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Beach groundwater dynamics

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Abstract

An understanding of the interaction between surface and groundwater flows in the swash zone is necessary to understand beach profile evolution. Coastal researchers have recognized the importance of beach watertable and swash interaction to accretion and erosion above the still water level (SWL), but the exact nature of the relationship between swash flows, beach watertable flow and cross-shore sediment transport is not fully understood. This paper reviews research on beach groundwater dynamics and identifies research questions which will need to be answered before swash zone sediment transport can be successfully modelled. After defining the principal terms relating to beach groundwater, the behavior, measurement and modelling of beach groundwater dynamics is described. Research questions related to the mechanisms of surface–subsurface flow interaction are reviewed, particularly infiltration, exfiltration and fluidisation. The implications of these mechanisms for sediment transport are discussed.

Keywords: Groundwater; Infiltration; Phreatic; Swash; Vadose; Watertable

1. Introduction

The section of beach above the still water level is perhaps the area of the nearshore environment about which least is known, yet it is of critical importance because it is the section of the foreshore where final wave energy dissipation occurs. Hughes and Turner (1999) identified five features of the swash zone that make it morphologically different from the rest of the beach. First, it involves a moving land–water boundary that travels across the intertidal beach at a range of frequencies from incident waves to tides. The second feature is that water depths in the swash zone are very shallow, particularly in the backwash where they are typically less than 5 cm. The nature of the hydrodynamic processes operating in the swash zone, particularly the small water depths, and high velocity, rapidly oscillating bi-directional currents, means that accurate measurement of swash processes can be difficult. Interference between the instruments and the very shallow flows can potentially result in measurement error of as much as 10% of the flow (Hughes, 1992). Third, these very shallow water depths mean that sediment in the swash zone is often transported as complicated, single-phase, granular-fluid flows rather than as the better-understood two-phase flows where the fluid and sediment in transport are readily distinguished. A fourth feature of the swash zone is that at least some part of the beachface is usually unsaturated, which means that infiltration into the beach and exfiltration from the watertable can play an important role in sediment transport. Recent

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research suggests that internal flow within the beach driven by hydraulic gradients due to watertable fluctuations and/or swash infiltration/exfiltration may influence the bed stability and thus sediment transport (see Section 4). Finally, Puleo et al. (2000) described several fundamental differences between the swash zone and the surf zone, pointing out that swash is fundamentally Lagrangian, so that fixed instruments will record discontinuous time series as they are alternately submerged and exposed. The discontinuous nature of swash zone processes means that concepts like wave period can become complicated by the fact that it is probably duration of fluid coverage rather than repeat time scale that may be most influential on sediment transport.

This paper reviews research on beach groundwater dynamics and identifies research questions that still require answers before swash zone sediment transport can be successfully modelled. The principal terms relating to beach groundwater are defined first. The behavior, measurement, and modelling of beach groundwater dynamics are then described. Research questions related to the mechanisms of surface-subsurface flow interaction are reviewed. Particularly attention is paid to infiltration, exfiltration and fluidisation. The implications of these mechanisms for sediment transport are discussed.

2. Background

Swash zone and beach groundwater processes are of interest to geomorphologists who wish to determine beach erosion or deposition or aeolian sediment transport, to marine biologists who are interested in intertidal fauna, and to engineers who require data on run-up, particularly on coastal structures such as breakwaters. Over the past few decades, data on the position of the shoreline, which is directly dependent on swash zone processes, have emerged as one of the principal sources of information for monitoring coastal change (USGS, 1998). In some cases, the shoreline position (identified as the maximum extent of run-up) is used to establish legal boundaries, setback lines or flood hazard zones (Morton and Speed, 1998).

Marine biologists have an interest in the swash zone (e.g., Jansson, 1967; Pollock and Humman, 1971; Reidl and Machan, 1972; McLachlan, 1989, 1996; McLachlan and Hesp, 1984; McLachlan et al., 1985; McArdle and McLachlan, 1991, 1992; Defeo et al., 1997), because the distribution and type of macrofauna inhabiting the intertidal zone of sandy beaches appears to be related to the swash climate. Both the interstitial fauna and the macrofauna of sandy beaches are directly affected by swash and groundwater processes: the former by infiltration, which is responsible for flushing oxygen and organic materials into the sand; and the latter by swash dynamics and the position of the seepage face, which influence tidal migrations and burrowing (McArdle and McLachlan, 1991). Differences in the spatial distribution and abundance of beach fauna have been explained in relation to sediment size and beach slope, but have not yet been related successfully to swash dynamics (McLachlan and Hesp, 1984).

Coastal engineers have long recognized the need for a better understanding of swash zone processes, largely concentrating on the measurement and modelling of run-up on structures such as breakwaters. Such studies are needed to establish design criteria, particularly the elevation of the structure required to prevent overtopping by the run-up of extreme waves. Other engineering applications where knowledge of beach groundwater dynamics is important include water quality management in closed coastal lakes and lagoons and the operation of water supply and sewage waste disposal facilities in coastal dunes (Turner, 1998), contaminant cycling in estuaries (Drabsch et al., 1999) and coastal water resource management issues such as salt water intrusion into coastal aquifers, wastewater disposal from coastal developments and pollution control (Nielsen, 1999). Recently the commercial possibility of modifying beach watertable elevation to control beach erosion has been recognized, and several studies have investigated the use of beach dewatering as an alternative to hard engineering practices. Beach dewatering involves lowering the watertable artificially through a system of buried drains and pumps. Although beach dewatering systems have been shown in some cases to be effective in inducing accretion and/or reducing erosion, the mechanism responsible is not clear (Turner and Leatherman, 1997). A beach drain system is generally thought to operate by reducing backwash sediment transport by lowering the beach watertable and thus increasing infiltration. However, there is some evidence to suggest that flow generated by the drain system may produce additional onshore transport and onshore bar migration (Oh and Dean, 1994; Sato et al., 1994). In order to operate a beach dewatering system in a cost-effective manner, coastal engineers need to know whether the drain enhances the rate of accretion during normally accretive wave climates or if the drain mitigates the effect of an erosive wave climate (Weisman et al., 1995). At present, engineers do not have sufficient physical understanding of the system to enable them to design the optimum system (Li et al., 1996). The effect of the location of the drain on the performance of the system is not clearly understood, in particular the depth below and the distance behind the still water line; nor is the dependence of system performance on discharge known (Sato et al., 1994; Weisman et al., 1995). Other sources of uncertainty include the effects of tidal range and stage of the tidal cycle, sediment size and sorting, beach slope, and the direction and frequency of storm events. Until these effects can be understood and quantified, it will not be possible to predict the performance, or even the success of a beach dewatering scheme, despite the potential of this 'soft' engineering technique. At present, the outcome of many beach dewatering schemes is inconclusive as to whether dewatering has had any net positive effect in mitigating local erosion problems (Turner and Leatherman, 1997).

An understanding of swash and beach groundwater dynamics is also important in the modelling of beach profile evolution. At present, most numerical models of shoreline change do not include sediment transport processes in the swash zone. This failure to model the swash zone means that sediment transport at the coastline will not be adequately represented. Cross-shore sediment transport models (e.g., Roelvink and Stive, 1989; Nairn and Southgate, 1993; Roelvink and Brøker, 1993; Larson, 1995; Winyu and Shibayama, 1996, etc.) have demonstrated considerable success in predicting eroding beach profiles on relatively fine sand beaches. Predictions of accretionary events and the behavior of coarser sediment beaches are generally not as good, particularly in the inner surf zone and swash zone (Seymour, 1986; Schoones and Theron, 1995). Existing cross-shore sediment transport models generally predict only offshore transport unless there is substantial tuning of the calibration coefficients (Hughes et al., 1997b; Cox and Hobensack, in review). Even under predominantly erosive conditions, the observed net transport in the swash zone can be onshore (Watanabe et al., 1980). This failure to predict beach profile behavior adequately is probably due at least in parts, to the lack of a realistic model for the hydrodynamics and sediment transport in the swash zone, as most models neglect or drastically simplify the swash hydrodynamics (Hamm et al., 1993; Roelvink and Brøker, 1993; Shah and Kamphuis, 1996). However, even recent models which include swash zone processes in some form do not predict accretionary features such as the berm correctly. Kobayashi and da Silva (1987) developed a time-dependent model of sediment particle motion in the swash zone that was successful in predicting run-up oscillations in the lab and field. This model was not able to predict net onshore transport beyond the still water level, possibly due to permeability effects (Cox and Hobensack, in review). Kawata and Kimura (2000) developed a numerical model of beach profile evolution including the swash zone in which the direction of transport predicted for the swash zone was the opposite of the transport direction predicted for the region offshore. However, the model still predicted overall net erosion, even with the swash zone included. The lack of detailed knowledge of swash and beach groundwater dynamics is probably an important factor in the ability of profile models to simulate accretionary events accurately, because an accretionary event is defined by the deposition of sediment above mean sea level.

Erosion and accretion of the beach profile, and the resulting movement of the position of the shoreline, are a direct result of sediment transport processes occurring in the swash zone and inner surf zone. Beach groundwater–swash dynamics provide an important control on swash zone sediment transport, which affects the morphology of the intertidal beach by controlling the potential for offshore transport or onshore sediment transport and deposition above the still water level. Cyclic erosion and accretion of the beachface as a result of relative elevations of the beach watertable and swash have been substantiated by researchers for many years (e.g., Bagnold, 1940; Shepard and LaFond, 1940; Emery and Foster, 1948; Longuet-Higgins and Parkin, 1962; Duncan, 1964;

Otvos, 1965; Strahler, 1966; Schwartz, 1967; Harrison, 1969, 1972; Waddell, 1976; Chappell et al., 1979; Kirk, 1980; Clarke et al., 1984; Eliot and Clarke, 1986, 1988; Nordstrom and Jackson, 1990; Turner, 1990; Ogden and Weisman, 1991; Turner, 1993c, 1995a; Oh and Dean, 1994; Weisman et al., 1995). Most of these studies suggest that beaches with a low watertable tend to accrete and beaches with a high watertable tend to erode. Recent observations indicate that flows in the swash zone can also affect the beach profile seaward of the intertidal profile, influencing sediment transport in the bar region (Oh and Dean, 1994; Sato et al., 1994).

Recent studies also indicate that longshore sediment transport is closely related to the hydrodynamic motion at the two boundaries of the surf zone—the break point and the swash zone, suggesting that correct estimates of mass and momentum fluxes inside the swash zone are important for predicting longshore transport (Brocchini and Peregrine, 1996; Wang, 1998). As with cross-shore sediment transport models, in most longshore sediment transport models the swash zone transport contribution is either completely ignored or included as part of the total sediment budget by applying a calibration coefficient to the transport model to allow for transport in the swash zone (van Wellen et al., 2000). This is likely to introduce significant error, as studies have indicated that the amount of longshore sediment transport in the swash zone may be equal to or greater than that in the surf zone (Bodge and Dean, 1987; Kamphuis, 1991). Modelling work by Van Wellen et al. (2000) suggested that 50–70% of total longshore sediment transport on a steep gravel beach occurred in the swash zone.

Swash and beach groundwater interaction may play a particularly important role in profile evolution and sedimentation patterns on macrotidal beaches. Many macrotidal beaches have two, and sometimes three, distinct beach zones: a flat, dissipative low-tide beach and a steeper, more reflective high-tide beach (Wright et al., 1982; Jago and Hardisty, 1984; Short, 1991; Horn, 1993; Masselink, 1993; Masselink and Short, 1993; Turner, 1993a,c; Masselink and Hegge, 1995). There is generally an abrupt decrease in beach slope on macrotidal beaches where the watertable intersects the beachface (Dyer, 1986; Turner, 1993c), which may be also marked by a change in sediment size between coarse and fine material (Carter and Orford, 1993). Turner (1995a) developed a simple numerical model that incorporated the interaction of the tide and the beach watertable outcrop. This model predicted the development of a break in slope resulting from landward sediment transport and berm development across the alternately saturated and unsaturated upper beach, while the profile lowered and widened across the saturated lower beach. Masselink and Turner (1999) concluded that macrotidal beach profiles could be divided into two distinct morphological domains: an upper intertidal region that is alternately saturated/unsaturated through the tide cycle, and a lower region within the intertidal profile that remains in a permanently saturated state. Hughes and Turner (1999) gave different empirical equations for equilibrium slope on unsaturated and saturated beachfaces.

Beach groundwater–swash interaction is also likely to play a role in sediment sorting processes. Carter and Orford (1993) suggested that the interaction between beach groundwater and swash flows may provide a mechanism for the shore-normal sorting of coarse and fine material that is often observed on macrotidal beaches, with dissipative, sandy low-tide terraces at the base of steep, reflective high-tide gravel ridges. Turner (1993c) surveyed 15 macrotidal beaches on the Queensland coast in Australia and found that a decrease in sediment size was strongly correlated to an increase in the relative extent of the lower gradient (saturated) lower region of the intertidal profile. Masselink and Turner (1999) suggested that the intersection of the saturated lower slope and intermittently unsaturated upper slope marks a point of divergent sediment transport, which may be reinforced by the selective sorting of coarser sediment upslope and finer sediment downslope. Hughes et al. (2000) suggested that infiltration might alter the critical entrainment stresses that contribute to heavy mineral sorting in the swash zone.

Common to all these studies is the observation that when the watertable outcrops above the tide, two zones are distinguished: a lower saturated zone that promotes downslope (offshore) sediment transport, and an upper region that alternates between saturated and unsaturated conditions, with upslope (onshore) sediment transport potentially enhanced by infiltration. However, the relative importance of infiltration is not yet known, and will be discussed further in Section 4.

3. Beach groundwater

3.1. Definitions

The beach groundwater system is a highly dynamic, shallow, unconfined aquifer in which flows are driven though saturated and unsaturated sediments by tides, waves and swash, and to a lesser extent by atmospheric exchanges (evaporation and rainfall) and exchanges with deeper aquifers. The complex interaction of surface and subsurface water in the swash zone means that it is useful to define the terminology (see Fig. 1). The still water level (SWL) is the water surface in the hypothetical situation of no waves. When the local water surface elevation is averaged over a time span much longer than incident and infragravity periods but shorter than the tidal period, the result is the local mean water level, which traces the mean water surface (MWS) (Nielsen, 1988). The mean water surface in the surf and swash zones generally has a gradient which balances the change in the radiation stress (Longuet-Higgins and Stewart, 1962, 1964). Changes in radiation stress are balanced by changes in hydrostatic pressure, in other words, by changes in water level. This difference is known as set-up or set-down. Set-up is a wave-induced increase in the MWS, whereas set-down is a wave-induced decrease in the MWS. Set-down occurs seaward of the breakers, where radiation stress is at its maximum. The positive gradient due to radiation stress is balanced by a negative water surface gradient, resulting in a lowering of the MWS to below SWL. Set-up occurs inside the surf zone, where the decrease in radiation stress due to energy dissipation is balanced by the raising of the MWS above SWL. As long as energy dissipation continues, set-up continues to increase in the onshore direction and is greatest at the shoreline.

The shoreline is the position where the MWS (including the set-up) intersects the beachface; in other words, the line of zero water depth. The shoreline represents the land-water boundary and the limits of shoreline excursion define the boundaries of the swash zone, which migrates up and down the foreshore of the beach over a tidal cycle. The swash zone migrates up and down the foreshore of the beach over a tidal cycle. The seaward and landward limits of the swash zone are, respectively, the point of collapse of the wave or bore as

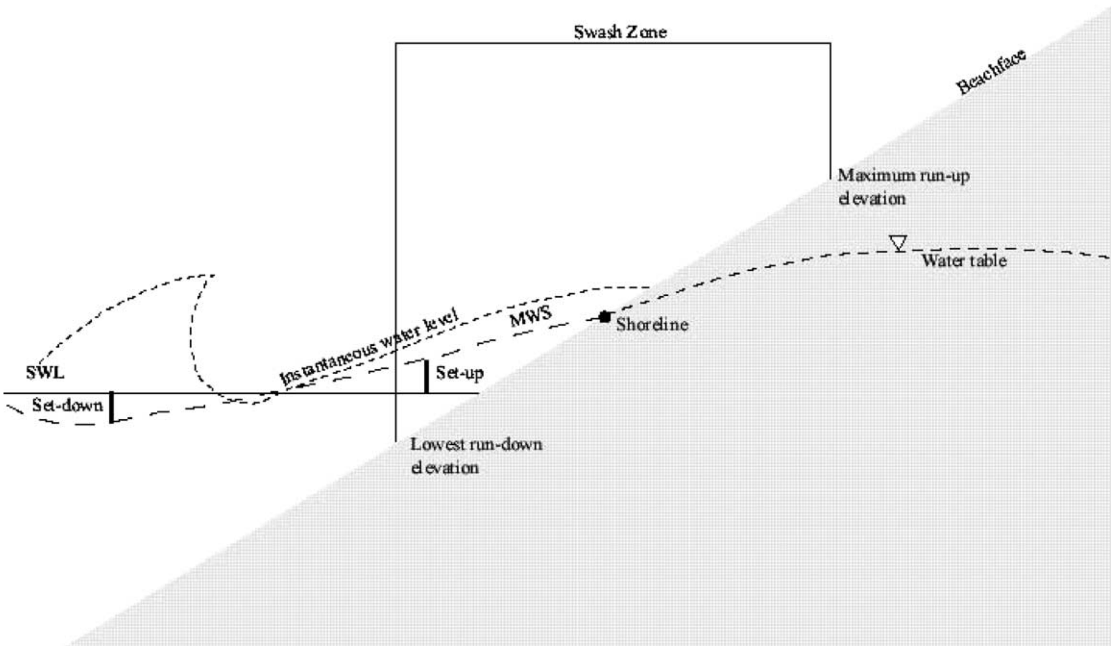

Fig. 1. Definition sketch of surface and subsurface water levels in the swash zone.

it reaches the shoreline, and the landward limit of wave action. There are two components to the water motions in the swash zone. The first is uprush, the landward-directed flow; the second component is the backwash, the downslope movement of the water which follows maximum run-up. The swash cycle is essentially an oscillation superimposed on the maximum MWS (including set-up) inside the surf zone. Total wave run-up represents the combined effect of set-up and swash at incident and infragravity frequencies. The maximum swash height, or maximum run-up elevation, is the maximum vertical height above SWL. Wave run-down elevation is the lowest vertical height reached by the backwash. The run-down elevation may be below SWL.

The beach watertable is generally considered to be the continuation of the MWS inside the beach, however, a more physically correct definition of the watertable is an equilibrium surface at which pore water pressure is equal to atmospheric pressure. The watertable is also referred to as the phreatic surface. Pore water pressure is the fluid pressure in the pores of a porous medium relative to atmospheric pressure. Below the watertable, pore water pressure is greater than atmospheric pressure; above the watertable, pore water pressure is less than atmospheric pressure. Hydrologists generally use the terms groundwater to refer to water below the watertable, and soil water to describe water above the watertable, where pore water pressures are negative (sub-atmospheric). However, to equate beach sediment with a soil would be misleading, so in beach hydrology the term groundwater is commonly used to mean any water held in the sand below the beach surface. The phreatic zone is the permanently saturated zone beneath the watertable (see Fig. 2). The vadose zone, also called the zone of aeration or the unsaturated zone, is the region of a beach sand body extending from the watertable to the sand surface. In the phreatic zone, pore spaces are filled with water and pore water pressures are equal to or greater than atmospheric pressure. In the zone of aeration, the pores are filled with both water and air and pore water pressures are less than atmospheric. For this reason, beach groundwater zones are better defined by pore water pressure distribution than by saturation levels. A capillary fringe develops immediately above the watertable as a result of the force of mutual attraction between water molecules and the molecular attraction between water and the surrounding sand matrix (Price, 1985). The capillary fringe may also be referred to as the tension-saturated zone. (Groundwater hydrologists often use the terms tension or suction—which can be

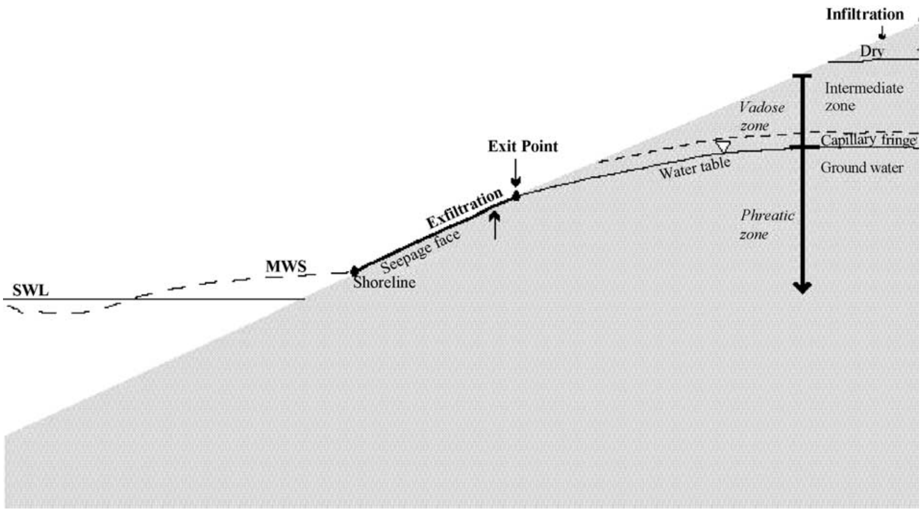

Fig. 2. Definition sketch of beach ground water zones when the water table is decoupled from the tide.

used interchangeably—to describe a pressure which is negative relative to atmospheric pressure.). In the capillary fringe, pore spaces are fully saturated, but the capillary fringe is distinguished from the watertable by the fact that pore water pressures are negative. The thickness of the capillary fringe in sand beaches may vary between a few millimeters to nearly a meter, and it may extend to the sand surface. Some workers (e.g., Turner, 1993b) also refer to an intermediate zone which may occur above the capillary zone where the degree of saturation may vary, but remains less than 100%.

3.2. Behavior of beach groundwater

A few studies have looked at groundwater movement in coastal barriers (e.g., Nielsen and Kang, 1995; Kang and Nielsen, 1996; Turner et al., 1997; Nielsen and Voisey, 1998; Nielsen, 1999), watertable fluctuations in estuarine environments (e.g., Nutttle and Hemond, 1988; Drabsch et al., 1999; Li et al., 1999) and gravel beaches (Ericksen, 1970; Carter and Orford, 1993) or sand moisture content effects on aeolian sediment transport (Jackson and Nordstrom, 1997; Nordstrom et al., 1996; Sarre, 1989; Sherman and Lyons, 1994; Sherman et al., 1998). However, most studies of coastal groundwater dynamics have concentrated on groundwater in beaches, and in particular, in the swash zone of sandy beaches.

Turner and Nielsen (1997) identified a number of mechanisms which have been associated with observed beach watertable oscillations: seasonal variations (e.g., Clarke and Eliot, 1983, 1987); barometric pressure changes associated with the passage of weather systems and storm events (e.g., Lanyon et al., 1982a; Eliot and Clarke, 1986; Turner et al., 1997); propagation of shelf waves (e.g., Lanyon et al., 1982a); and infragravity and incident waves (e.g., Waddell, 1973, 1976, 1980; Lewandowski and Zeidler, 1978; Hegge and Masselink, 1991; Kang et al., 1994b; Turner and Nielsen, 1997; Turner and Masselink, 1998). However, the majority of research has concentrated on tide-induced fluctuations of the beach watertable.

A number of studies since the 1940s have described the shape and elevation of the beach water table as a function of beach morphology and tidal state. The majority of these studies have been limited to measurements of watertable elevations across the beach profile, although Lanyon et al. (1982b) reported some limited observations of both longshore and cross-shore groundwater variations. The elevation of the beach water table depends on prevailing hydrodynamic conditions such as tidal elevation, wave run-up and rainfall, and characteristics of the beach sediment that determine hydraulic conductivity, such as sediment size, sediment shape, sediment size sorting, and porosity (Gourlay, 1992). Observations of beach watertable behavior show that the watertable surface is generally not flat. Several authors have showed that the slope of the watertable changes with the tide, sloping seaward on a falling tide and landward on a rising tide (e.g., Emery and Foster, 1948; Emery and Gale, 1951; Lanyon et al., 1982a,b; Turner, 1993a,b; Raubenheimer et al., 1998). The slope of the water surface has been found to be steeper on a rising tide than on a falling tide (Lanyon et al., 1982b). Other researchers have measured watertable elevations with a humped shape, with the hump near the run-up limit (Baird et al., 1998; Nielsen, 1999). Watertable oscillations have also been shown to lag behind tidal oscillations (e.g., Emery and Foster, 1948; Isaacs and Bascom, 1949; Pollock and Hummon, 1971; Lanyon et al., 1982a,b; Eliot and Clarke, 1986; Waddell, 1976, 1980; Lewandowski and Zeidler, 1978; Kang et al., 1994a; Nielsen and Kang, 1995). Observed watertable elevations are asymmetrical, as the watertable rises abruptly and drops off slowly compared to the near-sinusoidal tide which drives it. For a given geometry, the lag in watertable response is due mainly to the hydraulic conductivity of the beach sediment (Nielsen, 1990). With increasing distance landward, the lag between the watertable and the tide increases and the amplitude of the watertable oscillations decreases (Emery and Foster, 1948; Ericksen, 1970; Nielsen, 1990; Hegge and Masselink, 1991). However, Raubenheimer et al. (1999) found that fluctuations at spring-neap frequencies are attenuated less than fluctuations at diurnal or semi-diurnal frequencies. Wave run-up, tidal variation and rainfall may produce a super-elevation, or overheight, of the beach watertable above the elevation of the tide (e.g., Neilsen et al., 1988; Kang et al., 1994b; Turner et al., 1997).

Emery and Gale (1951) were among the first to recognise that the beach acts as a filter that only allows the larger or longer period swashes to pass. Both the amplitude and the frequency of the groundwater spectrum decrease in the landward direction. The further landwards the given groundwater spectrum, the narrower its band and the more it is shifted towards lower frequencies (Lewandowski and Zeidler, 1978). Waddell (1973, 1976) and Hegge and Masselink (1991) also showed that the beach acts to reduce the amplitude and frequency of the input swash energy. Based on observations such as these, the beach has often been described as a low-pass filter (meaning that only lower frequency oscillations are transmitted through the beach matrix). High frequency, small waves are damped and their effect is limited to the immediate vicinity of the intertidal beachface slope, whereas low frequency waves can propagate inland. Comparison of run-up and groundwater spectra shows a considerable reduction in dominant energy and also a shift in dominant energy towards lower frequencies (Hegge and Masselink, 1991).

Decoupling between the tide and the beach watertable occurs when the groundwater exit point becomes separated from the shoreline (shown in Fig. 2). This occurs because the rate at which the beach drains is less than the rate at which the tide falls, so the tidal elevation generally drops more rapidly than the watertable elevation and decoupling occurs, with the watertable elevation higher than the tidal elevation. The exit point is the position on the beach profile where the decoupled watertable intersects the beachface. After decoupling occurs, the position of the exit point is independent of the MWS until it is overtopped by the rising tide. Below the exit point, a seepage face develops where the watertable coincides with the beachface. The seepage face is distinguished by a glassy surface. The seepage face is different from the watertable in that its shape is determined by beach topography. However, water on the seepage face is at atmospheric pressure, as is water on the watertable. The extent of the seepage face depends on the tidal regime, the hydraulic properties of the beach sediment, and the geometry of the beachface. Thus, the degree of asymmetry in watertable response will vary between beaches. This asymmetry is due mainly to the fact that a beach tends to fill more easily than it drains. At high tide, a greater area of the beach is available for water to flow into the beach than at low tide, when the area of the beach from which groundwater flows is defined by the length of the beach under water (below the tide) and the extent of the seepage face. No matter how extensive, this length will always be less than the saturated area of beach at high water. The exit point is generally assumed to mark the boundary between a lower section of the beach which is saturated and an upper section which is unsaturated. However, this assumption is probably an oversimplification.

3.3. Measurement of beach groundwater

Techniques for the measurement of beach groundwater are discussed by Baird and Horn (1996) and Turner (1998), and the concepts summarised here are described more fully in these papers. Until quite recently, most investigations of beach groundwater behavior have concentrated on the measurement of beach watertables in response to low frequency tidal forcing. More recent studies such as Turner and Nielsen (1997), Horn et al. (1998) and Turner and Masselink (1998) have measured higher frequency fluctuations due to waves. The choice of monitoring system depends on the objectives of the monitoring program. The most important parameters in a beach groundwater system are the elevation of the beach watertable, pore water pressures, hydraulic conductivity, specific yield and moisture content.

The elevation of the beach watertable can be measured by using wells which are perforated for their entire length to allow water to flow freely between the sediment and the well at all depths (e.g., Baird et al., 1998). These wells are commonly made of PVC pipe and should be covered with a porous material to prevent sand from entering the holes. The surface of the water in the tube will be at atmospheric pressure and, by definition, will give the position of the watertable. The water surface can be measured manually with commercially available electronic dipmeters which emit a noise when the water surface is reached. However, Baird et al. (1998) noted that these electronic dipmeters were unreliable because films of salt water forming bridges between the co-axial elements on the sensing tip causing erroneous readings. An alternative is to measure the water level manually using a hollow graduated tube through which the observer blows to locate the water level. Manual measurements are generally sufficient for monitoring low frequency oscillations where watertable elevations only need to obtained every 15–20 min. However, pressure transducers should be used if a continuous time series is required. The transducers should be lowered to the very bottom of the well so that the sensor elevation remains constant even if the top of the well is disturbed. If higher frequency measurements are required, such as groundwater fluctuations in response to wave run-up, the pressure transducers should be buried directly in the beach. Turner (1998) recommended that the direct burial method should be used for measurement of oscillations at frequencies of less than 1 min. Pressure transducers which are buried directly in the beach should have a porous screen across the sensor port to stop sediment from touching the sensor, and must also have vented cable in order allow for changing atmospheric pressure.

Pore water pressures in the beach can be measured with a piezometer, which consists of a tube open only at the end, which may be fitted with a permeable tip. As with groundwater wells, the lower end needs to be screened to keep the piezometer from being filled with sand. Water in the beach sand will rise up the piezometer tube until it is at equilibrium with the pore water pressure in the sand around the piezometer tip. In hydrostatic conditions when the watertable is horizontal, the water level in the piezometer will correspond with the watertable. However, in hydrodynamic conditions, changes in hydraulic potential with depth in the beach may not be linear and a piezometer will no longer provide an adequate measure of watertable elevation. At least three piezometers are required to determine the direction and hydraulic gradient of groundwater flow. Two or more closely spaced piezometers with their lower ends located at different depths (nested piezometers) are used to measure vertical head gradients. This layout is essential in areas where rates of vertical flow may be significant (Turner, 1998). Piezometers have several uses. Pore water pressure measurements can be used to construct lines of equal hydraulic potential (the sum of pore water pressure and elevation potential) and therefore flow nets which can be used to estimate discharge from a beach (e.g., Fetter, 1994). Pore water pressure measurements can also be used to confirm whether the simplifying assumptions of groundwater flow models are met in practice. (For example, the Dupuit–Forchheimer approximation discussed in Section 3.4).

Baird and Horn (1996) noted that a key aspect of the design of both piezometers and wells is the response time of the instrument. A finite time is required for water to flow into or out of the piezometer or well to register a change in the pore water pressure and watertable elevation. Response time depends on the geometry of the instrument and the hydraulic conductivity of the beach sediment. With both piezometers and wells, instruments with a small bore require less exchange of water to register changes in the groundwater system, and in the swash zone where there will be rapid changes in watertable elevation and pore water pressures, relatively wide bore piezometers and wells may prove too slow in responding, giving a lagged and attenuated response.

Measurements of vertical pore water pressures in the swash zone are commonly made with pressure transducers buried at known vertical spacings (e.g., Turner and Nielsen, 1997; Turner and Masselink, 1998; Blewett et al., 2000). Buried pressure transducers generally measure pressures at spacings of 150–200 mm and depths of 10–340 mm below the beach surface. Near-surface hydraulic gradients in the beach can be measured at a much higher resolution by a technique developed by Baldock and Holmes (1996), which is described in detail in Horn et al. (1998) and Baldock et al. (2001). The probe tips can be arranged in a vertical array so that the pressures and vertical hydraulic gradients can be obtained just below the beach surface with spacings as small as 10 mm between probes. Vertical hydraulic gradients can therefore be obtained within the upper 30 mm of the beach sediment at a single position in the beach. Some of the results obtained with this system will be discussed in Section 4.3.

It is important to know the hydraulic conductivity of the beach sediment for modelling purposes. Hydraulic conductivity, K, may be defined as the specific discharge per unit hydraulic gradient. The hydraulic conductivity reflects the ease with which a liquid flows and the ease with which a porous medium permits the liquid to pass through it, and relates the mean discharge flowing through a porous substance per unit cross-section to the total gravitational and potential force. Hydraulic conductivity has units of velocity, usually m s^{-1} in the case of beach groundwater systems. Hydraulic conductivity should be distinguished from permeability (also referred to as intrinsic or specific permeability), denoted by k, which is the measure of the ability of a rock, soil or porous substance to transmit fluids and refers only to the characteristics of the porous medium and not to the fluid which passes through it. Permeability has dimensions of L^2.

Another important parameter in groundwater modelling is a dimensionless parameter known as specific yield, denoted by s. The specific yield, which is also known as the drainable porosity, is defined as the volume of water that an unconfined aquifer releases from storage per unit surface area of aquifer per unit decline in watertable (Freeze and Cherry, 1979). It should be noted that specific yield, or drainable porosity, is not the same as porosity, and the two terms should not be used interchangeably. Porosity is the volume of the voids in a sediment or rock divided by the total volume of the sediment or rock. Porosity is denoted by n, and is usually reported as a decimal fraction or percent. The moisture content or volumetric water content, θ, is defined as the volume of water in a sediment or rock sample divided by the total volume of the sediment or rock. In saturated conditions, where the pores are filled with water, the volumetric water content, θ, is equal to the porosity, n. In unsaturated conditions, where the pores are only partially filled with water, the volumetric water content is less than the porosity. There is also a difference between specific yield and specific storage, which is defined as the volume of water that a confined aquifer releases from storage under a unit decline in hydraulic head (Freeze and Cherry, 1979). The term specific storage refers to a unit decline in hydraulic head below the watertable, in a saturated aquifer. Releases from storage in unconfined aquifers (such as beach sediments) represents an actual dewatering of the pores, whereas releases from storage in confined aquifers represent only the secondary effects of water expansion and aquifer compaction caused by changes in the fluid pressure (Freeze and Cherry, 1979).

Most studies calculate hydraulic conductivity by collecting sediment samples and using one of the many empirical formulas that relate the hydraulic conductivity, K, to some measure of the representative grain size. A commonly used formula is that of Krumbein and Monk (1943) where permeability, k (in units of Darcies where 1 Darcy = 9.87×10^{-13} m^2), is given by:

$$k = 760D^2 e^{-1.31\sigma} \tag{1}$$

where D is the geometric mean grain diameter (mm), and σ is the sediment sorting (in phi units). Hydraulic conductivity is then given by:

$$K = \frac{kg}{v} \tag{2}$$

where k is the permeability, v is the kinematic viscosity of the beach groundwater ($L^2\ T^{-1}$) and g is acceleration due to gravity ($L\ T^{-2}$).

Although empirical equations such as Eqs. (1) and (2) are often used to calculate hydraulic conductivity from grain size, their use should be treated with caution. For example, Baird et al. (1998) found that there was an order of magnitude difference between measured and calculated hydraulic conductivity. The mean hydraulic conductivity of the sediment cores measured with a permeameter was 0.225 cm s^{-1}, while the mean hydraulic conductivity of the sand in the permeameter cores calculated using the empirical equation of Krumbein and Monk (1943) was 0.02 cm s^{-1}. Another weakness of empirical equations relating sediment size to hydraulic conductivity is that they are generally only applicable to sand-sized sediments and may not be appropriate for coarser sediments or mixed sediment distributions. A more accurate approach to obtaining hydraulic conductivity is to take samples of intact sediment from the beach using a coring device and then measure hydraulic conductivity using a laboratory permeameter. Another advantage of collecting sediment cores is that the porosity and specific yield of the sample can also be determined. Piezometers can also be used to estimate hydraulic conductivity by performing slug tests in which water is either added to or removed from the piezometer. The rate of recovery of the water level in the piezometer is then monitored and the hydraulic conductivity can be calculated.

Hydraulic conductivity can be extremely variable. For example, Baird et al. (1998) obtained a range of values of hydraulic conductivity from their laboratory permeameter measurements varying between 0.036 and 1.179 cm s^{-1}, with a coefficient of variation of 143%. Unfortunately, this degree of variability is not unusual for hydraulic conductivity. In many groundwater systems, hydraulic conductivity is known to vary by orders of magnitude even over short distances and the characterisation of hydraulic conductivity for use in models has presented major conceptual difficulties (Baird and Horn, 1996). As Freeze and Cherry (1979) remarked, there are few physical parameters that take on values over 13 orders of magnitude! In addition, some studies of beach groundwater behavior ignore the importance of variations in the hydraulic properties of the beach sediment. This may, perhaps, be due to the use of models which use a single "representative" value of hydraulic conductivity or the ratio of hydraulic conductivity to specific yield. Since sediment size across a beach is not constant, it is unlikely that hydraulic conductivity or specific yield will be constant, which reinforces the importance of collecting sediment cores to obtain an indication of the variability of hydraulic conductivity across a beach and with depth.

A potentially important, but poorly understood, consideration in beach groundwater studies is the role of air encapsulation during the wetting of beach sand. A large body of evidence within the soil physics literature suggests that few sediments below the watertable are fully saturated (e.g., Constantz et al., 1988; Fayer and Hillel, 1986; Faybishenko, 1995). Pockets of gas can be formed in a variety of ways. During rapid infiltration, air pockets may be trapped by infiltrating water and bypassed by a rising watertable. In sediments containing organic matter, biogenic gas production can also lead to gas pocket formation (e.g., Romanowicz et al., 1995). Air encapsulation is thought to have large effects on the hydraulic and storage properties of soils, particularly hydraulic conductivity and specific yield. Encapsulated gas will reduce hydraulic conductivity considerably below true saturation values if it blocks effective (i.e., water-conducting) pores. For example, in field and laboratory infiltration experiments on sand and gravel loam soils, Constantz et al. (1988) found that air encapsulation reduced hydraulic conductivities to between 0.1 and 0.2 of the value of the saturated hydraulic conductivity. The volume of air encapsulated in these soils ranged between 4% and 19% of total pore volume. Fayer and Hillel (1986) found that the watertable rose more rapidly when air was encapsulated than when it was not; the shallower the watertable, the more pronounced the effect of encapsulated air. In beach sediments, it is likely that the top few centimetres are not fully saturated even when the watertable is at the sand surface. Baldock et al. (2001) suggested that in beach sediments which are not fully saturated, dilation and contraction of encapsulated gas will slow the propagation of a pressure wave, causing hydraulic gradients to develop.

Several techniques are available to measure moisture content and thus air encapsulation, although none of them are entirely satisfactory. The most common technique is to collect sediment samples from the beach and calculate the gravimetric moisture. This technique is not capable of providing any time series data of moisture content values. Turner (1993a) used a neutron probe to measure moisture content above the watertable, with an estimated confidence interval for percentage saturation of $\pm$ 12%. The main disadvantage of this technique is that it averages moisture content over a relatively large volume of porous medium and therefore cannot measure steep moisture gradients or provide near-surface measurements of moisture content (Turner, 1993a; Atherton et al., 2001). An alternative technique was proposed by Baird and Horn (1996), who suggested that the presence of encapsulated air can be determined using time domain reflectrometry (TDR) to measure the volumetric water content of the soil. If the saturated water content of the sediment can be measured, the volume of encapsulated air can be calculated. TDR measures the apparent dielectric constant in the region between a pair of thin metal rods inserted into the sediment by measuring the speed of electromagnetic waves which travel in the waveguide formed by the two rods. The apparent dielectric constant of a partially saturated sediment can be related empirically to the volumetric water content (fresh or saline) of the sediment. Standard TDR techniques can estimate soil moisture content to an accuracy of $\pm$ 2% of total soil volume, which compares favourably with the thermal neutron technique used by Turner (1993a). They have the advantage that they do not need calibrating for each individual application (Baird and Horn, 1996). Recently, Atherton et al. (2001) used an instrument called a ThetaProbe to measure near-surface beach moisture content, which determines the impedance of a sensing rod array and relates voltage outputs to moisture content. The advantage of this device is that it is possible to measure relatively small volumes of sand (35 cm^3) rapidly, enabling Atherton et al. (2001) to make 445 measurements over part of a tidal cycle.

Although the method used by Atherton et al. (2001) made it possible to measure moisture content relatively rapidly (5–30 s), this rate of measurement is still significantly slower than the frequency at which moisture content in the swash zone may vary. Under the action of swash, the surface sediment is constantly re-worked and, during this process, air bubbles will be both trapped and released. The air content at a given point will vary between swashes, leading to highly dynamic variations in dilation, contraction and hydraulic conductivity. The degree of compaction, and hence porosity, of sand in the swash zone will also be highly variable, depending on the depth below the surface, tidal stage and wave conditions (Heathershaw et al., 1981). Horn et al. (1998) showed data indicating that regions which are fully and less fully saturated appear to develop below the sand surface, particularly a thin layer of totally saturated sediment which moves up and down the beach with the swash. These findings suggest a higher frequency method of measuring moisture content needs to be devised in order to improve understanding of beach groundwater–swash interactions.

3.4. Modelling beach groundwater dynamics

An aquifer is a saturated geologic unit that can transmit significant quantities of water under ordinary hydraulic gradients (Freeze and Cherry, 1979). An unconfined aquifer, or watertable aquifer, is one in which the watertable forms the upper boundary. Beach groundwater systems are generally treated as unconfined aquifers because commonly the upper boundary to groundwater flow is defined by the watertable itself rather than by some surface layer of impermeable material (Masselink and Turner, 1999). The beach groundwater system is underlain by an impermeable boundary at a depth which is often unknown. The rate of flow (or specific discharge) of water through unconfined aquifers, u, is given by Darcy's Law:

$$u = -K\frac{\partial h}{\partial x} \tag{3}$$

where h is hydraulic head (units of length, L), x is the distance (L) and K is hydraulic conductivity (L T^{-1}).

Darcy's Law is valid as long as flow is laminar, which is a reasonable assumption for sand beaches. This may not be the case for gravel beaches (Packwood and Peregrine, 1980). Darcy's Law shows that the rate of groundwater flow is proportional to the hydraulic gradient, or slope of the watertable. The hydraulic gradient ($\delta h/\delta x$) is the change in hydraulic head (h) over distance. Water flows down the hydraulic gradient in the direction of decreasing head. The hydraulic head (h) is the sum of the elevation head (z) and the pressure head (ψ), and is measured in length units above a datum. There is no standard datum used in beach hydrology, but many researchers use the elevation of an impermeable layer beneath the beach sediment, so that the vertical coordinate z is measured from the impermeable base. Some workers have considered the hydraulic head in a beach groundwater system to be the elevation of the free water surface, or watertable elevation. However, this is only true when there is no vertical component to the flow; in other words, when Dupuit–Forcheimer conditions apply (see below).

Groundwater hydrologists generally model water flow using Darcy's Law in combination with an equation of continuity that describes the conservation of fluid mass during flow through a porous medium. A common approach to modelling beach groundwater flow in response to tidal forcing in sandy beaches uses the one-dimensional form of the Boussinesq equation:

$$\frac{\partial h}{\partial t} = \frac{K}{s}\frac{\partial}{\partial x}\left(h\frac{\partial h}{\partial x}\right) \tag{4}$$

where h is the elevation of the watertable (L), t is time (T), K is hydraulic conductivity (L T^{-1}), s is the specific yield (dimensionless), and x is horizontal distance (L).

The main assumption in using Eq. (4) is that groundwater flow in a shallow aquifer can be described using the Dupuit–Forchheimer approximation. Dupuit–Forchheimer theory states that in a system of shallow gravity flow to a sink when the flow is approximately horizontal, the lines of equal hydraulic head or potential are vertical and the gradient of hydraulic head is given by the slope of the watertable (Kirkham, 1967). In effect, the theory neglects the vertical flow components. Using Dupuit–Forchheimer theory, two-dimensional flow to a sink can be approximated as one-dimensional flow, and the resulting differential equation (Eq. (4)) is relatively easily solved. In beaches that are underlain by relatively impermeable material, it is likely that Dupuit–

Forchheimer theory provides an adequate description of groundwater flow, and field studies such as those of Baird et al. (1998) and Raubenheimer et al. (1998) support this assumption.

Where Dupuit–Forchheimer assumptions do not apply, such as in artificially drained beaches (e.g., Li et al., 1996), the beach aquifer should be considered as a two-dimensional flow system. One approach is to assume that the watertable is a free surface or flow line so that

$$\frac{\partial h}{\partial t} = \frac{K}{s}\left(\frac{\partial H}{\partial z} - \frac{\partial h}{\partial x}\frac{\partial H}{\partial x}\right) \tag{5}$$

where H is the total or hydraulic head (L) and z is vertical distance (L). As in Eq. (4), h is the elevation of the watertable (L), t is time (T), K is hydraulic conductivity (L T^{-1}), s is the specific yield (dimensionless), and x is horizontal distance (L). Eq. (4) is much easier to solve than Eq. (5) and should be used whenever the assumption of near-horizontal flow through the beach sand is generally met.

A number of analytical and numerical models have been developed which are able to predict beach watertable fluctuations in response to tides (Nielsen, 1990; Turner, 1993a,b, 1995a,b; Kang and Nielsen, 1996; Li et al., 1996, 1997a; Baird et al., 1996, 1997, 1998; Raubenheimer et al., 1998, 1999). These Boussinesq models, based on solutions to Eq. (4), have been successful in reproducing observed fluctuations of the beach watertable at diurnal and higher tidal frequencies, and also reproduce observations such as the shape and slope of the beach watertable, the lag and landward attenuation of beach watertable oscillations, and seepage face development. However, these models generally underpredict the watertable elevations under conditions when wave effects are important.

Models of beach watertable fluctuations that incorporate wave effects have been developed only very recently (Kang and Nielsen, 1996; Li et al., 1997b; Li and Barry, 2000). Nielsen et al. (1988) and Kang and Nielsen (1996) proposed the use of a linearised version of the Boussinesq equation (Eq. (4)) with an additional term to model watertable fluctuations in the zone of run-up infiltration:

$$\frac{\partial h}{\partial t} = \frac{Kd_a}{s}\frac{\partial^2 h}{\partial x^2} + U_1(x,t) \tag{6}$$

where d_a is the aquifer depth and $U_1(x,t)$ is the infiltration/exfiltration velocity per unit area. As in Eq. (4), h is the elevation of the watertable (L), t is time (T), K is hydraulic conductivity (L T^{-1}), s is the specific yield or drainable porosity (dimensionless), and x is horizontal distance (L). Li et al. (1997b) and Li and Barry (2000) have developed more complicated models to predict wave-induced watertable fluctuations, which enabled them to incorporate capillarity effects and predict high-frequency watertable response to wave run-up landward of the swash zone. However, none of the models which include wave effects have yet been tested against field or laboratory data.

Finally, beach groundwater models have not yet been linked to swash hydrodynamic and sediment transport models, although Turner (1995a) modelled beach profile response to groundwater seepage using an equilibrium net transport parameter, and Li et al. (2002) modelled swash and beach groundwater flows to predict sediment transport and beach profile change in the swash zone. In particular, models of swash–groundwater interactions do not yet incorporate the physical processes such as infiltration and groundwater outflow which are thought to influence sediment transport in the swash zone. The relative importance of these mechanisms is where the greatest areas of uncertainty arise.

4. Mechanisms of surface–subsurface flow interaction

4.1. Infiltration and exfiltration

Several mechanisms have been suggested to explain why beaches with a low watertable tend to accrete and beaches with a high watertable tend to erode. The mechanisms which are proposed most frequently are infiltration and exfiltration. The terminology used to discuss these mechanisms requires some clarification, as different terms may be used by hydrologists, engineers and other coastal scientists. The physical process of interest is that of vertical flow within a porous bed and/or through a permeable boundary. Vertical flow exerts a force within the bed called seepage force. Seepage force is defined as a force acting on an individual grain in a porous medium under flow, which is due to the difference in hydraulic head between the front and back faces of the grain (Freeze and Cherry, 1979). The seepage force, F, is exerted in the direction of flow and is directly proportional to the hydraulic gradient, and is given by

$$F = \rho g \frac{\partial h}{\partial z} \tag{7}$$

where ρ is the density of the fluid (M L^{-3}), g is acceleration due to gravity (L T^{-2}) and $\delta h/\delta z$ is the hydraulic gradient (dimensionless). In the convention used here, a positive hydraulic gradient represents a downward-acting seepage force and a negative hydraulic gradient represents an upward-acting seepage force.

The vertical flows which produce this seepage force have been referred to in a number of different ways in the literature: injection (or blowing) and suction (Martin, 1970; Oldenziel and Brink, 1974; Willets and Drossos, 1975; Conley and Inman, 1994; Rao et al., 1994); influent and effluent (Watters and Rao, 1971); transpiration (Kays, 1972); and bed ventilation (Conley and Inman, 1992, 1994). The condition of fluidisation at the surface has been referred to as piping (Madsen, 1974; Higgins et al., 1988), seepage erosion (Hutchinson, 1968), and groundwater sapping (Higgins, 1982). In the case of beach hydrology, the terms infiltration and exfiltration are becoming common, and they will be used here. Infiltration is the process by which water enters into the surface horizon of a soil or porous medium, such as beach sediment, in a downward direction from the surface by means of pores or small openings. Infiltration is often used interchangeably with percolation, which more correctly refers to the flow of water through a soil or porous medium below the surface. Recently the term exfiltration has been used to describe outflow from the bed. Infiltration/exfiltration velocity may also be referred to as seepage velocity.

Grant (1946, 1948) was among the first to suggest a link between beach groundwater behavior and swash zone sediment transport, proposing a simple conceptual model which has been highly influential in beach hydrology research. Grant defined a dry foreshore as one with a low watertable and an extensive infiltration zone. On a dry foreshore, most of the water infiltrates rapidly into the sand above the watertable. This infiltration reduces the flow depth of the swash and thus the velocity, allowing sediment deposition. He suggested that near the swash limit, the velocity decreases below the lower critical limit and the flow will change from turbulent to laminar. Sediment is rapidly deposited when this flow transition occurs. When the backwash begins, the velocity is low and laminar flow prevails for a short period. This laminar flow decreases the likelihood of the backwash transporting sediment down the foreshore. He reasoned that laminar backwash persists for a longer time if the slope of the beach is small and if the depth of water is also small. Grant's conceptual model also described conditions on a wet foreshore, one whose watertable is high and contiguous with the surface of most of the foreshore. He reasoned that when the beach is in a saturated condition throughout all of the foreshore the backwash, instead of being reduced by infiltration, retains its depth and is augmented by the addition of water rising to the surface of what he called the effluent zone (the seepage face). This increased velocity and depth of the backwash produces a turbulent flow, which enhances offshore transport. Grant also noted that groundwater outcropping at the beach surface can cause dilation or fluidisation of the sand grains, allowing them to be entrained more easily by backwash flows.

The logic of Grant's conceptual model has led many researchers to concentrate on the effects of infiltration/exfiltration on beach accretion and erosion (e.g., Bagnold, 1940; Emery and Foster, 1948; Emery and Gale, 1951; Isaacs and Bascom, 1949; Longuet-Higgins and Parkin, 1962; Duncan, 1964; Strahler, 1966; Harrison, 1969, 1972; Waddell, 1976; Chappell et al., 1979; Heathershaw et al., 1981; Lanyon et al., 1982a,b; Carter and Orford, 1993; Turner, 1993a; Weisman et al., 1995; Turner and Nielsen, 1997; Turner and Masselink, 1998; Nielsen at al., 2000). Many of these authors suggested that infiltration losses during swash provide the main mechanism by which beach accretion occurs above the still water level. Because the swash and backwash are relatively shallow, a small change in water volume due to infiltration (or addition of water due to exfiltration) could influence uprush/backwash flow asymmetry and therefore the energy available for sediment transport. For example, Nelson and Miller (1974) showed that a reduction of swash volume due to infiltration into the sand matrix of a nonsaturated beach will decrease the energy of the swash. The resulting mass loss does not have to be great for it to have a significant effect on sediment transport. They found that losses due to infiltration become more critical as the waves become smaller or the beach slope lower. Work by Packwood (1983) suggested that backwash infiltration in particular could be important in affecting sediment movement in the swash zone. Within the swash zone, rapid watertable fluctuations due to swash infiltration into the capillary fringe may also influence sediment mobility (Li et al., 1997b, 2000; Turner and Nielsen, 1997; Turner and Masselink, 1998). Groundwater flow at deeper levels within the beach is also influenced by infiltration during swash uprush, although the hydraulic gradients developed tend to be small (Waddell, 1973; Lewandowski and Zeidler, 1978; Hegge and Masselink, 1991; Turner and Nielsen, 1997; Raubenheimer et al., 1998).

Although most studies have concentrated on infiltration/exfiltration and possible effects on swash/backwash asymmetry, researchers such as Nielsen (1992, 1997), Turner and Nielsen (1997), Turner and Masselink (1998) and Hughes and Turner (1999) have identified other mechanisms by which vertical flow through a porous bed could affect swash zone sediment transport. These include an alteration in the effective weight of the surface sediment due to vertical fluid drag, which will act to stabilise the bed under infiltration or destabilise under exfiltration, and modified shear stresses exerted on the bed due to boundary layer thinning due to infiltration or thickening due to exfiltration. Watters and Rao (1971) described a number of effects of vertical flow through a porous bed: the angle of attack at which the main flow contacts the particles is altered; 'dead' water (the nearly static fluid between adjacent particles) is flushed out of the top bed layer, increasing the exposed surface area of a particle to the main flow; and the changed wake behind a particle not only affects that particle but others in its lee. Turner and Masselink (1998) summarised the effect of these processes on the boundary layer, with streamlines being drawn closer to the sediment–fluid interface under infiltration and moved away from the sediment–fluid interface under exfiltration. The result is a vertical shift of the boundary layer velocity profile, with an increase of flow velocity and shear stress at the bed under infiltration and a decrease under exfiltration.

Experimental work on the influence of seepage flows within sediment beds provides conflicting results concerning their effect on bed stability. Most authors agree that infiltration increases shear stress and skin friction at the bed, whereas exfiltration decreases bed shear stress and friction. However, the effects of infiltration and exfiltration on entrainment and sediment transport is less clear. Martin (1970) concluded that infiltration could either enhance or hinder incipient sediment motion, depending on the sediment size and permeability, whereas exfiltration had no effect on incipient motion until the bed was fluidised. In contrast, Watters and Rao (1971) reached the conclusion that exfiltration inhibited the motion of bed particles while infiltration enhanced the motion. Oldenziel and Brink (1974) found that infiltration decreased the transport rate and exfiltration increased the transport rate. The experimental results of Willets and Drossos (1975) indicated that entrainment and transport rates were affected differently by infiltration. They observed that sediment particles in the infiltration zone were dislodged less frequently, but once entrained, travelled farther and faster than particles elsewhere in the flow. Willets and Drossos (1975) also argued that the transport path length had to be considered, as their observations suggested that transport rate would decay slowly with distance in the transport direction in a long zone of uniform infiltration, whereas under other conditions infiltration would increase the transport rate for a considerable length of the infiltration zone. Conley and Inman (1992) suggested that the sediment-mobilising properties of the flow would be diminished under exfiltration conditions due to decreased bed stress with turbulent kinetic energy removed from the bed, which would be characterised by thinner, less dense granular-fluid layers. Flow experiencing infiltration would be characterized by a more rapid and therefore distinct boundary layer, enhancing sediment mobilisation. They also suggested that different friction factors would be required for flow influenced by infiltration and exfiltration. Conley and Inman (1994) investigated the effect of seepage flows on oscillatory boundary layers in more detail, and suggested that infiltration tended to stabilise the flow and exfiltration tended to destabilise flow. Their experiments demonstrated that during infiltration, mean velocities throughout the boundary layer were uniformly greater than the unventilated velocities, with a greater vertical velocity gradient at the bed. The opposite was observed under exfiltration. Rao et al. (1994) found that seepage flow due to both infiltration and exfiltration could cause an increase or decrease in bed shear stress when compared to the no-seepage condition, depending on the relative magnitudes of the critical shear stress under the no-seepage condition, the sediment concentration, and the seepage rate.

These contradictory results may be because the effects of the seepage force and boundary layer thinning tend to oppose each other. While infiltration results in a stabilising seepage force, simultaneous boundary layer thinning has the opposing effect of enhancing sediment mobility, and vice versa for exfiltration (Hughes and Turner, 1999). Nielsen et al. (2001) suggested that the relative importance of these opposing effects depends on the density of the sediment and the permeability of the bed.

In a first attempt at quantifying these processes, Nielsen (1997) proposed a revised Shields parameter that includes the effects of infiltration/exfiltration:

$$\theta_{\mathrm{m}} = \frac{u_{*0}^2\left(1 - \alpha\frac{w}{u_{*0}^2}\right)}{gd_{50}\left(s - 1 - \beta\frac{w}{K}\right)} \tag{8}$$

where w is the seepage velocity (L T^{-1}, with infiltration negative), u_{*0}^2 is the shear velocity without seepage (L T^{-1}), s is relative density (dimensionless: ρ_s/ρ, where ρ_s is the density of the sediment and ρ is the density of the fluid), K is hydraulic conductivity (L T^{-1}), g is acceleration due to gravity (L T^{-2}), d_{50} is median grain diameter and α and β are constants, defined by Neilsen et al. (2001) as 16 and 0.4, respectively. The factor β is intended to quantify the increase of the particle's weight due to the vertical seepage velocity, and is 1 for particles in the bed but considerably smaller for particles on the surface (Nielsen et al., 2001). The modified Shields parameter in Eq. (8) was designed to account for the opposing effects of infiltration, as the extra term in the numerator represents the increase in shear stress due to the thinning of the boundary layer and the extra term in the denominator represents the effect of the downward seepage drag on the effective weight of the grains (Nielsen et al., 2001). Eq. (8) suggests that for a fixed sediment density, as grain size (and therefore hydraulic conductivity) decreases, the stabilising effect will increase. Therefore, finer quartz sands ($d_{50} < 0.58$ mm) are likely to be stabilised by infiltration, whereas the net effect of infiltration on beaches of coarser sediment may be destabilising (Nielsen, 1997). Nielsen et al. (2001) extended this analysis to show that infiltration is likely to enhance sediment mobility for dense, coarse sediment where $\alpha(s-1) > \beta[(u_{*0})/K]$ and impede sediment motion for light, fine sediment where $\alpha(s-1) < \beta[(u_{*0})/K]$.

Turner and Masselink (1998) also followed this approach, but included the effects of the seepage flow on the bed shear stress (e.g., Turcotte, 1960; Conley and Inman, 1994). They used their modified Shields parameter, which incorporated an additional through-bed term, to calculate the swash zone transport rate in the presence of infiltration/exfiltration relative to the case of no vertical flow through the bed. Their modelling showed that altered bed stresses dominated during uprush, indicating enhanced sediment mobility relative to the case of an impermeable bed. They found that altered bed stress effects were also dominant during backwash. The net effect of combined seepage force and altered bed stress was less pronounced during backwash than during uprush. These findings differ from those of Packwood (1983), whose model suggested that the effect of infiltration into a porous bed is felt much more in backwash than in uprush. However, Packwood (1983) considered only fluid loss into the beach due to infiltration. Turner and Masselink (1998) concluded that the effects of combined seepage force and altered bed stress enhanced net onshore sediment transport on a saturated beachface.

Although recent work by Baldock and Holmes (1998) showed that sediment transport over a fluidised bed in the presence of a steady current may differ little from that over a normal sediment bed, they also suggested that a seepage flow might have a significant effect on sediment transport during sheet flow. Sheet flow conditions are likely to occur during backwash (Bradshaw, 1982; Beach et al., 1992; Hughes, 1995; Osborne and Rooker, 1997; Hughes et al., 1997a; Masselink and Hughes, 1998) and probably also during the uprush. Consequently, the fluid flow within the near surface layers of a sand beach may significantly affect swash zone sediment transport characteristics.

Nielsen et al. (2001) conducted laboratory measurements to investigate the effects of infiltration on sediment mobility of a horizontal sand bed under regular non-breaking waves under conditions of steady downward seepage, and compared these to measurements without infiltration. Their experiments showed that infiltration had the effect of reducing the mobility of 0.2 mm sand. These experiments do not reproduce swash zone conditions, with irregular asymmetric waves alternately inundating and exposing a sloping bed of sediment which is unlikely to be uniformly sized. However, they suggested that infiltration effects on sediment mobility in the swash zone would be minor if infiltration rates are in the range reported by researchers such as Kang et al. (1994a,b) and Turner and Nielsen (1997), where $w<0.15$ K.

4.2. Fluidisation

Although infiltration and exfiltration are the primary mechanisms by which groundwater flow is thought to influence sediment transport in the swash zone, the potential of beach groundwater fluctuations to cause bed failure due to instantaneous fluidisation has also been considered. Fluidisation of sediment occurs when the upward-acting seepage force exceeds the downward-acting immersed particle weight (i.e., when the effective stress becomes zero). In particular, it has been suggested by a number of workers that tidally induced groundwater outflow from a beach during the ebb tide may enhance the potential for fluidisation of sand, and thus the ease with which sand can be transported by swash flows (e.g., Grant, 1946, 1948; Emery and Foster, 1948; Duncan, 1964; Chappell et al., 1979; Heathershaw et al., 1981; Turner, 1990; Turner and Nielsen, 1997). However, tidally induced groundwater outflow alone is unlikely to be sufficient to induce fluidisation, because hydraulic gradients under the sand surface will tend to be relatively small, generally of the order of the beach slope (1:100 to 1:10) (Baird et al., 1998). In addition, Turner and Nielsen (1997) found that, rather than fast watertable rise in the swash being the cause of upward flow (and hence potential fluidisation), rapid watertable rise within the swash zone resulted from a small amount of infiltration of the swash lens. However, upward-acting swash-induced hydraulic gradients that are capable of fluidising the bed have been measured within the top few centimetres of the beach. Horn et al. (1998) and Blewett et al. (1999) presented field measurements of large upward-acting hydraulic gradients which considerably exceeded the fluidisation criterion of Packwood and Peregrine (1980), who observed that, for many sands and fine gravels, fluidisation occurs when the upward-acting hydraulic gradient is greater than (i.e., more negative than) about -0.6 to -0.7 (in the convention used here). The mechanism responsible for these upward-acting hydraulic gradients is not clear.

Baird et al. (1996, 1997) argued that fluidisation is only generally possible in the presence of swash on a seepage face. As a swash flow advances over the saturated beach surface there will be a rapid increase in pore water pressures below the beach surface. When under swash flow, the beach sediment behaves like a confined aquifer. The sediment is saturated and movement of water into the beach is extremely limited since changes in porosity due to expansion and contraction of the mineral ‘skeleton’ will be minimal. However, water pressures will propagate rapidly through the sediment. As the swash retreats, there will be a release of pressure on the beachface, potentially giving large hydraulic gradients acting vertically upwards immediately below the surface. The resultant seepage force associated with these upward-acting hydraulic gradients could be sufficient to induce fluidisation of the sand grains at the surface. They showed theoretically how hydraulic gradients in the saturated sediment beneath swash can exceed, or at least come close to, the threshold for fluidisation.

Baldock et al. (2001) compared field measurements of swash-induced hydraulic gradients in the surface layers of a sand beach to the predictions of a simple 1D diffusion model based on Darcy’s law and the continuity equation. The model allows for dynamic storage within the sediment–fluid matrix due to loading/unloading on the upper sediment boundary. The model predicted minimal hydraulic gradients for a rigid, near fully saturated sediment which were in accordance with measurements close to the seaward limit of the swash zone. The model also provided a good description of the measured hydraulic gradients, both very close to the surface and deeper in the bed, for the region of the beach where the beach surface is frequently exposed between swash events. These model-data comparisons suggest that the surface layers of a sand beach store and release water under the action of swash, leading to the generation of relatively large hydraulic gradients, as suggested by Baird et al. (1996, 1997). However, the model is not able to predict the very large near-surface negative hydraulic gradients observed by Horn et al. (1998), although, for the same swash events, the agreement is good deeper in the bed. Baldock et al. (2001) concluded that the very large upward-acting hydraulic gradients observed in the upper part of the bed were not simply due to pressure propagation during swash loading/unloading or swash-generated 2D subsurface flow cells. Instead, they suggested that these very large negative hydraulic gradients are probably generated by alternative mechanisms, possibly due to non-hydrostatic pressures developing within the sheet flow layer that occurs during backwash.

4.3. Implications for sediment transport in the swash zone

The implications of vertical flows and seepage forces for sediment transport are not clear. Vertical seepage forces are not themselves capable of transporting sediment. They are also not likely to influence swash flows directly in sand beaches, because the flow across the sediment–fluid interface is insignificant in terms of fluid volumes, due to the low hydraulic conductivity of fine sand. Vertical seepage forces are, therefore, unlikely to influence the free stream surface flows in the swash zone (Baldock et al., 2001). However, seepage forces may act to provide readily entrainable material which is then available for transport, onshore under uprush or offshore under backwash. Madsen (1974) noted that the result of momentary failure caused by the flow within the bed was that the bed material is unable to resist any additional force. He suggested that the instability of the bed due to the flow induced in the bed may significantly influence the amount of bed material set in motion. Packwood and Peregrine (1980) showed that pressure gradients under bores or steep-fronted waves would produce an upward seepage flow capable of fluidising the bed under the bore face. They suggested that under such circumstances, the upward velocity just forward of the wave crest would be able to lift the fluidised material and inject the sediment into the flow.

Nielsen et al. (2001) noted that their experiments indicate only the effect of infiltration/exfiltration on sediment mobility and did not necessarily suggest anything about the direction of net sediment transport. This is likely to be affected by other factors such as the phase relationship between infiltration/exfiltration induced effects on sediment transport and swash flows. For example, Blewett et al. (1999) measured events in which large upward-acting hydraulic gradients occurred when the head of water at the surface, and therefore the uprush or backwash flow, was zero. Under these conditions, even if the sediment were to be fluidised, it would not be transported. However, in other data sets, Blewett et al. (1999) reported measurements with upward-acting hydraulic gradients of -1.7, which are more than sufficient to fluidise the bed. These hydraulic gradients lasted for approximately 4 s in waves with a period of 6.3 s under a falling head of water, initially as deep as 40 mm, and under offshore-directed flows of 0.7–1.4 m s^{-1}. This suggests a possible erosional mechanism under backwash. Clearly the phasing between these potentially destabilising hydraulic gradients and swash flows is critical to the potential for sediment transport.

Nielsen et al. (2001) argued that if the beachface tends to be fluidised during backwash as suggested by Horn et al. (1998), a mechanism must exist to enhance sediment transport during the uprush in order to balance this effect, otherwise the beach would rapidly disappear. Although the limited number of field measurements of swash zone sediment transport are not conclusive as to the relative magnitudes of sediment transport in uprush and backwash, most studies suggest that uprush transport is greater than backwash transport (Hughes et al., 1997a,b; Masselink and Hughes, 1998; Osborne and Rooker, 1999; Puleo et al., 2000). However, the mechanism responsible for this is not known. At least three mechanisms that may favor uprush transport have been suggested: (1) Nielsen et al. (2001) suggest that enhanced uprush transport might be produced by fluidisation due to strong horizontal pressure gradients near bore fronts as described by Madsen (1974); (2) Turner and Masselink (1998) demonstrated that the effect of altered bed stress dominated over the change in effective weight and that swash infiltration–exfiltration over a saturated beachface enhances the upslope transport of sediment; and (3) Hughes et al. (1997b), Masselink and Hughes (1998) and Osborne and Rooker (1999) suggested that onshore transport in the uprush is likely to be significantly influenced by turbulence and sediment advection from bores arriving at the beachface, with sediment mobilised by bore collapse at the initiation of uprush added to the sediment entrained by the instantaneous uprush velocities.

Little is known about the response of beach groundwater to bores. The large slope of the sea surface in the vicinity of a bore results in significant hydraulic gradients that may cause considerable beach groundwater flows locally. Infiltration/exfiltration rates are determined by the bore amplitude, the water depth at the bore front, and the thickness of the underlying aquifer. As a bore propagates across a beach, its amplitude and water depth will vary, as will the thickness of the underlying local aquifer, causing variations in local patterns of bore-induced groundwater flow (Li and Barry, 2000). Packwood and Peregrine (1980) studied bore-induced groundwater flows on a flat horizontal porous bed and found that infiltration occurred across the bed at the rear of the bore while exfiltration took place on the front side. Li and Barry (2000) modelled bores in the surf zone on a sloping beach, and obtained similar results, with infiltration occurring across the beachface on the back of the bore while exfiltration took place below the bore front. The rates of infiltration and exfiltration varied in response to changes in the bore amplitude, the front water depth and the aquifer thickness. However, the infiltration/exfiltration pattern remained unchanged and moved across the beachface with the bore. From their model results, Li and Barry (2000) inferred a pattern of groundwater circulation beneath the beachface in the vicinity of the bore. The bore caused large horizontal and vertical heads in the aquifer below it, with the heads at the back of the bore being higher than those at the front over the whole depth. The heads at the back of the bore decreased with depth from the beachface, causing a downward flow, while the heads at the bore front increased with depth, giving an upward flow. Li and Barry (2000) also considered cross-shore variations of the bore-induced beach groundwater flow, and patterns of beach groundwater flow in the swash zone. They found that exfiltration occurred as the bore approached and the front reached the location. Infiltration started subsequently as the bore centre moved shoreward. Without the bore nearby, the beach groundwater flow was relatively small. The hydraulic heads in the swash zone followed a similar pattern to the swash depth. While the swash lens covered the beachface, the vertical head gradients were downward. As the swash lens retreated and the swash depth reduced to zero, the heads likewise declined. Their model predicted that infiltration occurred at a steady rate for the whole of the time under the swash lens, while exfiltration occurred only for a short period when the swash depth was zero, and was of a smaller magnitude. Exfiltration during zero depth was observed in the field by Blewett et al. (1999), who reported measurements of large upward-acting gradients under zero water depths. This occurred only intermittently, whereas large upward-acting hydraulic gradients under backwash were measured more frequently. Li and Barry's model results suggested that infiltration is dominant in the swash zone. However, this has not been corroborated by the limited field data available on infiltration and exfiltration in the swash zone (Turner and Masselink, 1998; Blewett et al., 2001). Li and Barry's model also showed that vertical and horizontal groundwater flows are of similar magnitude. This result contrasts with the assumptions of Turner and Masselink (1998) and Baldock et al. (2001) that the horizontal gradients are several orders of magnitude smaller than the vertical head gradients, particularly during the latter stages of the backwash, where horizontal gradients will be minimal since the depth is typically relatively uniform in the cross-shore direction (Hughes, 1992; Baldock and Holmes, 1997; Blewett et al., 2001). Although Li and Barry (2000) concluded that the magnitude of the instantaneous beach groundwater flow in the swash zone was much less than the bore-generated groundwater flow in the surf zone, they also noted that the approach of the swash lens generated a hydraulic gradient similar to that due to a bore.

Hughes et al. (1997a,b) suggested an alternative mechanism for net onshore transport in the swash zone, reasoning that onshore transport in the uprush is likely to be significantly influenced by turbulence and sediment advection from bores arriving at the beachface. Masselink and Hughes (1998) argued that processes that affect uprush and backwash differently, such as flow acceleration/deceleration, infiltration/exfiltration and bore collapse, all appear to assist uprush transport more than backwash transport. Osborne and Rooker (1999) concluded that high concentrations measured during uprush are most likely to be associated with intense turbulence and high stresses associated with the front of onshore-propagating bores. However, they also noted that measurements with higher temporal and spatial resolution are needed in order to resolve the relative contribution of advection and locally generated suspension in swash events. Puleo et al. (2000) also identified the potential importance of bore turbulence in swash zone sediment transport. They suggested that bore-generated turbulence, which is concentrated in the leading edge of a bore and spreads downward towards the bed, differs from bottom shear turbulence, and that the fundamental difference is the ability of the bore turbulence to affect the bed directly and significantly influence the bottom boundary layer. Their data showed suspended sediment concentrations in the uprush as much as two times larger than those in the backwash near the bed, and up to seven times larger than the backwash suspended sediment concentrations at 5 cm above the bed. They also observed that backwash suspension was different from uprush suspension in that there was less sediment suspension up into the water column and the high suspended sediment concentrations were confined to just above the bed. In all of their data, the suspended sediment load within the first 1.5 s of the uprush (the motion associated with the leading edge) accounted for more than 60% of the total uprush load. Their analysis of the time-dependent cross-shore sediment flux indicated that net transport was always onshore. Puleo et al. (2000) also attempt to assess the importance of bore-generated turbulence to swash zone sediment transport and obtained a high correlation between bore-generated turbulence and suspended sediment transport. Although they were not able to say whether or not the uprush flow behind the leading edge was actively entraining sediment, they interpreted their results as supporting the laboratory work of Yeh and Ghazali (1988), who found that close to the shoreline, the turbulence associated with a bore is advected with the bore front and ultimately acts on a dry bed, while a relatively calm flow occurred behind the bore front. Puleo et al. (2000) suggested that this mechanism for suspension in the uprush may be responsible for onshore sediment transport in the swash zone.

At present little is known about the link between bore-generated turbulence and swash zone sediment transport, as very few measurements of either have been made to date. Puleo et al. (2000) argued that boundary layer shear stress may not be of primary importance to swash sediment transport, as the near-bed motion in the vicinity of the bore may be dominated by this bore-derived turbulence. This suggests that the work described in Section 4.1 on modified shear stresses exerted on the bed due to changes in the boundary layer by infiltration or exfiltration may need to be extended. In addition, there is some suggestion that infiltration and exfiltration may have an effect on turbulence, further complicating the picture. Watters and Rao (1971) found that infiltration had the effect of decreasing turbulence, while exfiltration increased turbulence. Conley and Inman (1994) investigated the effect of seepage flows on oscillatory boundary layers in more detail, and showed that turbulence levels near the bed were higher with infiltration than with exfiltration, and that the turbulence maximum was drawn closer to the bed under infiltration. However, although turbulence due to exfiltration was observed to be enhanced, the time required for this to occur led to a greater vertically averaged turbulence in the half-cycle of the oscillation where infiltration was occurring. Turbulence generated in the infiltration half-cycle was maintained in a compact layer much closer to the bed. Conley and Inman (1994) considered the implications of these findings for sediment transport (although not in the swash zone), reasoning that if sediment transport is approximated by the product of suspended sediment and local velocity, and the level of suspension is proportional to the instantaneous turbulence levels, transport would be in the direction of flow during infiltration. However, no studies of infiltration/exfiltration and turbulence in the swash zone have yet been reported, so the significance of studies such of these for swash zone sediment transport is not clear.

5. Conclusions

Despite the increased interest in swash zone processes in recent years, the exact nature of the relationship between swash flows, beach groundwater and sediment transport in the swash zone is not yet known. Most reviews of this topic have concluded pessimistically. Nielsen (1992) concluded that “it is beyond the present state of the art to model swash zone sediment transport. Too little is known, at present, about boundary layer flow and the corresponding shear stresses in the swash zone. . .and the details of the mechanisms by which flow perpendicular to the beach surface affects beach accretion. . .to even attempt a description of the basic sediment transport mechanisms in this area". Turner and Leatherman (1997) concluded that "at this time, even a basic description of the relative importance of infiltration/exfiltration-induced transport is beyond present understanding". As more measurements of swash zone sediment transport are made, it becomes clear that the basic physics of the processes have not yet been represented adequately. Hughes et al. (1997b) concluded that "the most important lesson learned so far is that existing sediment transport models do not adequately account for the processes governing swash zone sediment transport". Masselink and Hughes (1998) concluded that "the singular nature of swash flow confounded by interactions with the beach groundwater suggests that surf zone sediment transport concepts cannot automatically be transferred to the swash zone". Puleo et al. (2000) concluded that "Bagnold-type sediment transport equations are not adequate for describing sediment transport in the swash zone where complex fluid motions occur. This lack of success implies that not all the fluid physics are adequately described. . .".

A number of processes that require further investigation have been identified. These include sheet flow dynamics, accelerating vs. decelerating swash flows, interactions between swash and beach groundwater, excess suspended sediment present during the uprush due to bore collapse, the relationship between water depth, sediment advection and bore-generated turbulence, grain inertia, intensity levels and vortex structure of turbulence in swash flows, grain settling through turbulence, and the phase relationship between bed shear stress and horizontal flow velocity in the swash. Much work remains to be done before swash zone sediment transport can be modelled successfully.

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