Coastal Engineering 57 (2010) 160–175

Contents lists available at ScienceDirect

Coastal Engineering

journal homepage: www.elsevier.com/locate/coastaleng

Hydrodynamics of a headland-bay beach—Nearshore current circulation

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ARTICLE INFO

Available online 4 November 2009

Keywords:
Headland-bay beaches
Nearshore circulation
Shoal influence
Wave currents
Analytic solution
Mild-slope equation
Radiation stress tensor

ABSTRACT

This paper is devoted to the analysis of the hydrodynamic equilibrium of a headland or semi-elliptic shaped beach. It is shown that the state of equilibrium depends not only on the in- and outgoing sediment but also on the accommodation of the sediment within the embayment. The shape and relative depth of shoals, or settling zones, also directly affect the wave and current patterns inside the bay, within which the resultant breaker line almost stops wave-induced currents at some locations, whereas the magnitude of current increases at other locations. Several numerical tests are analytically conducted in a semi-elliptic beach with two symmetrical shoals of varying relative depth where circulatory current systems are detected and analyzed. Numerical modelling for wave climate and wave-induced current estimation is also presented in order to corroborate results and provide a tool for complicated and/or physical domains. The results lead to a redefinition of the concept of equilibrium for headland-bay beaches taking into account not only the net sediment transport but also the role of the formation and disappearance of settling zones, as well as sediment interchanges between the beach and shoals.

1. Introduction

According to Hsu et al. (2008) the stability of a headland-bay, or crenulated, beach, is determined by the balance between the incoming and outgoing sediment in the beach. It is this balance which allows beach classification, given as |, *static equilibrium*, and *unstable* due to a reduction in sediment supply and "reshaping beaches" caused by the installation of a coastal structure or modification to an existing wave diffraction point.

In crenulate beaches the sediment is partially or totally blocked at the headlands and they usually have therefore very little sand supply from neighbouring updrift areas. They show a characteristic curved coastline which is the result of a continuous adjustment to the incoming wave fronts. This gradual adaptation is mainly driven by the longshore sediment transport that takes place at locations where the waves break obliquely to the bathymetric contours. Under the prevailing wave climate, the circulatory system in a crenulate or headland-bay beach is relatively weak and net sediment transport is not significant. Therefore, the shoreline planform in such a beach closely reflects "equilibrium" conditions. However, the beach does not remain static, but responds to climatic changes, storms and calms; gradually adapting its planform to the prevailing wave conditions.

Understanding the morphodynamics of crenulate beaches is not possible without first examining the hydrodynamic phenomena that take place within the bay. The complexity and wide range of temporal and spatial scales of the processes involved make such a description very challenging.

Over a short period covering a few sea states, during a storm for example, the physical activity in a headland-bay is mainly dominated by wind waves and their transformation by bathymetric changes through shoaling, refraction, reflection and, at the headland, mainly diffraction. As the water depth becomes shallower, the wave profile becomes steeper and breaks when a threshold, approximately H_L/h, is reached (H_L being local wave height and h the still water level). This position is the beginning of the surf zone, where the height of the breaking wave falls, dissipating a great deal of energy, mainly through turbulence. As waves propagate into shallow water, the gradients of the excess momentum flux of the waves cause a rising and lowering of the mean water level (set-up and set-down) and may induce a complex circulatory system, particularly intense inside the surf zone, which broadly consists of undertow, longshore and rip currents. These wave-induced currents interact with the incoming waves modifying the wave propagation pattern. Finally, at the shoreline, the wave begins a landward run-up process in the form of a thin sheet of water which returns seaward by gravity.

In this general picture of beach dynamics, the alongshore drift (generated in the direction of the longshore component of the incident wave when wave breaking takes place oblique to the bathymetric contours) is the dominant motion averaged over the wind wave period. Although the current magnitude is often weak, it is however capable of transporting the sediment mobilised by wave action. This is the principal cause of morphological changes at the beach over short periods.

It is also easy to observe that inside the bay, outside the surf zone, shoals or zones of sediment accumulation develop, usually under moderate wave action and these are swept away under storm

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doi:10.1016/j.coastaleng.2009.10.003

conditions. The mechanisms involved in the generation and extinction of these morphological features are still not well understood. Generally speaking, with relatively low waves the sediment is brought shoreward in the propagation region, building a monotonic beach profile. The circulation induced within an embayment by wave breaking redistributes part of this sediment along the shore and also forms shoals at the end of the outgoing rips. The changing bathymetry modifies the wave propagation pattern and, therefore, the wave height and the angle at which the waves break, with subsequent changes in current field direction and intensity.

The concept of dynamic equilibrium in headland-bay beaches should not therefore isolate shoreline behaviour from shoal formation and the possible interrelations between these features should be examined. In this paper, the role played by shoals in the general circulation of a beach is analyzed. It is shown that the development of the shoals decreases the intensity of the velocity field and eventually stops it. Moreover, if the water depth over the shoals is shallow enough, wave breaking inside the bay can even invert this process, sweeping the shoals away.

2. Governing equations for mean flows

It is well known that the oscillatory motion in wind wave time scales can induce free surface motions and current fields of a longer time scale than the corresponding wave period. The equations governing these mean flows are the shallow water equations, obtained from time and depth averaged Navier–Stokes equations. In a Cartesian horizontal coordinate system with the (x,y) axes lying over the still water level and the z-axis pointing upwards, they can be written as

$$\frac{\partial\eta}{\partial t}+\frac{\partial(uh)}{\partial x}+\frac{\partial(vh)}{\partial y}=0 \tag{1}$$

$$\frac{\partial(uh)}{\partial t}+\frac{\partial(u^2h)}{\partial x}+\frac{\partial(uvh)}{\partial y}-\left(\frac{\partial(\varepsilon hu_x)}{\partial x}+\frac{\partial(\varepsilon hu_y)}{\partial y}\right) = \frac{\tau_{wx}-\tau_{bx}}{\rho}-gh\frac{\partial\eta}{\partial x}+hfv-\frac{1}{\rho}\left(\frac{\partial S_{xx}}{\partial x}+\frac{\partial S_{yy}}{\partial y}\right) \tag{2}$$

$$\frac{\partial(vh)}{\partial t}+\frac{\partial(uvh)}{\partial x}+\frac{\partial(v^2h)}{\partial y}-\left(\frac{\partial(\varepsilon hv_x)}{\partial x}+\frac{\partial(\varepsilon hv_y)}{\partial y}\right) = \frac{\tau_{wy}-\tau_{by}}{\rho}-gh\frac{\partial\eta}{\partial y}+hfu-\frac{1}{\rho}\left(\frac{\partial S_{yx}}{\partial x}+\frac{\partial S_{yy}}{\partial y}\right) \tag{3}$$

where η is the free surface elevation, u and v are the velocity vector components, respectively, h is the total depth ($h=h_s+\eta$) with h_s being the still water level. ε is the eddy viscosity, τ_{wi} the surface stresses, τ_{bi} is the bed friction stresses ($i=x,y$) and f is the Coriolis parameter. S_{ij} are the components of the radiation stress tensor that represent the excess momentum fluxes associated with the oscillatory wave motion.

The gradients of the radiation stress are the driving forces of the mean flows. These forcing terms depend on the wave propagation pattern and in a first approach they are usually specified and provided as an input, therefore neglecting the interaction between the mean and the instantaneous quantities. Even with this simplification, the resolution of the whole set of equations can be very complicated, especially over real domains with complex boundaries and bathymetric contours.

In the next section an overview is given of the analytical approaches to estimate circulation induced by wave breaking with emphasis on semi-elliptic beaches. Next, a numerical model that comprises both a wave propagation model (Silva et al. 2005) and the estimation of the mean flows valid for arbitrary bathymetric contours over complex domains is summarized.

3. Analytical solutions

Over half a century ago Iribarren (1947a) pointed out the importance of the alongshore variation of the set-up induced by waves to drive currents in a permanent regime and derived a formula based on that of Chezy to estimate the magnitude of the current. He applied these findings to explain the growth of a spit in the estuary of Fuenterrabía (Iribarren, 1947b) and the development of a crenulate beach bounded by a jetty at Cape Higuer. Elsewhere, Shepard and Inman (1950) described the influence of rip currents on incident waves, nearshore circulation of water, magnitude and direction of sediment transport and consequently the shape of shorelines. Later McKenzie (1958) underlined the relationship between rip current flows and wave-forcing parameters.

Since the introduction of the radiation stresses by Longuet-Higgins and Stewart (1964), several attempts have been made to estimate the circulation patterns induced by wave breaking. In this context, the analytical resolution of the simplified equations over certain regular domains, although less accurate than numerical models, has given insight into the behaviour of mean flows under different forcing conditions depending on the geometry of the considered domain.

With this aim, the solutions provided by Bowen (1969a) and Longuet-Higgins (1970) described the mean water level and the longshore current over an infinitely long beach with straight and parallel contours with the radiation stresses calculated for a monochromatic wave train approaching the shore at an oblique angle. Battjes (1974) addressed this case but with irregular waves instead. Their solutions showed that the obliquity of wave fronts at breaking is responsible for an alongshore component of the momentum flux into the surf zone.

Also on a straight beach Bowen (1969b) and Dalrymple (1975) showed that alongshore variations of the breaking wave height will cause a variation in wave set-up along the shore. Under such conditions feeder currents will flow away from zones of high waves and toward zones of low waves where they converge and move seaward as rip currents.

A more complex geometry was analyzed by O'Rourke and Le Blond (1972) who studied the circulation patterns in a semi-elliptic bay and found that the rip currents are mainly driven by the oblique angle of incidence. Also, Baquerizo et al. (2002) and Baquerizo and Losada (2002) presented a quasi-analytical approach for the circulation induced by wave breaking in a semi-elliptic beach, and showed that for this type of beach, the circulation inside the surf zone is mainly due to the angle at breaking, while the effect of the gradient of the wave height strengthens or weakens the velocity values without modifying the circulation pattern.

In this paper the analytical model of Baquerizo and Losada (2002) is applied to study the nearshore current field for headland-bays with a shoreline form resembling a semi-elliptic beach as formed within the shelter of two headlands. Appendix A summarises the model assumptions and the resolution procedure.

4. Numerical approach

For an arbitrary beach form with irregular bathymetric contours, the numerical resolution of equations requires the coupling of two numerical models. For the estimation of the wave climate a modified mild-slope equation, in its full form, is used given that reflection and

Table 1
Beach configuration and climatic parameters.

Beach configuration	x_s (m)	y_s (m)	δ	p	m_0	$s\phi_b$ (rad)
A	1200	1100	0.25	0.25	1	$\frac{\pi}{3}\cos v$
B	1200	1100	0.25	0.00	1	$\frac{\pi}{3}\cos v$
C	1200	900	0.25	0.25	1	$\frac{\pi}{3}\cos v$