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Sinking Depths of Sand Surface over an Intertidal Area Within a Tidal Inlet Channel

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ABSTRACT

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Observations from an intertidal area of a tidal inlet channel (during low water) show that, under a vertical pressure of around 2,500 kg m^{-2}, the bed surface can descend by between 0 and 9 cm (caused by different sediment packing patterns). An analysis of the relevant sediment samples indicates that such a range of variations is not caused by differences in sediment characteristics *i.e.*, grain size distribution, water content, mineral composition and shape. Hence, the observed variations are suggested to be hydrodynamically induced. The sinking depth is shown to be controlled by bed elevation, which is related to the tidal current speed and the rate of water-level changes immediately before the bed emerges during the ebb-tide phase. The presence of hydrodynamically induced variations in packing implies a difficulty in defining threshold of particle movement, on the basis of mechanics-based, semi-empirical approaches to sediment transport.

ADDITIONAL INDEX WORDS: *Sinking depth, tidal inlet channel sediment, depositional processes, threshold of motion, southern England.*

INTRODUCTION

The packing characteristics of non-cohesive, coarse-grained surficial sediments varies significantly in marine and coastal environments (for reviews, see ALLEN, 1982, p.137–177; and ROGERS *et al.*, 1994). Anyone walking on a beach, for example, may feel that the sediment surface is soft in some places, but hard in other places; this is indicated by the depth one sinks into the sediment. This phenomenon has been noticed for a long time (KINDLE, 1936). Such variations in packing are controlled by either geological properties of sediment particles (grain shape, grain size distribution, and water content), or sedimentary processes (deposition rate) (LEEDER, 1982, p.42–43).

Significant spatial variations in the sinking depth (*i.e.*, the distance of downward shift of the bed surface, under a constant vertical pressure), which contains information on sediment packing, was observed during a field survey, over an intertidal area adjacent to the entrance channel to Christchurch Harbour, southern England. The factors which control the observed sinking depth, together with some implications in sediment transport, are analysed in this contribution.

Christchurch Harbour is an estuarine/inlet system (Figure 1). Within the entrance, the long-term, cross-sectional mean current speed during the ebb-tide is around 0.56 m s^{-1}, with a maximum value of more than 2.25 m s^{-1} in response to a combination of spring tides and high freshwater discharges (GAO and COLLINS, 1994a). Within the harbour, the water depth is generally less than 2 m, with extensive intertidal flats and salt marshes (MURRAY, 1966). The northeasterly longshore drift of sand and gravel from Hengistbury Head has led to the development of two spits. Because of the periodic changes in the location and extension of the spits, the ebb tidal delta is considered unstable (ROBINSON, 1955). The location of the entrance was artificially stabilised due to the construction of the Mudeford Quay (along the northern side) in the early 1950s. At present, gravels are distributed within the entrance channel, whilst sand and gravel are present on the southern side of the entrance channel.

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FIELD OBSERVATIONS AND DATA ANALYSIS

On 20th February, 1992, surficial sediment mapping and a topographic survey were undertaken along the southern side of the entrance channel (Figure 2), as a part of a more general investigation into the sediment dynamics and morphological stability of the inlet system (GAO and COLLINS, 1994b). The survey was during a tidal phase between mean sea level and low water, on a spring tide, under calm weather conditions (with weak winds and small swells). At the beginning of the field work, waves of less than 0.2 m in height were propagating into the entrance over the ebb tidal delta. Towards low water, the water surface was almost flat.

For the topographic survey, a base line was established which was located above the high water mark, parallel with the axis of the channel (Figure 2). The origin and bearing of the profiles were fixed then, in relation to the base line. The bed elevation was determined by levelling, using a theodolite. The position of the sediment sampling sites were fixed, using the base line.

The type of surficial sediment (in terms of the FOLK (1980) classification scheme) varied significantly over a short dis-

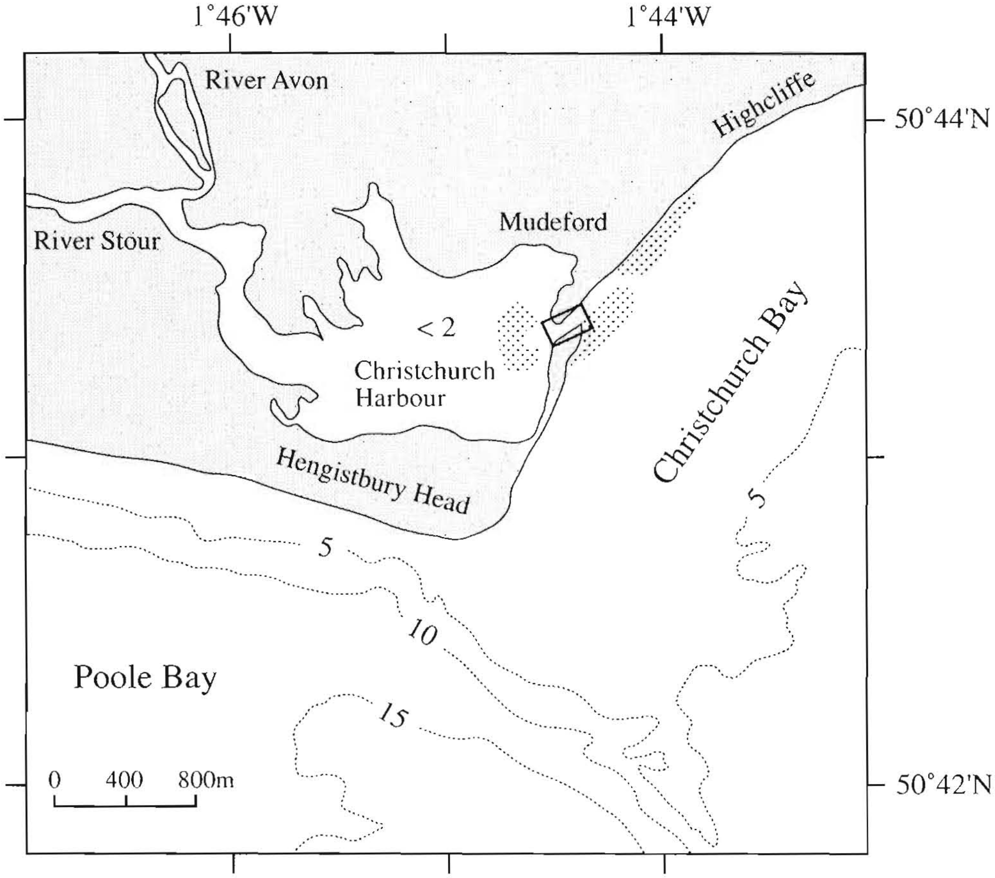

Figure 1. Location of the study area (boxed), in relation to Christchurch Harbour. Bathymetry in metres.

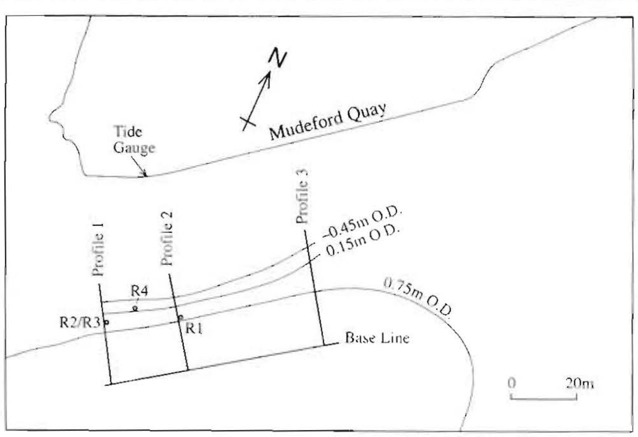

Figure 2. Sediment sampling sites (R1 to R4) and Profiles 1 to 3 for topographic surveys, for the boxed area shown in Figure 1.

tance, according to the result of the sediment mapping (Figure 3). However, a part of the survey area, across Profiles 1 and 2, was covered by well-sorted fine sand. Here, the thickness of the sand deposit varied between 0.2 and 0.3 m, as revealed by a trench dug across the intertidal zone, and was underlain by sandy gravel. No visible internal sedimentary structures were detected. The bed surface was smooth, with few ripples or other bedforms. When standing on the sand surface (this is equivalent to exert a vertical pressure of around 2,500 kg m^{-2}, as estimated using the ratio of one's body weight to the area of foot-bed contact), during low water when the beach emerged, the sinking depth varied between 0 and 9 cm, over only a 6 m distance.

Since the topographic data are available, it is convenient to show the distribution of the zones associated with different sinking depths. The spatial distribution of the sinking depth, determined by the field mapping on the basis of repeated measurements, was plotted along Profiles 1 and 2 (Figure 4). The sand bed along Profiles 1 and 2 is divided into soft, transitional and hard beds, with sinking depths of 3–9 cm, 1–3 cm and <1 cm, respectively. The bed type can be related to bed elevations, with significant zonation. Along Profile 1, the hard bed occurred either above 0.7 m O.D. (Ordnance Datum, Newlyn), which is the mean high water on springs (MHWS),

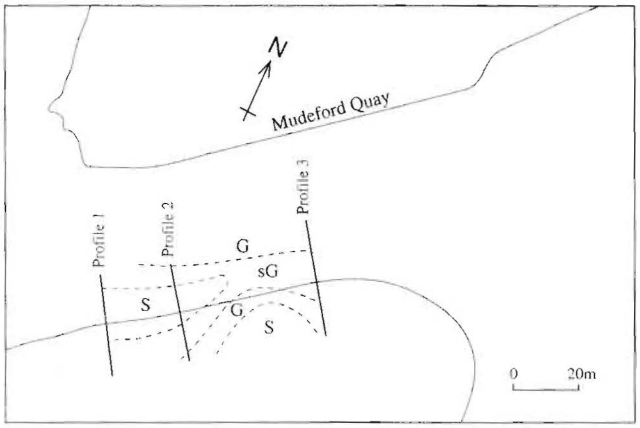

Figure 3. Surficial sediment distribution over the beach along the southern side of the entrance channel, determined by *in situ* mapping.

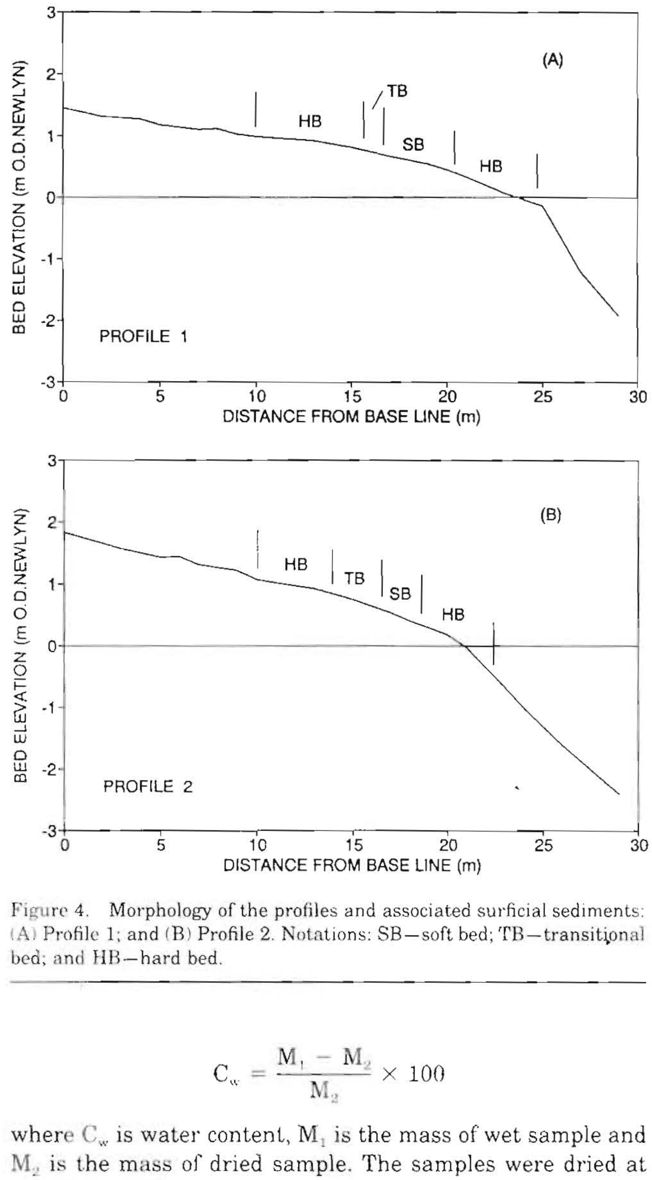

Figure 4. Morphology of the profiles and associated surficial sediments: (A) Profile 1; and (B) Profile 2. Notations: SB—soft bed; TB—transitional bed; and HB—hard bed.

or below 0.3 m O.D. (Newlyn). The transitional bed lay between 0.7 and 0.6 m O.D., with the soft bed between 0.6 and 0.3 m O.D. (Figure 4). Along Profile 2, the hard bed was located above 0.8 m O.D. and below 0.3 m O.D., with the transitional bed and the soft bed lying within 0.8–0.55 m O.D. and 0.55–0.3 m O.D., respectively. The sand bed was not present along Profile 3. In general, the hard bed lay either above MHWS, or around 0.15 m O. D. (this elevation represents the mean sea level here (GAO and COLLINS, 1994a)). The soft bed was located near the mean high water on neaps (MHWN).

Because the observation was made by chance, parameters relevant to sediment packing such as yield strength and porosity of the sand deposit were not obtained. Nevertheless, the sinking depths observed provide some information on the packing characteristics. In particular, such large variations in the sinking depth, over such a short distance, were striking. Thus, four of the sediment samples (i.e. R1 to R4 on Figure 2), which were collected from the locations representing different sinking depths, were used for further examination. The sampling sites for R1, R2/R3 and R4 were associated with bed elevations of 0.65 m, 0.55 m and 0.0 m O.D., respectively. Sample R1 was from the "transitional" bed, where the sinking depth ranged between 1 and 3 cm. Samples R2 and R3 were from a zone of the "soft" bed (with a sinking depth of 3–9 cm), where the maximum sinking depth (i.e. 9 cm) occurred, with R2 taken within the top 3 cm and R3 from the depths between 6 and 9 cm below the sediment surface. Finally, Sample R4 was collected from a "hard" bed, near to the low water mark, where the sinking depth was not detectable. These samples were placed in water-tight plastic bags (to maintain the *in situ* water content) for further analysis. No significant differences in sand colour and grain size could be identified visually.

In the laboratory, the water content (i.e. the ratio of water mass to dried sediment mass, in percentage) of the samples was obtained from:

$$C_w = \frac{M_1 - M_2}{M_2} \times 100$$

where C_w is water content, M_1 is the mass of wet sample and M_2 is the mass of dried sample. The samples were dried at 100°C, for 24 hr.

Grain size analysis of the samples was undertaken using dry sieving method and a 0.5 ϕ interval. Based upon the grain size distributions, dimensionless grain size parameters in terms of ϕ values (mean (μ_ϕ), sorting coefficient (σ_ϕ), skewness ($S_{k\phi}$) and kurtosis ($K_{u\phi}$)) were calculated using the moments method (MCMANUS, 1988). In addition, grain shape (roundness and sphericity) and mineral composition of the samples were examined, using a microscope.

RESULTS AND DISCUSSION

The results of the analysis revealed some distinctive characteristics. As shown in Table 1, the water content of the samples was similar for Samples R1, R2 and R3 (around 19%), with an exception for R4 (around 25%). The higher water content for R4 may be explained by the fact that this site was close to the low water mark and, therefore, exposed at a

Table 1. *Dimensionless grain size parameters and water content of the surficial sediment samples (for location, see Figure 2).*

Sample Number	$M_{d\phi}$	μ_ϕ	σ_ϕ	$S_{k\phi}$	$K_{u\phi}$	C_w (%)
R1	2.20	2.17	0.35	−0.26	0.50	19.2
R2	2.26	2.25	0.30	−0.16	0.43	19.5
R3	2.16	2.12	0.37	−0.37	0.62	19.5
R4	2.22	2.20	0.31	−0.22	0.44	25.2

later tidal stage. Further, the grain size parameters were similar for all the samples (Table 1).

No significant differences in mineral composition, roundness and sphericity were identified under microscope examinations. The sediment particles were dominated by quartz grains, with few (if any) heavy mineral particles. The roundness was between sub-angular and sub-rounded, with an intermediate sphericity, on the basis of visual determination (POWERS, 1953).

The analyses summarised above indicate that the sediment samples representing different sinking depths are highly similar in terms of their water content, grain size, mineral composition and shape characteristics. Hence, the observed variations in the sinking depth cannot be explained by differences in clast properties. Since variations in packing are controlled by either sediment characteristics or sedimentary processes (LEEDER, 1982, p.42–43), it is necessary to examine the hydrodynamic conditions associated with the depositional processes. In particular, because of tidal current domination within the channel during the field observations (see above), it is appropriate to use the conditions of tidal current speeds and rates of changes in water-level.

On the basis of simulated tidal data (for location of the tide gauge, see Figure 2), together with freshwater discharge and tidal basin hypsometric data, the cross-sectional mean current speeds were calculated (GAO and COLLINS, 1994b). Using the water-level and current speed information (Figure 5), typical speeds and rates of water-level change were determined for characteristic bed elevations (Table 2).

With the exception of Site R4, the data listed in Table 2 appear to suggest that the elevations with various sinking depths is related to mean current speed and rate of water-level change. The upper hard bed corresponded to low speeds and rates. Conversely, the soft bed was associated with high speeds and rates. The data for the lower hard bed elevation (from Site R4) did not fit into this generalisation. Here, the rate of water-level change was the highest, together with a relatively high value of the current speed averaged over the cross-section. Such a discrepancy may be explained by the following observation.

During the field survey, it was observed that when the water-level was high, currents near the southern shoreline were directed towards the sea; this was consistent with those at the central part of the channel. When the water-level fell, however, localised eddies were generated when a headland-like feature was exposed to the east (down-stream) of Site R4. Hence, although at this time the ebb currents were rapid within the central part of the channel, they reversed along its southern side. On the day of the survey, such reversed

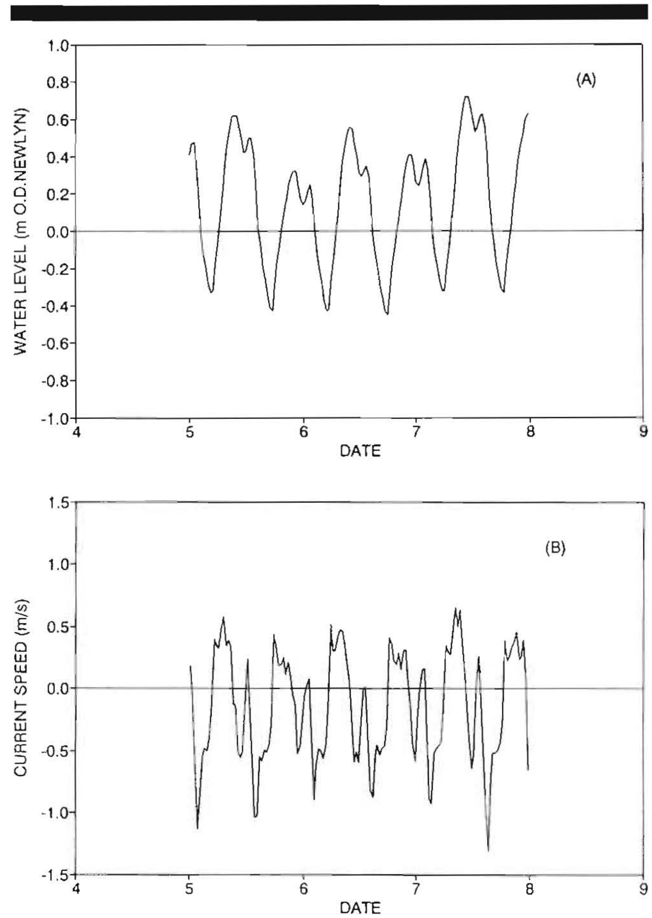

Figure 5. Water-level record (A) and the derived cross-sectional mean current speeds (B), for 5–7th January, 1992.

current speeds were estimated to be less than 0.3 m s^{-1}. This interpretation infers that *local* currents are important in the sediment packing patterns.

At Site R4, the rate of water-level change was high and the localised current speed was low at the time of the bed exposure. These observations suggest that a soft bed cannot be formed unless both the rate of water-level change and the localised current speed are sufficiently high, when deposition takes place.

Although the rate at which sand particles decelerate their motion was not evaluated explicitly in the present analysis,

Table 2. *Mean cross-sectional current speeds and rates of water level changes during the ebb, related to bed elevation and sediment packing zones (cf Figure 4).*

Sample Number/ Location	Elevation (m O.D.Newlyn)	dh/dt (cm hr^{-1})	U_c (m s^{-1})
HWLS	>0.7	<6	<0.4
R1	0.7–0.6	6–16	0.4–0.7
R2/R3	0.6–0.3	20–36	0.9–1.1
R4	<0.3	36–42	0.8–0.9

the effect of high current speeds followed by a rapid fall in water-level should be equivalent to a rapid deceleration (*cf.* LEEDER, 1982). Under such conditions, it would take only a short time for the sediment to change from a highly mobile state to a completely motionless state. For Site R2/R3, the current speed towards the final exposure was likely to exceed 0.8 m s^{-1}; thus, the sediment would be in suspension or under a transitional condition between bed-load and suspension (STERNBERG *et al.*, 1985). Consequently, rapid deposition would occur if sediment movement ceased suddenly.

There is a correlation between the strength of the sediment bed (represented by the yield stress) and the rate of bed formation. Therefore, both variables are likely to influence sediment transport processes, through modification of the critical shear stress at threshold of motion. Elsewhere, the critical shear stress for sediment motion has been identified on the basis of a series of "Shields' curves", depending upon the criterion adopted to define motion (LAVELLE and MOFJELD, 1987). Additionally, the observations reported here imply that packing characteristics (represented by the bed strength) can be hydrodynamically controlled, independent of sediment grain size and shape factors. Thus, on the basis of a linear relationship between critical shear stress and bed strength (DUNN, 1959), the same sedimentary material may be associated with different critical shear stresses for sediment motion. For strong tidal currents, which favour the formation of a soft bed, a lower threshold may result than if the currents are weak. Hence, the critical shear stress could be lower in high tidal energy environments than in low energy environments; this would further enhance the difference in sediment transport rates between the various environments. Flume experiments for the threshold of particle motion have demonstrated that sediment mobility depends upon the history of the shear stress exerted on the bed (TOMLINSON, 1993). This is another indication of the phenomenon that the bed mobility is related to packing patterns.

CONCLUSION

Significant variations in the sinking depth of sand surface have been observed on a beach adjacent to an entrance channel to a tidal basin. Over a 6 m distance, the sinking depth of the sediment surface ranged between 0 and 9 cm (under a constant vertical pressure of around 2,500 kg m^{-2}).

The grain size parameters, sphericity, roundness, mineral composition and *in situ* water content data do not indicate that such variations are controlled by sediment characteristics. The relationship between the sinking depth and bed elevation (and, therefore, localised current speed and rate of water-level change immediately preceding bed exposure) suggest a hydrodynamic control. Thus, for the same beach material, a soft bed can be formed if the localised speed and rate of water-level change are *both* sufficiently high when the bed is submerged; otherwise, a hard bed is formed. Likewise, for any sedimentary deposit which is never exposed in the air, a soft bed could be formed if the bed is subjected to strong currents followed by a rapid reduction in the current speed (and, hence, rapid accumulation).

Hydrodynamically-induced variation in grain packing implies a difficulty in defining the threshold of motion (on the basis of conventional approaches), in terms of the linear relationship between critical shear stress and shear strength of the sediment. For the same material, strong tidal currents which favour the formation of a soft bed may result in a lower threshold of motion, than if the currents are weak.

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