The Hydrodynamics of Surface Tidal Flow Exchange in Saltmarshes

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1 *Keywords:* tidal marsh, estuarine dynamics, shallow water tides, intertidal environment, flow
2 through vegetation, USA, Georgia, Savannah, Rose Dhu Island
3

Abstract

Modeling studies of estuary circulation show great sensitivity to the water exchange into and out of adjacent marshes, yet there is significant uncertainty in resolving the processes governing marsh surface flow. The objective of this study is to measure the estuary channel-to-saltmarsh pressure gradient and to guide parameterization for how it affects the surface flow in the high marsh. Current meters and high-resolution pressure transducers were deployed along a transect perpendicular to the nearby Little Ogeechee River in a saltmarsh adjacent to Rose Dhu Island near Savannah, Georgia, USA. The vertical elevations of the transducers were surveyed with static GPS to yield high accuracy water surface elevation data. It is found that water level differences between the Little Ogeechee River and neighboring saltmarsh are up to 15 *cm* and pressure gradients are up to 0.0017 *m* of water surface elevation change per *m* of linear distance during rising and falling tides. The resulting Little-Ogeechee-River-to-saltmarsh pressure gradient substantially affects tidal velocities at all current meter locations. At the velocity measurement station located closest to the Little Ogeechee River bank, the tidal velocity is nearly perpendicular to the bank. At this location, surface flow is effectively modeled as a balance between the pressure gradient force and the drag force due to marsh vegetation and bottom stress using the Darcy-Weisbach/Lindner's equations developed for flow-through-vegetation analysis in open channel flow. The study thus provides a direct connection between the pressure gradient and surface flow velocity in the high marsh, thereby overcoming a long-standing barrier in directly relating flow-through-saltmarsh studies to flow-through-vegetation studies in the open channel flow literature.

1. Introduction

Saltmarshes are valuable and productive ecosystems that serve as storm buffers, fish nurseries, and nutrient sources, and they play a significant role in the dynamics of estuary circulation. The limited amount of precise field data on the spatial variation of water surface elevation in estuarine creeks, channels, and in particular high marshes, is a critical gap in the understanding of saltmarsh hydrodynamics. The lack of data negatively impacts our ability to assess the effect of vegetation on surface water flow and accurately estimate the bottom stress in these settings. For instance, modeling results show that a channel-to-saltmarsh pressure gradient (i.e., differential water levels between the estuary main channel and the adjacent saltmarsh) is responsible for surface transport of water between the saltmarsh and the channel at high spring tide (Bruder et al., 2014). However, the modeling results are based on non-validated assumptions and the resulting pressure-gradient-driven surface flows show great sensitivity to the parameterization of the bottom stress, in agreement with earlier modeling results (Kjerfve et al., 1991). Therefore, detailed measurements of the pressure gradient and corresponding surface flows, together with application of appropriate hydrodynamic models, are needed to improve our understanding of the flow into and out of the high marsh.

In an effort to determine the effect of vegetation on the transition from channelized flow to sheet flow on the marsh platform, Vandenbruwaene et al. (2015) directly measured the velocity of the surface flow and estimated the water surface elevation in a high marsh. Their study is the only published field measurements we are aware of that show the presence of differential water levels between the high marsh and the adjacent main channel in a saltmarsh.

1 No study to our knowledge has attempted to quantitatively relate the corresponding pressure
2 gradient to the sheet flow through the high marsh vegetation.

3 Analyses of the effects of variable marsh friction (i.e., variable marsh vegetation) on the
4 mean flow in tidal creeks and deeper estuary channels are predominantly restricted to
5 modeling studies (Kjerfve et al., 1991; Rinaldo et al., 1999a; Rinaldo et al., 1999b; Bruder et al.,
6 2014). Attempts to quantify water surface elevations on high marsh platforms (and thus
7 quantify the pressure gradient between the high marsh and the creeks and channels) are also
8 chiefly limited to modeling studies (Kjerfve et al., 1991; Rinaldo et al., 1999a; Rinaldo et al.,
9 1999b). Unfortunately, due to the paucity of available field data, these models were unable to
10 validate their results in the high marsh. The principal gaps in current knowledge are the lack of
11 available water surface elevation measurements, the lack of directly observed flow patterns
12 (although see Temmerman et al., 2012; Vandenbruwaene et al., 2015), and an inability to verify
13 the model's parameterization of the bed/vegetation friction (e.g., through Manning's n or drag-
14 coefficient formulations; Kjerfve et al., 1991; Huang et al., 2008; Bruder et al., 2014). We note
15 that several recent studies have focused on the dissipation of wave energy by the marsh
16 vegetation rather than addressing the relationship between the mean surface flow and
17 bed/vegetation friction (Lowe et al., 2005; Augustin et al., 2009; Riffe et al., 2011; Wu, 2014).
18 Thus, the objective of this study is to determine the estuary channel-to-saltmarsh pressure
19 gradient and to evaluate how it affects the surface flow in the high marsh.

20 **1.1 Background**

21 Studies on water movement in saltmarshes generally have focused on one of three
22 areas: the surface flow in the network of branching tidal creeks (e.g., Bayliss-Smith et al., 1979;

French and Stoddart, 1992; Allen, 1994), the surface flow high in the marsh platform (Eiser and Kjerfve, 1986; Leonard and Luther, 1995), and the groundwater flow (Howes and Goehring, 1994; Gardner et al., 2002; Gardner, 2005; Wilson and Gardner, 2006). The flow in tidal creeks depends on the water surface elevation relative to the marsh bathymetry. For low relative water surface elevation, the creeks remain below bankfull and the flow velocity is small, $\sim 0.1 - 0.2 \text{ m/s}$ (Bayliss-Smith et al., 1979; Pethick, 1980; Healey et al., 1981; Dankers et al., 1984; French and Stoddart, 1992; Leopold et al., 1993; Allen, 1994; Pringle, 1995). The water surface elevation (i.e., pressure gradient) slopes into the tidal creek during flood and towards the ocean during ebb (Leopold et al., 1993). For high relative water surface elevation, the flow velocity in the creeks remains small while the creek is below bankfull and peaks at much larger values ($\sim 1 \text{ m/s}$) once the marsh is flooded (Bayliss-Smith et al., 1979; Pethick, 1980; Healey et al., 1981; Dankers et al., 1984; French and Stoddart, 1992; Allen, 1994; Pringle, 1995; Rinaldo et al., 1999a; Rinaldo et al., 1999b). Some studies observed “pulses” in the tidal creek velocity, which correspond to sudden changes in velocity as the marsh is flooded (e.g., Temmerman et al., 2005b; Torres and Styles, 2007). Similar pulses also have been observed as the creek itself initially flooded or became dry (Hazelden and Boorman, 1999). Consistent with low relative water surface elevation conditions, the pressure gradient in high relative water surface elevation conditions slopes into the tidal creek during flood and towards the ocean during ebb (Healey et al., 1981; French and Stoddart, 1992).

Regarding the hydrodynamics on the high marsh platform, Allen (2000) writes, “the local hydraulics of channels has undoubtedly been over-emphasized at the expense of what are in effect tidal floodplains.” Allen (2000) acknowledges that studies considering flow on the high

marsh platform face formidable challenges due to the difficulty in accurately measuring small gradients in the water surface elevation (Horstman et al., 2013) and the difficulty in making direct flow measurements in dense vegetation (Mazda et al., 2007). Flow velocities across the high marsh surface are much smaller than in the creeks ($\sim 0.01 - 0.1 \text{ m/s}$) and decrease in proximity of bathymetric obstacles (Eiser and Kjerfve, 1986; Wang et al., 1993; Allen, 1994). The flow is generally governed by the bathymetry of the high marsh (Eiser and Kjerfve, 1986; Davidson-Arnott et al., 2002; Temmerman et al., 2005a), the pressure gradient from the slope in the water surface elevation (Kjerfve et al., 1991; Bruder et al., 2014), and the vegetation characteristics (Leonard and Luther, 1995; Temmerman et al., 2012). At low relative water surface elevation, the flow essentially follows the high marsh bathymetry (Eiser and Kjerfve, 1986; Davidson-Arnott et al., 2002). As the relative water surface elevation becomes higher, currents in the high marsh begin to behave more analogously to sheet flow (Temmerman et al., 2005b; Vandenbruwaene et al., 2015) and are effectively forced by the water surface elevation slope between neighboring positions (Kjerfve et al., 1991; Bruder et al., 2014; Vandenbruwaene et al., 2015). This interplay between bathymetry-driven flow and sheet flow is also observed in estuaries dominated by mangrove swamps (e.g., Aucan and Ridd, 2000; Mazda et al., 2005; Horstman et al., 2013). In all cases, the flow is mediated by the characteristics of the vegetation and the relative water surface elevation (Leonard and Luther, 1995; Temmerman et al., 2012; Vandenbruwaene et al., 2015). When vegetation is partially submerged, the mean velocity depends on the plant morphology and density (Leonard and Luther, 1995). For deeply submerged vegetation, the flow is characterized by a two-layer velocity profile (Leonard and Luther, 1995). The bottom layer extends from the bed to the approximate height of the

1 vegetation stems, and the flow characteristics are similar to the partially-submerged case. The
2 upper layer is a logarithmic-law turbulent boundary layer that extends from essentially the top
3 of the vegetation to the free surface (Leonard and Luther, 1995).

4 Groundwater flow rates in marshes are typically substantially smaller than flow rates on
5 the surface (Wolanski and Elliott, 2015). However, the authors noted that groundwater flow is
6 a critical area of study because it influences the soil properties and fluid salinity. Salinity is
7 governed by a combination of marsh soil porosity, upland groundwater level, and tidal
8 inundation (Wolanski and Elliott, 2015). The water table and salinity respond rapidly to
9 precipitation and tidal inundation, as groundwater moves rapidly through saltmarshes (Gardner
10 et al., 2002; Gardner, 2005; Wolanski and Elliott, 2015). It is important to note that rapid
11 changes occur in response to the tide in areas that are not inundated due to tidal effects on
12 subsurface pore pressure gradients (Gardner et al., 2002; Gardner, 2005; Wilson and Gardner,
13 2006). Model results suggest that the majority of the direct groundwater interaction with
14 seawater occurs during the recharge and drainage of groundwater along the banks of tidal
15 creeks that branch into the marsh (Gardner 2005; Wilson and Gardner, 2006).

17 **2. Materials and Methods**

18 A field experiment was conducted in the tidal marsh adjacent to Rose Dhu Island near
19 Savannah, Georgia, USA (Fig. 1) from November 2nd to November 6th, 2014, coinciding with the
20 largest spring tide in November 2014. The project site was selected due to the availability of
21 numerical model data of water levels (Bomminayuni et al., 2012; Bruder et al., 2014) and
22 previously-collected bathymetric/topographic and vegetation data.

Three Onset HOBO pressure transducers (PTs), two Acoustic Doppler Velocimeters (ADVs, Nortek Vector – Nortek AS, Rud, Norway) and a current profiler (Nortek Aquadopp HR-profiler) were deployed along a transect perpendicular to the Little Ogeechee River in the tidal marsh. The ADVs measure the fluid velocity at essentially a point in the flow, whereas the Aquadopp measures the flow velocity in a series of evenly-spaced vertical bins. Instrument locations are shown in Fig. 1. The PTs were housed in custom-made PVC stands (Fig. 2). The stand's aluminum plate prevented instrument settling during the deployment cycle and the bolt/dive-weight assembly fixed the vertical elevation of the PTs within the stand. The three PTs were surveyed daily with static GPS for at least two hours. Additionally, a GPS base station, located on the nearby fixed dock, was surveyed daily for at least eight hours to improve GPS vertical elevation accuracy during post-processing. The GPS antennae were Ashtech Dual Frequency Marine Antenna (Ashtech S.A.S., Carquefou, France) that were connected to either Ashtech ProFlex 500 GS or Z-Surveyor receivers and were programmed to record internally at 10 sec intervals. Further, the ADV and Aquadopp locations were surveyed daily using a total station. The water salinity was measured near the surface from the dock (Fig. 1) at 4:15 PM on November 3rd, 2014 and 11:35 AM on November 6th, 2014 and was 25.7 and 26.3, respectively.

Fig.1: Map showing the locations of the six stations (courtesy of Google Earth). The instrument transect is shown as a dashed black line. Rose Dhu Island is outlined in red in the upper left image and shown as a yellow triangle on a map of the southeastern United States in the bottom right image. The lower left panel is the bathymetry (or topography, if exposed) along the transect in NAVD88. Green squares denote data points and the dashed gray line is the approximate bathymetry in the data gap. For reference, GPS coordinates for station D are Lat: $31^{\circ}55'11.62''\text{N}$ Long: $81^{\circ}8'20.99''\text{W}$. The highest and lowest high tides observed at station A during the experimental period are 1.40 m and 1.24 m (in NAVD88), respectively. Imagery data from November 20th, 2014.

Field data collected at this site also included *Spartina alterniflora* vegetation surveys at 25 locations (1 m x 1 m quadrats) across the marsh (within 0.6 km of the transect) conducted in December 2010. The data consisted of the number of stalks in each quadrat and the diameter of at least ten randomly-selected stalks. Visual comparison of the 2014 experiment and 2010 survey periods at ground level and via Google Earth images (20 November 2014 vs. 28 January 2011) indicate little change in the gross vegetation patterns.

2.1 Pressure to Water Surface Elevation

The PTs measured absolute pressure and temperature every 30 sec. To convert the absolute pressure to gauge pressure, atmospheric pressure readings from the nearby (~ 13 km away) weather station at Skidaway Institute of Oceanography were employed. Gauge pressure was converted to water depth above the PT by assuming the pressure variation in the vertical direction was solely due to hydrostatic effects. The water density was determined from the PT temperature measurement and the salinity measured at the dock, which was assumed to be equal to the salinity in the vicinity of the PTs. Temperature and salinity were assumed to be vertically homogeneous.

To improve the accuracy of the absolute pressure readings (and thus the estimated water surface elevation), the three PTs were calibrated in a 1 m deep tank at seven temperatures, ranging from 12 to 26 °C. Linear calibration curves were developed to calculate the pressure difference between two of the PTs (stations A and D) and the third reference PT (station F) as a function of temperature. These curves were used during data processing to adjust the pressure data to a common reference pressure. The standard deviations of the mean pressure difference between the PTs at stations A and D and the reference PT at station F

1 were calculated for each calibration temperature. The difference was found to be less than 10
2 Pa in all cases, which corresponded to less than 1 mm of water surface elevation difference.
3 Note that the ADVs and the Aquadopp are equipped with built-in PTs. Unfortunately, the
4 accuracy of the built-in PTs is one order-of-magnitude worse than the accuracy of the Onset
5 HOBO PTs and is not sufficient to resolve sub- cm water level differences.

6 The water surface elevation was calculated relative to the NAVD88 vertical datum by
7 adding the water depth data to the PT elevation using the dimensions of the stand (shown in
8 Fig. 2) and the GPS measurement of the antenna position. To arrive at sufficiently accurate
9 water surface elevation measurements, the GPS measurements were post-processed using
10 GrafNet (NovAtel, Calgary, Alberta, Canada), treating the GPS system as a stationary closed loop
11 network. The system included four GPS antenna/receivers (one at each PT and one on the
12 dock) and several Continuously Operating Reference Stations (CORS) operated by the U.S.
13 National Geodetic Survey (NGS). The estimated vertical accuracy from GrafNet was
14 approximately 0.5 cm for all stations. This level of accuracy was obtained using simultaneous
15 GPS measurements. The vertical positions of the PT stands did not vary during the
16 experimental periods.

Fig. 2: (a) Diagram of a pressure transducer stand, (b) in situ photograph of a pressure transducer stand during the GPS survey, and (c) in situ photograph of an ADV during deployment.

2.2 ADV Data Processing

The ADV measurements have a resolution of 1 mm/s and were collected continuously at 1 Hz. Data collected during low tide (determined by instances of the PT at station D being dry) were discarded. Following the recommendations of Chanson et al. (2008) and Wilson et al. (2013), data for which the average correlation coefficient of the three beams was less than 70% for 45 consecutive samples, or for which the average over 300 samples of the mean correlation coefficient of the three beams dropped below 70%, were discarded. This filtering predominately removed data collected while the instrument was not submerged, and less than 2% of the submerged instrument data were removed due to filtering. The dominant flow direction (if one was present) was determined for each ADV. An ellipse was fit to a scatter plot of the measured east and north velocity components for each ADV, and the major axis of the resulting ellipse is defined as the “along-stream” flow direction. Correspondingly, the minor axis is defined as the “cross-stream” direction. Fig. 3 illustrates this process for the ADV at station E.

Fig. 3: Scatter plot of the north and east velocity components at station E. Symbols indicate the measured data, and the ellipse is the best fit to the data. There are 24,100 velocity samples included in the plot.

2.3 Aquadopp Data Processing

Aquadopp measurements were collected at 1 Hz. The Aquadopp returned the velocity within 10 cm vertical bins, equally spaced from 0.3 to 2.2 m above the instrument. Dry bins were identified using the built-in PT and temperature gauges to estimate the water depth above the instrument. All bins in which the water depth had not reached at least the midpoint of the next highest bin were discarded. The velocity measurements in each bin were averaged over 30 sec intervals to smooth out turbulent velocity fluctuations. The flux-per-unit-width (q) was estimated by summing the velocity in each bin multiplied by the bin height. Flux-per-unit-width therefore corresponded to the fluid flow through a vertical line extending from the Aquadopp to the surface. The ellipse fitting technique (Section 2.2) combined with polar histograms of the flux vectors were used to determine the along-stream flow direction in the Crooked Creek tributary. The horizontal flux vector was rotated to correspond to the defined along-stream and cross-stream directions.

3. Results and Discussion

Tidal flow in and out of the saltmarsh is more complex than the typical behavior of estuarine channels due to the highly variable bathymetry and vegetation across the marsh. Further, the comparatively strong Little-Ogeechee-River-to-saltmarsh pressure gradient significantly affects the marsh flow characteristics, as presented below. A typical tidal cycle in this saltmarsh proceeds as follows (illustrated in Fig. 4, which shows the relative water surface elevation at five instances in the tidal cycle):

(1) Early Flood: Starting from low tide, the water begins to rise as the tide comes in. At the earliest stages of rising tide, the water in the Crooked Creek tributary and the nearby Little Ogeechee River rise at roughly the same rate. The water progresses into the marsh along the Crooked Creek tributary and the flux-per-unit-width (at station B – see Fig. 1) begins to increase towards a maximum value as the water rises (see Fig. 5). The ADV at station C becomes inundated after the Aquadopp (at station B) due to the station's higher elevation. Water at station C initially flows into the marsh towards the northwest along a small non-vegetated depression (see Fig. 1).

(2) Mid Flood: The water level begins to rise more rapidly in Little Ogeechee River than in the Crooked Creek tributary. As the water level in Little Ogeechee River rises above its banks, the high marsh floods from Little Ogeechee River towards the Crooked Creek tributary. The flow of water from Little Ogeechee River briefly (i.e., 20 to 30 minutes) forces the flow at station C out of alignment with the small northwesterly-oriented depression and shifts the flow direction towards the northeast along the instrument transect (i.e., towards the Crooked Creek tributary). Station E becomes inundated and the initial velocity at the instrument is similarly

1 towards the northeast along the instrument transect. The flux-per-unit-width at station B,
2 which has been increasing with the rising tide, begins to slow towards a local minimum value
3 (Fig. 5) as the water flooding into the marsh from Little Ogeechee River briefly restricts the
4 water flooding along the Crooked Creek tributary.

5 (3) Late Flood: Before high tide is reached, the water level in the Crooked Creek
6 tributary (station A) catches up to and exceeds the water level in Little Ogeechee River (station
7 F). Simultaneously, the flux-per-unit-width in the Crooked Creek tributary increases towards a
8 local maximum value (smaller than the previous maximum described in (1) Early Flood, Fig. 5),
9 and the velocity at station E switches direction, flowing to the southwest (towards Little
10 Ogeechee River) along the instrument transect. The velocity at station C returns to the
11 direction aligned with the depression towards the northwest.

12 (4) Early Ebb: The water level in Little Ogeechee River begins to drop slightly before the
13 water level in the Crooked Creek tributary. The water level in Little Ogeechee River drops
14 substantially faster, such that the water level in the Crooked Creek tributary remains higher for
15 the remainder of the tidal cycle. The water level in the marsh becomes sufficiently higher than
16 that in Little Ogeechee River to once again temporarily force the flow at station C out of
17 alignment with the small depression. The flow at station C shifts towards the southwest along
18 the instrument transect (towards Little Ogeechee River). The flow at station E remains strongly
19 southwestern towards Little Ogeechee River until the ADV becomes dry. Concurrently, the flow
20 in the Crooked Creek tributary slows and switches to ebb – upon which it follows a more
21 standard tidal stage-flux pattern (Fig. 5).

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- 1 (5) Mid-Late Ebb: As the water level in Little Ogeechee River drops below its banks, the
- 2 flow at station C returns to alignment along the small depression, flowing towards the
- 3 southeast.
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Fig. 4: Illustration of the tidal flow sequence. The blue shade indicates the relative water surface elevation at the position along the transect for each stage. The bathymetry (or topography) along the transect is shown as a dashed grey line. The compass roses with velocity arrows indicate the velocity direction and magnitude at the respective instruments (lack of an arrow indicates the instrument was not submerged during that stage).

Fig. 5: Tidal stage (NAVD88) – along-stream-flux-per-unit-width (q) diagram at station B for a November 5th, 2014 tidal cycle. Positive values of flux-per-unit-width denote flood (inundation) and negative values represent ebb (draining). The color transition indicates time progression from the beginning (green) to end (blue) of the tidal cycle. The dashed grey line marks the elevation of the bank of the Little Ogeechee River (1.1 m).

3.1 Water Levels and Pressure Gradients

As the narrative of the tidal cycle sequence made clear (illustrated in Fig. 4), the differential water level between the high marsh and Little Ogeechee River strongly affects the flow characteristics at all instrument locations. To quantify the differential water levels (and corresponding pressure gradients), the water surface elevations at each PT are shown in Fig. 6 for a representative tidal cycle.

Fig. 6: Water surface elevation (NAVD88) for the tidal cycle shown in Fig. 5 for each pressure transducer (PT). Flat segments indicate dry periods for the instrument. If the pressure gradient is the dominant forcing term, then water flows from the location with higher water surface elevation to the location with lower water surface elevation, as indicated by the black arrows.

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7 2 The water levels at the two PTs in the marsh (stations A and D) are nearly identical over
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10 3 much of the tidal cycle, but the data also reveal some differences. The water level in Crooked
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12 4 Creek tributary (station A) rises slightly faster and falls more quickly than the water level in the
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15 5 high marsh (station D), which is to be expected from the reduced tidal wave propagation speed
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17 6 in the high marsh (Parker 1991).

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20 7 In contrast, the water level in Little Ogeechee River (station F) is substantially different
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22 8 than the water level at either of the two marsh PT stations. The water level difference between
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25 9 Little Ogeechee River and the marsh is approximately 7 – 8 *cm* at high tide and becomes as
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28 10 large as 15 *cm* during rising and falling tide. The water level difference is similar to that
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30 11 observed in Vandenbruwaene et al. (2015) during flood tide. Given that the distance between
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33 12 stations D and F is less than 90 *m*, the observed water surface elevations translate to a
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36 13 substantial Little-Ogeechee-River-to-saltmarsh pressure gradient.

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38 14 Another interesting feature in Fig. 6 is that the water level in the Crooked Creek
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41 15 tributary and Little Ogeechee River (stations A and F) are very similar at the beginning of flood
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43 16 tide prior to the inundation of the high marsh. Once the water level in the Crooked Creek
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46 17 tributary (station A) exceeds the level of the bank, approximately 0.4 *m* (Fig. 1), its rate of rise
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49 18 decreases as the marsh is flooded until it matches the water level in the high marsh (station D).
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51 19 During ebb tide, the water surface elevation in the Crooked Creek tributary (Station A) and the
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54 20 high marsh (Station D) initially decrease at the same rate, then the water level in the Crooked
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56 21 Creek tributary (Station A) starts to decrease more rapidly. The water surface elevation in the
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1 Crooked Creek tributary (station A) is greater than that in the Little Ogeechee River (station F)
2 throughout the ebb tide, which indicates the tributary drains at a slower rate.

3 ***3.2 Effects of the Little-Ogeechee-River-to-Saltmarsh Pressure Gradient on Flow***

4 The effect of the Little-Ogeechee-River-to-saltmarsh pressure gradient on the flow is
5 more subtle at stations B and C than at station E (see Fig. 1), largely due to the bathymetry. The
6 effect of the Little-Ogeechee-River-to-saltmarsh pressure gradient at station B is apparent in
7 the stage versus along-stream flux-per-unit-width (q) diagram shown in Fig. 5. Typical estuary
8 channel stage-flux (or velocity) diagrams appear as a tilted ellipse (with the angle depending on
9 the standing versus progressive nature of the tidal wave). Therefore, a typical flow is
10 characterized by a single maximum value of flux during flood and a single maximum value of
11 flux during ebb. Fig. 5 shows an obvious “pulse” in the flood stage, where the flux-per-unit-
12 width reaches a maximum value, then temporarily decreases and subsequently increases to a
13 local maximum value before returning to a more standard diagram shape. As described in the
14 tidal sequence, this brief slowing and local minimum value of flux-per-unit-width coincided with
15 flow of water from Little Ogeechee River towards the Crooked Creek tributary due to the higher
16 water level (in Little Ogeechee River) overtopping the Little Ogeechee River bank. This flow of
17 water essentially retards the flux of water up the Crooked Creek tributary until the water level
18 in the tributary rises above the water level in Little Ogeechee River and the tributary flux begins
19 to increase again. Torres and Styles (2007) observed similar stage diagrams for a tidal creek
20 near Winyah Bay (Georgetown, South Carolina, USA), in which marsh inundation was strong
21 enough to temporarily reverse the tidal creek flow, thus producing a tidal stage diagram in the
22 shape of a figure eight.

1 The flow at station C appears to be predominately confined to a natural mud-flat
2 depression (see Fig. 1) roughly parallel to Little Ogeechee River. However, during both rising
3 and falling tide, the Little-Ogeechee-River-to-saltmarsh pressure gradient becomes sufficiently
4 large to redirect the velocity at station C to be perpendicular to the alignment of the natural
5 depression. The flow direction in these instances is aligned with the instrument transect
6 towards either the Crooked Creek tributary (if the Little Ogeechee River water level is higher) or
7 Little Ogeechee River (if the marsh water level is higher).

8 The flow at station E is obviously affected by the pressure gradient between stations D
9 and F. The angle between the along-stream flow direction (i.e., major axis of the best fit ellipse
10 in Fig. 3) and the instrument transect is less than 9° , which indicates the adjacent PTs were
11 well-placed to capture the pressure gradient driving the flow at this location. Fig. 7 plots the
12 along-stream velocity, averaged over 30 sec intervals, at station E against the corresponding
13 pressure gradient (i.e., the water surface elevation differential between the PT at station D and
14 the PT at station F divided by the distance between stations) for all measured tidal cycles.
15 Along-stream velocity is defined as positive to the northwest, roughly along the PT transect,
16 and positive pressure gradient is defined as higher water level in the high marsh (PT at station
17 D) than in the Little Ogeechee River (PT at station F). Because of the direct correlation of the
18 sign and magnitude of the surface flow and pressure gradient, Fig. 7 strongly indicates the
19 pressure gradient between stations D and F governs the flow characteristics at this location,
20 regardless of tidal cycle or water level.

Fig. 7: Along-stream velocity at station E (red stars) and depth-based Reynolds number (Re_h - green pluses) plotted as a function of the corresponding pressure gradient (m of water surface elevation drop per m of linear distance). The blue line represents the model for $d_{stalk} = 1 \text{ cm}$ and $N_{stalk} = 93$. 96% of the variance in the velocity parameter is explained by the model.

3.3 Predicting Velocity Given the Pressure Gradient

It is desirable to accurately predict the surface flow velocity given the slope of the water surface elevation by defining an equation (or set of equations) that relate these quantities. The data shown in Fig. 7 afford an excellent opportunity to explore this relationship since the flow at station E appears unconfined by bathymetry and thus is driven by the pressure gradient between stations D and F. A simple control volume analysis yields the following force balance (Järvelä, 2004):

$$\rho g A_b h \frac{\partial \zeta}{\partial x} = \frac{1}{2} \rho C_d A_p V^2 \quad (1)$$

The primary forcing is due to the pressure gradient (from the slope in the water surface elevation, $\frac{\partial \zeta}{\partial x}$, which accounts for the component of the fluid weight in the flow direction due to the bed slope and the net force on the control volume due to the variable water depth, i.e., the unequal hydrostatic pressure distributions). In the formulation on the left-hand-side of Eqn. 1, x is the coordinate direction along the instrument transect, ζ is the water surface elevation, h is the total water depth (i.e., $h = h_0 + \zeta$), h_0 is the still water depth, A_b is the bed area, and g is the acceleration due to gravity. The primary retarding force is the drag force due to the bed and vegetation (i.e., the right hand side of Eqn. 1). The drag force is modeled with a drag coefficient, C_d , formulation where ρ is the fluid density, V is the fluid velocity, and A_p is the projected (frontal) area of the vegetation (Munson et al. 2013). The assumptions leading to the force balance in Eqn. 1 include neglecting the local (temporal) acceleration (because the tidal variation occurs over comparatively large time scales, e.g., hours), neglecting the convective acceleration (because the velocity magnitude in Fig. 7 is small in comparison to the marsh

spatial scales, i.e., $V \sim 10 \text{ cm/s}$ versus spatial scales of $\sim 100 \text{ m}$), and neglecting the cross-stream flow (because the flow at station E is nearly aligned with the transect).

Much discussion in the flow-through-vegetation literature is devoted to the optimum parameterization of the drag coefficient and effective areas (A_b and A_p in Eqn. 1) under a variety of circumstances (e.g., Lindner, 1982; Koch and Ladd, 1997; Wu et al., 1999; Nepf, 1999; Tsihrintzis, 2001; Järvelä, 2002; Stone and Shen, 2002; Lee et al., 2004; Tanino and Nepf, 2008; Cheng and Nguyen, 2011). The literature is generally in agreement that the effective areas must depend on marsh plant morphology and density (e.g., the stalk diameter and the spacing between the stalks), but the authors have not come to a consensus on the form of the dependence. Additionally, the drag coefficient's dependence on Reynolds number is unresolved. Some studies exclude the Reynolds number dependence entirely (e.g., Lindner, 1982; Järvelä, 2002), whereas others include the drag coefficient as a function of the depth-based Reynolds number ($Re_h = \frac{vh}{\nu}$, e.g., Wu et al., 1999; Tsihrintzis, 2001; Stone and Shen, 2002; Lee et al., 2004) or the stalk-diameter-based Reynolds number ($Re_d = \frac{Vd_{stalk}}{\nu}$, e.g., Koch and Ladd, 1997; Nepf, 1999). The most recent studies incorporate the pore velocity between the cylinders/stalks as the relevant velocity scale (Tanino and Nepf, 2008) or the pore velocity as the velocity scale and the vegetation-related-hydraulic-radius ($r_v = \frac{\pi}{4} \frac{1-\lambda}{\lambda} d_{stalk}$, where λ is the stem density) as the relevant length scale (e.g., Cheng and Nguyen, 2011) in the formulation of the Reynolds number. It is also important to note that the two-layer velocity profile described for deeply submerged saltmarsh vegetation in Leonard and Luther (1995) has previously been shown to exist in atmospheric turbulent boundary layers above crops or forests

(Raupack et al., 1991; Kaimal and Finnigan, 1994), in flume studies of flow through submerged plants (e.g., Murota et al., 1984; Gambi et al., 1990), and in flume studies of flow through roughened cylinders (Stone and Shen, 2002). The vegetation in our study remained emergent in the vicinity of the ADV stations during spring tide; therefore we restrict our discussion to the partially submerged case.

The data in this study are best modeled by a Darcy-Weisbach friction factor (f) approach, making use of Lindner's equation (Lindner, 1982) to estimate f based on local vegetation-related parameters. This modeling approach has been successfully applied in numerical studies of river floodplains (Stoesser et al., 2003), as well as flume studies with natural vegetation (Järvelä, 2002). Originally intended for pipe flow, the Darcy-Weisbach equation (Eqn. 2) adopts a shear velocity, $u^* = \sqrt{\left(gh \frac{\partial \zeta}{\partial x}\right)}$, to account for the pressure gradient and parameterizes the velocity as follows:

$$\frac{V}{u^*} = \sqrt{\frac{8}{f}} \quad (2)$$

By comparing Eqns. 1 and 2, f can be written in terms of C_d as $f = 4C_d \frac{A_p}{A_b}$ (Lindner, 1982).

Lindner (1982) employs the plant stalk diameter (d_{stalk}) multiplied by the water depth (h) as a surrogate for projected vegetal area (A_p) and the longitudinal and lateral plant spacing (a_x and a_y , respectively) as a surrogate for bottom area (A_b):

$$f = 4C_d \frac{d_{stalk}h}{a_x a_y} \quad (3)$$

Flume experiments by Järvelä (2002) indicate that the drag coefficient falls in the range $1.43 \leq C_d \leq 1.55$ for Lindner's equation (Eqn. 3).

To apply the model, the pressure gradient in Eqns. 1 and 2 $\left(\frac{\partial \zeta}{\partial x}\right)$ is taken as the absolute value of the measured quantity, and the direction of the velocity is assumed to be opposite the sign of the measured pressure gradient. The water depth (h) is assumed to be equivalent to that at the nearby station D. The vegetation parameters (d_{stalk} , a_x , a_y) are determined from the vegetation survey of the grass within the quadrat located closest to station E (see Fig. 1): the average stalk diameter is $d_{stalk} = 1.0 \pm 0.07 \text{ cm}$ [mean $\pm$ standard error of the mean] and the number of stalks in the quadrat is $N_{stalk} = 93$. The parameters a_x and a_y are both assumed to be equal to the mean spacing (s) between individual stalks given by $s = \frac{\sqrt{A_s} - \sqrt{N_{stalk}d_{stalk}}}{\sqrt{N_{stalk}-1}}$ for a square vegetation survey area (A_s). The value of C_d that results in the best fit is 2.0, which is slightly larger than the value range reported by Järvelä (2002). This discrepancy is discussed in greater detail below.

3.4 Discussion of the Darcy-Weisbach/Lindner Equation Fit

The model results, employing Eqns. 2 and 3, are shown in Fig. 7. The model effectively predicts the along-stream velocity at station E for the majority of the dataset. The model performs particularly well near the change in flow direction (i.e., velocity = 0) as well as in the range of large negative velocity values (which forms the majority of the data). The model does not predict as well the apparent transition in slope at velocity values around -6 cm/s (and pressure gradients near $2 \times 10^{-4} \text{ m}$ of water surface elevation drop per m of linear distance).

When applying the Darcy-Weisbach equation to pipe flow as originally intended, the dependence of C_d (or more properly, f) on Reynolds number is given by the Moody diagram (Moody 1944). Lindner's formulation effectively assumes f is independent of depth-based

Reynolds number, analogous to the “fully rough” regime in pipe flow or the flow at stalk-based Reynolds numbers greater than 200 described in Nepf (1999), and thus may be entirely determined from vegetative properties. In contrast, some studies in the flow-through-vegetation literature (e.g., Wu et al., 1999; Tsihrintzis, 2001; Lee et al., 2004) parameterize C_d as a function of the depth-based Reynolds number ($Re_h = \frac{\rho V h}{\mu}$). Tsihrintzis (2001) in particular raises the possibility that the dependence on Re_h may change with Re_h regime, analogous to the regime dependence of the drag coefficient for flow past cylinders.

The apparent transition in slope at pressure gradients near $2 \times 10^{-4} \text{ m/m}$ in Fig. 7 raises the possibility of a Reynolds number transition. Thus, the corresponding depth-based Reynolds number is also plotted in Fig. 7, revealing a transition in the slope of the Re_h curve also near pressure gradient values of $2 \times 10^{-4} \text{ m/m}$ (a similar transition is observed for the stalk-diameter-based Reynolds number, although not shown in Fig. 7). The transition occurs at a much higher Reynolds number than that observed for the transition to unsteady flow described in Koch and Ladd (1997) or Nepf (1999) and could be associated with the velocity becoming sufficiently large such that the stem flexibility is no longer negligible (note that dissipation of wave energy by saltmarsh vegetation is affected by stem flexibility, Riffe et al., 2011). As a result, a two-equation approach was considered to model these data using the Tsihrintzis (2001) formulation for C_d ($C_d = \gamma Re_h^{-k}$, where γ and k are parameters determined via the procedures in Lee et al., 2004) for the low Re_h regime and the Darcy-Weisbach/Lindner approach for the high Re_h regime. This approach was ultimately discarded in favor of applying the Darcy-Weisbach/Lindner formulation to all data because γ and k must be determined from a multi-variable least-squares fit to an existing dataset and the resulting model performed only

marginally better in the low Re_h regime than the Darcy-Weisbach/Lindner model. Formulations incorporating the stalk-diameter-based Reynolds number (Tanino and Nepf, 2008) or the hydraulic radius-based Reynolds number (Cheng and Nguyen, 2011) were similarly discarded.

3.5 Parameter Sensitivity

It is beneficial to evaluate the sensitivity of the Darcy-Weisbach/Lindner formulation to the various input parameters (i.e., N_{stalk} , d_{stalk} , h , and the pressure gradient). The effect of the expected level of variability or uncertainty in each of the four input variables on the Darcy-Weisbach/Lindner model is shown in Fig. 8. The variability observed in N_{stalk} and d_{stalk} among the 25 locations of the vegetation survey quadrats provides insight to the expected range of values of these parameters. Among the data from these quadrats, $N_{stalk} = 80.4 \pm 18.1$ and $d_{stalk} = 0.77 \pm 0.13$ cm [mean $\pm$ standard deviation], hence one standard deviation corresponds to approximately 20% for each parameter. Therefore, the potential variability in N_{stalk} and d_{stalk} is taken to be up to 20% higher or lower than the representative values used in the original formulation (which are specific for the quadrat location nearest to station E). The potential uncertainty in the water depth and pressure gradient estimates is based on the constraints of the measurements. The uncertainty in the water depth is assumed to be ± 0.54 cm by combining the uncertainty in the PT water level measurements and the accuracy of the GPS survey elevation estimates. Similarly, combining the resolution of the physical measurements yields an uncertainty in the pressure gradient values of $\pm 8.9 \times 10^{-5}$ m of water surface elevation difference per m of linear distance.

Fig. 8: Sensitivity of Darcy-Weisbach/Lindner model to variation in (a) N_{stalk} , (b) d_{stalk} , (c) h , and (d) pressure gradient. The upper and lower bounds for each parameter are 20% higher/lower for N_{stalk} and d_{stalk} , $\pm 0.54 \text{ cm}$ for h , and $\pm 8.9 \times 10^{-5} \text{ m}$ of water surface elevation drop per m of linear distance for pressure gradient. Note that the curves are obscured by near perfect overlap in (c).

Fig. 8 (a) and (b) indicate that variability in N_{stalk} and d_{stalk} influences the estimated velocity magnitude for a given pressure gradient, which highlights the need to accurately estimate these parameters via vegetation surveys or other means. Fig. 8 (c) reveals that the net

effect of water depth variation is negligible, as expected based on Eqns. 2 and 3. Lastly, despite the comparatively small expected uncertainty in the pressure gradient (relative to the vegetation-related quantities), the effect on the estimated velocity magnitude can be quite significant, particularly near the predicted change in flow direction. At large values of the pressure gradient (regardless of sign), the impact on the predicted velocity magnitude is smaller because the velocity is proportional to the square root of the water surface elevation slope ($\sqrt{\frac{\partial \zeta}{\partial x}}$).

The results shown in Fig. 8 help explain the need to increase the value of C_d relative to the range recommended by Järvelä (2002) when modeling the flow velocity at station E. In addition to the uncertainties in the vegetation parameters already discussed, one must consider that the vegetation varies substantially along the transect between stations D and F (i.e., the PTs that bracket station E - see Fig. 1). It is possible that the use of representative value of N_{stalk} and d_{stalk} near the site of the flow measurements is insufficient to adequately capture the effects of the variation in N_{stalk} and d_{stalk} along the transect. Similarly, the reported values of the pressure gradient at station E are the best estimate of the slope of the water surface elevation based on the information available (i.e., the slope of the line connecting two neighboring water surface elevation locations at stations D and F). As shown in Fig. 8, even small errors in the pressure gradient estimates can have strong effects on the estimated velocity magnitude. Great care was taken in the current measurements to minimize the uncertainty in pressure gradient, and the results show that the change in sign of the calculated pressure gradient coincides very well with the change in velocity direction (see Fig. 7). However, it is possible that the water level measurements at stations D and F are capturing

1 the macro-scale pressure gradient switch properly, but overestimating the magnitude at station
2 E due to localized variation in water level between the two PTs from effects induced by the
3 variable bathymetry (Fig. 1). Additionally, although the saltmarsh groundwater flow rate is
4 typically much smaller than the surface flow rate (Wolanski and Elliott, 2015), this could be
5 another contributing factor in the increase in the drag coefficient, e.g., if the pressure gradient
6 is driving groundwater flow unaccounted for in our force balance. Further, groundwater
7 intrusion can affect the surface water salinity and temperature (Wolanski and Elliott, 2015),
8 introducing an error in our calculated water density and thus the water surface elevation
9 measurements. As a result, one can reasonably suggest that the need to increase C_d is an
10 artifact of the uncertainty in quantifying the vegetation and/or pressure gradient parameters or
11 potentially due to unaccounted groundwater flow.

4 Conclusions

The results of this study offer strong support for the existence of a Little-Ogeechee-River-to-saltmarsh pressure gradient, corroborating similar observations in the numerical study of Bruder et al. (2014) and field study of Vandenbruwaene et al. (2015). Fig. 6 clearly shows the existence of large differential water levels between Little Ogeechee River and the adjacent tidal marsh for the majority of the tidal cycle. Further, data from the current meters indicate that the resulting Little-Ogeechee-River-to-saltmarsh pressure gradient has a significant effect on the flow throughout the marsh.

At station E (on the high marsh near the bank of Little Ogeechee River) the flow characteristics are governed by the pressure gradient between the neighboring stations (Fig. 7). At this site, the flow velocity is quantitatively related to the pressure gradient between stations D and F via modeling techniques adapted from the open-channel-flow-through-vegetation literature. The model is based on a balance between the force of the pressure gradient and the drag force due to the vegetation and bottom stress. Specifically, the Darcy-Weisbach/Lindner equations effectively model the velocity based on specific vegetation-related parameters and the pressure gradient between stations D and F.

This study demonstrates the ability to quantify the water surface elevation measurements to sufficient accuracy to directly relate the resulting pressure gradient to the flow velocity in the high marsh. The study, therefore, overcomes a long-standing barrier in directly relating flow-through-saltmarsh studies to quantitative relationships developed in flumes for flow through vegetation. Many of these formulations, such as the Darcy-Weisbach/Lindner model, may prove superior for estimating the drag coefficient in the high

marsh (i.e., re-arranging Eqn. 3) than the current Finite Volume Coastal Ocean Model (FVCOM) technique of taking the greater of a user-entered drag coefficient or a logarithmic-boundary-layer-fit-estimated drag coefficient (Bruder et al., 2014). The Darcy-Weisbach/Lindner model may likewise prove superior to single-coefficient equations that assume a logarithmic-law velocity profile, such as Manning's equation (e.g., Wolanski and Elliott, 2015), which is known to be ineffective in modeling flow through emergent vegetation (Kadlec, 1990; Jadhav and Buchberger, 1995; Nepf, 1999). Estimating the drag coefficient via Eqn. 3 with spatial variability in the vegetation characteristics (and thus the drag coefficient) across the high marsh based on vegetation surveys and/or satellite image estimates of grass distributions will significantly enhance models of surface flows in the marsh.

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