth progressed (Fig. 13A). Conversely, bed shear stress ratio initially exhibited a negative correlation with the sapling growth (R = -0.75, p < 0.05), and then transitioned to a positive correlation (R = 0.78, p < 0.01). This trend demonstrated a shift in sedimentation dynamics, conversing from an acceleration phase to deceleration as saplings matured (Fig. 13B).

Notably, the breakpoints in correlation fittings for both tidal asymmetry and bed shear stress were nearly coincided, occurring when plant heights reached approximately 100 cm and canopies reached up to 20 cm (Fig. 13). This aligns with the transport patterns and sedimentation states of the actual restoration, fostering conditions favorable for the overall accretion observed in Fig. 9. As the bed level elevated, mangrove plants gained the ability to establish roots and disperse within the coastal environment (Huang et al., 2023). However, excessive deposition in some cases may lead to forest dieback by burying the propagules or pneumatophores (Nardin et al., 2021). The drops in sediment import and deposition following the peak would likely help retain a reasonable accretion rate that supports plant growth in later stages (Fig. 13).

In comparison, the influences of saplings growth at a sparse planting density $(4/m^2)$ on both flood-dominant asymmetry and bed shear stress ratio were relatively subtle (Fig. 13). That is probably because sparsely distributed young plants had limited ability to alter the velocity structure with ample space for inflow (Sidik and Friess, 2020). Furthermore, beyond individual morphology, the reproductive behavior of the plants themselves could affect sediment dynamics, particularly with the emergence of randomly distributed young saplings with low-lying canopies (Fig. 4E, F). Thus, continuous observation is essential to monitor this co-evolution process under specific bio-geomorphological conditions.

4.4. Schematic diagram of bio-hydromorphodynamic processes within the restored mangrove-pond system

Considering all the observed and simulated elements, a schematic was presented to illustrate the continuous hydro-sediment dynamics within a pond-to-mangrove system following initial human involvement (Fig. 14). In this schematic, four essential components contribute to the initial success of the bio-hydro-morphological restoration in the afforested pond. One key aspect is the sediment availability from the outer channel for the pond. The outer channel transports riverine sediment and locally resuspended sediment due to high bed shear stress, serving as the sediment source for the inner pond system. Upon reaching the pond's water inlet, the abruptly decreased bed shear stress and the distorted flood-dominant asymmetry promote the sedimentation process (Figs. 5-8). These exterior elements provide the pond with favorable conditions for deposition.

The other three aspects are reflected in the planted pond, mainly under the influence of human involvement and natural plant growth. Within this inner system, the hydro-sediment dynamic processes, consisting of horizontal sediment import and vertical deposition, serve as a pivotal linkage between these two key factors and morphodynamics. First, the closure reconstruction of the pond could exert a nonlinear impact on the flood dominance and sedimentation. When the closure degree is lower than 0.9, it has a positive impact on bed accretion, and vice versa (Fig. 12). This highlights the necessity of the optimal topographic closure selection for the successful initiation of restoration. Second, the natural growth of the planted mangrove saplings exerts dynamic effects on tidal asymmetry and bed shear stress, eventually reaching a relatively stable state for sediment accumulation (Fig. 13). This represents a progressively beneficial form of geomorphological adaptation to vegetation growth. Third, another human involvement-selection of the planting density-contributes to hydro-sediment



Fig. 14. Schematic diagram on bio-morphodynamics within a pond covered by restored monoculture mangrove. Lines without arrow represent co-occurrence; arrow lines without circled plus or minus signs represent direct causes or elaboration; circled plus and minus signs denote positive and negative impacts, respectively; a large solid-line circular icon represents direct effects on the object, while smaller one surrounded by dashed line represents the effects on the effect. In this diagram, different colored boxes represent different meanings: blue boxes indicate processes related to hydrodynamics; yellow boxes represent states related to sediment and geomorphology; gray ones represent human intervention; green ones denote natural growth; and the gray-green mixed ones signify a combination of both. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

dynamics by affecting the topographic and vegetative factors (Figs. 12, 13). Simulation results suggested that denser planting perform well in promoting deposition, by enhancing the effects of closure decline and plant growth.

Considering all the factors, the actual combination of restoration strategies has provided relatively favorable conditions for effective accretion, which was also observed in the field (Fig. 9). Undoubtedly, this moderate overall accretion supports a favorable substrate for plant growth, advancing the restored system toward a more natural state. This is crucial for the successful bio-geomorphologic development of the ecosystem (Samsmito et al., 2023; Day et al., 2024).

4.5. Limitations

Numerical simulations provide valuable insights into the effects of restoration-related factors that are challenging to measure directly, as summarized above. As noted in Section 2.5, despite the model's underestimation of current velocities, the two key dimensionless parameters used to characterize horizontal and vertical motions—tidal asymmetry and bed shear stress ratio—can still be reliably calculated due to the strong linear correlation between observed and simulated velocities (Fig. S3). However, certain limitations remain regarding the applicability of these simulations. The model is constrained to a relatively small spatial and temporal scale, making it suitable for analyzing intra-pond dynamics but not for capturing broader estuarine processes. Additionally, plant growth is only examined during the early stages of restoration, constrained by the limited duration of actual measurements, before the accumulation of organic sediment and the onset of interplant

competition. Furthermore, while our discussion of topographic modifications focuses on the degree of closure, other geomorphic factors—such as bed slope and inner channel depth—also warrant further investigation. To advance a more comprehensive understanding of bio-hydromorphodynamic interactions within mangrove-restored ponds, future studies should incorporate extended field measurements and refined numerical modeling approaches.

5. Conclusions

Artificial planting of mangroves in abandoned aquaculture ponds has emerged as an effective strategy for restoring tidal ecosystems. However, our understanding of the associated hydro-sediment dynamics remains limited. In this study, we combined in-situ observations and numerical simulations to investigated bio-hydro-morphodynamics within a recently afforested mangrove pond in an estuarine setting. Particular attention was given to the early-stage restoration features through modeling, including dike closure reconstruction, planting density, and natural vegetation development. The main findings are summarized as follows:

- (1) Changes in tidal current asymmetry and bed shear stress cause the outer channel to function as a sediment source for the inner restored pond system.
- (2) Spatial variations in suspended sediment concentration and transport patterns contributed to the overall accretion within the pond.

(3) During the early restoration stage, increasing topographic closure and natural vegetation growth exerts a non-linear influence in shaping on the hydro-morphodynamics of the pond, with higherdensity afforestation contributing more significantly.

This study addresses a critical gap in understanding hydromorphodynamic processes in pond-mangrove ecosystems, providing scientific insights into reconstructing closure structures, optimizing planting density, and achieving near-natural restoration. Given that many estuarine mangrove habitats worldwide are adjacent to abandoned aquaculture ponds, these new insights provide important theoretical implications for ecological restoration in such regions. However, several key variables during restoration, such as topographic slope, inlet channel depth, and the morphology of mature vegetation over time, remain unexplored. Additionally, long-term observations and modeling are scarce in the current research. Future studies should focus on refining and analyzing these factors to gain a more comprehensive understanding.

CRediT authorship contribution statement

Jiejun Luo: Writing – review & editing, Writing – original draft, Software, Methodology, Formal analysis. Zhijun Dai: Writing – review & editing, Writing – original draft, Visualization, Validation, Supervision, Software, Project administration, Methodology, Investigation, Funding acquisition, Formal analysis, Data curation, Conceptualization. Xixing Liang: Visualization, Formal analysis, Data curation. Wenjun Zeng: Formal analysis, Data curation. Riming Wang: Investigation, Data curation. Hu Huang: Investigation, Data curation. Jaap Nienhuis: Writing – review & editing, Investigation, Data curation.

Declaration of competing interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

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Appendix A. Supplementary data

Supplementary data to this article can be found online at https://doi.org/10.1016/j.jhydrol.2025.133396.

Data availability

Data will be made available on request.

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