1. Introduction

Salt marshes are highly productive ecosystems and are often the first line of defense against coastal storm impacts, protecting shorelines and attenuating waves (Gedan et al., 2011). Understanding the trajectory of salt marshes—whether they expand or retreat, accrete or erode—is key to predicting their future evolution. Suspended-sediment supply has long been associated with the persistence of salt marshes (Redfield, 1972), with a sediment surplus indicating an accreting or laterally expanding marsh, and a deficit an eroding or retreating marsh. (Other processes like organic production are also relevant to maintaining marsh elevation Nyman et al., 2006.) Determining the sediment trajectory requires accurate sediment-flux measurements. Both water (Lane et al., 1997; Simpson & Bland, 2000) and sediment (French et al., 2008; Townend & Whitehead, 2003) fluxes, however, are difficult to measure, particularly in determining a residual flux from large and opposing flood and ebb values. Employing simpler parameters that capture net flux direction and magnitude, then, is an appealing prospect.

Idealized wetland geomorphic models (e.g., D’Alpaos & Marani, 2016; Mariotti & Fagherazzi, 2010) require synthesized, distilled field observations to serve as model boundary conditions. Constant and uniform SSC is specified in many wetland models, including the elevation-based Marsh Equilibrium Model (e.g., Byrd et al., 2016; Schile et al., 2014) and WARMER models (Swanson et al., 2014), as well as idealized 3D models (Kirwan & Murray, 2007). Other idealized models may allow spatial variability in suspended-sediment concentration (SSC), but specify constant sediment supply at the model boundary. Additional parameters, including time-variable SSC and different SSC values corresponding to floods and ebbs are logical metrics for model development because temporal variability may be crucial to flux dynamics.

Metrics based on SSC, including the flood–ebb SSC differential and mean SSC, have been shown to reflect the sediment-flux trajectory for a variety of salt-marsh systems (Ganju et al., 2013, 2015, 2017; Suk et al., 1999). Much of the previous work demonstrating a linear relationship between the SSC differential and area-normalized sediment flux has been applied at timescales several weeks to several months in length, leaving unclear how or whether these metrics apply to shorter timescales, as well as the mechanistic driver of the relationship. In the present work, we show that these metrics are robust at timescales as short as a diurnal tidal cycle. We further show that for sites with a standing tidal wave, defining flood and ebb using the temporal water-surface gradient instead of velocity is sufficient to compute the SSC differential, enabling less-intensive measurements that capture fundamental sediment-flux parameters. Distilling the sediment-flux trajectory into simple metrics improves sediment-budget assessment, drives geomorphic model development, and clarifies field observations.

1.1. Study Sites

We consider tidal-channel timeseries data previously collected and published by the U.S. Geological Survey in seven salt-marsh complexes on the U.S. Atlantic and Pacific coasts, and present new data from two sites.
Figure 1. Base map showing the nine sites in this study (circles) and three of the four sites from other studies (squares) described in Figure 4.

on the U.S. Gulf coast (Figure 1). Sites re-analyzed here include Seal Beach, California (Ganju et al., 2016; Rosencranz et al., 2016), Blackwater and Fishing Bay, Maryland (Ganju et al., 2013, 2015), Reedy Creek and Dinner Creek, New Jersey (Suttles et al., 2015), and Ogunquit, Maine (Montgomery et al., 2015); full details of those studies are given in the preceding citations. All sites had well-defined creeksheds (i.e., drainage areas landward of the measurement location) constrained by topographic boundaries and were drained by a primary tidal channel, where sediment-flux measurements were made. Maximum tidal range was 0.35 to 2.0 m at the sites, which featured a mixture of salt-marsh vegetation species, climate, elevation, and relative sea-level rise (Ganju et al., 2017).

Additional sites for this study are located in Grand Bay, a subembayment of Mississippi Sound in the Gulf of Mexico. Straddling the Alabama/Mississippi border, Grand Bay is a 30-km² microtidal bay surrounded by 20 km² of salt marsh with lateral erosion rates up to 5 m y⁻¹. We measured sediment fluxes and water-quality parameters from August 2016 to January 2017 (Nowacki et al., 2018) in two tidal creeks, Bayou Heron and Bayou Middle, which incise a Juncus roemerianus salt marsh and drain to Grand Bay. Bayou Heron, with a channel width of 140 m and thalweg depth of 4 m at the measurement location, is larger than Bayou Middle, which has a width of 50 m and is 3 m deep. Maximum tidal range is 0.7 m in both channels.

In addition to the nine sites already described, in Section 4.2 we consider data from four previously published studies at salt-marsh sites in the United States (Figure 1) and the Netherlands.

2. Methods

We use two primary metrics: flood-ebb SSC differential (ΔC), defined as mean flood SSC minus mean ebb SSC, with flood and ebb determined using velocity direction, and mean SSC (⟨C⟩). Later, we consider an additional method of determining flood and ebb by using the temporal water-surface gradient (ΔCws) in locations with a standing tidal wave. These metrics can be applied over any timescale equal to or longer than a tidal cycle. The timescale used historically has been the longest possible for a given data set, capturing as many events as possible in an attempt to represent the long-term trajectory of the system. Here we apply the metrics over shorter timescales: the diurnal tidal cycle (24.8 hr), spring-neap cycle (14 days), one and two lunar months (28 days and 56 days), and full deployment (mean 76 days, range 25–147 days). We have time-weighted the ΔC for the diurnal interval by multiplying the mean concentrations on flood and ebb by the duration of the flood and ebb, respectively, and dividing the difference of these numbers by the total duration of flood and ebb. The relationships described at the diurnal interval hold without the time weighting, but with weaker correlations, because small differences in the actual length of the tidal cycle and the averaging interval may introduce errors in the computed quantities. A brief discussion of error sources and their potential impact on these metrics is available in the supporting information.
Figure 2. Example timeseries from Bayou Middle, Grand Bay. (a): SSC and channel depth; (b): ⟨C⟩ at 24.8 hr and 14 d averaging intervals; (c): ΔC at the same intervals; (d): sediment flux with estimated uncertainty in gray; (e): sediment flux averaged over the same intervals.

We compute the total sediment flux as the product of the water discharge and SSC: \( Q_t = QC \). We normalize these fluxes by the landward drainage area of each channel:

\[ Q_s = \frac{Q}{A_t} \text{ kg m}^{-2} \text{ y}^{-1}. \]

Positive values of \( Q_s \) indicate sediment import to the marsh complex. This approach assumes that mass is not exchanged across creekshed boundaries, and that fluxes derived from the channel measurements are representative of the creekshed; this assumption is discussed in the supporting information.

3. Results
3.1. Computing the Metrics
Inspecting data collected at Bayou Middle provides an example of computing ΔC and ⟨C⟩ (Figure 2). During the 1-month example timeseries, SSC ranged from about 10 mg L\(^{-1}\) to greater than 100 mg L\(^{-1}\) because of tidal variability and wind-wave resuspension from daily sea breezes (Figure 2a). The 15-min SSC data are averaged into tidal (24.8 hr) and spring–neap (14 days) intervals (Figure 2b). These different ⟨C⟩ averaging intervals illustrate variability at different scales (e.g., diurnal, storms, fortnightly) and show a reduction in variability with longer timescales. The tidal interval shows greater ⟨C⟩ during spring tides and lessened
Figure 3. Flood-ebb SSC differential versus sediment flux per unit area for tidal (24.8 hr), fortnightly (14 d), four-week (28 days), eight-week (56 days), and full deployment (mean 76 days) averaging intervals. Values of ΔC for the 24.8 hr interval have been time-weighted, and Blackwater has not been included in computing the best-fit line or other statistics. Including Blackwater in the best-fit line for the 24.8-hr interval results in $r^2 = 0.35$. The predictive power of ΔC also tends to improve with lengthening of the averaging interval (Table 1). For example, at the tidal timescale, directional agreement is 76%, but at the 2-month interval, the agreement is 83%. As above, restricting $|\Delta C|$ to values greater than 1 mg L$^{-1}$ improves the agreement to 84% at the tidal timescale.

3.2. Predicting the Flux Direction and Magnitude
At all locations and averaging intervals, ΔC is correlated with $Q_s$ (Figure 3), predicting net flux directionality over the timescale of interest. For example, if SSC and velocity are measured for 14 days, one can be reasonably confident that the sign of the net sediment flux over those 14 days is the same as that of ΔC—that is, that positive ΔC indicates sediment import, and negative ΔC indicates export.

A challenging aspect of accurately measuring net sediment fluxes is that two large values, one positive and one negative, are combined, resulting in a residual value often near zero. Indeed, predicting net sediment-flux direction using ΔC has greater success when ΔC is relatively large. For example, when considering all timescales, we have 75% confidence in predicting $Q_s$ direction when the absolute value of the flood-ebb SSC differential ($|\Delta C|$) is less than 1 mg L$^{-1}$. When $|\Delta C|$ is greater than 1 mg L$^{-1}$, this rises to 83%; when $|\Delta C|$ is greater than 5 mg L$^{-1}$, our confidence is 97% that the direction of $Q_s$ is the same as ΔC. This relatively easy-to-measure metric quantifies uncertainty in the sediment budget: the sign of a large ΔC value is more likely to indicate the true flux direction.
Table 1

| Interval | $r$   | $m$              | $b$   | $f$ | $n$ | $f_{|\Delta C|>1}$ | $n$ | $f_{|\Delta C|>5}$ | $n$ | $f_{|\Delta C_{ss}|>1}$ | $n$ |
|----------|-------|------------------|-------|-----|-----|-------------------|-----|-------------------|-----|-------------------|-----|
| 24.8 h   | 0.73  | 0.190 ± 0.006    | 0.131 | 0.76 | 882 | 0.84              | 607 | 0.97              | 186 | 0.72              | 485 |
| 14 d     | 0.77  | 0.098 ± 0.011    | 0.048 | 0.78 | 59  | 0.83              | 42  | 0.90              | 10  | 0.77              | 31  |
| 28 d     | 0.70  | 0.070 ± 0.015    | 0.065 | 0.85 | 27  | 0.95              | 19  | 1.00              | 2   | 0.87              | 15  |
| 56 d     | 0.75  | 0.060 ± 0.018    | 0.059 | 0.83 | 12  | 0.90              | 10  | 1.00              | 1   | 1.00              | 6   |
| Deployment (mean 76 d) | 0.73  | 0.057 ± 0.018    | 0.100 | 0.67 | 12  | 0.88              | 8   | 1.00              | 2   | 1.00              | 5   |

Note. Definitions for sign agreement include all data ($f$), those with absolute value of $\Delta C$ greater than 1 mg L$^{-1}$ ($f_{|\Delta C|>1}$), and for standing-wave sites, using the water-surface gradient to determine flood and ebb, with $|\Delta C_{ss}| > 1$ mg L$^{-1}$ ($f_{|\Delta C_{ss}|>1}$). Sample sizes $n$ are shown for each of the fractional agreement values. Values of $\Delta C$ for the 24.8-hr interval have been time-weighted, and Blackwater has not been included in computing the best-fit line.

In addition to the flux direction, $\Delta C$ can be used to predict the magnitude of the residual sediment flux. We omitted Blackwater in developing the best-fit line between $\Delta C$ and $Q_t$ because it has a large internal source of sediment from marsh deterioration and subsequent conversion to open water (Stevenson et al., 1985; Ganju et al., 2015). Including Blackwater at the 24.8-hr interval reduces the goodness of fit, which we speculate may be related to the strong subtidal flux at that location. Across all timescales, least-squares slopes range from 0.057 to 0.190 (Table 1). As timescales lengthen, the slope of the relationship between $\Delta C$ and $Q_t$ decreases, but remains within a factor of three. In Table 1, we have chosen a single regression at timescale and 90% at the 2-month interval. The $y$-intercept of the regression (Table 1) can be used to assess the applicability of this metric for a given site, as $\Delta C$ values less than the intercept may result in an incorrect inferred flux direction.

In addition to the flux direction, $\Delta C$ can be used to predict the magnitude of the residual sediment flux. We omitted Blackwater in developing the best-fit line between $\Delta C$ and $Q_t$ because it has a large internal source of sediment from marsh deterioration and subsequent conversion to open water (Stevenson et al., 1985; Ganju et al., 2015). Including Blackwater at the 24.8-hr interval reduces the goodness of fit, which we speculate may be related to the strong subtidal flux at that location. Across all timescales, least-squares slopes range from 0.057 to 0.190 (Table 1). As timescales lengthen, the slope of the relationship between $\Delta C$ and $Q_t$ decreases, but remains within a factor of three. In Table 1, we have chosen a single regression at the diurnal timescale; individual slopes for each site are presented in Supporting Information Table S1, and variation among the individual slopes is discussed in Section 4.

3.3. Defining Flood and Ebb by Water-Surface Gradient

For sites with a standing tidal wave, flood and ebb can be approximately defined using the temporal water-surface gradient instead of flow velocity. The phase differences between the water-level and velocity for the primary tidal constituent at Blackwater and Fishing Bay are $1^\circ$ and $36^\circ$, respectively, indicating that they both have roughly progressive tides, so we remove them from this analysis. The other sites had phase differences of $75^\circ$ (range 54–84$^\circ$), representative of standing-wave tides. Using the gradient to compute the flood-ebb SSC differential ($\Delta C_{ss}$) for these sites maintains the relationship with $Q_t$ for $\Delta C_{ss} > 1$ (Table 1). At the tidal timescale, there is 72% agreement in directionality. At the 2-month and longer interval, the agreement is 100%. For these standing-wave sites, simply measuring the water level and SSC is sufficient to predict whether a marsh is importing or exporting sediment over a given averaging interval.

4. Discussion

4.1. Flood–Ebb SSC Differential as a Predictive Tool

An idealized model specifying sinusoidal water discharge and different functional forms for SSC illustrates how $\Delta C$ can predict net sediment flux. Assume $Q = \sin \omega t$, $C = c_0 + c_1 \sin \omega t$, and $Q_t = QC$, where $\omega$ is the angular tidal frequency, $c_0$ is the mean SSC, and $c_1$ is the amplitude of the SSC variability such that $C \geq 0$. Here SSC is a scaled and shifted version of water discharge. Integrating $Q_t$ over a tidal cycle $\tau$ gives

$$\int_0^\tau \sin \omega t (c_0 + c_1 \sin \omega t) \, dt = \frac{c_1 \tau}{\omega}.$$ 

The flood–ebb SSC differential is

$$\Delta C = \frac{\tau}{2} \int_0^{\tau/2} (c_0 + c_1 \sin \omega t) \, dt - \frac{\tau}{2} \int_{\tau/2}^{\tau} (c_0 + c_1 \sin \omega t) \, dt = \frac{4c_1}{\omega}.$$ 

In this scenario, net water flux is zero, but net sediment flux need not be because $c_1$ drives variability in $C$ over the tidal cycle. Both $\Delta C$ and net $Q_t$ scale with $c_1$, the amplitude of the SSC curve, and as a result, $\Delta C$ varies linearly with $Q_t$. A simpler scenario models SSC as a square wave: $C = c_0 + c_1 \text{sign} (\sin \omega t)$. Here $\Delta C = 2c_1$, and the integrated sediment flux is

$$\int_0^\tau \sin \omega t (c_0 + c_1 \text{sign} (\sin \omega t)) \, dt = \frac{4c_1}{\omega}.$$
Again ΔC and Q_s scale with c_1. In both scenarios, the sign of ΔC indicates the sign of Q_s, and furthermore a larger ΔC (via greater c_1) drives a larger Q_s.

For less-idealized field observations, we can implement a flux decomposition approach (e.g., Dyer, 1974) such that

\[ F = \langle Q_s \rangle = \langle uaC \rangle = \langle \langle u \rangle \langle a \rangle \langle C \rangle \rangle + \langle \langle u' \rangle \langle a' \rangle \langle C \rangle \rangle + \text{small terms}, \quad (1) \]

where \( u \) is the mean channel velocity, \( a \) is the channel area, \( C \) is SSC, angle brackets indicate tidally averaged values, and primes denote deviations from the tidal average. The first term on the right-hand side is the advective flux, arising from the subtidal terms; the second is the dispersive flux, attributable to the tidal-scale temporal correlation between velocity and SSC; and the third is the Stokes-drift flux, from the correlation between velocity and depth when tides are progressive (this term is zero for standing-wave tides). For the purely tidal system described in the idealized example above, the advective component is zero because \( \langle u \rangle = 0 \).

The advective and Stokes components depend on \( \langle C \rangle \), and the dispersive component is dependent on temporal (e.g., flood-ebb) SSC fluctuations, similar to \( \text{ΔC} \). At all sites, \( \text{ΔC} \) scales with \( \langle C \rangle \) (Figure 4 and supporting information Figure S5), so much of the sediment-flux signal is captured using these parameters. In general, systems with large \( \langle C \rangle \) have considerable sediment sources (e.g., river input, eroded marsh material) nearby, and these sources may preferentially influence one phase of the tide. As a result, when \( \langle C \rangle \) increases, the absolute value of \( \text{ΔC} \) is likely to increase as well.

### 4.2. Comparison to Other Studies

Here we consider four additional studies (Dankers et al., 1984; Settlemyre & Gardner, 1977; Suk et al., 1999; Ward, 1981) that included tabular data from which \( \text{ΔC} \), \( \langle C \rangle \), and \( Q_s \) can be computed. These studies, three in the United States and one in the Netherlands, measured fluxes in salt-marsh tidal channels over periods ranging from several weeks to nearly a year. The relationship between \( \text{ΔC} \) and \( Q_s \) was moderate to strong in all four studies, with \( r \) values ranging from 0.71 to 0.98, and diurnal \( \text{ΔC} \)-\( Q_s \) slopes ranging from 0.019 to 0.314 (Supporting Information Table S1). These slopes are comparable to those of the nine sites considered so far.

A wide range of sedimentary environments is captured in the combined 13-site data set: the maximum \( \langle C \rangle \) of 257 mg L\(^{-1}\) is about 64 times greater than the minimum of 4 mg L\(^{-1}\) (Supporting Information Table S1). The absolute value of \( \text{ΔC} \) ranges from 0.05 to 110 mg L\(^{-1}\), with six of the 13 sites having negative \( \text{ΔC} \). When considering both importing and exporting sites, \( \text{ΔC} \) remains correlated with both \( \langle C \rangle \) and \( Q_s \) (Supporting Information Figures S4 and S5), continuing the trends identified in Figures 3 and 4.

### 4.3. Relating \( \text{ΔC} \) to Geomorphic Development of Salt-Marsh Systems

Computing \( \text{ΔC} \) provides a metric by which to infer net sediment flux for a given time period. Over a sufficiently long period, this quantity is representative of the salt marsh’s geomorphic trajectory because long-term sediment flux ultimately drives growth or decay. At the longest temporal scales—seasons, years, and decades—remotely sensed quantities like the ratio of unvegetated to vegetated marsh coverage (UVVR) are indicative of marsh health (Ganju et al., 2017). Vertical accretion and its relation to \( \langle C \rangle \) also are often used to assess the health of a marsh, but \( \langle C \rangle \) misses a fundamental aspect of sediment in suspension—the direction in which it is being fluxed. A valuable alternative is found in \( \text{ΔC} \), which incorporates the flux direction via its relationship to \( Q_s \). Even though the indicated flux direction and magnitude is limited to that of the monitoring period, shorter measurement campaigns remain valuable. If a large storm—a hurricane, for example—is captured during a relatively short record, the \( \text{ΔC} \) method enables inference of the directionality of marsh-channel sediment flux during that event, which can be compared to remotely sensed data collected prior to and after the storm. Similarly, if the sediment supply is dominated by tidal action, then a shorter campaign may indicate as much with \( \text{ΔC} \). Measurement campaigns can be timed to when a setting is most active and expected to have the largest flux signals, for example, winter storms or summer sea breezes.
A sediment budget representative of the period of interest can in this way be developed via application of \( \Delta C \). This approach can be simplified further if a standing tidal wave is present: in this case, a pressure and turbidity sensor may be deployed instead of using velocity instruments.

For the systems considered in this study, we can use representative values of \( \Delta C \) for the longest timescales to infer marsh resilience. Using the long-term \( \Delta C-Q_s \) relationship in Table 1 and assuming a \( \Delta C \) value of 1 mg L\(^{-1}\) (reasonable for the marshes in this study; Supporting Information Table S1) and a marsh-sediment density of 373 kg m\(^{-3}\) (Ganju et al., 2017) results in a depositional thickness of 0.375–0.472 mm on a theoretical marsh over the course of a year. (This estimate assumes no lateral expansion of the marsh and represents an upper bound on deposition.) Given that the average rate of local sea-level rise at the nine sites considered here is 3.2 mm y\(^{-1}\), a \( \Delta C \) value of 1 mg L\(^{-1}\) is not sufficient to prevent marsh drowning. On average, \( \Delta C \) must be at least 15–30 mg L\(^{-1}\) to prevent submergence for the sites considered here. The mean \( \Delta C \) for the sediment-importing sites (Bayou Heron, Dinner Creek, Fishing Bay, Ogunquit, and Seal Beach 1) is approximately 2 mg L\(^{-1}\), indicating that sediment deficits are pervasive even for those sites not losing sediment.

### 4.4. Collapsing the \( \Delta C-Q_s \) Slope Using Creekshed Area

Developed using creekshed areas spanning two orders of magnitude (0.2–70 km\(^2\)), site-specific diurnal regression slopes range from 0.042 to 0.314 (Supporting Information Table S1). As a result, using the slope from Table 1 (0.190 ± 0.006) instead of the site-specific slopes could result in a predicted sediment flux that is low by a factor of two or high by a factor of four. The approximately eightfold increase between the minimum and maximum slopes is considerable, but shows that normalizing the flux by drainage area is an important step in constraining this relationship. Without normalizing by the area, the relationship with directionality would hold, but with less skill in predicting the flux magnitude. Although area-normalizing improves the prediction skill, the remaining range in slopes hints at what additional parameters may be important in further collapsing the relationship. The smallest slopes correspond to two sites—Blackwater and the channel studied by Dankers et al. (1984)—that are relative outliers. Blackwater has by far the largest drainage area and the second-highest mean SSC, about 64 mg L\(^{-1}\). The Dankers et al. (1984) channel had a relatively small drainage area but had the greatest mean SSC by a factor of four over Blackwater, and a factor of 12 over the nine-site average (Supporting Information Table S1). Similarly, the slope between \( \Delta C \) and \( \langle C \rangle \) begins to flatten for large values of \( \Delta C \) (Figure 4 and Figure S5). That the slopes appear to level off at high \( \Delta C \), and \( \langle C \rangle \) may suggest a change in the relationship among these three variables at sites with high concentrations, and that an additional geomorphic or sedimentary parameter could be used to collapse the relationship further.

### 5. Conclusions

Previous work has shown the importance of the flood-ebb SSC differential (\( \Delta C \)) as an indicator of sediment-flux direction and, by inference, tidal wetland vulnerability, over monthly and longer timescales. By extending this approach to shorter timescales using data from 13 salt-marsh channels, we showed the relationship holds down to the 24.8-hr tidal timescale. In addition to directionality, flux magnitude can be predicted with \( \Delta C \). For sites with a standing tidal wave, flux direction can also be inferred by defining flood and ebb using the water-surface gradient. The magnitude of \( \Delta C \) can be predicted with \( \Delta C \). For sites with a standing tidal wave, flux direction can also be inferred by defining flood and ebb using the water-surface gradient. The magnitude of \( \Delta C \) can be predicted with \( \Delta C \).

### References

Byrd, K. B., Windham-Myers, L., Leeuw, T., Downing, B., Morris, J. T., & Ferner, M. C. (2016). Forecasting tidal marsh elevation and habitat change through fusion of Earth observations and a process model. Ecosphere, 7(11), 1–27. https://doi.org/10.1002/ecs2.1582

D’Alpaos, A., & Marani, M. (2016). Reading the signatures of biologic-geomorphic feedbacks in salt-marsh landscapes. Advances in Water Resources, 93, 265–275. https://doi.org/10.1016/j.advwatres.2015.09.004
NOWACKI AND GANJU

Geophysical Research Letters
10.1029/2019GL083819

Dankers, N., Binsbergen, M., Zegers, K., Laane, R., & van der Loeff, M. R. (1984). Transportation of water, particulate and dissolved organic and inorganic matter between a Salt Marsh and the Eems–Dollard Estuary, The Netherlands. *Estuaries and Coastal Shelf Science*, 19(2), 143–165. https://doi.org/10.1016/0302-3524(77)90017-6

Dyer, K. R. (1974). The salt balance in stratified estuaries. *Estuaries and Coastal Marine Science*, 2(3), 273–281. https://doi.org/10.1016/0302-3524(74)90017-6

French, J. R., Burningham, H., & Benson, T. (2008). Tidal and meteorological forcing of suspended sediment flux in a muddy mesotidal estuary. *Estuaries and Coasts*, 31(5), 845–859. https://doi.org/10.1007/s12237-008-9072-5

Ganju, N. K., Defne, Z., Kirwan, M. L., Fagherazzi, S., D’Alpaos, A., & Carniello, L. (2017). Spatially integrative metrics reveal hidden vulnerability of microtidal salt marshes. *Nature Communications*, 8, 14156. https://doi.org/10.1038/ncomms14156

Ganju, N. K., Dickhudt, P. J., Montgomery, E. T., & Brosnahan, S. M. (2016). Oceanographic and water quality measurements in two Southern California coastal wetlands, 2013-2014. https://doi.org/10.5066/P7805OFZ

Ganju, N. K., Kirwan, M. L., Dickhudt, P. J., Guntenegspergen, G. R., Calhoon, D. R., & Kroeger, K. D. (2015). Sediment transport-based metrics of wetland stability. *Geophysical Research Letters*, 42, 7992–8000. https://doi.org/10.1002/2015GL065980

Ganju, N. K., Nizzielko, N. J., & Kirwan, M. L. (2013). Inferring tidal wetland stability from channel sediment fluxes: Observations and a conceptual model. *Journal of Geophysical Research: Earth Surface*, 118, 2045–2058. https://doi.org/10.1002/jgrf.20143

Gedan, K. B., Kirwan, M. L., Wolanski, E., Barbier, E. B., & Silliman, B. R. (2011). The present and future role of coastal wetland vegetation in protecting shorelines: Answering recent challenges to the paradigm. *Climatic Change*, 106(1), 7–29. https://doi.org/10.1007/s10584-010-0003-7

Kirwan, M. L., & Murray, A. B. (2007). A coupled geomorphic and ecological model of tidal marsh evolution. *Proceedings of the National Academy of Sciences*, 104(15), 6118–6122. https://doi.org/10.1073/pnas.0700958104

Lane, A., Prandle, D., Harrison, A. J., Jones, P. D., & Jarvis, C. J. (1997). Measuring fluxes in tidal estuaries: Sensitivity to instrumentation and associated data analyses. *Estuaries and Coastal Shelf Science*, 45(4), 433–451. https://doi.org/10.1006/ecss.1996.0220

Mariotti, G., & Fagherazzi, S. (2010). A numerical model for the coupled long-term evolution of salt marshes and tidal flats. *Journal of Geophysical Research*, 115, F01004. https://doi.org/10.1029/2009JF001326

Montgomery, E. T., Ganju, N. K., Dickhudt, P. J., Borden, J., Martini, M. A., & Brosnahan, S. M. (2015). Oceanographic and water-quality measurements in Rachel Carson National Wildlife Refuge, Wells, Maine, 2013. https://doi.org/10.5066/F7ST7MWS

Nowacki, D. J., Suttles, S. E., Ganju, N. K., Montgomery, E. T., & Martini, M. A. (2018). Oceanographic and water quality measurements collected in Grand Bay, Alabama/Mississippi, August 2016–January 2017. https://doi.org/10.5066/F9UG9F9Q

Nyman, J. A., Walters, R. J., Delaune, R. D., & Patrick, W. H. (2006). Marsh vertical accretion via vegetative growth. *Estuarine, Coastal and Shelf Science*, 69(3-4), 370–380. https://doi.org/10.1016/j.ecss.2006.05.041

Redfield, A. C. (1972). Development of a New England salt marsh. *Ecological Monographs*, 42(2), 201–237.

Rosencranz, J. A., Ganju, N. K., Ambrose, R. F., Brosnahan, S. M., Dickhudt, P. J., Guntenegspergen, G. R., et al. (2016). Balanced sediment fluxes in Southern California’s Mediterranean-climate zone salt marshes. *Estuaries and Coasts*, 39(4), 1035–1049. https://doi.org/10.1007/s12237-015-0056-y

Schile, L. M., Callaway, J. C., Morris, J. T., Stralberg, D., Thomas Parker, V., & Kelly, M. (2014). Modeling tidal marsh distribution with sea-level rise: Evaluating the role of vegetation, sediment, and upland habitat in marsh resiliency. *PLoS ONE*, 9(2), e87670. https://doi.org/10.1371/journal.pone.0088760

Settlemyre, J. L., & Gardner, L. R. (1977). Suspended sediment flux through a salt marsh drainage basin. *Estuarine and Coastal Marine Science*, 5(3), 653–663. https://doi.org/10.1016/0302-3524(77)90090-1

Simpson, M. R., & Bland, R. (2000). Methods for accurate estimation of net discharge in a tidal channel. *IEEE Journal of Oceanic Engineering*, 25(4), 437–445. http://ieeexplore.ieee.org/xpls/abs_all.jsp?arnumber=895351

Stevenson, J. C., Kearney, M. S., & Pendleton, E. C. (1985). Sedimentation and erosion in a Chesapeake Bay brackish marsh system. *Marine Geology*, 67(1-4), 215–235. https://doi.org/10.1016/0025-3227(85)90093-3

Suk, N. S., Guo, Q., & Puxty, N. P. (1999). Suspended solids flux between salt marsh and adjacent bay: A long-term continuous measurement. *Estuarine, Coastal and Shelf Science*, 49, 61–81. https://doi.org/10.1006/ecss.1999.0486

Suttles, S. E., Ganju, N. K., Montgomery, E. T., Dickhudt, P. J., Borden, J., Martini, M. A., & Brosnahan, S. M. (2015). Oceanographic and water-quality measurements in Barnegat Bay, New Jersey, 2014-2015. https://doi.org/10.5066/P7CH712Z

Swanson, K. M., Drexl, J. Z., Schoellhamer, D. H., Thorne, K. M., Casaza, M. L., Overton, C. T., et al. (2014). Wetland Accretion Rate Model of Ecosystem Resilience (WARMER) and its application to habitat sustainability for endangered species in the San Francisco Estuary. *Estuaries and Coasts*, 37(2), 476–492. https://doi.org/10.1007/s12237-013-9694-0

Townend, I., & Whitehead, P. (2003). A preliminary net sediment budget for the Humber Estuary. *Science of the Total Environment*, 314–316(January), 755–767. https://doi.org/10.1016/S0048-9697(03)00082-2

Ward, L. G. (1981). Suspended-material transport in marsh tidal channels, Kiawah Island, South Carolina. *Marine Geology*, 40(1-2), 139–154. https://doi.org/10.1016/0025-3227(81)90047-5