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Hydrodynamics and sediment transport in a southeast Florida tidal inlet

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Abstract

A three-dimensional ocean circulation model is used to investigate the hydrodynamics of a tidal inlet and deltas system in Southeast Florida, and to understand the consequences for suspended and bedload sediment transport patterns. The model reproduces observed tidal currents and provides insight about residual currents caused by spatial asymmetries in the inlet throat and tidal deltas during ebb and flood flows. A particle-tracking approach for suspended and bedload sediment transport is used to simulate deposition patterns for different particle sizes. The simulation results qualitatively correlate with the distribution of sediment characteristics within the tidal inlet and deltas system and demonstrate sensitivity to the choice of advection velocities (e.g., near-bottom versus depth-averaged) and regions of sediment origin. Furthermore, the distinction between suspended and bedload transport as a function of particle size indicates significant differences in deposition patterns and their potential connection to geomorphologic features of the tidal inlet and deltas system.

Keywords

tidal inlets and deltas; tidal currents; residual flow; sediment transport; sediment distribution; numerical circulation model

1. Introduction

Flows in tidal inlets have been extensively studied, particularly along the east coast of the United States (e.g., Kana et al., 1999) and in the North Sea (e.g., van de Kreeke et al., 1997). Since tides are typically the dominant hydrodynamic forcing, considerable effort has been invested in understanding the impact of tidal asymmetry and residual currents in inlets and estuaries, as well as their linkages to geomorphology and sediment transport (e.g., van Leeuwen et al., 2003; van de Kreeke and Himba, 2005). Hydrodynamic and sediment transport models have further demonstrated the importance of spatial asymmetries and residual flows associated with tidal currents on sediment relocation in inlets and estuaries (e.g., Ridderinkhof, 1988, 1989; Mallet et al., 2000; van Leeuwen and de Swart, 2002). In parallel, dynamical models have been used to investigate initial seabed changes and long-term morphological development in tidal inlets (e.g., Wang et al., 1995). Simulations of barotropic (i.e., depth-integrated) tidal flow in idealized and realistic inlet systems have also been used to quantify the temporal and

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spatial variability of along- and cross-stream momentum balances (Hench and Luettich, 2003).

Since previous modeling studies have generally concentrated on the two-dimensional characteristics of the tidal currents and used depth-averaged velocities to compute sediment transport, the focus of the present study is to incorporate the full three-dimensional characteristics of the tidal and wind-driven currents into the computation of suspended and bedload transport mechanisms. As such, this work contributes to emerging efforts targeted towards the use of three-dimensional hydrodynamic models to predict sediment transport and morphodynamics in realistic configurations (e.g., Lesser et al., 2004). Here, the fundamental hydrodynamic characteristics of a Southeast Florida tidal inlet are investigated using a three-dimensional coastal ocean model. The spatial and temporal variability of the simulated fields is analyzed to describe the full vertical structure of the flow. In addition, a Lagrangian particle-tracking approach is used to model suspended and bedload sediment transport as a function of grain size attributes. Deposition patterns from the particle transport model are qualitatively compared to observed distributions of sediment characteristics and geomorphologic features in the tidal inlet and deltas system.

2. Study domain

Located on the southeastern coast of Florida just offshore Miami, Bear Cut is a naturally occurring inlet, formed within a break of the Key Largo Limestone between two barrier islands, Virginia Key and Key Biscayne (Fig. 1A). The tidal inlet and associated deltas are situated on a narrow continental shelf, 7 km landward of the deep Straits of Florida. Landward of the barrier island complex is the broad (ca. 6 km near Bear Cut), shallow (2 to 3 m on average) lagoon of Biscayne Bay. Historically, while no significant southward migration of the inlet has been observed, sediment accretion has narrowed Bear Cut by ca. half of its width over the past two centuries. Erosion most noticeably occurred near the northeastern tip of Key Biscayne and along the seaward coastline of Virginia Key. Erosion related to dredging, filling, and seawall construction activities near the southern tip of Virginia Key have also contributed to the creation of a new flood channel.

The Bear Cut tidal inlet (Fig. 1B) includes the geomorphic elements analogous to those documented in other tidal inlet and deltas systems: flood delta, inlet throat, and ebb delta (terminology of Hayes, 1980). The flood delta includes numerous broad sub-lobes cut by narrow channels, whereas the ebb delta has one large central lobe with a main channel and a marginal channel on each side of the lobe. The inlet throat comprises a main channel and two sinuous channel margin bars rising 2 to 4 m above the bottom. The length (measured from flood delta to ebb delta) and width (measured across the narrowest point) of the inlet throat are ca. 2800 and 600 m, respectively. Depths in the inlet throat reach ca. 6 m in the main channel.

The sediment or grain type available for potential transport varies throughout the tidal inlet system. The surficial sediments within the flood tidal delta and lagoon are almost exclusively carbonate grains produced within the tidal inlet system. With distance from the flood tidal delta towards the ebb delta, siliciclastic quartz grains become more abundant, reaching up to 70% in the ebb delta. Historically, these sediments were supplied by southward alongshore drift. Since the construction of jetties just north of the study area, the alongshore transport has decreased, limiting the supply of siliciclastic grains to Bear Cut and contributing to erosion along the seaward coasts of Virginia Key and Key Biscayne.

3. Inlet circulation model

3.1. Model implementation

The tidal inlet numerical model is an implementation of the Princeton Ocean Model (POM; Mellor, 2003) on a curvilinear grid. The model domain is centered on the tidal inlet and deltas with the outer boundaries extended inside the lagoon (i.e., Biscayne Bay) and onto the continental shelf of Southeast Florida (Fig. 1C). The highest horizontal grid resolution (60 to 80 m) is located in the inlet throat and decreases away from the throat to an average spacing of ca. 200 m (Fig. 1D). The vertical resolution, provided by 15 sigma (terrain-following) levels, ranges between a few centimeters to a few meters depending on the water depth. The bottom topography is based on a 3-second (ca. 90 m) gridded water depth dataset from the National Geodetic Data Center (NGDC). Since wetting and drying was not included, the minimum water depth in the model is set to 1.5 m to prevent “dry” cells.

The model is forced at its eastern boundary by hourly sea surface height data from the Virginia Key tide gauge (for location, see Fig. 1E). During the study period (1 to 14 June, 2004), the maximum semi-diurnal amplitudes were ca. 90 cm during spring tide and 45 cm during neap tide. The southern and northern boundaries on the shelf have radiation conditions applied, and all boundaries inside the lagoon are closed. This domain configuration is obviously not entirely realistic, but it significantly reduces the complexity in dealing with the full extent of Biscayne Bay. Also, as there are no other permanent tide gauges inside Biscayne Bay, having open boundaries in the lagoon would require running a larger domain tidal model to properly specify sea surface height at these boundaries. Hourly winds from the Fowey Rocks C-MAN station (located approximately 15 km south and 6 km offshore of the study area) are used to compute surface wind stress forcing. The study period was dominated by northwestward winds at speeds ranging from 2 to 6 ms^{-1} on average, which is typical of summer conditions in Southeast Florida. To simplify the model, the influence of density stratification is neglected (i.e., temperature and salinity are held constant during the simulations). Since fresh water inputs are the dominant source of stratification during summer conditions, and since the largest salinity gradients are concentrated on the western side of Biscayne Bay (Wang et al., 2003), assuming a constant density in the model is a reasonable first approximation.

The model is run for the first fifteen days of June 2004, which includes both spring and neap tide conditions. This time period coincides with the deployment of a bottom-mounted acoustic Doppler current profiler (ADCP) in the inlet throat (5 to 6 m depth; for location, see Fig. 1E) and, hence, provides a useful benchmark for validating the model against observations.

3.2. Model-observations comparison

Since the sea surface height from the Virginia Key tide gauge is used for open boundary forcing, the model is first checked for its ability to reproduce the tidal signal. A direct comparison between the simulated sea surface elevation at the tide gauge location and that imposed at the eastern boundary confirms that the signal propagates correctly into the domain (RMSE is 17% of standard deviation, with the model lagging observations by ca. 45 min). While the lag between the tide gauge position and the eastern boundary is partly explained by tidal wave travel time and retardation by bottom friction, the artificially closed boundaries inside the lagoon are probably exaggerating the delay. The impact of boundary specification and domain extent for the lagoon need to be investigated further to understand their implications on the timing accuracy of tidal wave propagation through the inlet.

To assess the ability of the model to predict horizontal velocities versus depth, time series of along-channel velocities in the inlet throat from the model and the bottom-mounted ADCP are compared (Fig. 2A and B) (velocities are positive (negative) for flood/westward (ebb/eastward)

flow). The velocity fluctuations are dominated by the lunar semi-diurnal tide (M2), with some contribution from the solar diurnal tide (K1). During spring tide, the diurnal strengthening and weakening of the semi-diurnal current peaks reaches 20 cm s^{-1} near the surface (1 m depth) and 10 cm s^{-1} near the bottom (4 m depth). The spectra of depth-averaged simulated and observed along-channel velocity (Fig. 2C) confirm the dominant semi-diurnal variability at ca. 2 cycles per day (cpd) and the secondary diurnal variability at ca. 1 cpd. The M2 overtides are also present and dominated by the even components (M4, M6), with peaks at ca. 4 and 6 cpd comparable in magnitude to the diurnal (K1) contribution. Since the interactions between the semi-diurnal tide and its even overtides are known to contribute to tidal asymmetry and, thus, to drive net sediment transport (van de Kreeke and Robaczewska, 1993; van de Kreeke et al., 1997), the suspended and bedload deposition patterns computed in the present study will be indicative of both spatial and temporal asymmetries in the tidal currents.

Further comparisons between along-channel velocities at 1 m and 4 m depths reveal that the model reproduces the vertical structure of the tidal flow, but exhibits slightly higher speeds (5 to 10 cm s^{-1}) near the surface during flood flow. The depth-averaged, along-channel velocity is also similar between model and observations (RMSE is 20% of standard deviation, with the model leading observations by ca. 30 min). Since near-bottom and depth-averaged flows are used in the sediment transport model, the above agreement between simulated and observed velocities in both time and frequency domains provides an important first level of validation.

3.3. Simulated tidal and residual currents

The results are described in terms of the mean flood, ebb, and residual depth-averaged velocities (mean velocities represent averages of floods and ebbs over 15 days (one spring-neap tidal cycle)). Overall, the magnitude of the surface and bottom velocities is, respectively, ca. 20% larger and 50% smaller than the magnitude of the depth-averaged velocities. For flood flow, mean depth-averaged speeds are in the range of 50 to 70 cm s^{-1} in the inlet throat, and 20 to 40 cm s^{-1} in the tidal deltas (Fig. 2D). Maximum depth-averaged current speeds reach ca. 100 cm s^{-1} in the inlet throat and 40 to 60 cm s^{-1} in the tidal deltas. Mean and maximum ebb velocities have a slightly lower speed than the flood velocities and their spatial patterns are somewhat different, especially in the main channels of the tidal deltas (Fig. 2E). Depth-averaged residual velocities (Fig. 2F) indicate significant current asymmetries: flood flow dominates the northern part of the inlet throat, the two southern channels of the flood delta, and the southern channel of the ebb delta; ebb flow dominates the southern part of the inlet throat, the northernmost channel of the flood delta, and the northern channel of the ebb delta. The regions of no net residual velocity between the northern channel of the flood delta and the inlet throat, and between the southern channel of the ebb delta and the inlet throat, apparently coincide with the locations of the channel margin bars.

To further relate residual currents and sediment deposition, the mean depth-averaged relative vorticity of the flow ($\zeta = dv/dx - du/dy$, where u and v are the eastward and northward depth-averaged components of the velocity vector) normalized by the local planetary vorticity ($f = 2\Omega_e \sin\theta$, where Ω_e is the Earth's angular velocity and θ the latitude) is analyzed. Relative vorticity patterns suggest that the generation of vorticity by differential bottom friction dominates along the sides of the inlet throat, which is consistent with previous numerical studies in shallow estuaries (Ridderinkhof, 1989). Due to the spatial asymmetries between ebb and flood flows, the relative vorticity does not average out over a tidal cycle and excess negative (anticyclonic/clockwise) vorticity is present along the parts of the Key Biscayne and Virginia Key coastlines bordering the inlet throat. The regions of large negative vorticity present in both the northern part of the flood delta and the center of the ebb delta and are due to the advection of the relative vorticity generated along the sides of the inlet throat into the deltas. In tidal inlets, this mechanism of vorticity advection is hypothetically linked to the formation of a pair

of counter-rotating residual eddies in the flood and ebb deltas. The geometry of Bear Cut being non-symmetrical, the basic quadrupole eddy pattern is highly distorted, but still recognizable, especially in the flood delta with a weak cyclonic (counterclockwise) eddy in the southern part and a stronger anticyclonic (clockwise) eddy in the northern part. Finally, a maximum of positive residual relative vorticity is present in the deep part of the inlet throat and is presumably caused by vortex stretching (i.e., generation of cyclonic vorticity when depth is increasing; e.g. Robinson, 1981).

Since the residual depth-averaged relative vorticity is on average 5 to 10 times larger than the planetary vorticity, residual eddies are expected to induce sediment trapping (i.e., trapping occurs if $\zeta/f < -2$ or $\zeta/f > 0$; Pingree, 1978) and, deposition in the tidal inlet and deltas is instead controlled by a balance between sediment trapping and bottom shear stress (i.e., regions with low erosion potentials will be more conducive to deposition). For instance, sediment trapping might occur in the residual eddy located near the central lobe of the ebb delta (separating the main and marginal ebb channels), which coincides with a region of weaker ebb and flood currents.

4. Sediment transport model

The purpose of the model is to investigate the relocation of sediments due to tidal and wind-driven currents as a function of grain sizes by computing motion thresholds for suspended and bedload transport and advecting sediments using a Lagrangian particle-tracking method. While more complex formulations (e.g., van Rijn in Lesser et al., 2004) suitable for morphodynamic modeling are available to compute sediment exchange with the bottom through sources and sinks, the present model provides a simple approach to investigate the basic processes of entrainment, transport, and deposition by using empirical formulae to compute advection velocities for suspended and bedload transport and, thus, to identify prevalent mechanisms responsible for sediment relocation in the tidal inlet and deltas system. Since these empirical relationships relate sediment transport to bottom shear stress and frictional velocity, the model is also useful to assess motion thresholds and advection based on either depth-averaged flow velocities with a specified bulk drag coefficient (c_d), or near-bottom velocities with a prescribed roughness height (z_0). Considering that sediment characteristics do not necessarily dictate c_d or z_0 , the particle-tracking approach provides a framework to study the sensitivity of the results to these parameters, and their spatial and temporal variability based on bottom type and sediment redistribution. For instance, in Bear Cut, current ripples and small bioherms (0.2 to 0.5 m in height/width) in the inlet throat or seagrass beds in some parts of the deltas will presumably define the roughness height and drag coefficient in these regions.

Based on sediment samples collected in the tidal inlet and deltas (Steffen, personal communication), five categories (following the Wentworth scale) of sand diameters are considered: very fine ($d = 0.0625$ to 0.125 mm), fine ($d = 0.125$ to 0.25 mm), medium ($d = 0.25$ to 0.5 mm), coarse ($d = 0.5$ to 1.0 mm), and very coarse ($d = 1.0$ to 2.0 mm) particles. For each grain size, critical and advection velocities are computed as described below.

4.1. Suspended particle transport

Suspended transport is the dominant mode of motion for high flow velocities and/or small grain sizes, and its threshold is reached when the fall velocity (w_s) of a given particle equals the upward turbulent velocity component of the flow. Introducing the frictional velocity (u_*), this criterion can be approximated by the following relationship (Soulsby, 1997):

$$u_* = w_s \quad (1)$$

An empirical formula relating settling velocity and grain size for sand particles (see also Hallermeier (1981) and van Rijn (1984)) is used to relate grain size and settling velocity:

$$w_s = \nu d^{-1} \left[\left(10.36^2 + 1.049 D_*^3 \right)^{0.5} - 10.36 \right], \quad (2)$$

where ν is the kinematic viscosity of water, d the grain size diameter, and D_* the dimensionless grain size given by:

$$D_* = \left[g (s - 1) \nu^{-2} \right]^{1/3} d, \quad (3)$$

with g being the gravitational acceleration, and s the ratio of the grain density to the water density. The grain density is set to a default value of 2650 kg m^{-3} , representative of quartz sediments. An expression for the friction velocity is computed for both the depth-averaged (u_{bar}) and near-bottom (u_{bot}) flow conditions according to:

$$\text{depth-averaged flow: } \tau = \rho u_*^2 = \rho c_d u_{\text{bar}}^2 \quad (4)$$

$$\text{near-bottom flow: } u_{\text{bot}} = u_* \ln(z_b/z_0) \kappa^{-1}, \quad (5)$$

u_{bot} is the velocity at the first grid point above the bottom, and z_b the distance of the first grid point above the bottom. The drag coefficient c_d and roughness height z_0 are parameters specified in the ocean circulation model as 2.5×10^{-3} and 0.01 m, respectively. Finally, κ is Von Karman's constant ($\kappa = 0.4$). Substituting eqs. (4) and (5) into eq. (1), critical velocities for suspension as a function of grain size diameter for depth-averaged and bottom flow are given by:

$$u_{\text{bar,cr}} = w_s c_d^{-0.5} \quad (6)$$

and

$$u_{\text{bot,cr}} = w_s \ln(z_b/z_0) \kappa^{-1}, \quad (7)$$

where the fall velocity w_s is computed from eqs. (2) and (3). In the sediment transport model, the particles are simply advected by the flow velocity (u_{bar} and u_{bot}) when the critical threshold conditions are met (or exceeded) and remain stationary otherwise.

4.2. Bedload particle transport

Bedload transport is the dominant mode of motion for low flow velocities and/or large grain sizes, and its threshold is a function of the bottom shear stress. Introducing the Shield's

parameter, $\theta = \tau[\rho g(s-1)d]^{-1}$, an empirical relationship (Soulsby and Whitehouse, 1997) is used to relate grain diameter and critical bottom shear stress:

$$\theta_{cr} = 0.3(1 + 1.2D_*)^{-1} + 0.055 \left(1 - e^{-0.02D_*}\right), \quad (8)$$

where D_* is the dimensionless grain size (see eq. (3) above) and θ_{cr} the critical Shield's parameter defined as:

$$\theta_{cr} = \tau_{cr} [\rho g(s-1)d]^{-1} \quad (9)$$

Combining eqs. (8) and (9) to determine τ_{cr} and substituting into eqs. (4) and (5), expressions for the critical depth-averaged and near-bottom velocities are given by:

$$u_{bar,cr} = c_d^{-0.5} [g(s-1)d\theta_{cr}]^{0.5} \quad (10)$$

and

$$u_{bot,cr} = \ln(z_b/z_0) \kappa^{-1} [g(s-1)d\theta_{cr}]^{0.5} \quad (11)$$

Once threshold conditions are reached, an empirical formula (here, Bagnold, 1963) is used to compute the dimensionless bedload transport:

$$\Phi = F\theta^{0.5} (\theta - \theta_{cr}) \quad (12)$$

with $F = 0.1c_d^{-0.5}(\tan\phi + \tan\beta)^{-1}$, where ϕ is the angle of repose and β the angle of bed slope. For simplicity, the angle of repose is held constant with a value of $\tan\phi = 0.6$ (van de Kreeke and Robaczewska, 1993) and a flat bottom is assumed ($\beta = 0$). This approximation is generally valid for tidal deltas where topographic gradients are small, but may not necessarily apply to steeper slopes along the sides of the inlet throat. The dimensional bedload transport per unit width is typically obtained by multiplying Φ by $[g(s-1)d^3]^{0.5}$ and, similarly, a bedload transport velocity can be defined as:

$$u_b = \Phi [g(s-1)d]^{0.5} \quad (13)$$

(i.e., the vertical length scale for bedload transport is the grain size diameter, which is fairly realistic in Bear Cut since no significant bedform motion is observed in the inlet throat; this assumption may not be entirely valid in other inlet systems with different geological characteristics). Finally, by combining eqs. (4), (5), and (9) into eq. (13), the particle advection velocities for the bedload transport model for depth-averaged and near-bottom flow conditions are given by:

$$u_{b,bar} = F_{bar} u_{bar} (u_{bar}^2 - u_{bar,cr}^2) \quad (14)$$

and

$$u_{b,bot} = F_{bot} u_{bot} (u_{bot}^2 - u_{bot,cr}^2) \quad (15)$$

where

$$F_{bar} = 0.1 c_d \tan \phi^{-1} [g(s-1)d]^{-1}$$

and

$$F_{bot} = 0.1 [\kappa / \ln(z_b/z_0)]^3 c_d^{-0.5} \tan \phi^{-1} [g(s-1)d]^{-1}$$

In the model, the particles are advected by the bedload velocity ($u_{b,bar}$ and $u_{b,bot}$) when the critical threshold conditions are met and remain stationary otherwise.

Sediment transport due to surface gravity wave orbital velocities is not considered in the model because the shallow continental shelf attenuates waves propagating into the inlet (during the study period, surface gravity waves were at least an order of magnitude smaller than the tidal amplitude). However, their contribution is certainly important for suspending and transporting sediments over the ebb delta and during periods of high winds associated with cold front (winter) or tropical cyclone (summer) passages. The Stokes drift associated with tidal wave distortion in shallow water has also been neglected, but its potential contribution to net sediment transport should be investigated and compared to that of the residual tidal current (Stokes drift velocities (see LeBlond and Mysak, 1978) in the tidal inlet and deltas are of the same order of magnitude ($O(10 \text{ cm s}^{-1})$) as the residual currents).

4.3. Simulated sediment transport

To concentrate on transport through the main flow channels, particles are initially placed uniformly in the inlet throat and tidal deltas where the depth is greater than 3 m and are released every hour during the first tidal cycle to remove the influence of flow direction at the time of release. The influence of tidal currents on suspended and bedload sediment transport patterns is studied by analyzing the mean particle position averaged over the last tidal cycle as a function of grain size (particles transported from delta to delta during the averaging period are excluded as the mean position is not indicative of relocation).

For suspended sediment transport (Fig. 3A–C), the deposition patterns are strongly influenced by grain sizes. Very fine particles are advected beyond the flood delta and deposit in the lagoon. Fine particles remain within the flood delta and aggregate near the lobes on the edges of the flood-dominated (i.e., southern and center) channels. The critical threshold of motion for medium-size sand is never reached in the tidal inlet, so the particles never leave their initial location. Consequently, all particles larger than fine sand will not be displaced by suspended sediment transport due to tides alone. Overall, depth-averaged and near-bottom flows result in similar deposition patterns, implying that the two approaches are mutually consistent.

Since only very fine and fine particles are transported in suspension, bedload motion is the main transport mechanism (Fig. 3D–F) for larger particles (i.e., medium, coarse, and very coarse sand sizes). The deposition patterns for medium-size sand indicate that most particles are carried away from the inlet throat into the deltas. In the flood delta, medium sand aggregates into the channels. In the ebb delta, the particles that do not remain in the throat and channels (mostly medium and coarse sand sizes) typically aggregate along the margin bar, as well as near the edges of the channels. While the overall deposition patterns based on depth-averaged and near-bottom flow conditions are similar, differences occur where the bottom topography changes rapidly (e.g., the channel margin bars), suggesting that the two approaches are not entirely consistent in these regions (i.e., depth-averaged flow may not adequately represent effects of abrupt changes in topography).

Since suspended sediments are the principal source of scatterers in the tidal inlet, the signal-to-noise ratio (SNR) from the ADCP (expected to be proportional to the amount of scatterers in the water column) should contain information about waterborne sediment content. The demeaned (removal of acoustic signal attenuation with distance) 14-day SNR time series indicates that the values are mostly uniform through depth, with higher values occurring episodically in the upper part of the water column over the duration of the record (Fig. 4A). The depth-averaged SNR values typically exhibit larger fluctuations during the first part of the record (2 to 6 June, coinciding with spring tides) and for a 2-day interval around 12 June (Fig. 4B). While mostly uniform with depth, the near-bottom SNR values are significantly (i.e., one standard deviation) larger than the near-surface SNR values during the first part of the record (Fig. 4C), indicating that larger suspended particles are present near the bottom during periods of higher flow speeds (i.e., spring tide). These increases in SNR values occur only once a day, which suggests that the additive contribution of the diurnal (K1) tidal current to the semidiurnal (M2) tidal current might play a significant role in creating conditions where heavier sediments can be transported in suspension in the inlet throat. Since the suspended sediment transport model includes a motion threshold (see eqs. (6) and (7) in Section 4.1), advection speeds at the ADCP location can be computed for different particle sizes. The results indicate that fine sand particles ($d = 0.2$ mm) will be in suspension at some point during ebb and flood flow throughout the 14-day period (Fig. 4D). Medium-to-fine particles ($d = 0.25$ mm) are in suspension for each ebb and flood flow during spring tide, but, as the currents weaken, their threshold of motion is only reached diurnally for a few days around 12 to 14 June (Fig. 4E), coinciding, interestingly, with the higher depth-averaged SNR values observed in the ADCP record during the same period. Medium sand particles ($d = 0.3$ mm) are expected to be in suspension four times diurnally during spring tide (Fig. 4F), which is probably related to the increased near-bottom SNR values over the first six days of the ADCP record. These striking similarities between observed SNR values and simulated suspended sediment characteristics, as well as the large variations in transport conditions (i.e., motion threshold) over a narrow range of grain size diameters (0.2 to 0.3 mm), indicate a need for conducting detailed suspended sediment measurements in the Bear Cut tidal inlet to clearly establish the relationship between sediment size, SNR values, and simulated particle transport.

5. Discussion

The results from the sediment transport model can be qualitatively compared with sediment samples collected from the inlet system (Steffen, personal communication). In the flood tidal delta, the simulated deposition locations for suspended transport of very fine and fine sand sizes suggest that the smaller particles are transported into Biscayne Bay, and the larger particles are deposited on the lobes and along the edges of the channels. The bedload sediment transport patterns indicate that medium size particles remain in the channels of the flood delta, which is consistent with the observed distribution of sediment characteristics in that region. The throat and the channels of the ebb delta are predicted to be mostly dominated by coarse

and very coarse sand sizes, with some medium size particles on the lobes and along the edges and margin bars. Again, a qualitative comparison with sediment samples collected in the same regions indicates reasonable similarities. Overall, the combined simulated suspended and bedload sediment deposition patterns demonstrate a general trend of decreasing sand sizes with increasing distance away from the throat and into the deltas (Fig. 5A). This finding is consistent with the fact that particles are advected by the flow, or a fraction thereof, and that the currents are typically stronger in the throat (i.e., only larger particles tend to remain there) and weaker in the deltas (i.e., smaller particles tend to deposit there). A similar gradation pattern in grain sizes is observed in the sediment samples collected in the tidal inlet system (Fig. 5B). Qualitatively, the mean grain size is greatest in the inlet throat and decreases outward towards both flood and ebb tidal deltas.

Since the model is simply forced by tides with no alongshore propagation, the results do not account for the influence of alongshore tidal flow, offshore boundary currents, and storm-driven currents and waves. Coupling the high-resolution tidal inlet model to a coarser (1 to 2 km horizontal resolution) regional model for the Straits of Florida might improve the realism of the simulated flow in Bear Cut by allowing the tidal currents to interact with the Florida Current and associated mesoscale variability (e.g., frontal eddy passages (Fiechter and Mooers, 2003)). Surface radar observations (Haus et al., 2000) indicate that cyclonic disturbances translating along the Southeast Florida shelfbreak may induce a near-shore equatorward current, which could modulate the tidal and wind-driven flows. However, the recurrence period (ca. 1 week) and velocities (max. 50 cm s^{-1}) associated with frontal eddy passages are, probably, of lesser importance to sediment transport within the inlet and deltas system than the strong, semidiurnal tidal currents. Valuable information on the modulation of tidal flow by wind- and wave-driven currents would also certainly be gained by adding a surface gravity wave model and repeating this study during cold front passages in the winter months to investigate the complex geomorphologic processes linked to wave action at the offshore edge of the ebb delta. Similarly, including wetting and drying might also improve the simulated currents and sediment transport. Further work should include additional validation of the Lagrangian suspended and bedload particle transport approach, as well as some sensitivity studies related to the specification of the roughness height and drag coefficient.

6. Conclusions

Despite some discrepancies related to the oversimplification of boundary conditions, the three-dimensional coastal ocean circulation model is an effective tool for investigation of the horizontal and vertical structure of the currents in a tidal inlet and deltas system. The simulated flow fields reveal significant spatial asymmetries and preferential pathways in the inlet system and provide some insight on the balance between flood and ebb tides in the different channels. The results also suggest different generation mechanisms for residual relative vorticity, such as differential bottom friction and vortex stretching.

The Lagrangian particle-tracking model for suspended and bedload sediment transport provides a simple and useful method to investigate sediment relocation and dominant transport mode (i.e., suspended vs. bedload) as a function of grain size attributes. While the deposition patterns calculated from depth-averaged and near-bottom velocities are in general similar, slight differences in bedload transport are noticeable in regions of abrupt topographic changes. In general, the simulated deposition patterns qualitatively correlate with the observed distribution of sediment characteristics in the inlet throat and tidal deltas, and are linked to some of the geomorphologic features.

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Fig. 1.

Bear Cut tidal inlet and deltas system. (A) Location map of the southeastern coast of the United States, Florida, and Miami-Dade County (Bear Cut is located just offshore of Miami between Virginia Key and Key Biscayne). (B) Schematic map of the general geomorphic attributes for Bear Cut and associated ebb and flood tidal deltas. (C) Model domain and computational grid (subsamped by 5 for clarity) with bottom topography (grayscale: depth in meters) for Bear Cut tidal inlet system and surrounding region. (D) Details of computational grid (full resolution) in Bear Cut tidal inlet with bottom topography (grayscale: depth in meters). (E) Locations of in-situ observations: Virginia Key tide gauge (square) and bottom-mounted ADCP (triangle).

Fig. 2. Observed and simulated velocities in the tidal inlet system. (A, B) Fourteen-day time-series of horizontal along-channel velocity in cm s^{-1} from simulation (black line) and bottom-mounted ADCP observations (gray line) at 1 m depth (A) and 4 m depth (B) (velocity is positive for flood flow and water depth at this location is ca. 5 m; depths are referenced to mean sea level). (C) Raw spectra (Fast Fourier transform) of simulated (black line) and observed (gray line) depth-averaged velocities at bottom-mounted ADCP location. (D–F) Horizontal maps of depth-averaged velocities (grayscale: speed in cm s^{-1}) for mean flood (D), mean ebb (E), and mean residual (F) flows (contours indicate depth in meters).

Fig. 3.

Simulated deposition patterns for suspended and bedload sediment transport (particles are initially located in the main channels (i.e., deeper than 3 m) of the inlet throat and tidal deltas). (A–C) Suspended transport using depth-averaged (white particles) and near-bottom (black particles) velocities for very fine ($d = 0.0938$ mm, A), fine ($d = 0.1875$ mm, B), and medium ($d = 0.375$ mm, C) sand sizes; critical suspension speeds for very fine, fine, and medium particles are 14, 47, 113 cm s^{-1} for depth-averaged flow and 4, 14, 32 cm s^{-1} for near-bottom flow. (D–F) Bedload transport using depth-averaged (white particles) and near-bottom (black particles) velocities for medium ($d = 0.375$ mm, D), coarse ($d = 0.75$ mm, E), and very coarse ($d = 1.5$ mm, F) sand sizes; critical bedload speeds for medium, coarse, and very coarse particles are 28, 38, 58 cm s^{-1} for depth-averaged flow and 8, 11, 17 cm s^{-1} for near-bottom flow.

Fig. 4.

Simulated suspended particle characteristics (at location of bottom-mounted ADCP) and signal-to-noise ratio (SNR) from bottom-mounted ADCP (averaged over three beams and mean removed to account for expected decay with distance). (A) SNR values versus depth (grayscale: dB). (B) Depth-averaged SNR values (dB). (C) Near-bottom SNR minus near-surface SNR values in excess of mean plus one standard deviation (dB). (D–F) Simulated depth-averaged velocity (cm s^{-1}) of suspended sediments at ADCP location for sand size of 0.2 mm (D), 0.25 mm (E), and 0.3 mm (F) diameters.

Fig. 5. Simulated and observed spatial distributions of sediments in the Bear Cut tidal inlet and deltas system. (Left) Deposition patterns from suspended and bedload sediment transport models. (Right) Distribution of mean grain size from surficial sediment samples (Ikonos image; ©spaceimaging.com).