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In situ measurements of shear stress, erosion and deposition in man-made tidal channels within a tidal saltmarsh

Aline Pieterse [a, *], Jack A. Puleo [b], Thomas E. McKenna [c], Jens Figlus [d]

[a] Department of Geological Sciences, University of Delaware, Newark, DE, United States
[b] Center for Applied Coastal Research, University of Delaware, Newark, DE, United States
[c] Delaware Geological Survey, University of Delaware, Newark, DE, United States
[d] Department of Ocean Engineering, Texas A&M University, Galveston, TX, United States

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ABSTRACT

A field study was conducted in man-made ditches in a tidal saltmarsh in Lewes, Delaware, USA. Ditches are prevalent throughout tidal marshes along the Atlantic US coast, and influence hydrodynamics and sediment transport. The field study focused on measuring near-bed velocity, shear stress, sediment concentration, and bed level variability at 5 stations over a 3-week period. Velocities in the ditch (2–5 m wide, 1 m deep) peaked between 0.4 and 0.6 m/s and were slightly ebb dominated. Velocity and shear stress were maximum during a storm event, with peak shear stresses of 2 N/m^2. Bed levels were estimated from acoustic amplitude return of a downward-looking velocity profiler. The bed level in the ditch at the landward locations increased ~ 0.03 m over 3 weeks, while there was ~ 0.01 m bed level decrease at the most seaward site suggesting a net import of sediment into the channel. At all sites, erosion (~ 0.005–0.015 m) occurred during the accelerating phase of the flood tide, and accretion of a similar magnitude occurred during the decelerating phase of the ebb tide. This erosion-deposition sequence resulted in small net changes in bed level at the end of each tidal cycle. The intratidal behavior of the bed level was simulated using erosion and deposition flux equations based on shear stress, critical shear stress, and suspended sediment concentration. Erosion was predicted well with RMS errors on the order of $2 \cdot 10^{-3}$ m. The bed level during the deposition phase could not be reproduced using the simple approach. Model inaccuracies for deposition were attributed to advection and variations in fall velocity due to flocculation that were not modeled due to lack of ground-truth observations.

1. Introduction

1.1. Marsh ditching

Humans have altered the tidal wetlands of the USA East coast since the arrival of European settlers in the 1600's (Philipp, 2005). A major episode of alteration was the ditching of salt marshes in the 1930's along the Atlantic coast of the USA (Gedan et al., 2009; Lesser, 2007), including the majority (90%) of marshes from New England to Virginia (Bourn and Cottam, 1950). Maintenance of the ditches continued until the 1960's and 1970's. Ditches were typically oriented in parallel or in a grid pattern at 30–100 m spacing (Tonjes, 2013). Marsh ditching was performed for a variety of reasons, but most commonly in an effort to control mosquito populations. Ditches were intended to drain water from the marsh platform, thereby reducing the amount of standing water available for mosquito breeding.

Ditches have a variety of effects on the marshes, ranging from lowering the water table (Singh and Nathan, 1965; Turner and Lewis, 1996), changing vegetation patterns and bird habitat (Adamowicz and Roman, 2005; Bourn and Cottam, 1950; Clarke et al., 1984; Daiber, 1986), and potentially retaining sediment (Corman et al., 2012; LeMay 2007). Previous research focused largely on the ecological impact of ditching marshes including enriched concentrations of inorganic nutrients and dissolved and particulate organic nitrogen and carbon (Koch and Gobler, 2009), and the effects mentioned earlier. Some studies have shown that the ditches provide a sink for sediment that could otherwise be transported onto the marsh platform (Corman et al., 2012; LeMay 2007), which could result in a lower marsh elevation. In addition,

* Corresponding author.
E-mail address: alinepieterse@gmail.com (A. Pieterse).

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the interior regions of ditched marshes typically flood first and remain flooded longer than marsh regions farther away from the ditched areas (LeMay 2007). Man-made levees along the ditches were created when the excavated material was placed adjacent to the ditch, thereby elevating these areas. These levees influence marsh inundation duration and patterns affecting sediment delivery across the marsh platform (Kennish, 2001). However, the alteration of hydrodynamics, turbidity, erosion, and deposition within ditched marshes has not been investigated in detail.

Changes in the morphology of tidal channels, such as ditching, alter the movement of water in the channels, the tidal asymmetry, and can influence the residence time of tides in ditched marshes (Zheng et al., 2003). Additionally, both high and low marsh areas are ditched, allowing water to spread to high marsh areas that would otherwise be unaffected by average tidal exchange (Kennish, 2001; Tonjes, 2013). High marshes could then transition to low marshes due to increased inundation frequency and potential changes in flora species (Fitzgerald et al., 2008).

1.2. Hydrodynamics

Velocities in tidal channels within tidal marshes are generally asymmetric between flooding and ebbing tide, and vary between spring and neap tide due to the difference in tidal amplitude (Dronkers, 1986; Fagherazzi et al., 2008; Nidzieko and Ralston, 2012; Pethick, 1980). In addition, channel - marsh platform interaction influences velocities in the channels when the high tide level is above the marsh platform elevation (Lawrence et al., 2004; Torres and Styles, 2007). Rapid velocities induce large shear stresses that increase the potential for sediment suspension (Pope et al., 2006; Van Prooijen and Winterwerp, 2010). Shear stresses are maximum at the bed and generally decrease upward (Biron et al., 2004; Rippeth et al., 2002). Wind stress can cause larger fluid shear stress at the top of the water column (Christiansen et al., 2006; Dyer et al., 2000; Mariotti and Fagherazzi, 2012). However, in relatively small tidal channels that are sheltered from the wind by marsh vegetation, the influence of local wind stress on the water column is minimal. Shear stress magnitudes are largest during maximum flow conditions, and minimum around slack tide, when velocities are near zero (Biron et al., 2004; French and Clifford, 1992; Korotenko et al., 2013; Pacheco et al., 2009; Verney et al., 2006). In its simplest approach, shear stress (τ) is scaled as velocity squared (U^2) divided by water depth ($\tau \sim U^2/\rho h$), suggesting that the largest shear stress occurs during the largest flow velocities and shallowest depths (French and Clifford, 1992; Traynum and Styles, 2007). Peak shear stress values found in field experiments in tidal channels range from 0.2 to 3 N/m^2, depending on field site and flow conditions (French and Clifford, 1992; Korotenko et al., 2013; Pacheco et al., 2009; Pieterse et al., 2015; Rippeth et al., 2002; Verney et al., 2006; Wiles et al., 2006). These generalizations are for natural tidal channels, while detailed hydrodynamics in man-made ditches have not been measured previously.

1.3. Sediment suspension and deposition

The erosion of cohesive sediment depends on both the erodibility of the sediment and the shear stress induced by the flow (Andersen et al., 2007; Winterwerp and van Kesteren, 2004). The erodibility of cohesive sediment is influenced by consolidation and cohesion of the sediment, and biological structures present within the sediment (Andersen et al., 2007; Grabowski et al., 2011). These properties change in time, on the scale of hours, days and seasons (Andersen et al., 2006; Grabowski et al., 2011), and with depth into the sediment bed (Amos et al., 1992). A distinction is made between two different types of erosion: a) Type I erosion, which decreases exponentially with depth and time as the shear strength of the sediment increases, also called depth-limited erosion, and b) Type II erosion, which is constant with depth and time, because the shear strength of the sediment remains constant or below the applied shear stress, commonly referred to as unlimited erosion (Amos et al., 1992; Mehta and Partheniades, 1982; Sanford and Maa, 2001; Van Prooijen and Winterwerp, 2010; Winterwerp and van Kesteren, 2004). In the top layer of natural cohesive sediment beds, depth-limited erosion is generally expected. The threshold for erosion is described by the critical shear stress, which in the case of depth-limited erosion, increases into the bed (Winterwerp and van Kesteren, 2004; Winterwerp et al., 2012).

It has been suggested that erosion of cohesive sediment in natural channels occurs primarily in the accelerating phase of the tide, due to the increasing critical shear stress with depth into the bed (Maa and Kim, 2002; Maa et al., 2008). Since erosion in tidal channels is a relatively fast process, the less consolidated layers with smaller critical shear stress are eroded during the accelerating phase of the tide and subsequently the critical shear stress exceeds the shear stress such that erosion does not occur during the decelerating phase of the tide.

Erosion and deposition rates in tidal channels and on tidal flats can vary considerably over a range of time scales. Many studies have focused on time scales of days to months (Andersen et al., 2006; Bassoullet et al., 2000; Christie et al., 2000; Deloffre et al., 2006; Fan et al., 2002; Houwing, 1999; Yang et al., 2001, 2003), while few have investigated intratidal erosion and deposition (Andersen et al., 2007; Deloffre et al., 2005; Shi et al., 2014), and those studies occurred on tidal flats. Net erosion or deposition is often small, O(mm/day), on day to month time scales (Andersen et al., 2006; Deloffre et al., 2007; Shi et al., 2014). However, when considering erosion/deposition within a tidal cycle, erosion rates of 1 mm per hour or more on tidal flats have been documented (Andersen et al., 2007; Deloffre et al., 2005; Shi et al., 2014). These rapid erosion rates indicate that the upper layer of the sediment bed is active during a tidal cycle, while net bed level changes are small. The difference between intratidal sediment movement and net bed level change is in part caused by the existence of a high porosity fluffy layer in tidal regimes. This dilute sediment layer is eroded at the onset of each tidal cycle and redeposited during slack tide (Gust and Morris, 1989; Winterwerp and van Kesteren, 2004; Winterwerp et al., 2012).

Erosion rates are not normally measured directly, but are calculated based on measured velocity and shear stress time series. One method to calculate the erosion flux is from the shear stress and the critical shear stress (Amos et al., 1992; Parchure and Mehta, 1985; Sanford and Maa, 2001; Winterwerp and van Kesteren, 2004; Winterwerp et al., 2012) as

$$E = M\left(\frac{\tau_b - \tau_c}{\tau_c}\right)^n \text{ for } \tau_b > \tau_c, \tag{1}$$

where E is the erosion flux, M is an erodibility parameter, τ_b is the shear stress, τ_c is the critical shear stress, and n is usually taken to be 1. In the case of depth-limited erosion, both M and τ_c change with depth into the bed (Amos et al., 1992; Winterwerp et al., 2012). The depositional flux depends on the suspended sediment concentration and the fall velocity of the sediment (Maa et al., 2008; Shi et al., 2014), which in turn depends on flocculation of the sediment and the density of the flocculated particles (Winterwerp and van Kesteren, 2004). Often, it is assumed that deposition occurs only when the shear stress is less than a critical shear stress for deposition (Krone, 1962; Maa et al., 2008; Mehta and Partheniades, 1975), suggesting that deposition and erosion do not occur at the same time.This paper focuses on the hydrodynamics in man-made