Advances in swash-zone research: small-scale hydrodynamic and sediment transport processes

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Abstract:

Extensive research efforts in the last decade have improved knowledge of swash-zone processes. This paper reviews and synthesizes from 2004 onward the research advances of small-scale hydrodynamics and sediment transport processes considering field, laboratory and numerical modeling efforts. Swash-zone hydrodynamics are examined by means of shoreline motions, fluid velocities, pressure gradients, flow structure, bed shear stress, friction coefficients and turbulence. Subsequently, advances related to sediment transport mechanisms and morphological changes are described. Detailed descriptions of measuring techniques, novel findings, and the strengths and weaknesses of the different approaches are discussed. The review also acknowledges the major advances in the development of instrumentation to collect highly resolved flow and sediment concentration measurements in direct proximity of the bed and the

experimental framework for analyzing local dynamics on a wave-by-wave time scale (e.g., dam break).

Keywords: Foreshore; Morphology; Sediment transport; Shear stress; Swash-zone dynamics; Turbulence.

1. Introduction

The active swash zone (Fig. 1) is a dynamic region separating land from inner surf zone coastal processes. There are many reasons why the swash zone is important: it is a primary area for beach recreation, it is a region of active beach accretion and erosion, beach nourishment activities often extend into the swash zone, nutrient/chemical species cycling between groundwater/oceanic flows occurs across the foreshore, and certain types of fauna are prevalent in or use the swash zone (e.g. birds, horseshoe and sand crabs). The swash zone is an integral part of the nearshore and represents a landward boundary condition for large-scale numerical models. Hence, there is a general scientific interest of the physical processes taking place within this region. Swash-zone processes occur across the foreshore with the cross-shore extent modulated by individual waves, infragravity energy and tidal fluctuations. This region is where wave energy is finally dissipated or reflected implying that the swash zone serves in a buffer capacity for hydrodynamic forcing.

The swash zone on most beaches is readily accessible, but this accessibility does not translate into a wide knowledge of the underlying physical processes. This zone exhibits some of the most challenging attributes in the coastal ocean with regard to measurement and modeling campaigns.

46 Flows in the swash zone reverse direction with the reversal timing varying across the foreshore
47 extent. Flows are rapid and contain large sediment loads and void fraction. Turbulence is
48 acknowledged as important for energy considerations and sediment mobility but quantification is
49 hampered through non-stationarity and signal discontinuity (for instance due to intermittent
50 submergence of instrumentation and/or to flows laden with bubbles and sediment).
51 Fluid/sediment processes are further complicated by a highly variable sea bed with elevation
52 variability occurring across a wide range of frequencies.

53 Increased societal/scientific interest and technological advances have contributed to the more
54 recent advances in understanding of swash-zone processes. The number of peer-reviewed papers
55 related to swash-zone processes was quantified (Fig. 2) through Google Scholar
56 (scholar.google.com) using the search terms of: swash, uprush, backwash, backrush, runup and
57 foreshore in the paper title and including the seminal works of Meyer and his colleagues. There
58 have been over 560 publications with more than 19,500 total citations. The number of published
59 papers has grown considerably since 1950. The corresponding total citations for papers
60 published in a given year also shows considerable growth with the exception of the most recent
61 papers that do not yet have a published life extensive enough to garner a high number of
62 citations. These numbers are certainly an under prediction of the true number of swash-zone
63 peer-reviewed publications based on the limited search parameters used. But, this simple
64 presentation shows the increasing importance of swash-zone research through time. Papers
65 related to swash-zone research increased more in the late 1990's and early 2000's (Nearshore
66 processes workshop report released in 2000; Thornton et al., 2000). The 1st International
67 Workshop on Swash-Zone Processes was held in 2004 where researchers discussed the most
68 pressing issues related to swash-zone research. Many new findings/approaches have resulted

from the “way ahead” discussions that took place there and were documented (Puleo and Butt, 2006). A similar workshop was held in 2014 (Puleo and Torres-Freyermuth, this issue) to again identify the most pressing needs for swash-zone research.

The increased interest in swash-zone research resulted in the publication of several review papers (Bakhtyar et al., 2009a; Brocchini and Baldock, 2008; Butt and Russell, 2000; Elfrink and Baldock, 2002; Longo et al., 2002; Masselink and Puleo, 2006) related to hydrodynamics, sediment transport and morphodynamics. New approaches and scientific results through laboratory, field and numerical efforts have emerged since the most recent review papers. For example, numerical models have been developed that explicitly include swash-zone processes either through long wave, short wave or both types of motions. New instrumentation has been developed/used to provide high-resolution (space/time) velocity profiles, sheet flow sediment concentration and time-dependent bed level elevations. The present work provides an overview of the most recent advances (mainly since 2004; the 1st International Workshop on Swash-Zone Processes) related to swash-zone research. This review paper addresses short time-scale hydrodynamics (Section 2), and small-scale sediment transport and morphology (Section 3). Some aspects of numerical modeling related to those processes occur in each section. Online supplementary material is included that contains information related to recent laboratory and field efforts. The description, in tabular form, is meant to provide researchers a means to determine easily the types of recent swash-zone studies that have been conducted and the parameters associated with each study. This review paper does not address in detail longer term processes (e.g., Ruggiero and Cohn, this issue), runup (e.g., Hughes et al., 2010; Ruggiero et al., 2004; Stockdon et al., 2006; Vousdoukas, 2014), nor address groundwater aspects of the swash

zone (e.g., Austin and Masselink, 2006a; Heiss et al., 2015, 2014; Horn, 2006; Pintado-Patiño et al., 2015; Sous et al., 2013; Steenhauer et al., 2011; Turner and Masselink, 1998).

2. Swash-zone hydrodynamics

Natural beaches are dynamic and complex with high temporal and spatial variability, leading to difficulties in quantifying swash-zone hydrodynamic processes. Numerous research efforts have reduced their temporal scope of interest to focus on processes occurring on a wave-by-wave time scale. In laboratory studies, the isolation of such processes occurs for a single swash event via solitary wave (e.g., Alsina et al., 2009; Barnes et al., 2009; Barnes and Baldock, 2007; Sumer et al., 2011), dam break forcing (Kikkert et al., 2013, 2012; O'Donoghue et al., 2010; Othman et al., 2014; Steenhauer et al., 2011; Yeh and Ghazali, 1988), or ensemble averaging of swash events under regular or irregular wave forcing (e.g., Cowen et al., 2003; Mihoubi et al., 2012; Pedrozo-Acuña et al., 2011; Sou and Yeh, 2011). In field studies, the analysis may be simplified by using only a single swash event or a “typical” swash event through ensemble averaging. A single event in the field is identified by periods of zero water depth, zero up crossings in a velocity time series, or the time between submergence and emergence of an elevated current meter (e.g., Austin et al., 2011; Baldock and Hughes, 2006; Butt et al., 2004; Hughes and Baldock, 2004; Lanckriet et al., 2014; Masselink et al., 2009; Masselink and Russell, 2006; Raubenheimer et al., 2004). Individual swash events have received attention in part due to repeatability in laboratory studies. Repeatability 1) allows for spatial coverage of the cross-shore swash extent by laterally and/or vertically shifting measurement instrumentation, 2) increases confidence of measurements and 3) enables quantification of the turbulence field via ensemble-averaging techniques.

The evolution of hydrodynamic properties (i.e., water depths, velocities, bed shear stresses, turbulence) in the swash zone varies throughout the swash event. The different flow conditions are sometimes categorized using the Froude number ($Fr = u/\sqrt{gh}$) in terms of the cross-shore flow velocity (u), a length scale given by the water depth (h), and the gravitational acceleration (g). The different phases of the swash motion are: (i) bore arrival at the seaward boundary where the velocity is onshore-directed with $Fr > 1$; (ii) an uprush phase where the flow decelerates with $Fr < 1$; (iii) a flow reversal phase where the velocity reduces until it eventually reverses direction ($Fr < 1$); and (iv) a backwash phase where the velocity is offshore-directed and accelerating, evolving from $Fr < 1$ to $Fr > 1$ late in the backwash phase (Baldock and Hughes, 2006; Brocchini and Baldock, 2008; Butt et al., 2004; Guard and Baldock, 2006; Masselink et al., 2005; Mory et al., 2011), until the flow decelerates during the last portion of the event.

In the following subsections, the main hydrodynamic properties are described during the different flow phases, together with descriptions of new measuring techniques and numerical model findings. Quantitative results are provided where possible. However, it is acknowledged that the reported values may be applicable only to a specific study. No mathematical descriptions of the numerical schemes are given. Instead, the reader is referred to Briganti, et al. (this issue) for in-depth description regarding numerical formulations.

2.1 Water level and shoreline motions

Fig. 3 depicts a typical “free” swash event including water depth, cross-shore velocity, and sediment transport. The term “free” is used here to describe a swash event with little interaction between the backwash and ensuing uprush (i.e., dam break scenario). Water depth for a swash event (Fig. 3a) is asymmetric, rising rapidly upon initial inundation and reaching a maximum

137 shortly before flow reversal (Barnes et al., 2009; Hughes and Baldock, 2004; Kikkert et al.,
138 2012; O'Donoghue et al., 2010). The maximum water depth is related to variations in
139 momentum of the different parcels of fluid. The inflow of additional momentum is important as
140 it shifts the location of the maximum water depth experienced at a given location to later instants
141 after bore arrival (Baldock et al., 2005). Fluid parcels at the bore front experience larger friction
142 and converge at the bed while the parcels behind the bore front with higher momentum are
143 continuously injected into the front (Baldock et al., 2014; Barnes and Baldock, 2010). Water
144 depth decreases during backwash at a slower rate relative to the increase during uprush and
145 ultimately pinches to zero at the end of the backwash if the free swash event is unencumbered by
146 a subsequent uprush. Water depth time series are more symmetric in the landward direction as
147 the bore head and overall momentum of the uprush are progressively reduced.

148
149 Water depth can be quantified using *in situ* or non-intrusive approaches. *In situ* pressure
150 transducers measure water depth based on the assumption of hydrostatic pressure distribution
151 (see Section 2.4). These transducers are commonly positioned just below or in direct vicinity of
152 the bed (see online supplementary material for papers utilizing this approach). Capacitance and
153 resistance gauges measure water depth, but are commonly restricted to laboratory settings (see
154 online supplementary material). *In situ* resistance runup wires are used to measure the runup
155 excursion across the foreshore (see online supplementary material). Non-intrusive techniques
156 used to measure free surface elevations consist of ultrasonic displacement meters (UDM), laser-
157 induced fluorescence (LIF), light detection and ranging (LIDAR), laser Doppler anemometry
158 (LDA or LDV) (see Section 2.3), and video imagery for shoreline position (see online
159 supplementary material). UDMs measure free surface or bed level when the water either covers

or uncovers a particular cross-shore location. Therefore, they are often deployed in cross-shore arrays. LIF techniques are used in laboratory studies because the water must be dyed and illuminated to enhance water reflection. LIDAR is a continuously scanning laser beam that detects swash-zone water surface and exposed bed when the incident angle of the beam is approximately perpendicular to the surface (Blenkinsopp et al., 2010a; Vousdoukas et al., 2014). Video imagery through the side wall of a laboratory flume can be used to determine water levels (Erikson et al., 2005; Sou et al., 2010; Sou and Yeh, 2011).

Water depth measurements over time and space allow the estimation of hydrodynamic parameters such as bore front velocity (see Section 2.2) and cross-shore pressure gradients (see Section 2.4). The relative position of the instrumentation within the swash zone can be estimated in terms of the percentage of time the bed is inundated in the measured signal over a given period (e.g., tide) or by considering the mean water level (Aagaard and Hughes, 2006; Hughes and Moseley, 2007; Masselink et al., 2005; Masselink and Russell, 2006). Hence, estimates of shoreline minima and maxima (e.g., the 2.5 and 97.5 percentiles of inundation time; Hughes and Moseley, 2007) can be used to delimit the swash zone extent over time. Additionally, free surface elevation and shoreline motion time series enable the identification of swash-swash interactions (Alsina and Cáceres, 2011; Brocchini and Baldock, 2008; Jackson et al., 2004).

Shoreline motions are driven by incident and infragravity waves. Incident-band related motions result from wind waves and swell (roughly $f > 0.05$ Hz; where f is frequency). Water depth and shoreline motions in the incident band are largely related to foreshore slope where the energy in the surf zone is typically saturated (Guza and Thornton, 1982; Hughes et al., 2014; Ruessink et

al., 1998; Senechal et al., 2011). Infragravity related motions correspond to long-waves (roughly $f < 0.05$ Hz) often associated with edge waves (Holman and Sallenger, 1985; Hughes et al., 2014; Ruggiero et al., 2004). Water depth and shoreline motions in the infragravity band increase with increasing offshore energy (Guza and Feddersen, 2012; Holland et al., 1995; Hughes et al., 2014; Ruggiero et al., 2004).

The analytical solution of Shen and Meyer (1963) (referred to as SM63 hereafter) provides an asymptotic description of the incident-band shoreline motion driven by a collapsing bore. The SM63 study serves as a starting point for numerical models applied to swash-zone flows (Brocchini and Baldock, 2008) as it leads to estimates of the runup distance given an initial bore definition and a foreshore slope (Peregrine and Williams, 2001; Pritchard and Hogg, 2005). Many limitations have been identified regarding the SM63 solution (readers are referred to (Brocchini and Baldock, 2008; Guard and Baldock, 2006; Power et al., 2011; Pritchard, 2009). Here, it is mentioned that the SM63 solution is singular and results in shallower depths, reduced flow volumes (Guard and Baldock, 2006) and neglected flow momentum behind the bore front entering the swash domain (Baldock et al., 2005). These limitations mainly relate to the seaward boundary condition which is only defined by an invariant bore velocity at the initial shoreline (Baldock et al., 2005; Guard and Baldock, 2006; Mory et al., 2011; Pritchard and Hogg, 2005). Guard and Baldock (2006) introduce a time-variant boundary condition through an additional term that denotes a rate of increase for the mass and momentum flux into the swash zone. This numerical treatment showed that the water depth and velocity asymmetry mainly depend on the shape and wavelength of the incoming bore in addition to the foreshore slope.

Other depth-integrated models that solve the non-linear shallow water equations (NLSW) have been formulated and employed to predict shoreline motions and water depths (Baldock et al., 2008a; Briganti et al., 2011; Briganti and Dodd, 2009; Nielsen et al., 2005). Their work is significant as the numerical framework is less computationally demanding than other numerical models, yet provides accurate predictions. However, the turbulence at the bore front is not resolved, usually resulting in over prediction of initial water depths (Briganti et al., 2011). Another related limitation refers to the bore collapse and resulting shoreline excursion. Yeh and Ghazali (1988) observed that a small water wedge is formed as the bore collapses that is pushed up by the bore front as it propagates over the foreshore. NLSW models cannot produce this behavior, and its description is provided by studies with depth-resolving models. Zhang and Liu (2008) used a depth-resolving Reynolds Averaged Navier Stokes (RANS) numerical model and showed that the water wedge in front of the bore acts as a "buffer" that reduces the intensity of the bore collapse. Multiphase depth-resolving model simulations indicated that the bore collapse occurs beyond the initial shoreline for different bore conditions (Mory et al., 2011). Their findings suggest that the SM63-based models may not properly scale the energy contained in the forcing bore that drive swash flows (Guard and Baldock, 2006; Peregrine and Williams, 2001; Pritchard and Hogg, 2005). The inclusion of these types of bore features results in shorter runup distances than predicted by NLSW solvers.

Shoreline motions, composed of wave setup and swash, are influenced by offshore wave characteristics, foreshore slope and morphology. The maximum elevation is reached during storm events influenced by storm surge and infragravity components within the surf zone. Prediction of runup height (including both wave setup and swash) is of particular interest in

engineering applications (i.e., improve coastal morphology response and flooding hazard models and design of coastal structures). Although the aim of this review paper is to focus on small-scale swash-zone processes, a brief overview of statistical and spectral analysis of runup is given. Readers are directed to the cited papers for in-depth description.

Runup on dissipative beaches is linearly proportional to the offshore significant wave height with variability in the empirical coefficients (e.g., Guza and Thornton, 1982; Ruessink et al., 1998; Ruggiero et al., 2004; Ruju et al., 2014). Statistical approaches identified a dependence of the 2% runup exceedence ($R_{2\%}$) (Stockdon et al., 2006) to dimensional forms of the Iribarren number (Iribarren and Nogales, 1949) relating the foreshore slope to offshore wave steepness. The $R_{2\%}$ provides a quantitative method to predict flooding hazards and morphodynamic changes (Stockdon et al., 2007). Stockton et al. (2006) evaluated a wide range of beach types and non-storm conditions to develop an empirical formulation for $R_{2\%}$. $R_{2\%}$ correlates well with Iribarren number on reflective/intermediate beaches but only with offshore significant wave height and wavelength on dissipative beaches. Stockdon et al. (2007) used numerical modelling to extend the applicability of the Stockton et al. (2006) empirical model to storm conditions. Runup heights can also be affected by offshore frequency and directional spread of incoming waves (Guza and Feddersen, 2012). Hughes et al. (2014) performed spectral analyses on numerous time series from a variety of beaches with foreshore slopes ranging from 1:6 to 1:60 and offshore wave heights of 0.5 to 3.0 m. They found that swash energy in the incident band is dominant on steep and intermediate foreshores whereas that in the infragravity band is dominant on shallow sloping foreshores. The results are cast into a morphodynamic beach state model (Wright and Short, 1984) such that there is a shift in the spectral signatures under increasing or decreasing wave

energy as: 1) the incident band spectrum is fairly constant for all beach states with an f^{-4} roll off and controlled by the foreshore slope, 2) the width of the energy roll off varies with offshore energy and inversely with foreshore slope, and 3) the infragravity portion of the spectrum varies for all beach states and is related to offshore energy and beach morphology.

2.2 Cross-shore and alongshore swash-zone velocities

Swash-zone flows are shallow, short-lived and regularly laden with bubbles and sediment. Cross-shore swash-zone flow (Fig. 3b) is typically initiated by a collapsing bore that moves a pulse of turbulent water up the foreshore inducing a sudden increase in velocity. During the majority of the uprush the flow decelerates due to bottom friction, gravitational acceleration and the pressure gradient (see Section 2.4) force acting in tandem (Baldock and Hughes, 2006). Backwash flow accelerates under favorable gravity and pressure gradient force with bottom friction acting in opposition. These forces acting in the same direction (uprush) or in opposition (backwash) cause asymmetry in the swash-zone velocity time series. The asymmetry is particularly evident in the latter stages of backwash where the bottom friction begins to dominate the flow causing the velocity to decrease (Barnes and Baldock, 2007; O'Donoghue et al., 2010).

Many efforts have focused on measuring swash-zone velocities to quantify flow structure over the foreshore enabling the estimation of hydrodynamic parameters such as bed shear stress (Aagaard and Hughes, 2006; Inch et al., 2015; Kikkert et al., 2013, 2012; Masselink et al., 2005; O'Donoghue et al., 2010; Puleo et al., 2012), friction coefficients (Barnes et al., 2009; Inch et al., 2015; Puleo et al., 2014a; Raubenheimer et al., 2004) and turbulence (Kikkert et al., 2013, 2012; Lanckriet and Puleo, 2013; O'Donoghue et al., 2010; Raubenheimer et al., 2004; Sou et al., 2010). Velocities are obtained using *in situ* sensors and non-intrusive techniques. *In situ* sensors

consist of impeller current meters, electromagnetic current meters (EMCM), acoustic Doppler velocimeters (ADV), and acoustic Doppler profiling velocimeters (ADPV) (see online supplementary material). Impeller current meters measure the signed velocity magnitude; usually cross-shore but alongshore motions cannot be separated due to the mechanics of the sensor. The sensor diameter (~ 0.04 m) does not allow deployment in close proximity to the bed (Puleo et al., 2000). EMCMs provide two-dimensional velocity measurements (usually u , v corresponding to the cross-shore, x , and alongshore y , directions) at a fixed elevation with low-sampling rate (≤ 8 Hz). EMCMs have a small measuring volume in the vertical direction enabling sensor deployment within a few centimeters of the bed. ADVs and ADPVs utilize the Doppler shift of acoustic returns and are capable of sampling at high frequency (100 Hz or more). Side-looking ADVs can be deployed just above the bed. Other ADVs are deployed in the downward-looking direction providing all three velocity components (including w in the vertical, z , direction). ADPVs (e.g. Vectrino II by Nortek; Craig et al., 2011) deployed downward-looking can profile all the three velocity components over a range of 0.03 m at 0.001 m bin spacing. The minimum elevation for downward-looking acoustic sensors is about 0.05 m. One difficulty with acoustic sensors is the inability to return robust velocities when the sensor is initially inundated by bubbly flow. EMCMs and impellers have fewer difficulties under intermittent submersion conditions but impellers may be slowed under sediment-laden flow. All *in situ* current meters require elevation adjustment under conditions with active morphological change if the distance relative to the bed is meant to be similar throughout a particular study. In addition, data are only obtained when the water depth reaches or exceeds the sensor elevation. Thus, elevated (relative to the bed) sensors cause artificial truncation of swash event duration (see the line pattern areas in Fig. 3 for example).

There are a variety of non-intrusive sensing techniques that can be used to measure or infer swash velocities. Particle image velocimetry (PIV) is a non-intrusive flow measurement technique that relies on optical sensors to capture variations in image texture usually provided by seeded particle patterns (Adrian, 1991; Stevens and Coates, 1994). PIV uses statistical techniques to quantify the shift of pixel intensity patterns captured in sequential frames of video imagery. The inferred velocity varies depending on the camera position, whether pointed through a glass flume wall or bed or from above imaging the water surface. In laboratory experiments, the camera is typically located outside the wave tank and captures successive images of sediment particles, fluorescent particles or silver-coated hollow glass spheres. The water is illuminated by a laser system that delivers a light sheet (e.g., Kikkert et al., 2013, 2012; O'Donoghue et al., 2010) along a particular plane, generally x, z . Kikkert et al. (2013, 2012) and O'Donoghue et al. (2010) reported that velocities cannot be measured near bore arrival due to the high percentage of entrained air. Cowen et al. (2003) used fluorescent particles and optical filters such that only the particles were imaged overcoming the difficulty with bubbles. Puleo et al. (2003a) imaged the swash-zone free surface in the field using cameras deployed on a building roof top. Uprush velocities were quantified via PIV using natural aeration and bubbles as tracers, but backwash velocities were difficult to quantify due to insufficient surface texture. Other non-intrusive techniques use laser Doppler anemometry or velocimetry (LDA or LDV) that measures fluid velocities by detecting the frequency shift of laser light scattered by particles suspended in the translucent and non-aerated fluid (Shin and Cox, 2006). These sensors are positioned outside the wave tank and emit the laser pulse through the glass side wall. The authors are unaware of LDA or LDV techniques being used in swash-zone field studies.

Swash-zone velocity measurements can be collected across the foreshore but are often divided into regions for simplicity. The outer swash zone is the region with wave-swash interactions, while the inner swash zone consists of solely swash motion (Hughes and Moseley, 2007). These regions are relative cross-shore locations on the foreshore and vary with tidal modulations. Sensor elevation may also dictate whether or not the potential for wave-swash interaction is captured. For example, velocities may be recorded for a large portion of the swash event duration if deployed close to the bed, but only a short portion of the swash event duration if farther away (note that the difference between “close to the bed” and “farther away” may only be 0.1 m; for example). The relative cross-shore position and elevation vary from study to study. The reader is cautioned that the following comparisons from multiple studies do not account explicitly for the potential spatial differences in sensor deployment. Rather, maximum observed values are given where appropriate.

Maximum uprush velocities measured on moderate to gently sloping natural sandy foreshores varied from 1.0 m s^{-1} to more than 2.0 m s^{-1} , while backwash velocities varied from 0.5 to 2.0 m s^{-1} (e.g., Aagaard and Hughes, 2006; Butt et al., 2009; Masselink et al., 2009; Masselink and Russell, 2006; Puleo et al., 2014a, 2012; Tinker et al., 2009). Meanwhile, on steep sandy foreshores maximum uprush velocities of 1.5 to 3.0 m s^{-1} and 1.0 to 2.2 m s^{-1} for backwash velocities were measured (Conley and Griffin, 2004; Houser and Barrett, 2010; Shanehsazzadeh and Holmes, 2007). Maximum uprush velocities of 1.5 m s^{-1} to 2.0 m s^{-1} and backwash velocities exceeding 2.0 m s^{-1} have been reported on steep foreshores composed of fine gravel (Austin et al., 2011; Masselink et al., 2010). It is generally found that maximum uprush velocities are either similar to or slightly larger than the maximum backwash velocities. Swash-zone uprush and backwash velocities recorded in laboratory dam break experiments vary from 1.5 to 2.0 m s^{-1} and

1.0 to 1.5 m s⁻¹ respectively on immobile rough beds under the same forcing conditions (Barnes et al., 2009; Kikkert et al., 2012; O'Donoghue et al., 2010). Maximum velocity magnitudes exceeding 1.5 m s⁻¹ were recorded in laboratory experiments with swash motion forced by irregular waves (e.g., Alsina et al., 2012; Briganti et al., 2011).

Alongshore flows are mostly unidirectional and have received less attention compared to cross-shore flows (Austin et al., 2011). Alongshore flows on gentle and steep foreshores can reach maximum velocities of 1.5 m s⁻¹, often exceeding cross-shore velocities (Austin et al., 2011; Puleo et al., 2014a). Therefore, it may be critical to incorporate alongshore velocities into calculations of the bed shear stress and predictions of swash-zone sediment transport.

2.3 Depth-averaged cross-shore velocity

The previous section highlighted the difficulty in obtaining swash-zone velocity time series for the full duration of a swash event. The volume continuity method using spatial and temporal measurements of water depth can partially overcome this difficulty. The method enables estimation of the depth-averaged cross-shore velocity by dividing cross-shore volume flux by local water depth (for example using UDM or LIDAR). The volume continuity method has been used in numerical models (e.g., Hughes and Baldock, 2004; Turner and Masselink, 1998), laboratory studies (Baldock and Holmes, 1997), and field studies (e.g., Baldock et al., 2004; Blenkinsopp et al., 2010a, 2010b; Houser and Barrett, 2010, 2009). Cross-shore velocities estimated in this manner compare well with *in situ* measurements in terms of velocity magnitude (root-mean square velocity differences of 0.2 m s⁻¹) and the timing of flow reversal (Blenkinsopp et al., 2010b; Hughes and Baldock, 2004). The estimated velocity may be larger at the initiation

of uprush and end of backwash relative to *in situ* velocity measurements due to the spacing of the *in situ* or non-intrusive sensors (EMCM or UDM). Differences in the sampling rates between sensors may also cause phase inconsistencies in the signals.

The advantage of the volume continuity technique is the ability to provide cross-shore velocity time series for an entire swash event and at multiple locations across the foreshore. However, 1) no information about the vertical velocity structure is obtained, 2) a fixed bed level must be assumed, and 3) uncertainties from LIDAR during the extremely shallow backwash flows may yield exaggerated velocity asymmetry. Additionally, this method does not account for fluid infiltration/exfiltration and alongshore flows.

2.4 Pressure gradients

The fluid momentum at a given swash location is related to the local balance between the horizontal pressure gradient, gravity, and resistive forces. The pressure gradient contribution may be substantial (i.e., 1/3 of the total bed shear stress; Kranenburg et al., 2012) over a horizontal bed under wave action. The contribution of the pressure gradient force over a sloping bed in the swash zone is an order of magnitude smaller than the total force acting on the bed (gravity, bed shear stress, and pressure gradient) (Barnes et al., 2009). Despite this, the horizontal pressure gradient has been identified as a potentially important hydrodynamic force driving swash-zone sediment dynamics (Baldock and Hughes, 2006; Puleo et al., 2007). Assuming hydrostatic pressure ($\frac{1}{\rho} \frac{\partial P}{\partial x} = g \frac{\partial \eta}{\partial x}$, where ρ is the fluid density, p is the pressure and η is the free surface elevation) and small shear stress at the free surface, the free surface slope is related to the horizontal pressure gradient and directly proportional to the total fluid acceleration as

$$\rho \frac{DU}{Dt} = -\frac{\partial P}{\partial x} = -\rho g \frac{\partial \eta}{\partial x}, \quad (1)$$

where U is the flow velocity (free stream or depth-averaged), and $\rho \frac{DU}{Dt}$ denotes the flow momentum. Eq. (1) implies that the total (local + convective) acceleration may be assessed through the horizontal pressure gradient if only water depth information is known. The Euler's approach (in hydrostatic form) has led to improvements of sediment transport formulations applied in the surf zone (Drake and Calantoni, 2001; Hoefel and Elgar, 2003) where the local fluid acceleration is well-correlated to the pressure gradient. However, the local acceleration in the swash zone cannot be used as a proxy for the pressure gradient (Baldock, 2012) as confirmed by field (Baldock and Hughes, 2006), laboratory (Pedrozo-Acuña et al., 2011), and numerical (Puleo et al., 2007; Torres-Freyermuth et al., 2013) studies.

Baldock and Hughes (2006) used video images of the water surface and fixed horizontal reference string lines (stacked vertically at 0.05 m spacing) to determine the horizontal pressure gradient in the swash zone of a natural beach. Just before bore collapse, the flow has a maximum negative pressure gradient ("landward dipping") while the leading edge maintains nearly horizontal or positive pressure gradient ("seaward dipping") throughout the remaining swash cycle. Thus, onshore flow is generally hindered by horizontal pressure gradients whereas offshore flow occurs under favorable horizontal pressure gradients. Moreover, the total acceleration is always offshore-directed (Baldock and Hughes, 2006; Pedrozo-Acuña et al., 2011). The lack of correlation between local acceleration and pressure gradient is further supported by RANS modeling investigations (Pedrozo-Acuña et al., 2012; Puleo et al., 2007), regardless of the elevation in the water column or cross-shore location (Torres-Freyermuth et al., 2013).

Pressure gradient measurements or estimates in the swash zone have largely focused on the horizontal gradients but vertical gradients can also exist. Masselink et al. (2005) observed the temporal variability of the vertical pressure gradients on a dissipative beach. Vertical pressure gradients were 1 N m^{-3} during initial uprush and -0.5 N m^{-3} late in the backwash. Vertical pressure gradients alter infiltration and exfiltration of fluid, and can enhance or weaken bed shear stress (Butt et al., 2001; Karambas, 2003; Nielsen et al., 2001; Turner and Masselink, 1998).

2.5 Boundary layer

The boundary layer structure differs between the uprush and backwash phase of swash motions. The initial development of the uprush boundary layer (Fig. 3a) occurs at bore collapse. The boundary layer grows until reaching the water surface and subsequently is governed by the water depth until flow reversal (Barnes and Baldock, 2010). The cross-shore velocity, bed shear stress and turbulence intensity decay (Cowen et al., 2003; Lanckriet and Puleo, 2013; Sou and Yeh, 2011; Zhang and Liu, 2008) and the boundary layer vanishes during flow reversal (Briganti et al., 2011; Cowen et al., 2003; Masselink et al., 2005; Torres-Freyermuth et al., 2013). The swash-zone flow close to the bed reacts to gravity earlier than the flow in the upper water column during flow reversal (Desombre et al., 2013; O'Donoghue et al., 2010). In this transition, there is flow convergence at the bed (Sou and Yeh, 2011) and the boundary layer develops again during offshore-directed motion. The boundary layer grows quasi-linearly during the initial backwash phase (Pintado-Patiño et al., 2015) with a duration that is progressively shortened in the landward direction (Kikkert et al., 2013, 2012). During the remaining backwash, the flow gains momentum, evolves from critical to supercritical (Barnes and Baldock, 2010) and the boundary layer continues growing until it becomes again bounded by the water depth (Barnes

and Baldock, 2010; Briganti et al., 2011; Pintado-Patiño et al., 2015; Torres-Freyermuth et al., 2013).

Velocities collected with an ADPV or vertical arrays of EMCs or ADVs provide the vertical structure of the flow. The difficulty of array deployment is that they rely on selecting the minimum offsets between the stacked sensors in order to avoid the generation of wakes that may introduce errors in the measurements (Austin et al., 2011; Puleo et al., 2014a; Raubenheimer et al., 2004), and may not resolve fully the boundary layer. Boundary layer thickness is estimated from the vertical profile of velocities in the lower water column. The logarithmic law, if valid, is appealing for swash-zone flows as the velocity profile can be determined from a single characteristic velocity from an elevation inside the boundary layer, as

$$u(z) = \frac{u_*}{\kappa} \ln\left(\frac{z}{z_o}\right), \quad (2)$$

where u_* is the friction velocity, κ is the von Karman constant ($= 0.4$), and z_o is the bed roughness. Friction velocity can be estimated by calculating the slope (m) of the least squares regression between the velocity profiles and $\ln(z)$ as

$$u_* = m\kappa. \quad (3)$$

Alternatively, if only two velocity time series from different elevations are collected, u_* can be calculated as,

$$u_* = \kappa \frac{\Delta u}{\ln(z_u/z_l)}, \quad (4)$$

where Δu is the velocity difference between the upper and lower sensor, located at z_u and z_l , respectively (Austin et al., 2011). If only one velocity time series from a single elevation is

collected, u_* may be calculated by fitting a logarithmic model and assuming a value for the bed roughness z_0 ($= 0.004$) (Raubenheimer et al., 2004).

$$u_* = \frac{\kappa u}{\ln(z/z_0)}. \quad (5)$$

The logarithmic law can be applied from the top or within the interstices (Middleton and Southard, 1984) of roughness elements until the limit of the log layer. The vertical limit of the log layer is normally defined as the elevation in which the vertical gradients in velocity become less than some critical value or until the velocity reaches some percentage of the free-stream velocity (e.g., Dean and Dalrymple, 2001; Nielsen, 1992; Puleo et al., 2004; Schlichting and Gersten, 2000). The logarithmic law was not developed for accelerating or decelerating shallow flows in which the concept of a free-stream velocity may no longer be viable. However, various swash zone studies have assessed the logarithmic law with highly resolved cross-shore velocity profiles (e.g., Inch et al., 2015; O'Donoghue et al., 2010; Puleo et al., 2012).

Most studies focused on estimating boundary layer thickness use the vertical profile of cross-shore velocity (Inch et al., 2015; Masselink et al., 2005; Puleo et al., 2014a, 2012; Raubenheimer et al., 2004). These authors report maximum boundary layer thickness between 0.02 to 0.07 m. Austin et al. (2011), estimated the thickness of the boundary layer between 0.06 to 0.1 m using alongshore velocity only. One possible explanation for the difference in thickness depending on which velocity component is used is that the alongshore flow is unidirectional and the boundary layer has more time to develop during the swash event. However, further analysis is needed to identify the influence of alongshore flows on the boundary layer structure.

Laboratory studies have analyzed the evolution of the boundary layer over varying bed permeability. The uprush boundary layer becomes thicker on rough impermeable beds where the

bore and swash flow propagates over a ‘dry’ foreshore (Barnes and Baldock, 2010; Kikkert et al., 2012; O’Donoghue et al., 2010). Infiltration retards boundary layer growth over permeable beds resulting in a thinner boundary layer compared to an impermeable bed under similar forcing conditions (Kikkert et al., 2013). The differences related to permeability indicate that on natural beaches, the infiltration and exfiltration effects are important to estimate accurately the boundary layer evolution and thickness (Kikkert et al., 2013; Sous et al., 2013; Turner and Masselink, 1998). Nevertheless, to the authors knowledge, results from a detailed study of boundary layer evolution and simultaneous beach saturation and infiltration/exfiltration processes are lacking in the scientific literature.

Various boundary layer theories related to the different momentum integral approaches have been incorporated into NLSW models (Barnes and Baldock, 2010; Briganti et al., 2011). These modifications allow for improved predictions of bed shear stress with respect to measurements and logarithmic law estimates (see Section 2.6), and the ability to predict the velocity profile even from a depth-averaged model framework. The boundary layer evolution has also been investigated using RANS models. It was found that the velocity structure may be subjected to horizontal staining during uprush that hampers the concept of a depth-limited boundary layer in a strict sense (Torres-Freyermuth et al., 2013). Indeed, the boundary layer thickness deviates from a depth-limited state in the presence of flow divergence/convergence due to differential acceleration/deceleration at the surface. Additional modulations of the boundary layer are associated with infiltration/exfiltration processes, where the local saturation of the bed plays a major role defining whether the boundary layer growth rate is sustained or hampered by decreasing depths (Pintado-Patiño et al., 2015).

2.6 Bed shear stress and friction coefficients

Bed shear stress and friction coefficients are important hydrodynamic parameters associated with sediment transport processes. Different methods have been proposed to estimate these parameters in the swash zone, such as the quadratic drag law (Barnes et al., 2009; Masselink et al., 2009, 2005; Puleo et al., 2012; Raubenheimer et al., 2004), the von Karman-Prandtl relationship using near bed velocity profiles (e.g., Austin et al., 2011; Inch et al., 2015; Miles et al., 2006; Puleo et al., 2014a, 2012), and direct measurements using a shear plate or thermal techniques (Barnes et al., 2009; Conley and Griffin, 2004). The shear plate approach relates bed shear stress from a linear relationship to plate displacement. Hot film anemometry quantifies the shear stress based on temperature changes within the sensor.

The quadratic drag law is given as

$$\tau = \frac{1}{2} \rho c_f u |u|, \quad (6)$$

where τ is the shear stress, c_f is the friction coefficient, and $| |$ indicate absolute value. Eq. (6) is widely applied in studies where the velocity was collected at a single elevation above bed. The embedded friction coefficient should mostly depend on grain size and flow conditions (Swart et al., 1974) but frictional resistance may also depend on the sediment transport regime (Wilson, 1987). Using Eq. (6) in the swash zone often relies on a constant friction coefficient (Barnes et al., 2009; Barnes and Baldock, 2007; Conley and Griffin, 2004; O'Donoghue et al., 2010; Puleo et al., 2014a; Puleo and Holland, 2001). But, c_f may change with time and space (Barnes et al., 2009; Inch et al., 2015; Puleo et al., 2014a, 2012; Raubenheimer et al., 2004). When velocity profile data exist, c_f is estimated by

$$c_f = \frac{2u_* |u_*|}{u |u|}, \quad (7)$$

where u_* is the friction velocity (can be determined via Eq. 3, 4 and 5) and the fundamental definition of u_* is

$$u_* = \sqrt{\tau/\rho}. \quad (8)$$

Hence, the bed shear stress is obtained by algebraic rearrangement of Eq. (8).

Furthermore, when bed shear stress (i.e., hot film and shear plate) and velocity profile data are known, c_f can be estimated using Eq. (6) as (Conley and Griffin, 2004)

$$c_f = \tau \frac{2}{\rho u^2}, \quad (9)$$

The friction coefficient has a local minimum after uprush initiation and at mid stages of backwash (Fig. 3d), while, a local maximum c_f occurs near flow reversal. Friction coefficient estimates determined from Eq. (7) range from 0.003 to 0.06 for flows on natural beaches (e.g., Puleo et al., 2014a, 2012; Raubenheimer et al., 2004). On rough, impermeable beds, c_f ranges from 0.01 to 0.1 (Kikkert et al., 2012; O'Donoghue et al., 2010) and from roughly 0.002 to 1.0 on smooth beds (Cowen et al., 2003; O'Donoghue et al., 2010). Over rough, permeable beds, c_f ranges from 0.009 to 0.5 (Kikkert et al., 2013). Friction coefficients estimated from Eq. (9) under field conditions range from 0.002 to 0.004 (Conley and Griffin, 2004), 5 to 10 times smaller than estimates obtained from Eq. (7). Meanwhile, c_f estimates using Eq. (9) vary from 0.003 to 0.04 on smooth, impermeable beds to 0.002 to 0.06 on rough, impermeable beds (Barnes et al., 2009). The large range of c_f can be attributed to differences in cross-shore measurement location, instrumentation, bottom type and the elevation above the bed from which the velocity was collected. For example, errors in the assumed elevation of a velocity measurement can lead to

errors in c_f by 40 % (Raubenheimer et al., 2004). This error, consequently, has a direct effect on the bed shear stress estimated by Eq. (6) (Puleo et al., 2014a).

Studies have shown the same general evolution of the bed shear stress within an individual swash event regardless of the method used (Fig. 3c). Bed shear stress is maximum at uprush initiation and decreases to zero through flow reversal (Barnes and Baldock, 2007). Bed shear stress peaks again near the mid- to late-backwash (Austin et al., 2011; Inch et al., 2015; Masselink et al., 2005; Puleo et al., 2012) and then decreases to zero at the end of the backwash (e.g., Barnes et al., 2009; Barnes and Baldock, 2007; Conley and Griffin, 2004).

Maximum bed shear stress estimates using the von Karman-Prandtl approach vary from 4 to 50 N m^{-2} on dissipative beaches (Austin et al., 2011; Inch et al., 2015; Masselink et al., 2005; Puleo et al., 2014a). On moderate sloping foreshores, bed shear stress can reach 150 N m^{-2} (Miles et al., 2006). Weaker forcing conditions (smaller swash-zone velocities) are, not surprisingly, associated with smaller bed shear stresses. Under low wave-energy conditions, Conley and Griffin (2004) using hot film anemometry, estimate a maximum bed shear stress of 3.2 N m^{-2} and 1.2 N m^{-2} during the uprush and backwash, respectively. Uprush bed shear stress exceeds backwash bed shear stress in most cases (Austin et al., 2011; Barnes et al., 2009; Barnes and Baldock, 2007; Butt et al., 2005; Conley and Griffin, 2004; Miles et al., 2006; Puleo et al., 2014a). An exception was in the Puleo et al. (2012) study, but the acoustic sensor was often unable to obtain velocities or allow for bed shear stress estimation for much of the short uprush duration due to aeration. Thus, larger bed shear stresses expected in the uprush were not measured. Maximum bed shear stresses over rough, impermeable beds can reach $O(10^2)$ N m^{-2} (Barnes et al., 2009; Kikkert et al., 2012; O'Donoghue et al., 2010), while the bed shear stress is $O(10)$ N m^{-2} over smooth beds (Barnes et al., 2009; O'Donoghue et al., 2010). Maximum bed

shear stresses over permeable beds (Kikkert et al., 2013) were similar to those reported by Kikkert et al. (2012) over impermeable beds. The aforementioned field and laboratory studies have focused on estimating bed shear stress variability from cross-shore velocities. However, bed shear stress on sandy beaches estimated using alongshore velocities can have similar magnitudes (12 to 15 N m⁻²) to those estimated from cross-shore velocities only (Austin et al., 2011; Puleo et al., 2014a).

A Lagrangian NLSW model using the momentum integral approach for a turbulent flat-plate bottom boundary layer has provided further knowledge of bed shear stress variability (Barnes et al., 2009; Barnes and Baldock, 2010). The model was validated with direct bed shear stress measurements over smooth beds. The model can replicate the overall variability of the bed shear stress evolution with large (8 N m⁻²) short-lived bed shear stress during uprush, and longer-lived smaller bed shear stress (2 N m⁻²) values during the backwash (Barnes et al., 2009). Briganti et al. (2011) presented a similar NLSW model that solves the momentum integral method for the turbulent boundary layer of Fredsøe and Deigaard (1992) in an Eulerian reference frame. The model predictions show good overall accuracy with respect to logarithmic law bed shear stress estimates from laboratory data (Kikkert et al., 2012).

RANS models can also reproduce the skewed bed shear stress behavior where bed shear stress is related to the velocity gradients and eddy viscosity of the fluid from anisotropic turbulence models (Mory et al., 2011; Torres-Freyermuth et al., 2013). The occurrence of maximum bed shear stress is at the transition region from the bore collapse to swash initiation and its magnitude relates well with Fr (Mory et al., 2011). Moreover, the peak bed shear stress magnitude and location is dependent on the beach slope and bed roughness (Torres-Freyermuth et al., 2013). The peak bed shear stress magnitude is reduced and shifts its position onshore due to an increase

in foreshore slope and/or an increase in bed roughness. One of the more difficult aspects of predicting bed shear stress relates to the inability to validate numerical models with the additional characteristics of mobile, permeable beds. The vertical influx/outflow of momentum and turbulence into the bed is known to induce larger/reduced bed shear stress and the thinning/thickening of the boundary layer and stabilization/destabilization of sediments may occur (e.g., Butt et al., 2005, 2004, 2001; Horn, 2006; Karambas, 2003; Kikkert et al., 2013; Turner and Masselink, 1998). The reduced/additional flow volumes at the sediment/water interface may have implications for the momentum balance (Baldock and Nielsen, 2010).

2.7 Turbulence structure

Turbulence in the swash zone arises from plunging breakers and wave-driven bores collapsing/arriving at the shoreline (externally advected, referred to as bore-related), and due to the bottom presence (locally generated, referred to as bed-related) (Aagaard and Hughes, 2006; Cowen et al., 2003; Longo et al., 2002; Petti and Longo, 2001; Puleo et al., 2000; Sou et al., 2010; Zhang and Liu, 2008). The bed-related source of turbulence is present during uprush and backwash. The bore-related turbulence exists during the uprush phase of the flow only. Turbulence characterization in the swash zone is related to turbulent kinetic energy (TKE) or the turbulent dissipation rate per unit mass (ε).

2.7.1 Turbulent kinetic energy

The separation of velocity fluctuations from the mean flow in the swash zone is difficult (Petti and Longo, 2001) and standard approaches used in the surf zone are challenging to apply in the swash zone owing to flow discontinuities and accelerations (Longo et al., 2002). Turbulence

generation due to vortex shedding around sensors can also lead to difficulty when estimating TKE. One approach used to investigate TKE is via spectral analysis of the velocity components, where the macro (at similar wave frequencies) and micro (shorter frequencies) vortices are related to the energy transfer and energy dissipation, respectively (Petti and Longo, 2001). Butt et al. (2004) estimated TKE using ADV time series collected near the surf-swash zone transition to minimize discontinuities in the time series. The maximum TKE was related to the interaction between a receding backwash and incident bore with TKE values on the order of $O(10^{-1}) \text{ m}^2 \text{ s}^{-2}$. The contribution of each component of the velocity to the total velocity variance accounted for 65 %, 28 % and 7 %, corresponding to the cross-shore, u , alongshore, v , and vertical, w , velocity components respectively, suggesting that the use of u and v may suffice for TKE estimates. The cutoff frequency separating turbulence motions may affect the estimated TKE levels. For instance the TKE increased by 37 % using a 0.5 Hz cut off, or decreased by 23 % using a 2 Hz cut off.

Aagaard and Hughes (2006) assumed that TKE scales with the vertical velocity variance ($\text{TKE} \propto w'/w' \propto w |w|$; Svendsen, 1987) as obtained from ADV measurements at three locations across the foreshore. The reported values were mostly negative with values in the range $-0.35 \text{ m}^2 \text{ s}^{-2} < w |w| < 0.05 \text{ m}^2 \text{ s}^{-2}$ suggesting the formation and down bursting of coherent structures at the bore front and strong shear in the water column near the swash-swash interaction region (i.e., hydraulic jumps). Aagaard and Hughes (2006) also concatenated swash events, essentially removing portions of the time series when the sensor was exposed, before performing spectral analysis. An $f^{-5/3}$ roll off was found in the inertial subrange ($0.1 \text{ Hz} \leq f \leq 2.0 \text{ Hz}$) with turbulence accounting for up to 95 % of the vertical velocity fluctuations. The existence

of a characteristic $f^{-5/3}$ roll off in the inertial subrange indicates that the turbulence is concentrated in this frequency range, accounting for 74 % to 95 % of the total w variance.

Another approach to estimate TKE is through phase or ensemble averaging (Longo et al., 2002) of swash events with a high degree of similarity. This approach is most easily applied for laboratory studies under controlled forcing (dam break, regular or solitary waves) and fixed impermeable (Cowen et al., 2003; Kikkert et al., 2012; O'Donoghue et al., 2010; Sou and Yeh, 2011) or fixed permeable (Kikkert et al., 2013) beds. Velocity fluctuations used in the TKE estimate are obtained as the deviations from the ensemble average. In particular, O'Donoghue et al., (2010) report increases in TKE on the order of $O(10^1) \text{ m}^2 \text{ s}^{-2}$ due to increases in roughness for dam break-driven swash events over smooth and rough beds. Moreover, Kikkert et al. (2013) document diminished turbulence levels (TKE and turbulent shear) due to infiltration that depends on grain size.

Findings from laboratory and field efforts indicate TKE during uprush is bore-related (Aagaard and Hughes, 2006; Butt et al., 2004; Butt and Russell, 1999; Jackson et al., 2004; Kikkert et al., 2013, 2012; Masselink et al., 2005; O'Donoghue et al., 2010; Petti and Longo, 2001; Puleo et al., 2000) with magnitudes ranging from 0.02 to $0.4 \text{ m}^2 \text{ s}^{-2}$. Consequently, the production of TKE appears to result from shear forces inside the flow during bore collapse (Sou et al., 2010) and the main TKE mechanism during uprush is production (Zhang and Liu, 2008) and sinking of TKE into the bed (Kikkert et al., 2013; Pintado-Patiño et al., 2015). Most of the TKE content during uprush is dissipated by the time of flow reversal, decaying homogeneously (Kikkert et al., 2013, 2012; O'Donoghue et al., 2010) and similar to grid turbulence (Cowen et al., 2003; Sou et al., 2010). Backwash begins from rest with minimal initial TKE content. TKE is continuously generated at the bed (Cowen et al., 2003; Desombre et al., 2013; O'Donoghue et al., 2010; Sou et

al., 2010) until the end of the backwash . The formation of a backwash bore (Zhang and Liu, 2008) and hydraulic jumps due to swash-swash interactions (Aagaard and Hughes, 2006; Butt and Russell, 2005) contribute as additional sources of TKE that can be advected back into the swash zone and/or exported to the surf zone (Butt et al., 2004; Petti and Longo, 2001) . The infiltration and exfiltration of water at the sediment water interface may also increase TKE on a permeable bed (Kikkert et al., 2013; Pintado-Patiño et al., 2015). It is anticipated that enhanced understanding of swash-zone TKE will come from more elaborate laboratory studies focusing on swash-swash interactions.

2.7.2 Turbulence dissipation

Turbulence dissipation is quantified to estimate the rate at which TKE is converted to smaller scales and ultimately thermal energy. Different methods (direct and indirect) exist to calculate ε using field and laboratory measurements. An example of the direct method is based on PIV-generated data from repeatable swash events in the laboratory (Sou et al., 2010). The turbulent dissipation rate is quantified from

$$\varepsilon = \nu \left[4 \left\langle \left(\frac{\partial u'}{\partial x} \right)^2 \right\rangle + 4 \left\langle \left(\frac{\partial w'}{\partial z} \right)^2 \right\rangle + 3 \left\langle \left(\frac{\partial u'}{\partial z} \right)^2 \right\rangle + 3 \left\langle \left(\frac{\partial w'}{\partial x} \right)^2 \right\rangle + 6 \left\langle \left(\frac{\partial u'}{\partial z} \frac{\partial w'}{\partial x} \right) \right\rangle + 4 \left\langle \left(\frac{\partial u'}{\partial x} \frac{\partial w'}{\partial z} \right) \right\rangle \right], \quad (10)$$

where ν is the kinematic viscosity and $\langle \rangle$ denote phase averaging. The time evolution of ε exhibits a decay rate in the form of a power law with an exponent of 2.3 during the majority of the swash event and consistent with expectations for grid turbulence (Sou et al., 2010). Dissipation rate magnitudes of $O(10^{-2}) \text{ m}^2 \text{ s}^{-3}$ were observed for plunging waves ($H_0 = 0.246 \text{ m}$; $\xi = 0.56$) on an 1:20 slope.

Indirect methods employ the Kolmogorov hypothesis and rely on the frequency spectrum (Aagaard and Hughes, 2006; Raubenheimer et al., 2004), wavenumber spectrum (Cowen et al., 2003), or the structure function (Lanckriet and Puleo, 2013). The frequency spectrum requires a special treatment of the spectral domain due to the non-periodicity of the swash signal and related noise (Sou et al., 2010). Raubenheimer et al. (2004) pre-set to null values the velocity measurements during times of sensor exposure. This procedure served to eliminate the spectral contribution of the exposure times when performing spectral calculations. Swash-zone velocity spectra showed an $f^{-5/3}$ roll off in the inertial sub range denoting the dissipation of energy. The analysis resulted in turbulence dissipation rates of $O(10^{-2}) \text{ m}^2 \text{ s}^{-3}$ and $O(10^{-1}) \text{ m}^2 \text{ s}^{-3}$ for the inner surf zone and swash zone respectively.

Lanckriet and Puleo (2013) and Brinkkemper et al. (2015) estimated ε using the structure function (Pope, 2000; Wiles et al., 2006) applied to swash-zone ADPV data. The structure function relies on vertical profiles of w' , where the vertical component is typically used due to a lower noise floor. The second-order longitudinal structure function (D) is defined as

$$D(z, r, t) = \langle \{w'(z + r, t) - w'(z, t)\}^2 \rangle = C\varepsilon(z, t)^{\frac{2}{3}} r^{\frac{2}{3}}, \quad (11)$$

where r is the separation distance between velocity bins in the profile, $C = 2.0$ is an empirical constant, and $\langle \rangle$ denotes time averaging. Curve fitting and algebraic rearrangement of Eq. (11) provides the dissipation rate. Lanckriet and Puleo (2013) and Brinkkemper et al. (2015), applied this technique to data obtained from 3 ADPVs co-located in the cross-shore direction and at different elevations. Dissipation rates range between $O(10^{-5}) \text{ m}^2 \text{ s}^{-3}$ to $O(10^{-3}) \text{ m}^2 \text{ s}^{-3}$ and increase towards the surface during uprush and towards the bed during backwash corroborating expectations of turbulence generation and dissipation related to the dominant forcing (bore-

related and bed-related respectively). Uprush dissipation away from the surf-swash transition is bed-related and does not show an obvious increase towards the surface as was found by Lanckriet and Puleo (2013). Dissipation rates tend to increase with decreasing water depth (Lanckriet and Puleo, 2013; Raubenheimer et al., 2004; Sou et al., 2010).

2.7.2 Numerical modeling of the turbulence field

A detailed description of TKE or ε is difficult to obtain from NLSW models as the models cannot simulate the large vertical gradients in velocity (Brocchini and Baldock, 2008). Examples of large vertical gradients in velocity are the region below the (onshore-directed) bore at the zone of interaction with the remaining (offshore-directed) backwash flow (Butt et al., 2004). Hence, the assessment of TKE largely comes from depth-resolved numerical models. Different approaches have been employed: $k - \varepsilon$ (Bakhtyar et al., 2009b; Pintado-Patiño et al., 2015; Sou and Yeh, 2011; Torres-Freyermuth et al., 2013; Zhang and Liu, 2008) where k is the turbulent kinetic energy); $k - \omega$ (Desombre et al., 2013; Mory et al., 2011); where ω is the specific dissipation rate); and re-normalized group theory (Bakhtyar et al., 2013) to resolve the higher order non-linear turbulence terms, Reynolds stresses, and computation of eddy viscosities. Most numerical approaches have been validated and yielded improved understanding of swash-zone turbulence. However, as discussed by Longo et al. (2002), the biggest limitation is that the nature of turbulence is three-dimensional (3D) and most turbulence methodologies currently employed come from parameterized closure models in a two-dimensional numerical framework. It is anticipated that the use of 3D Large Eddy Simulation (LES) (Briganti et al., this issue) and direct multiphase solvers will provide a better representation of the turbulence field as high vorticity is produced in the roller region after wave breaking and during bore propagation (Watanabe et al.,

2005). Moreover, the air bubble content contributes to the TKE due to energy transfer during bubble rising (Longo et al., 2002) and turbulent dissipation due to air release (Bakhtyar et al., 2009b).

3. Sediment transport and morphology

3.1 Sediment fluxes

Sensors do not exist to measure directly sediment flux. Instantaneous fluxes are derived from the product of concentration and velocity measurements (Aagaard and Hughes, 2006; Alsina and Cáceres, 2011; Butt et al., 2009; Cáceres and Alsina, 2012; Masselink et al., 2005; Masselink and Russell, 2006; Puleo, 2009). Total load fluxes are obtained via vertical integration of the instantaneous fluxes. Net sediment transport at a particular horizontal location is quantified via time integration of the total load sediment flux. Net sediment transport over some time duration can also be obtained using sediment traps (Horn and Mason, 1994; Masselink et al., 2009), sediment tracers (e.g., Masselink and Russell, 2006) or pump sampling, truncated foreshores (Alsina et al., 2009; Othman et al., 2014), or spatially and temporally dense morphology measurements (Blenkinsopp et al., 2011; Masselink et al., 2009; Weir et al., 2006).

3.1.1 Suspension measurements

The main approach for measuring suspended sediment concentration (SSC) in the swash zone uses optics (see online supplementary material). Optical Backscatter Sensors (OBS; D&A and Campbell Scientific; Downing, 2006) and Fiber Optical Backscatter Sensors (FOBS; custom built; University of Washington) sensors emit infrared light that is backscattered off of particles

suspended in the water column and received by a photodetector. Some researchers have developed their own optical sensors analogous to the FOBS (e.g. Miniature optical backscatter sensors; Butt and Russell, 1999 and subsequent work by these authors). The FOBS and its variation provide SSC measurements with as small as 0.01 m vertical resolution. Individual OBS can also be nearly co-located with vertical resolution near 0.01 m, but with a larger footprint within the flow field. Current meters or acoustic velocimeters (Section 2.2) provide the advection velocity that when multiplied by the concentration at each elevation yields the instantaneous suspended sediment flux.

Optical sensors must be calibrated using sediment within the water column where they are located. Calibration can be difficult for a variety of reasons. Sediment from the bed is usually collected and assumed to be the material that is in suspension. This assumption is generally valid unless the sediment distribution is poorly-sorted (well-graded), bi-(multi) modal, or contains a significant fines or shell fraction. Calibrations are typically performed in a re-circulating vessel. Only certain size fractions may be suspended under the natural conditions whereas laboratory calibrations may suspend all size classes yielding an overestimate of the true sediment concentration. For coarse sediments, the grains settle quickly making homogenous suspensions difficult to generate in the laboratory. Errors in suspension concentration manifest directly in the calibration. These errors can be circumvented if a viscous liquid, such as glycerol, (Butt et al., 2002) is used. Relating the optical qualities of the viscous liquid to those of water is needed in a second calibration step. Additionally, there is still some debate as to the effect of bubbles on optical sensors in the swash zone (Downing, 2006; Puleo et al., 2006) with bubble size likely a key factor. Puleo et al. (2006) suggest that bubble presence could render concentrations from optical backscatter sensors in error by 25 %. Still, optical sensors are probably the most

appropriate for use in the swash zone because they are rugged and can provide reasonable concentration estimates over a fairly wide range up to several hundred g l^{-1} . Acoustic sensors are not generally used to measure SSC in the swash zone due to signal attenuation via bubbles. However, new, acoustic sensors (multi-frequency acoustic concentration and velocity profiler; ACVP; Chassagneux and Hurther, 2014; Hurther et al., 2011) show promise in quantifying sediment concentration near the bed.

3.1.2 Bed load / sheet flow sediment measurements

Until recently, nearly every swash-zone sediment transport study focused on sediment suspension or transport derived from morphology or overtopping (e.g., Baldock et al., 2008b, 2005; Jensen et al., 2009; Mull and Ruggiero, 2014; Weir et al., 2006). A few older works (Horikawa et al., 1982; Yu et al., 1990) made some initial sheet flow measurements in the swash zone that provided the impetus for future work. Near bed sediment concentration is quantified using conductivity as a proxy. The two types of sensors used are the conductivity concentration meter (CCM/CCM+; Ribberink and Al-Salem, 1995; Sánchez-Arcilla et al., 2011; van der Zanden et al., 2015) and the conductivity concentration profiler (CCP; Lanckriet et al., 2013; Puleo et al., 2010; Table 1). The original CCM contains a single probe with 4 platinum electrodes in a horizontal array (electrode diameter = 0.3 mm; center-to-center spacing = 0.6 mm). The sensor operates by supplying an alternating current to the outer 2 electrodes and measuring the voltage across the inner 2 electrodes. The sensor measures at a single elevation and requires manipulation to raise or lower it in between laboratory wave runs. Thus, the measurement position within the sheet flow layer and relative to the initial at-rest bed level varies through time as the sheet flow layer thickness and bed level vary. Concentration profiles

were obtained by performing replicate runs of the same conditions under a laboratory setting by varying the vertical position of the sensor for each particular run. This approach is not feasible under irregular wave trains or under field conditions. The CCP was developed to obtain multiple measurements simultaneously across a vertical range. The CCP has 32 electrodes in a vertical array and uses the same conductivity measurement approach. Multiplexers are used to sequentially select individual sets of 4 electrodes to generate a sheet flow concentration profile at a specified time. The CCP can sample concentrations every 0.001 m in the vertical but only over a limited vertical span of 0.029 m. Multiple adjacent sensors can be used to increase vertical coverage (Lanckriet et al., 2014; Puleo et al., 2014a, in review; Ruessink et al., in review) under the assumption of alongshore uniformity over a short alongshore separation distance. The sensors cannot be adjusted vertically during wave action as they are deployed buried with only the measurement section extending above the at rest bed. Thus, if the active sheet flow layer is above or below the extent of the electrodes (through accretion or erosion respectively), no sheet flow concentration measurements are collected. The CCM+ partially circumvents this issue by having multiple CCM probes that are controlled via stepper motors contained in a large pressure housing buried below the bed. One of the CCMs is used to “find” a specified concentration and track it as a function of time with the other CCM probes adjusted vertically according to the required shift. The CCM+ can respond to bed level fluctuations on the time scale of wave groups, but not incident band forcing (van der Zanden et al., 2015). The CCM+ can measure concentrations at a few vertical elevations but does not yield a highly resolved profile unless measurements are collected under repeatable conditions. To our knowledge, the CCM+ has not yet been utilized in the field but has been used under large-scale laboratory conditions with success.

A benefit of the CCM and CCM+ is the high sampling rate capability (1000 Hz). The CCP cannot sample faster than 8 Hz (Lanckriet et al., 2013) and thus restricts its ability to quantify rapid fluctuations in the sheet flow concentration or use the signal for estimating velocity. High sample rates and adjacent sensors in the cross-shore direction enables cross-correlation of synchronized time series for quantifying velocities in the sheet flow layer (e.g. McLean et al., 2001). The cross-correlation approach is generally applied to ensemble averages since the individual cross-correlations can be contaminated by noise (McLean et al., 2001). Required use of ensemble averaging may render the approach difficult under irregular waves and/or field conditions. There are few other options available currently to quantify velocity in the sheet flow layer, but these velocities are critical if a flux estimate is to be made. Other options include quantitative imaging using a boroscopic technique (Cowen et al., 2009), imagery from a partially buried high speed camera or imagery through a flume side wall and possibly the ACVP (Chassagneux and Hurther, 2014; Hurther et al., 2011). The boroscope requires careful calibration as the imager field of view is sediment concentration dependent. However, the images may also provide concentration profiles and hence the sediment flux. The submerged camera housing can cause additional scour or if used with a long false wall may induce unnatural wave reflections. The ACVP may suffer from acoustic attenuation and has not yet been fully validated in the swash zone but has been shown to be robust in the surf zone. A potential benefit to the ACVP is the ability to capture sediment concentration and velocity at the same time using a single instrument. In short, much progress has been made in quantifying sheet flow concentrations. Obtaining sheet flow sediment flux estimates in the swash zone is still problematic.

3.2. *Sediment concentration and transport within a swash event*

Sediment concentrations, for a “free” swash event, spike when a particular location is inundated through a rapid rise in water level (Fig. 3f). Uprush suspended sediment concentrations can approach or exceed 100 g l^{-1} on steep and dissipative beaches (Aagaard and Hughes, 2006; Alsina et al., 2012; Alsina and Cáceres, 2011; Cáceres and Alsina, 2012; Hughes et al., 2007; Masselink et al., 2005; Masselink and Russell, 2006; Puleo, 2009; Puleo et al., 2014a) and may be more homogenous over the lower portion of the water column (Masselink et al., 2005; Puleo et al., 2014a, 2000, in review; Ruju et al., in review). Meanwhile, sheet flow concentrations increase rapidly as the bed is sheared and dilates (Lanckriet et al., 2013; Puleo et al., 2014a). Concentrations extend from the packed bed maximum or the enduring contact region typically defined as a volumetric concentration of 0.51 (1350 g l^{-1} ; Bagnold, 1966a) to a dilute suspension typically defined as a volumetric concentration of 0.08 (212 g l^{-1} ; Bagnold, 1966a). Sheet flow thickness may extend over several centimeters whereas sediment suspension tends to extend across the swash-zone water column. Without measurements, velocities in the sheet flow layer are assumed to vary linearly from a near bed velocity to zero at the immobile bed location (Puleo et al., 2014a, in review; Ruju et al., in review) similar to research suggesting the velocity decays to the $\frac{1}{2}$ (Wang and Yu, 2007) to $\frac{3}{4}$ (Sumer et al., 1996) power with distance into the sheet flow layer. Corresponding sediment transport rates also peak at uprush initiation (Fig. 3e). Instantaneous sediment fluxes in the sheet layer exceed those in the water column (Puleo et al., in review; Ruju et al., in review). Sheet load is quantified by vertical integration from the immobile bed to top of the sheet layer. Suspended load is quantified by vertical integration from the top of the sheet layer to the free surface or highest measurement location. The vertical range

of integration causes the suspended load transport to dominate sheet load transport during the majority of uprush (Puleo et al., in review; Ruju et al., in review).

The contribution of both transport modes is similar near flow reversal as sediment settles out of the water column and the sheet flow layer thickness decreases. Suspended sediment concentrations increase during backwash (Hughes et al., 2007; Masselink et al., 2005; Masselink and Russell, 2006; Puleo et al., in review; Ruju et al., in review) but generally to less magnitude than during uprush. Sheet flow sediment concentrations also increase during backwash (Puleo et al., in review; Ruju et al., in review) where the concentration profile is well-described by a linear profile with a power law tail (Lanckriet et al., 2014; O'Donoghue and Wright, 2004) regardless of the sheet flow thickness. The self-similarity of the sheet flow concentration profile may enable future efforts to utilize a near bed sediment concentration to infer the entire sheet flow profile if only the sheet flow thickness via multiple CCMs or CCM+ (van der Zanden et al., 2015) with individual probes separated for vertical coverage or from a predictive model (Dong et al., 2013; Lanckriet and Puleo, 2015; Malarkey et al., 2003; Wilson, 1987) is known. Sheet load transport exceeds suspended load transport rates throughout backwash (Puleo et al., in review; Ruju et al., in review). The importance of sheet load transport is likely manifest by the dominance of bed-related rather than surface-related turbulence (Section 2.7). Suspended load does not occur near the end of the backwash as the thinning and decelerating flow loses capacity to suspend sediment. Instead, near bed sediment transport likely occurs as bed load in the last stages of backwash before the event terminates.

The previous description of sediment transport was for a “free” swash event. Swash interaction on beaches is a common occurrence. The general description of the swash motion is similar but the level and type of interaction on the foreshore plays a role in the sediment suspension

processes (Blenkinsopp et al., 2011; Brocchini and Baldock, 2008; Cáceres and Alsina, 2012; Erikson et al., 2005; Hughes and Moseley, 2007; Jackson et al., 2004; Masselink et al., 2009). Irregular wave forcing causes irregular swash interactions and alters the surf-swash boundary condition (Brocchini and Baldock, 2008; Guard and Baldock, 2006; Pritchard and Hogg, 2005) for effects on varying the seaward swash zone boundary conditions). Cáceres and Alsina (2012) (Fig. 4) categorize interaction conditions as:

a) Swash overtake (capture) where a large (faster) bore overtakes the swash tip. An uprush has not had sufficient time to reach flow reversal before the subsequent bore is able to propagate to and overtake the original swash edge extending the uprush motion. It is expected that sediment in transport from the original uprush will be incorporated into the overtaking bore and be transported onshore.

b) Weak backwash / bore interaction where the bore overrides a preceding backwash. Backwash flow in these cases is weak and may have only been active for a short duration. Eventually the bore overtakes the backwash landward edge and initiates the ensuing uprush event. Large-scale laboratory results indicate weak backwash / bore interaction occurs most frequently when related to the “high” SSC events (Cáceres and Alsina, 2012) and generally experienced net onshore transport. These types of interactions are common in the transition zone of dissipative beaches where multiple incident band swash events within an infragravity scale motion have numerous uprushes with onshore-directed sediment transport (e.g., Masselink et al., 2005).

c) Strong backwash / bore interaction where the ensuing bore may be partially or fully slowed down or move offshore depending on flow conditions (Froude number; see also Elfrink and Baldock, 2002). Strong backwash / bore interaction could lead to hydraulic jumps that are

efficient sediment mobilizers (Butt and Russell, 2005). These strong events produce the largest instantaneous SSC signals and net offshore suspended sediment transport especially when occurring in the trough of infragravity waves (Cáceres and Alsina, 2012). High sediment concentrations result from flow separation under the bore and associated turbulence that effectively homogenizes the water column (Elfrink and Baldock, 2002). Strong backwash / bore interactions, like those for the weak interaction, are also common on dissipative beaches. In fact, a long duration backwash following multiple incident band weak bore interaction events leads to high SSC and strong interaction at the surf-swash boundary (e.g., Masselink et al., 2005).

3.3. Sediment advection and the importance of bore turbulence

Sediment transport rates are generally related to local conditions whereby the assumption of cross-shore advection is neglected. However, advection of sediment from the surf zone and or cross-shore advection due to the level of swash interactions (Section 3.2) can alter the sediment concentration at a particular location (Blenkinsopp et al., 2011; Cáceres and Alsina, 2012; Jackson et al., 2004; Masselink et al., 2009). Thus, both mechanisms alter actual sediment transport rates. The sediment arriving at the swash zone from seaward is sometimes referred to as “pre-suspended” sediment (Hughes et al., 2007; Pritchard and Hogg, 2005). Sediment trap data reveal that the maximum sediment transport load occurs just landward of the base of the swash zone for weak swash interactions and closer to mid swash locations for stronger swash interactions (Jackson et al., 2004). These data suggest that sediment at the inner surf / swash zone transition are carried through the swash zone and may not have a swash zone origin. Masselink and Russell (2006) came to a similar conclusion using field data on dissipative and reflective beaches where the beaches accreted even though the mean flow was offshore-directed.

924 The asymmetry of swash-zone velocities would seem to bias the net sediment flux offshore. This
925 overshadows strongly sediment transport predictions as the net flux is commonly related to the
926 mean velocity of the flow (e.g., Shields parameter) or to the velocity to some power (i.e.,
927 Bagnold type model). Some of the processes counteracting an offshore flux erosive tendency are
928 the advection of pre-suspended sediment into the swash zone and settling lag effects of the
929 sediment. Indeed, defining a settling lag parameter (e.g., ratio between the settling velocity of
930 sediments and flow velocity) in numerical predictions together with an initial sediment
931 concentration available for advection, can generate onshore net fluxes for a single swash event
932 (Pritchard, 2009; Pritchard and Hogg, 2005). The time-dependent suspended sediment supply at
933 bore arrival and just after is a critical factor that has not been thoroughly examined. There is
934 general agreement (Alsina et al., 2009; Brocchini and Baldock, 2008; Guard and Baldock, 2006;
935 Hsu and Hanes, 2004; Hsu and Raubenheimer, 2006; Pritchard, 2009; Pritchard and Hogg, 2005)
936 that the contribution of bore-related turbulence to sediment stirring capacity and sediment
937 advection is dominant and must be investigated further. However, this contribution is rarely
938 included in standard sediment transport models, and the turbulence field is absent or
939 parameterized in most numerical model approaches for swash-zone sediment transport (Bakhtyar
940 et al., 2009a).

941 Bore turbulence is expected to affect the bottom when the bore height is less than half the local
942 water depth (Hsu and Raubenheimer, 2006; Svendsen et al., 2000). It follows that the active zone
943 where the bore influence becomes relevant for sediment pickup and advection into the swash
944 zone may be defined by physical characteristics of the bore (e.g., bore height and momentum)
945 and a given (local) slope. The Pritchard and Hogg (2005) Lagrangian transformation of the
946 SM63 solution, naturally allows the inspection of water parcel trajectories (that may also

represent gross sediment motion) that are advected into the swash zone. Baldock et al. (2008a) propose an advection length that is bounded within 0.4 to 0.5 times the runup distance seawards of the bore collapse limit. The advection length defines the pickup region from which fluid parcels can be carried into the swash zone. The ratio obtained from the analysis of laboratory and field data holds regardless of the forcing conditions (Baldock et al., 2008a). Alsina et al. (2009) conducted a novel laboratory experiment to address the concept of pre-suspended sediment. They utilized a fixed beach slope with a mobile bed near the bore collapse point. Traps could be installed at different locations on the foreshore to collect the pre-suspended sediment and determine how that sediment distributes across the swash zone during uprush. They found that 25 % of the pre-suspended sediment reaches the mid-swash position further demonstrating the importance of non-local sediment transport processes in the swash zone. The experimental results were also used to test a NLSW solver with an advection-diffusion equation for turbulence and suspended sediment in a Lagrangian reference frame. The advection-diffusion equation includes a settling velocity and eddy viscosity (following Kobayashi and Johnson, 2001). A pre-suspended sediment concentration based on energy dissipation and stirring efficiency due to wave breaking and bed friction is supplied at the point of initial advection. The model although limited, (e.g. friction and turbulence efficiency factors are held constant), is able to predict an inflection point in the cross-shore distribution of sediment flux leading to beach accretion in the landward swash zone. Similar efforts are still needed in this regard under more resolved numerical approaches with less assumptions in the turbulence field (e.g. LES) or more appropriate parameterizations for the evolution of turbulence (e.g., rapid distortion theory vs. turbulent viscosity hypothesis, see molecular time scale responses for turbulent flows in Pope, 2000). Moreover, the numerical modeling of swash-swash interactions and bound wave

oscillations (e.g., surf beat) affecting the pre-suspended sediment loads is still an open topic for future research.

3.4. Sediment transport predictions

Predicting instantaneous or swash-event-integrated sediment transport has been attempted historically via energetics-type formulations (e.g., Bagnold, 1966b; Meyer-Peter and Müller, 1948). The energetics formulations predict the total, suspended or bed load transport that is generally related to the velocity to some power. In the case of a power of 3 the transport is related to a shear stress velocity product where the shear stress arises from the quadratic drag law. Multiple recent papers have shown that this approach has limited applicability in the swash zone (e.g., Aagaard and Hughes, 2006; Butt et al., 2005; Hughes et al., 2007; Masselink and Russell, 2006), although a few indicate moderate predictive capability (Masselink et al., 2009, 2005; Othman et al., 2014; Puleo et al., 2003b; some related to net sediment transport obtained via sediment traps or for ensemble-averaged events). For example, some studies have shown the simple formulations fail to even predict the net sediment transport direction correctly, let alone the magnitude (Holland and Puleo, 2001; Masselink and Russell, 2006). Other studies have shown the calibration coefficients between measured and predicted transport rates vary between uprush and backwash (Aagaard and Hughes, 2006) and discussed by (Masselink and Puleo, 2006) and may vary depending on cross-shore location within the swash zone (Aagaard and Hughes, 2006). On the one hand the inability to predict transport rates via the energetics approach is not totally surprising since the formulations were originally developed for steady, unidirectional flow; conditions not met in the swash zone. On the other hand, incomplete

992 measurements (coarse vertical resolution, lack of sheet flow measurements, incomplete temporal
993 sampling for an entire swash event due to instrument difficulties) may hamper true testing of
994 these simple formulations. The best validation may come from carefully controlled laboratory
995 studies with high-resolution (vertical, cross-shore and time) measurements of the velocity
996 profile, sheet flow transport, suspended sediment transport and morphological evolution.

997 A better representation of the processes controlling sediment transport can occur in more robust
998 process-based numerical models. However, the time scale of interest will determine the degree of
999 coupling (i.e., uncoupled, weakly or fully coupled) between hydrodynamic and sediment
1000 transport/morphological modules. The possibility for fully or weakly coupled approaches is
1001 suitable for individual swash events but the sediment transport estimates (Kelly and Dodd, 2010;
1002 Zhu and Dodd, 2013) may not differ from an uncoupled approach (Postacchini et al., 2014)
1003 given the short time scales involved. The improvements made from a fully coupled numerical
1004 approach become important under longer time scales. Recently, Zhu and Dodd (2015) developed
1005 a numerical model, coupling one-dimensional shallow water and bed evolution equations, with
1006 sediment advection (bed and/or suspended sediment load), to examine bed and suspended load
1007 transport. The numerical modelling of two swash events, a solitary wave and one event
1008 (overtopping wave) of Peregrine and Williams (2001), show the model capability to estimate
1009 swash-zone morphodynamics. The numerical solutions reveal that bed and suspended load have
1010 distinct morphodynamic signatures and that coupling both types of sediment transport does not
1011 affect significantly the solution. It is important to recall that these model results were based on
1012 idealized conditions where no flow interaction occurred and no bore turbulence effects were
1013 considered. Thus, more sediment would be entrained if these other processes were included.
1014 Other sediment transport models based on kinetic theory of collisional grain flow (e.g., Amoudry

et al., 2008) may be suitable for predicting swash-zone sediment dynamics where sediment transport often occurs under large bed shear stress and large sediment concentrations. Under these conditions the particle-particle interactions become important and strongly coupled with fluid kinematics (Bakhtyar et al., 2009a).

3.5 Modifications to simple transport formulations

Some studies acknowledged additional processes that are not included in the original energetics-type formulations that may enhance sediment transport. Differences between uprush and backwash calibration coefficients when calibrating the energetics-type formulations are more prevalent on steeper beaches. Bore turbulence (Aagaard and Hughes, 2006; Butt et al., 2004; Puleo et al., 2000) will enhance sediment transport and when included in the formulation as an additional mechanism increases the predictive ability (Aagaard and Hughes, 2006; Alsina et al., 2009). Turbulence itself is not a force and thus must be included in conjunction with other terms. For instance, Aagaard and Hughes (2006) included the vertical velocity that was assumed to be largely related to turbulent fluctuations in their sediment transport estimate. Their approach quantifies bed shear stress as a product between the cross-shore velocity and the vertical velocity rather than through the typically used cross-shore velocity squared. The approach by Aagaard and Hughes (2006) is similar to that by Masselink et al. (2005) in that sediment transport predictions improved if a shear stress parameterization other than the quadratic drag law was used. Estimating the bed shear stress from a logarithmic velocity profile also caused there to be little variability between uprush and backwash calibration coefficients (Masselink et al., 2005). Pedrozo-Acuña et al. (2006) found improvements to the predicted morphological response of a gravel beach, if friction factors dependent on flow phase are used. In their work however, the

variation of the transport efficiency coefficients also served to parameterize other mechanisms such as infiltration effects. Hence the relative importance of varying friction coefficients was obscured by other more dominant processes as demonstrated in Jamal et al. (2014). The inclusion of a simple infiltration model (e.g., Dodd et al., 2008; Packwood, 1983; Stoker and Dodd, 2006) resulted in significant improvements of the morphological response for the same case as Pedrozo-Acuña et al. (2006).

Butt et al. (2004) assumed the dissipated fluid power was related to turbulence and shear stress such that the sediment transport prediction was related to the cross-shore velocity times the sum of the shear stress and turbulence estimate. Although simple, the inclusion of turbulence in this manner improved the sediment transport prediction by 55 % when considering only bore events. However, more recent data from a large-scale laboratory study indicates that the suspended sediment concentration does not correlate well with TKE suggesting the empirical inclusion of TKE in swash-zone sediment transport formulations requires further study (Alsina and Cáceres, 2011).

Cross-shore pressure gradients, mostly associated with a sloping sea surface, improved sediment transport predictions for coarse sand but not fine sands (Othman et al., 2014). However, the transport coefficient was $O(10)$ times larger than previously suggested values in the literature indicating that this effect may really serve as a proxy for some other process. Local accelerations, possibly related to pressure gradients, have also been hypothesized as enhancing sediment transport in the swash zone (Nielsen, 2002; Puleo et al., 2007) showing some improvement in predictions. But, local onshore-directed accelerations occur for only a limited portion of uprush near the location of swash initiation (Puleo et al., 2007) and cannot be related

to the sloping sea surface pressure gradient that is seaward dipping for nearly the full duration of the swash event (Baldock and Hughes, 2006) as previously discussed (Section 2.4).

3.6. Total load from morphology

Section 3.2 described difficulties in quantifying sediment transport rates due to incomplete coverage throughout the water column. The sediment continuity equation can be used to infer net total load sediment transport rates across the swash zone from time-dependent morphology measurements (Alsina et al., 2012; Blenkinsopp et al., 2010b; Brocchini and Baldock, 2008; Masselink et al., 2010, 2009). UDMs have been used to determine the bed level in between swash events (e.g., Blenkinsopp et al., 2011; Houser and Barrett, 2010; Masselink et al., 2009; Turner et al., 2008). These sensors are appealing because they are inexpensive, robust, simple to use and can be deployed easily in an array to obtain cross-shore and alongshore variations in bed levels. LIDAR (Section 2.1) can also be used in this capacity to provide a dense cross-shore profile. Either data set, when differenced in time, yields a profile of the cross-shore gradient in sediment transport assuming alongshore uniformity. When measurements extend landward of the maximum runup region, the gradients are related to local total load sediment transport rates via discretization.

Masselink et al. (2009) and Blenkinsopp et al. (2011) used an array of UDMs across a high-energy beach to infer net sediment transport rates. Net sediment transport for a swash event could exceed 100 kg m^{-1} of beach. The inferred net sediment transport is only an order of magnitude smaller than the total sediment transport for the swash zone over an entire tidal cycle. Masselink et al. (2010) and Austin and Masselink (2006b) found similar results for a steep beach fronted by a beach step. The fact that an individual swash event can transport such a large

quantity of sediment relative to the total change over a tidal cycle is cause for concern given the current inability to predict well swash-zone sediment transport rates. Large errors in any available sediment transport formulation will cause the predicted morphology to diverge rapidly from the true morphology.

3.7. Morphologic variability

Puleo et al. (2014b) provides an overview of the different techniques to quantify bed level change. UDMs, CCPs and LIDAR were already identified in earlier sections. The most straightforward approach to quantify bed level changes is the use of stakes embedded in the foreshore and sampled at some specific time interval (Austin and Buscombe, 2008; Austin and Masselink, 2006b; Kulkarni et al., 2004). Time-sequenced imagery through stereo-metric intersection (Holland and Puleo, 2001) has been used to map foreshore elevations as identified through successive shoreline position digitization. Data were compiled to identify the foreshore morphology every 15 minutes since errors in geometrical image rectification and the procedure were not conducive to identifying the bed level change for a particular swash event. Laboratory swash-zone studies using stereo-metric approaches provide highly accurate and dense morphology measurements in between swash events (Astier et al., 2012; Astruc et al., 2012). This method is probably the most accurate of the non-intrusive methods with elevation errors < 1 mm. However, the technique requires careful instrument set up and imager positioning. It is not clear the technique can be used under field conditions and like other non-intrusive methods can only measure the morphology when the beach surface is exposed. Other approaches use a combined imaging/LIDAR system to monitor the foreshore elevation on a wave-by-wave time scale (Vousdoukas et al., 2014). The inclusion of LIDAR reduced image registration errors by an

1106 order of magnitude. The advent of UDMs, CCPs and LIDAR may indicate the imaging approach
1107 obsolete for short term field investigations of high-frequency bed level change in the swash zone.

1108 Section 3.6 discussed a method to extract sediment transport rates from dense swash-zone
1109 morphology measurements. There are also interesting signals in the morphologic variability
1110 obtained from these detailed measurements. On beaches with little to moderate net beach change
1111 the local bed level can fluctuate up (accretion) or down (erosion) by up to several centimeters for
1112 a particular swash event (Blenkinsopp et al., 2011; Houser and Barrett, 2010; Masselink et al.,
1113 2010, 2009; Puleo et al., 2014b; Ruessink et al., in review). The ability for a particular swash
1114 event to cause these changes means that perhaps only one or a few swash events are able to alter
1115 the final morphological configuration of the foreshore over a tidal cycle (Blenkinsopp et al.,
1116 2011; Cáceres and Alsina, 2012; Masselink et al., 2009; Puleo et al., 2014b).

1117 Blenkinsopp et al. (2011) analyzed swash-zone bed level variability using UDM data collected
1118 from a high energy steep beach. The foreshore accreted at a particular location by 60 mm over
1119 one tidal cycle. Roughly 60 % of the individual bed level changes for individual swash events
1120 were within 2 mm relative to the initial bed level. Over 90 % of the bed level changes for a
1121 particular event were within 10 mm of the initial bed level. Distributions of single swash event
1122 bed level change for all measured tides were found to be nearly Gaussian regardless of whether
1123 or not the beach exhibited net erosion or accretion. Puleo et al. (2014b) found similar nearly
1124 Gaussian distributions of individual swash event bed level variability on a high energy
1125 macrotidal beach. The net change over a tidal cycle in the Puleo et al. (2014b) study was less
1126 than 10 mm but some individual swash event bed level changes exceeded 20 mm erosion or
1127 accretion. Puleo et al. (2014b) further categorized individual swash events as having negligible
1128 or “large” (7 mm) net elevation change. Swash events that exhibited negligible net elevation

change had velocity time series that were symmetric with respect to uprush and backwash duration and uprush and backwash maximum cross-shore velocity. Large accretion events had a shorter uprush duration relative to backwash with weaker uprush vs. backwash cross-shore velocities. In contrast, large erosion events had longer uprush vs. backwash duration and stronger uprush velocities. These findings may be opposite of those expected. However, the cross-shore location where data were sampled likely plays a key role in overall bed level changes and how they may be related to underlying hydrodynamics (Aagaard and Hughes, 2006; Blenkinsopp et al., 2011; Houser and Barrett, 2010; Puleo et al., 2014b).

UDMs sample the bed level in between swash events when the bed becomes exposed. CCPs have the ability to also sample the bed level at high resolution during a swash event providing insight into intra-swash bed level variability (Puleo et al., 2014b; Ruessink et al., in review). Instantaneous bed level elevations were nearly normally-distributed similar to findings for inter-swash bed level changes. Spectral calculations of the bed level time series showed no significant peaks at the same frequencies as the peaks for incident or infragravity hydrodynamic parameters. The spectrum displayed an f^{-2} roll off across 2 decades of frequency with no inertial subrange. The f^{-2} roll off corroborates the roughly Gaussian distribution of bed level change observed in this and the other studies of bed level change using UDMs cited earlier.

The foreshore variability also has an interplay with the beach step dynamics (Austin and Buscombe, 2008; Austin and Masselink, 2006b; Kulkarni et al., 2004; Masselink et al., 2010). The step is more prominent on the rising tide and migrates landward as the tide rises. Its amplitude decreases and it migrates offshore during the falling tide. More detailed analyses suggest the step and foreshore evolve in different phases. Kulkarni et al. (2004) indicated an initial erosion phase up the foreshore upon tidal inundation before the foreshore accretes

1152 throughout the rest of the rising tide. There is then slight erosion with offshore step migration
1153 while the rest of the foreshore remained relatively stable. The different phases were tied largely
1154 to water table fluctuation in the coarse-grained beach. Austin and Buscombe (2008) indicated the
1155 initial change on the rising tide is step growth and foreshore erosion. The step continues to
1156 increase in volume with rising tide while the foreshore region experienced little to no change in
1157 volume. At high tide, the step and foreshore are in near equilibrium with the waves and
1158 experience little change in volume with some slight foreshore accretion. Both the step and
1159 foreshore experience a decrease in volume as the tide levels drops. The study also indicated that
1160 grain size distribution at a particular location might change on the wave time scale. The grain
1161 size variability could enhance or hinder sediment transport for similar forcing conditions
1162 suggesting that knowledge of the time-dependent sediment distribution may be needed (e.g.
1163 friction factor) for accurate sediment transport predictions as was discussed in Section 3.4.

1165 **4. Summary**

1166 In the last decade, numerous research efforts have expanded our knowledge of small-scale
1167 swash-zone hydrodynamics and sediment transport processes. Recent descriptions of the
1168 complex and dynamic system have been possible thanks to advances in sensor technology and
1169 novel approaches in field, laboratory and numerical modeling studies. This review paper
1170 discussed recent advances that have occurred but there are still numerous knowledge gaps, for
1171 example:

- 1172 1) Advances have been made regarding sediment concentration specifically in direct
1173 proximity of the bed. Sediment flux estimates near the bed and in the water column are

less advanced due to coarse resolution or general lack of ability to quantify velocity in the sheet flow layer. Little advancement in predicting swash-zone sediment transport from simple formulations has occurred. This aspect of swash-zone research is viewed by the authors as one of the most pressing problems facing swash-zone researchers.

2) The approach of reducing the temporal scale to a single swash event (e.g. dam break) to focus on processes occurring on a wave-by-wave time scale has helped highlight the important processes occurring in the swash zone. However, less attention has been paid to the inherent interactions that take place near the surf-swash transition or through swash-swash interaction.

3) The majority of swash-zone research has focused on cross-shore processes including sediment transport, bed shear stress and turbulence. There has been limited progress on how alongshore flows affect these processes.

4) Recent research has focused generally on small-scale processes over short temporal durations. Upscaling understanding/predictions of these small-scale processes to longer term events such as storms is a major challenge.

5) Swash-zone numerical models span a broad range in degree of sophistication (e.g., depth-averaged, depth-averaged boundary layer, depth-resolving, LES). Selecting a numerical model may be more related to the subject of interest, time constraints and computational effort rather than to model limitations. The common numerical models have been validated and may be used to assist in understanding hydrodynamic parameters. However, the prediction of net sediment transport fluxes and morphological response are still based on limited formulations. Potential improvements may come from fully (or weakly) coupled approaches but have yet to be tested for longer time scales and/or more

1197 realistic settings. Few depth-resolving models have focused on sediment transport
1198 prediction and their contribution may be limited to short time scales (several swash
1199 events).

1200

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Table 1 CCM and CCP capabilities.

Sensor	Profile or repeatable conditions (RC) needed	Vertical profiling range (mm)	Max sample rate (Hz)	Adjustable during wave activity	Vertical sample volume (mm)	Velocity capability
CCM	RC	0	1000	No	1 to 2	Yes if paired
CCP	Profile	29	8	No	~ 4	No
CCM+	RC	280	1000	Yes	1 to 2	Yes

Fig. 1. Schematic of the cross-shore beach profile showing surf and swash zones.

Fig. 2. Time history of the number of papers published (black circles) and the sum of the citations for papers published within a given year (gray squares). Source for data: Google Scholar (scholar.google.com) using the search terms: swash, uprush, backwash, backrush, foreshore, runup.

Fig. 3. Schematic of hydrodynamic and sediment transport processes occurring within a single “free” swash event. The gray area depicts the duration of *in situ* measurements generally acquired in field and laboratory studies. The line pattern area depicts the duration of measurements acquired with numerical approaches or a few other techniques (*in situ* or remote) that can sample the entire swash duration.

Fig. 4. Schematic of interactions in the swash zone: a) swash overtake where a bore initializes a new swash event before the preceding swash event has experienced backwash; b) weak backwash / bore interaction where the bore propagates shoreward largely unimpeded by the preceding backwash; c) strong backwash / bore interaction where a rapid backwash flow collides with an ensuing bore possibly slowing the bore or causing a hydraulic jump. The thin black curve is the foreshore. The dotted line is the still water level. The thick black curve is the initial swash event. The thick gray curve is the ensuing bore motion. Arrows indicate the direction and scale of expected sediment flux. Adapted from Cáceres and Alsina (2012).

Figure 1

Figure 2

Figure 3

Figure 4

a) Swash overtime

b) Weak backwash / bore interaction

c) Strong backwash / bore interaction